Soil Survey Papers, No. 13

SOIL SURVEY AND

THE STUDY OF

WATER IN

UNSATURATED SOIL

J. Bouma

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SIMPLIFIED THEORY AND SOME CASE STUDIES

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SOIL SURVEY INSTITUTE, WAGENINGEN, THE NETHERLANDS — 1977

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PREFACE ON TERMINOLOGY

The SI system ("Système International d'Unités") is globally accepted and its use is now required for expressing physical processes (Hesse, 1975; ISSS, 1976). These units, but also more conventional units, are used in this publication, which is intended as an introductory text for soil scientists involved with applied field work. Exclusive use of SI units would, in this context, probably restrict its informative value. Of particular relevance here is the expression for pressure which can, according to the SI system, no longer occur in terms of centimetre of water, millimetre of mercury, bars etc., but which must be expressed in pascal (Pa). A recent ISSS committee report on soil physics terminology concluded: "it is a sheer impossible task to suggest a set of symbols which on the one hand corresponds with common usage and, on the other hand, is completely consistant" (ISSS, 1976). The committee therefore still uses mbars in their definitions and includes a definition of soil moisture pressure in terms of a "pressure head" (m). The latter, in particular, has the advantage for this publication that the important hydraulic conductivity (K) can be simply expressed in terms of m s^{-1} (SI units) or in the common: cm day^{-1} (cm/day). This avoids complex expressions in terms of m^2 $Pa^{-1}s^{-1}$ (SI units), which are, of course, more acceptable from a strictly physical point of view. To allow transformation into SI units, where needed, the following summarized information is pertinent.

The basic units of the IS system, relevant for this publication, are: metre (m) (length), kilogramme (kg) (mass) and second (s) (time). Pressure is expressed in pascal (Pa) (kg $m^{-1}s^{-2}$). A pascal is one newton per square metre (1 N m^{-2}) and a newton is defined as the force necessary to give an accelleration of one metre per second squared to a mass of one kilogramme.

A pressure of 1 cm of water = 0.098 kN m^{-2}
A pressure of 1 mm of mercury = 0.133 kN m^{-2}
A pressure of 1 bar = 100 kN m^{-2}
Bulk density or particle density of soil (b and ϱ) (kg m^{-3})
Viscosity η (kg $m^{-1}s^{-1}$)
Surface tension σ (N m^{-1})

1. INTRODUCTION

Understanding and predicting water movement in unsaturated soil above the groundwater is of great practical significance for a wide variety of soil uses. Physical measurement techniques and flow theory — which are indispensible to reach such understanding — have been developed, but they are complicated and may not always be adequate to describe flow through heterogeneous, anisotropic soils as found in the field. In fact, a review of current literature shows that the emphasis in soil physical studies is still on well controlled laboratory experiments with known boundary conditions which are not necessarily representative of field conditions (KOENIGS and BOLT, 1973). This approach is, of course, significant for increasing the basic understanding of flow processes in soil. But lately there has also been an increased emphasis in soil physics on more practical studies determining field variability of soil physical constants (NIELSEN et al., 1973; KLUTE, 1973) and on using large and undisturbed, rather than artificially packed soil columns for physical experiments (CASSEL et al., 1974). This tendency will be strongly encouraged by the development of physical simulation techniques which are being used more and more to calculate soil moisture regimes, not only in the soil at one location (DE WIT and VAN KEULEN, 1972; VAN DER PLOEG, 1974) but also in entire regions with many different soils (DE LAAT et al., 1975; DUFFY et al., 1975). The latter type of simulation in particular requires representative soil physical data in terms of hydraulic conductivity and moisture retention for representative soils within a given area. To obtain such data it may be more relevant to apply a large number of relatively simple, yet accurate, measurement procedures to many soils than to obtain a few, very precise measurements in, by necessity, a very limited number of soils. This aspect in the development of the physical study of water movement in soil clearly represents a tendency to broaden the scope of research, thereby limiting its depth.

Research activities in soil survey, morphology and classification have resulted in new soil classification systems which are more quantitative and less subjective than before (SOIL SURVEY STAFF, 1975; DE BAKKER and SCHELLING, 1966) and in new microtechniques for studying soil morphology (BREWER, 1964; JONGERIUS, 1974). But in addition, emphasis is increasingly placed on improving soil survey interpretations not only by emphasizing soil limitations or suitability for a given use at the current technology level, but also by defining soil potential for a wide variety of uses utilizing specified new technology (MCCORMACK, 1974, BOUMA, 1974). These improved interpretations require specific soil physical data, and represent a tendency to focus the research (thereby increasing its depth) towards selected pedons which are to be characterized by modern techniques rather than by estimates based on experience.

These tendencies in the disciplines of soil physics and soil survey, occurring as a result of modern demands, appear to suggest a narrowing of a historic and perhaps somewhat unfortunate communication gap between specialists in both disciplines. This publication was written as an attempt to provide a simple, if possible nonmathematical, discussion of unsaturated flow phenomena, to be useful for field soil scientists. Some basic understanding is necessary on their part if they are to provide significant physical data by either direct measurement or by estimation. The practical need for such data is acute, because soil maps are increasingly being used as basic data for simulation programmes, since they present a directly available, systematic display of the occurrence of different soils in the landscape (DE LAAT et al, 1975). Several good textbooks on soil physics are now available but there is a need to combine an essentially non-mathematical approach with examples relating to concepts used in soil survey and morphology! This consideration is based on the author's experience in Wisconsin (USA), when trying to explain unsaturated flow phenomena during liquid waste disposal, to sanitarians and field soil scientists.

This publication also represents an attempt to demonstrate to the soil physicist that useful data, in terms of defining the occurrence of different soils in the field and of using soil morphology to define anisotropic soils, can be derived from soil survey data. An integrated discussion which avoids mutual points of irritation, such as complex mathematics and opaque terminology, and which attempts to combine basic elements of both disciplines, is thought to be useful in advancing practical, applied field work studying water movement in soil.

2. SOIL POROSITY

2.1 A morphological characterization of soil pores

Water in the soil can only move through the pores. Every soil material is composed of solid particles and the remaining spaces are called voids or pores. All voids are interconnected, so soil materials can be regarded as having a single void of intricate shape which varies considerably in its dimensions throughout its extent. The shape of the space which results from the regular packing of spheres is known, but voids in many soil materials show that factors additional to simple packing have influenced their shape, size and arrangement. In many soil materials a significant proportion of the void space consists of voids of comparable size interconnected by very much smaller voids, so that the larger ones can be treated as entities. Other soil materials may have systems of interconnected planar voids which meet at relatively sharp angles so that, again, they can be treated as entities, even though the actual extent of a single individual may be a matter of judgement. Even the interconnected voids that result from the packing of individual grains can be treated as entities if the rule is adopted that the significant constrictions where the walls of the solid individuals come closest together are the boundaries of individual voids, which are then connected through relatively narrow "necks". This very concept has been used by MARSHALL (1958) and others, to calculate the permeability of porous media. If this concept of voids as entities which are interconnected, either through voids of dissimilar size and shape, through narrow necks, or through intersection with voids of similar size and shape, is adopted, they can be described and classified in terms of their size, shape and arrangement (BREWER, 1964).

A discussion of detailed morphological studies is beyond the scope of this publication which mainly focusses on physical techniques and theory as far as they are relevant to field pedologists. However, in discussing physical terms, such as porosity and bulk density, and in reviewing methods of determination, it is essential to have in mind a picture of the real constituton of the soil material. A brief review will therefore be presented of major pore types and of soil materials in which they occur. This review generally follows criteria defined by BREWER (1964).

Soil structure, as used in this text, refers to the physical constitution of a soil material as expressed by the size, shape and arrangement of the solid particles and voids, including both the primary particles forming compound particles and the compound particles themselves. The primary particles consist of the clay, silt and sand fraction, which are defined on the basis of size. The clay fraction covers particles smaller than 2 micron; the silt fraction the range between 2-50 micron and the sand particles are coarser than 50 micron. In addition, organic matter may occur in various forms, but these compounds are not considered part of the soil texture, which expresses the particle-size distribution of a given soil material (SOIL SURVEY STAFF, 1951). The term "soil texture" is so common in soil science, that a re-definition as proposed by Brewer seems unrealistic.

The primary particles may or may not form compound particles. If they do, such compound particles may qualify as peds, which are individual natural soil aggregates consisting of a cluster of primary particles and separated from adjoining peds by surfaces of weakness which are recognizable as natural voids. Soil materials with peds are called pedal (Fig. 1, right drawing), and those without peds apedal (left drawing). In both cases, the primary particles constitute the bulk volume of the soil, which is called the S-matrix. This S-matrix is therefore the material within the simplest peds, or composing apedal soil material. The structure of the S-matrix can be described and measured by noting size, shape and arrangement of the solid particles and the voids. It is convenient to refer in this context to a description of the basic structure of the soil material. STOOPS and JONGERIUS (1975) have proposed a number of terms that can be used. Generally, clayey soils (which swell upon wetting and shrink upon drying) have peds and their basic structures contain more clay (plasma) than basic structures of apedal sandy soil materials (Fig. 1). The following major types of voids are distinguished: (i) Simple packing voids. These voids are due to random packing of single grains. (ii) Compound packing voids. These voids result from the packing of compound individuals, such as peds, which do not accommodate each other. (iii) Vughs. These voids are significantly larger than simple packing voids, and they appear as discrete entities at the magnification at which they are recognized. (iv) Channels. These voids are significantly larger than simple packing voids and generally have a cylindrical elongated shape. (v) Planes. These voids are planar according to the ratios of their principle axes. By virtue of their shape and length, they constitute an obvious deviation from the normal packing of primary soil particles. These pores are schematically shown in Figure 1 for pedal and apedal soil materials.

Morphometric techniques are available for measuring pore-size distributions and shapes of pores in thin sections. A detailed discussion of these methods is beyond the scope of this text. The reader is referred to BREWER (1964), JONGERIUS (1974) and ISMAIL (1975). However, the general principles of quantitative soil morphological measurements, whether using conventional point-count techniques or modern electro-optical image analysis, need to be discussed.

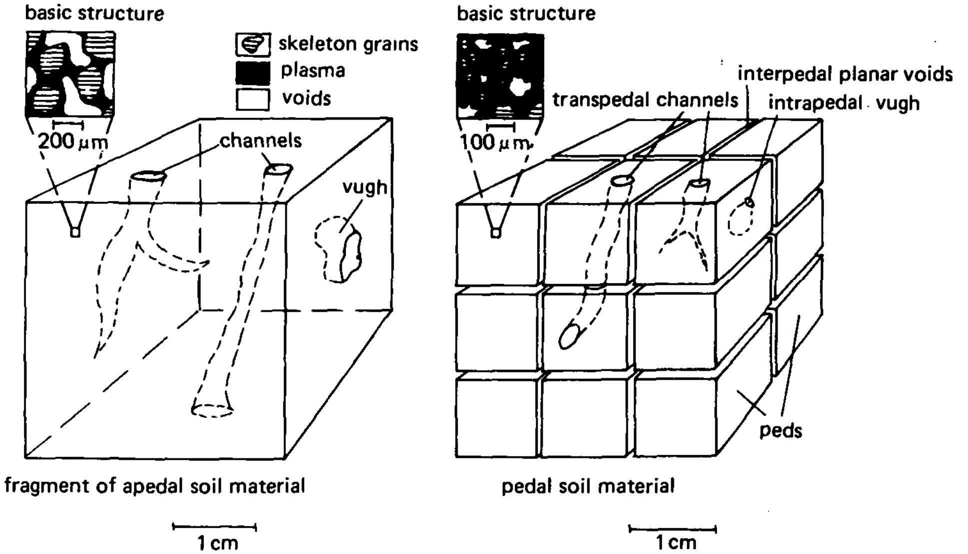


Fig. 1. Simplified structure models in apedal and pedal soil materials.

Detailed observations of soil morphological features are made on thin sections, which are thin slices of soil (20 micron or so), impregnated with plastic (BREWER, 1964; JONGERIUS and HEINTZBERGER, 1975).

A two-dimensional cross section through the soil is thus used to measure quantities and sizes of soil constituents. Data from such measurements are only significant if they apply to the three-dimensional soil body from which the thin slice was taken. Statistical analyses have proved that this extrapolation from a two-dimensional sample to a three-dimensional body is justified when the proper sampling technique is used (ANDERSON and BINNIE, 1961). A commonly used technique in the past was point-counting. Imagine a thin section with voids and solid particles. A point grid is superimposed upon the image and the number of points that fall within voids are expressed as a percentage of all points to obtain an area-percentage for voids which is representative of the percentage (by volume) within the soil body. Pore-size distributions can be determined by measuring the size of the pore at each point, which falls within a void, and the occurrence of different types of voids can be expressed by also noting the type of pore.

Statistical techniques should be used to determine the minimum number of points necessary to obtain reliable counts (VAN DER PLAS and TOBI, 1965). Modern electro-optical image analysis which has recently been developed, allows a much faster and more accurate determination of pore volumes and pore-size distributions (JONGERIUS, 1974, ISMAIL, 1975).

Generally, transmittent light is used to study the arrangement and properties of the soil constituents with a light microscope. When very fine pores, equal to or slightly greater than the thickness of the section, are measured, a "neck" effect is introduced. A pore is seen because it is black under crossed polarizers and remains so when the microscope table is turned. Only a few pores whose walls make right angles with the surface of the section show their true cross-sectional size, while all others, with oblique walls, show their smallest dimensions in the plane of the thin section. When the pores are larger, the relative error is much smaller.

Generally only pore diameters larger than 30 micron are reported. Distinction of different types of pores is best possible when dealing with larger pores. The possibility of distinguishing different types of pores is perhaps the most significant contribution that can be made by morphological analyses.

2.2 The physical characterization of soil porosity

Physical analyses do not allow the distinction of different types of pores: a pore is a pore is a pore, and they all add up to the total pore volume. Equivalent pore-size distributions can be derived from the moisture retention curve for some soils but that aspect will be discussed in Chapter 3. The following example will be used to illustrate the calculation of soil porosity.

Assume a soil sample with a volume of A m^3 which is composed of the solid phase, the liquid phase and the gas phase in the soil. The mass of the soil (dried at 105° C, which by definition corresponds with a moisture content $w = 0\,\%$) is M kg. To know the volume of the pores, we need to know the volume of the solid phase within the sample of A m^3. This volume is found by dividing the mass M by the solid phase density of the soil (ϱ_s) which is defined as:

$$\varrho_s = \frac{M}{V}\ (\mathrm{kg/m^3}) \qquad \text{(ISSS, 1976).}$$

In this expression, V = volume solid phase, ϱ_s can be determined with standard techniques (BLAKE, 1965), and has been commonly expressed and recognized up to now in terms of g/cm³. For clay soils in riverine areas of the Netherlands the following empirical relationship was developed:

$$\varrho_s = \frac{100}{\frac{\%\ \text{organic matter}}{1.47} + \frac{\%\ \text{clay}}{2.88} + \frac{\%\ \text{mineral fraction-clay}}{2.66}} \quad (\text{g/cm}^3).$$

Other expressions can probably be developed for different soils (POELMAN, 1975). The volume of the solid phase within the sample is now M/ϱ_s m³ and the volume of the gas phase (the pores) is $(A - M/\varrho_s)$.m³. For the porosity (%) we find: $100.(A - M/\varrho_s) / A$.

Another very common characteristic used in soil science is the bulk density (b) which is a measure for the amount of soil particles within a certain volume, as follows: $b = \frac{M}{A}$ (kg/m³) (abbreviated from ISSS, 1976). Also this value has been commonly expressed up to now in terms of g/cm³. As before, M = mass of soil (dried at 105° C) and A = total volume of the solid, liquid and gas phase within the sample.

A numerical example follows, which uses g and cm³ to allow better recognition of existing data.

Mass of soil dried at 105° C = 140 gramme;
Volume of sample = 105 cm³ (for example, a SARAN coated clod, to be described in Chapter 3);
Particle density = 2.65 g/cm³.

The calculation of porosity starts with the mass of the stove-dry soil (140 g), with a volume of 140/2.65 = 52.8 cm³. The volume of the non-solid phase is then 52.2 cm³ or $\frac{52.2}{105} \times 100\,\% = 49.7\,\%$. The bulk density (b) is 140/105 = 1.33 g/cm³. This is all very simple. Occurrence of swelling and shrinkage in clayey soils requires a different approach which will be discussed in Chapter 3.

2.3 A comparison of morphological and physical techniques

As already discussed, porosities can be determined by physical and morphometric methods. A comparison of data, obtained with both methods applied to identical samples, is useful to demonstrate limitations and advantages of both methods. Measurements of total porosity were made on different mixtures of sand and gypsum (BOUMA and DENNING, 1974) and sand and clay (BOUMA and ANDERSON, 1973). Pores in these samples were all simple packing voids (Fig. 2). Results, shown in Table 1, indicate that physical porosities are always higher than those obtained by morphometric methods, except for the sand where good agreement is found. The relatively fine pores that exist between the clay plates and the gypsum crystals cannot be seen in the thin section but do, of course, contribute to total porosity. This result would also be obtained if modern electro-optical image analysis were used, because it is due to limitations imposed by using images obtained by transmitting light through a 20 micron thick thin section (see section 2.1). Of course, incident rather than transmittent light could be used to avoid these errors. Then, the true size of the pores is shown at the upper

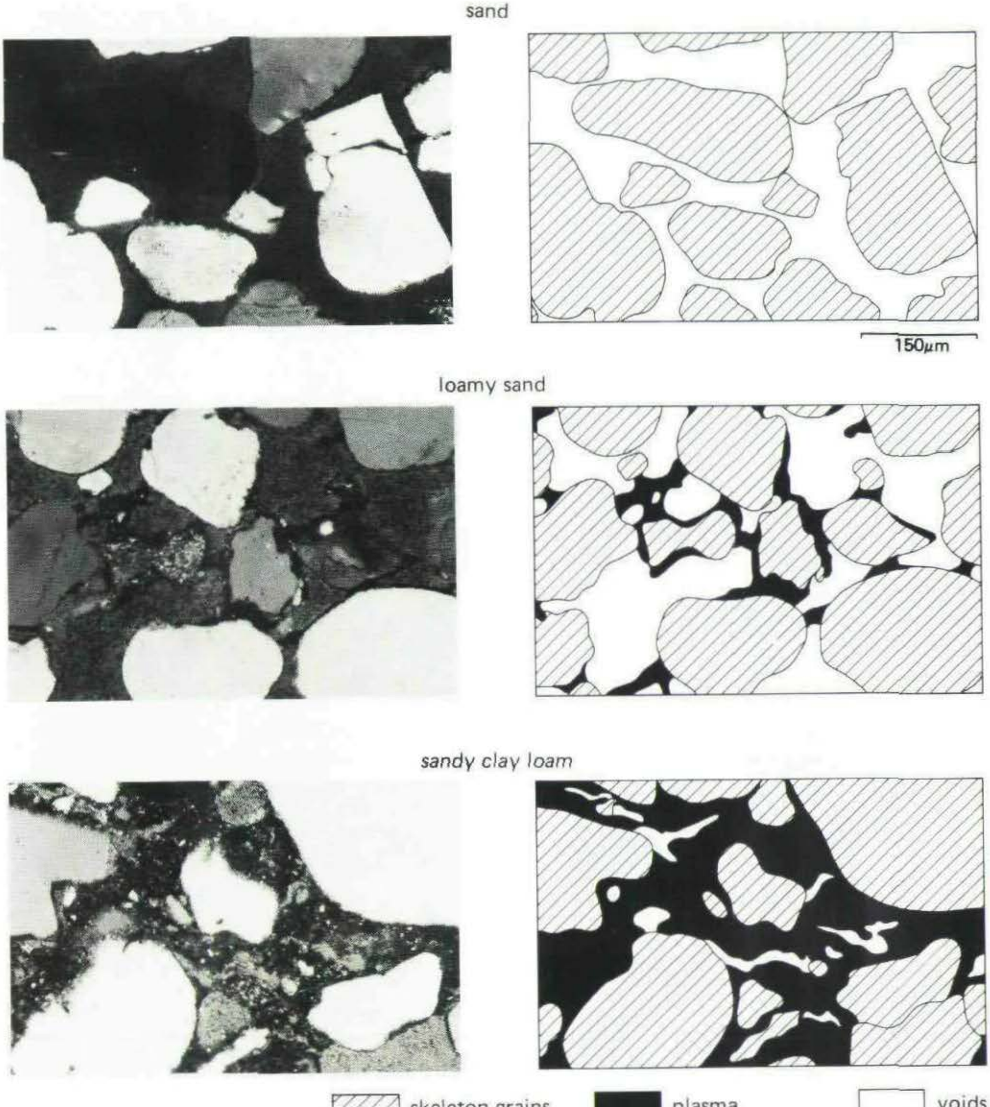


Fig. 2. Three examples of soil fabrics used to calculate porosities with physical and morphometric methods (see Table 1).

surface of the section where the incident light is reflected. This technique offers more technical problems and is uncommon.

Use of *both* physical and morphological data in analysing pore volumes of a 100 cm³ cylinder filled with aggregates was reported by BOUMA (1969). Pictures from thin sections through the aggregate-filled cores are shown in Figure 3. A point count in the thin section yielded a porosity of 39 % for sample A, entirely composed of inter-

Table 1. A comparison of porosities determined with physical and morphometric methods.

Soil material	% clay fraction	Porosity in % according to	
		physical method	morphometric method (point count)
Sand	—	39	40
Sand	4.6	36	31
Loamy sand	9.3	35	21
Sandy loam	13.8	37	14
Sandy clay loam	20.6	37	12
Sand + 12 % gypsum	—	32	23
Sand + 20 % gypsum	—	31	23
Sand + 50 % gypsum	—	21	16

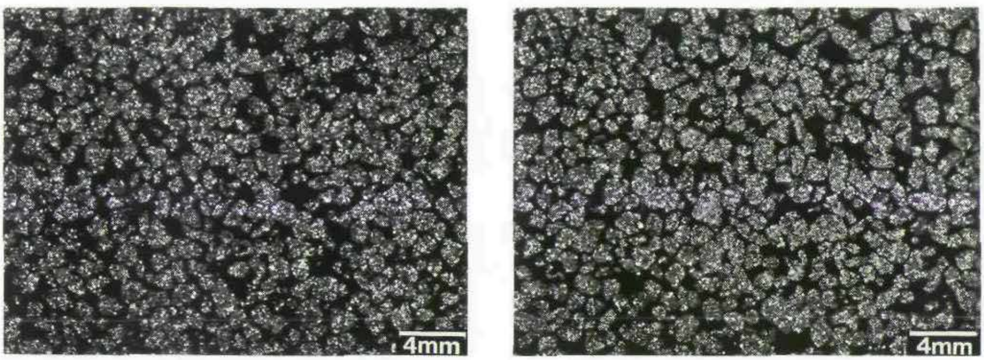


Fig. 3. Part of thin sections through 100 cm^3 cores which are filled with small aggregates (1-2 mm) of two sandy loam soils, with a high (left) and a low (right) organic matter content. These images were used to compare porosities measured with physical and morphometric techniques.

aggregate pores (compound packing voids). The internal porosity of the aggregates was determined in á separate experiment with the use of the apolar liquid kerosine (MCINTYRE, 1954; KUIPERS, 1961). The result was 36.6 ± 0.4 %. The core as a whole had 39 % interaggregate pores (point-count) and therefore 100 — 39 = 61 % aggregates. In these aggregates, 36.6 % of the volume was occupied by pores and so 63.4 % by the solids. These values can be modified to apply to the entire 100 cm^3 cylinder, as follows: 61/100 x 36.6 % = 22.3 % pores and 61/100 x 63.4 % = 38.7 % solids. Total pore space within the 100 cm^3 core would thus be 22.3 + 39 = 61.3 %. For comparison, the porosity was also determined by direct physical measurement, using the procedure described in section 2.2. The result was a porosity of 63.6 %. Agreement between the values derived by the two procedures, is considered reasonable. The second example (sample B) had aggregates with an internal porosity of 33.2 % (kerosine method), whereas the morphometric point count yielded an interaggregate porosity of 43 %. The calculated total pore volume, using the point count data, in the 100 cm^3 core was 61.9 % and the real measured figure was 62.4 %. Again good agreement.

These calculations indicate that the morphometric data combined well with those derived physically and that they can be used together in characterizing soil porosity. The morphometric values allow a more specific characterization of the larger pores. However, the physical methodology is much faster and simpler and therefore preferable for routine determinations of soil porosity. Morphometric analyses have a much more specific role in determining real pore-size distributions (Chapter 3) and, in particular, the occurrence of different pore types.

3. WATER AND SOIL IN EQUILIBRIUM

3.1 The moisture content

The soil moisture content can be expressed as a percentage by weight (w) or by volume (θ). These values can be determined as a percentage as follows:

$$w = \frac{100. (M_m - M)}{M}$$

where: M_m = mass of the moist soil and M = mass of the dry soil (105° C)

and $$\theta = \frac{w.b}{\varrho_w}$$

where: ϱ_w = density of water (usually = 1 g/cm³) (abbreviated from ISSS, 1976).
A numerical example follows:

Mass of wet soil	=	160 gramme
Mass of soil dried at 105° C	=	140 gramme
Volume of sample	=	105 cm³

$$w = \frac{100. (160-140)}{140} = 14.28 \%$$

$$\theta = \frac{14.28 . (140/105)}{1} = 19.04 \%$$

This method, which involves sampling, transporting to the laboratory and repeated weighings entails some inherent errors. It is laborious and time consuming since a period of at least 24 hours is usually allowed for complete drying. The method is also somewhat arbitrary: some clays may still contain appreciable amounts of adsorbed water even at 105° C. On the other hand, some organic matter may oxidize and decompose at this temperature so that the weight loss may not be due entirely to the evaporation of water. In addition, the sampling method is destructive and may disturb an experimental plot sufficiently to distort the results. For these reasons, many workers prefer in situ methods which permit frequent or continuous measurements at the same points, and, once the equipment is installed and calibrated, with much less time and labor. These methods, using measurements of electrical resistances with nylon blocks, neutron scattering or gamma-ray attenuation will not be discussed here. Many specific references are available in the literature (ROSE, 1966; HILLEL, 1971).

3.2 The soil moisture potential

3.2.1 Introduction

The study of the occurrence and movement of water in the soil relies completely on the basic concepts of the soil moisture potential, which are essentially based on thermodynamic principles. Very thorough discussions are presented by ROSE (1966), HILLEL

(1971), CHILDS (1969), BAVER et al. (1972) and ISSS (1976). Any reader wishing to thoroughly understand these concepts is referred to these publications. However, application of simulation techniques often implies the use of simple expressions of the total potential in terms of gradients of the hydraulic head H. These can be explained in rather simple terms, which do, however, often omit details which may be quite relevant from a strictly physical point of view. The following summary should therefore be read and interpreted with some caution.

It is common experience that soil with a high clay content may feel dry to the touch while a sandy soil with the same water content may appear quite moist. This is due to the energy of retention of soil water being greater in the former than in the latter. This means that more mechanical work would have to be used to remove a small amount of water from the clayey soil than from the sandy soil at the same moisture content. Thus, it is essential to have information on the energy associated with soil water - whether we are interested in the availability of soil water for plant growth or in the flow of moisture. Many processes that involve water in soil and in plant systems may be dealt with through consideration of potential energy and, particularly, differences in potential energy.

Potential energy is the energy a body has by virtue of its *position* in a force field, and is determined by the force required to move a body directly against the force field and is, like any energy, the product of a force and a distance (dimension: $kg.m.s.^{-2}.m = kg.m^2.s^{-2}$). One basic law of thermodynamics states that energy cannot disappear. In other words, the energy used to transport the liquid is stored in the liquid in its new position as potential energy (we assume that the transport process as such did not consume nor yield any energy and that temperature is constant). This theoretical definition, which defines the potential of what may be a static quantity of water in terms of a hypothetical dynamic process, has occasionally offered conceptual problems. Examples will be used later to illustrate the theory. But first it is useful to recall that the potential energy of a body was defined as energy present by virtue of its position in a force field. Different force fields are acting upon the water in the soil, and they will be discussed separately.

New definitions (ISSS, 1976) provide more clarity in what used to be a rather complicated theoretical set of criteria. The total potential is now defined in terms of three *experimentally accessible parameters* which characterize the gravitational, the osmotic and the tensiometer-pressure potential (or briefly: pressure potential). Definition of the latter component potential appears to be quite helpful for improving the practical applicability of potential theory.

3.2.2 The gravitational potential (ψ_g or z)

Each body on the earth's surface is attracted towards the centre of the earth by a gravitational force equal to its weight. To raise the body against this attraction, work must be done, and this work is stored in the raised body as gravitational potential energy (Z) which is determined at each point by the elevation of the point relative to some arbitrary reference level. Therefore $Z = M.g.z.$ (dimension: $kg.m.s.^{-2}.m = kg.m^2.s^{-2}$) where Z is the gravitational potential energy of a mass m of water at a height z above a reference level and g = acceleration of gravity. This potential expressed per unit weight, becomes: z (dimension: m) ($kg.m^2.s^{-2}/kg.m.s^{-2}$).

The definition of the gravitational potential can be expressed per unit *mass* (ψ_g) and this is physically most acceptable (ISSS, 1976). However, expression in terms of unit

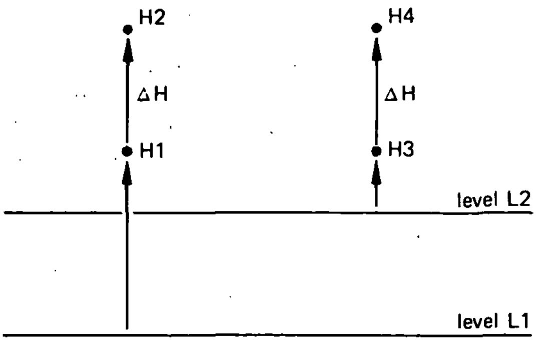


Fig. 4. The difference in gravitational potential between two points is identical for different reference levels.

weight is simple and attractive for applied work. Assume a reference level L_1 (Fig. 4) and the presence of one gramme of water at different heights H_1 cm and H_2 cm above that level. The gravitational potential (z) is now H_1 cm and H_2 cm respectively. The reference level, which is arbitrary, can be shifted to level L_2. Then the z potentials are H_3 and H_4 cm respectively. Still, the difference between the two (H_3 - H_4) is equal to the earlier difference when the reference level was at L_1 (H_2 - H_1) (Fig. 4). Differences in potential are generally more important than their absolute values when studying water movement in soils.

3.2.3 The tensiometer-pressure potential (pressure potential) (ψ_p or h)

Soil wetness refers solely to the total amount of liquid in a soil sample. In addition, it is important to ascertain the distribution of water in the soil at different moisture contents and to understand the natural laws that govern it. *As the moisture content of a soil sample decreases, water leaves the larger soil pores but remains in the finer ones.* This can be explained by considering the basic phenomena of liquid surface tension and capillarity.

Surface tension occurs typically at the interface of a liquid and a gas. Molecules in the liquid attract each other from all sides. In the surface areas the molecules are attracted into the denser liquid phase by a force greater than the force attracting them into the gaseous phase. The resulting force draws the surface molecules downward, which results in a tendency for the liquid surface to contract.

Capillarity refers to the well-known phenomenon of the rise of water into a capillary tube inserted in water, due to its surface tension (Fig. 5). The finer the tube, the higher the capillary rise and the greater the negative pressure below the water meniscus in the tube. The negative pressure (p) is a result of the curvature of the meniscus (α), which increases as tubes become smaller, and can be calculated (in Pa) as follows (assuming that the contact angle between water and tube is zero):

$$p = \frac{2\sigma}{r} \tag{3.1}$$

where; σ = surface tension of the water (N/m), and r = radius of the capillary (m).
The height of capillary rise (r) is

$$r = \frac{2\sigma}{\varrho g r} \tag{3.2}$$

where: ϱ = density of the water (kg/m^3) and g = gravitational constant (m/s^2).

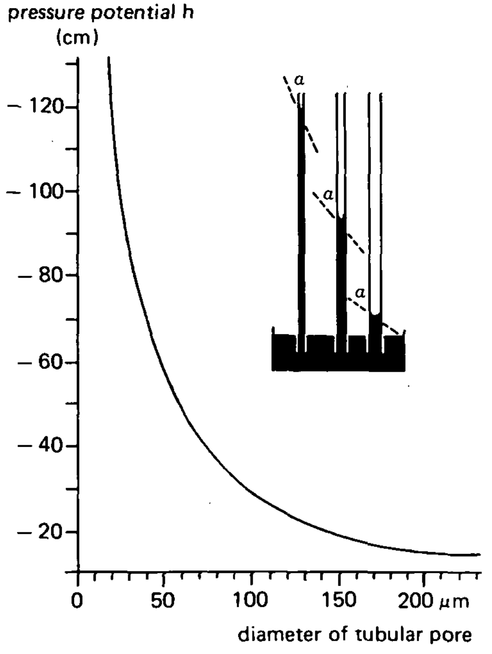


Fig. 5. The relationship between tubular pore size and corresponding pressure potential.

Function 3.1 can be pictured as a continuous graph, relating capillary diameter (2r) to corresponding negative pressure (Fig. 5). The negative pressure below the meniscus in the water (in Pa) can also be expressed in terms of the equivalent height of a column of water (cm) that can be "pulled" from a cup of water by the capillary tube (since 1 mbar = 100 Pa $\simeq$ 1 cm water). Figure 5 illustrates that fine pores can exercise a larger "pull" than large pores. For example, a cylindrical pore radius of 100 μm corresponds with a relatively low capillary rise of 28 cm water (pressure below meniscus = —28 cm water), a radius of 30 μm with a relatively high rise of 103 cm (pressure: —103 cm water). These figures also imply, of course, that it takes a larger force (more energy) to remove water from a small pore than from a large one. It is physically correct to define such conditions in terms of a negative pressure. However, the minus sign may be inconvenient and then conditions are often described in terms of "tensions" or "suctions". In other words, a moisture pressure of —20 cm water is equivalent to a "suction" or "tension" of +20 cm water. The negative notation is preferred and will generally be used in this text. To represent the porosity of a certain soil material as a bundle of capillaries with a characteristic size range is, of course, unrealistic because real pores in the soil have a much more complex configuration (section 2.1). This representation can nevertheless be helpful for visualizing the energy condition of water in soil and to explain flow phenomena.

Returning now to the discussion of the tensiometer-pressure potential, we have seen that the pressure in the soil water in unsaturated soil is less than that of the local atmosphere (see Fig. 5). The pressure reduction is associated with the location of water in the fine pores and the associated potential, which can be measured with tensiometers, is called the tensiometer-pressure potential ψ_p, which is defined per unit *mass* of pure water (ISSS, 1976).

The definition of this potential, in terms of pressure equivalent, is as follows: "The tensiometer-pressure p is the gauge pressure (in Pa or mbar) relative to atmospheric pressure, to which a sample of the soil solution (with identical pressure and temperature) must be subjected to be in equilibrium via a membrane impermeable to the soil matrix with the soil water at the point under consideration" (ISSS, 1976). For applied studies it is again useful to define the pressure potential in terms of a "pressure head" h, which is the pressure potential per unit weight (assuming ϱ to be constant) as follows:

$$h = \frac{p}{\varrho . g} = \frac{\varrho . g . h}{\varrho . g} \text{(dimension: } \frac{kg.m^{-3}.m.s^{-2}.m}{kg.m^{-3}.m.s^{-2}} = m).$$

These definitions imply the use of a tensiometer. A description and discussion of this instrument is useful not only because of its significance in soil physics but also because it allows a further discussion on the concept of pressure potential.

3.2.4 Tensiometers

Tensiometers measure the magnitude of the pressure potential h. The simplest form as shown in Figure 6A, consists of a porous cup, mostly of a ceramic material connected with a water-filled tube which has a U-shape and is open at the end (manometer). When the cup is placed in the soil, the bulk water inside the cup comes into hydraulic contact with the liquid phase in the soil through the water-filled small pores in the ceramic walls. When initially placed in the soil, the water in the tensiometer is at atmospheric pressure. Soil water in unsaturated soil has a negative pressure and therefore exercises a suction which draws out a certain amount of water from the rigid and air-tight tensiometer, thus causing a drop in the water level at the open end of the U-tube. The drier the soil, the higher the suction that the soil exercises on the water in the porous cup and the lower the water level at equilibrium in the U-tube. The

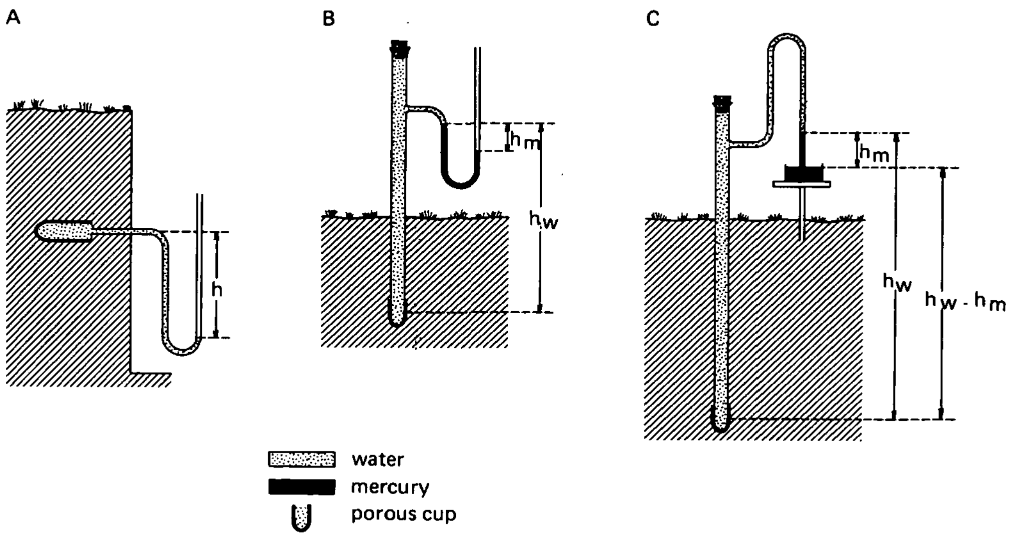


Fig. 6. Three types of tensiometers for measuring pressure potentials.

height of the liquid column that has moved into (has been "sucked into") the soil (h cm in Fig. 6A) is therefore an index of the magnitude of the potential.

This type of tensiometer is very simple and useful to illustrate the basic principles involved. However, practical applications often do not allow the use of the water manometer because the U-tube extends below the level of the tensiometer cup and measurements thus require inconvenient, deep pits. With mercury manometers these problems do not arise. Two common examples are shown in Figure 6B and C. Pressures can be expressed as before. For conditions pictured in Figure 6B (ROSE, 1966) where the continuous water-body in the tensiometer functions as a "link" between the water in the soil and the mercury, the pressure potential in soil surrounding the porous cup is equal to:

$$h = \frac{\varrho_m \cdot g \cdot h_m}{\varrho_m \cdot g} + \frac{\varrho_m \cdot g \cdot h}{\varrho_w \cdot g}$$

where: ϱ_m = density of mercury (13.55 g/cm³).
Using mercury implies that a relatively short height indicates a relatively large pressure difference in the manometer (1 cm of mercury corresponds to 13.55 cm of water). Assume that $h_m = 10$ cm and $h_w = 100$ cm. The pressure potential (h) of the soil water surrounding the porous cup is then: $(-10 \times 13.55) + 100 = -35.5$ cm (or the soil water tension or suction $= +35.5$ cm).

The arrangement shown in Figure 6C is often very convenient because the mercury level in the cup can be used as a fixed reference. The pressure potential (h) measured for the conditions shown in Figure 6C, is identical to that calculated for Figure 6B! However, by calculating $(-h_m \times 12.55) + (h_w - h_m)$ the fixed value of $h_w - h_m$ is introduced in the calculation and then varying values of h_m can more easily be transformed into pressure potentials. For the numerical example based on Figure 6B it follows:
$h = (-10 \times 12.55) + 90 = -35.5$ cm.

Specific limitations do exist when using tensiometers and their performance and possible malfunctioning may be explained by a further analysis of the system:

(i) Water within the tensiometer should be continuous throughout the system to allow a correct transfer of pressure from the soil water to the mercury. Occurrence of gas-bubbles disrupts this continuity and makes the system inoperative. The fine porous cup has the function of *not* allowing penetration of air from the unsaturated soil into the water-filled tensiometer tube, even though water can and should move through it. The fine pores inside the wall of the ceramic cup have a high "air-entry value", which is the pressure needed to remove the water from the pores in the cup, replacing it by air. As discussed, fine pores exercise a relatively high capillary pressure (section 3.2.3) and a fine-porous water-filled disc has therefore a high "air-entry value". Even without such air entry, breakdown of the system occurs due to entrapped air within the tensiometer tube or to air coming out of solution at the reduced pressure. Whatever the cause, liquid continuity fails when negative pressures exceed about 850 cm of water or perhaps lower, depending on the experimental arrangement. Use of tensiometers in the field is therefore only possible when pressures do not exceed this value. Care should be taken to fill the tensiometers with air-free water to retard formation of air-bubbles within the system.

(ii) The measurement of pressure potentials, as discussed, implies movement of water from the tensiometer into the soil. However, the purpose of the measurement is to characterize the existing moisture potential in the soil. This potential is changed (increased) when water moves from the cup into the soil. The observer has to wait until this added water has flowed away in the soil and the original moisture condition is re-established. Only then does the tensiometer indicate the potential that really existed. This may take a considerable time in slowly permeable soil (KLUTE and GARDNER, 1962; BAKKER, 1975). In this context the type of tensiometer used is also very important. The use of a water manometer would be very unwise because a relatively high quantity of water must flow into the soil before equilibrium occurs. The mercury manometers are much better, certainly when small diameter plastic tubing is used. Still, often a considerable volume of water must be absorbed by the soil. A very convenient modern device, the electronic transducer, can be used which reacts to very small changes in pressure and converts these changes in a small electrical current which can be registered and amplified by a voltmeter (KLUTE and PETERS, 1962, 1966). This system is very accurate but also very sensitive to, for example, the occurrence of small air bubbles in the tensiometer system. One major problem in this context has yet to be discussed. The filling of the tensiometer, when done in situ after the porous cup has already been installed, may result in large additions of water to the soil before the system is closed off and the mercury (or the transducer) will start to respond to the potential which exists at the interface of the porous cup and the soil. Under these conditions a transducer will promptly report the gradual decrease of the potential that occurs as all the added water moves away in the soil until the original potential is reached again. This may take a long time, and if the initial moisture content was quite low, this point may not be reached within an acceptable period. If possible, tensiometers should therefore be filled outside the soil to be placed *in situ* when they are entirely filled with water. Otherwise, filling should be done very quickly to restrict water movement into the soil.

(iii) Contact between the porous cup and the surrounding soil is essential for proper functioning of the tensiometer. Generally, the porous cup is pushed into a hole with a slightly smaller diameter to ensure good contact. If the soil is initially rather dry and hard, pre-wetting of the hole may be necessary (BOUMA et al., 1974a, BAKER et al., 1974). Stony soil cannot be tested without some special measures. A small excavation should be made and filled with very fine sand into which the tensiometer can be placed.

3.2.5 Other potentials

The potential energy of a body was defined as energy present by virtue of its presence in a force field. Gravitational forces and forces which can be measured with a tensiometer have been discussed so far and they are the most important when water movement under field conditions is considered in non-saline soils. The effect of solutes on the total potential of soil water, expressed by the osmotic potential ψ_o, becomes of primary significance if the water is separated by a membrane whose permeability to water molecules differs from that to solute. This potential is of importance in water movement into and through plant roots, in which there are layers of cells which exhibit different permeabilities to solvent and solute. For further discussions the reader is referred to ISSS (1976).

Some final remarks must be made regarding the pressure potential ψ_p. This potential characterizes the state of water in the soil (together with ψ_g and ψ_o) including the effect of the presence of an external gas pressure different from atmospheric and/or the presence of a mechanical envelope pressure (also called an overburden pressure).

The pressure potential has some sub-components (ISSS, 1976). The *pneumatic potential* ($\psi_p{}^a$) is due to excess gas pressure; the *envelope-pressure* potential ($\psi_p{}^e$), or *overburden potential,* is due to matrix pressures, for example in swelling soils. Using the two subcomponents, which can often *not* be measured specifically, two additional potentials are defined. These are: (i) the *matric potential,* which is a pressure potential (ψ_p) with a given envelope potential but with $\psi_p{}^a = o$, and (ii) the *wetness potential,* which is a pressure potential (ψ_p) without envelope or pneumatic potential ($\psi_p{}^a = \psi_p{}^e = o$). However, the pressure reading of a tensiometer installed in situ should be seen as one of three parameters (together with ψ_o and ψ_g) characterizing fully the state of water in soil under the prevailing conditions. Their gradients are the basis for transport theory (ISSS, 1976). As stated, expresssion per unit quantity weight is useful for applied work because this results in simple expressions (see section 3.2.6).

3.2.6 The hydraulic head (H)

The hydraulic head, previously called the hydraulic potential (ROSE, 1966), is the sum of the gravitational and pressure potentials (expressed per unit quantity weight as defined earlier):

$$H = z + h$$

where: H = hydraulic head, z = the height of the point under consideration above a reference level and h = the height of a vertical water column which would exert a pressure at its base numerically equal to the soil-water pressure (negative in unsaturated soil, zero or positive in saturated soil). Gas pressures in field soils are generally atmospheric and are reflected in h, as discussed, because h is measured by tensiometry.

The hydraulic head is a very convenient concept for describing water movement and will be frequently referred to in the remainder of this text. Figure 7 illustrates hydraulic heads as they occur during static equilibrium in a 60 cm high soil core.

3.3 The moisture retention curve, the soil moisture characteristic or the retentivity curve

In a saturated soil at equilibrium with free water at the same elevation, the actual pressure is atmospheric and the soil liquid pressure is zero. If a slight negative pressure

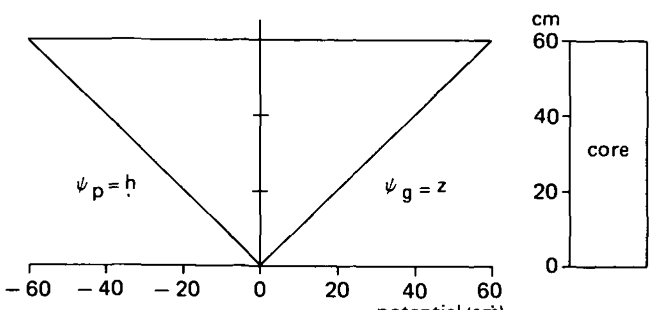


Fig. 7. Static hydraulic equilibrium (H = 0 at all points) in a drained soil core with a height of 60 cm.

is applied to this water, no outflow may occur until, as the pressure is further decreased, a certain critical value is exceeded at which the largest pore begins to empty. This critical pressure, already mentioned in section 3.2.4, is called the air-entry value. This value is generally very small in coarse textured or in well-aggregated soils where large pores will immediately lose their water when only a very small negative pressure is applied. However, non-aggregated fine sandy or loamy soils with few or no larger channels may exhibit air-entry phenomena. As the pressure is decreased (suction increased) more water is drawn out of the soil and more of the relatively large pores, which cannot exercise an adequate capillary force to retain their water against the pressure applied, will empty (see section 3.2.3). A gradual decrease in pressure will result in the emptying of progressively smaller pores until, at high negative pressures, only the very narrow pores retain water. A decreasing pressure is thus associated with decreasing soil wetness. The rate of decrease is characteristic for any particular porous medium because it is a function of the pore-size distribution. The curve which shows the relationship between the negative pressure and the water content is a very important soil physical characteristic and is known as the *soil moisture retention curve,* the *soil moisture characteristic* or the *liquid retentivity curve* (ISSS, 1976). The first two terms are frequently used, the latter was recently proposed. Examples for a number of soils are presented in Figure 8 (HAANS and VAN DER SLUIJS, 1970).

Retentivity curves have been determined for many years. Before discussing methodology it is necessary to review the formal definitions of ISSS (1976). The water retentivity curve is defined as the curve relating the matric or wetness potential (expressed per unit volume in Pa or mbar) to the volume fraction θ of the soil liquid phase. As discussed in section 3.2.5, the matric potential *includes* the overburden potential, which is not part of the wetness potential. Traditionally, retentivity curves have been determined on small samples for which the pneumatic and the overburden potentials (the latter in swelling soils) are low. Therefore it seems justified to define the large body of existing retentivity curves (pF curves) in terms of curves relating the pressure potential (h) to θ (moisture content). Whether such small samples are representative of soil in the field, particularly in swelling soils, is a realistic question which cannot be solved here (ISSS, 1976).

In attempting to express the negative pressure potential of soil water in terms of an equivalent hydraulic head, we must contend with the fact that this head may be as

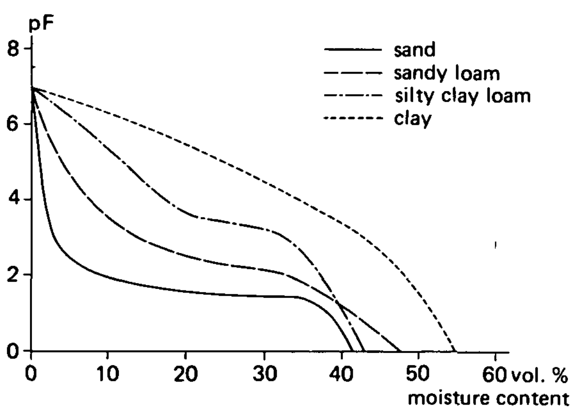


Fig. 8. Examples of pF curves of different soil materials (from HAANS and VAN DER SLUIJS, 1970).

much as — 10 000 cm or even — 100 000 cm of water. To avoid the use of such cumbersomely large numbers, the negative logarithm of the equivalent hydraulic head may be used instead. This value is called the pF. The pF concept is widely used but has not recently been included in soil physics terminology (ISSS, 1976) because use of equivalent hydraulic heads has the disadvantage that the variable accelleration of gravity (g) is included. A pF of 1.0 represents a pressure of —10 cm H_2O; a pF of 3 a pressure of —1000 cm, etc. The curves of Figure 8 demonstrate the effect of pore-size distributions on the shape of the pF curve. In sandy soil, most of the pores are relatively large and once these pores are emptied at a given pressure, only a small amount of liquid remains. The greater the clay content and the amount of fine pores, the greater the moisture content at any particular pressure and the more gradual the slope of the curve.

These differences are schematically illustrated in Figure 9 for a "sand", a "sandy loam" and a "clay" soil, which are represented as a set of interconnected capillary tubes. At a low pressure of —30 cm (pF 1.5) relatively large pores in the "sand" lose their water, whereas the finer pores in the "sandy loam" and "clay" remain filled. At a lower pressure of —100 cm (pF 2.0), most of the water in the "sand" is gone, but the moisture content of the "sandy loam" has decreased somewhat and the "clay" has lost no water. Figure 10 illustrates the principle of the method for determination of the retention curve for low pF values. Samples are placed on a porous plate (or on a fine porous sand-bed), which is subjected to varying negative pressures. The air-entry value of the porous plate or the sand should not be exceeded to assure adequate liquid contact and thus the possibility for flow. Very fine sand can be used up to pressures corresponding with pF 2.0. Kaoline clay is used for pressures up to pF 2.7. With an arrangement as shown in Figure 10, the soil sample — or a part of it — does not have to be removed for determination of the water content at each equilibrium suction (Baker et al., 1974). Evaporative losses are avoided in this system and the contact between the soil sample and the porous disc is maintained during the entire experiment. Repeated removal of samples for weighing, as occurring in other experimental systems, may result in a loss of capillary contact between water in the soil and in the porous plate (or sand), certainly at lower pressures and this may lead to erroneous measurements.

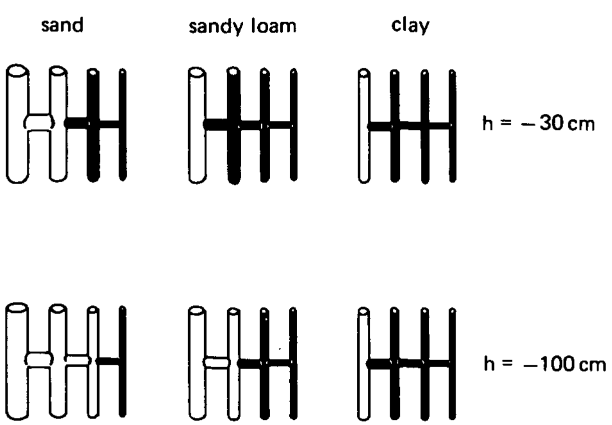


Fig. 9. Moisture retention in three schematic soil materials at pressure potentials of —30 and —100 cm water (pF 1.5 and 2.0 resp.).

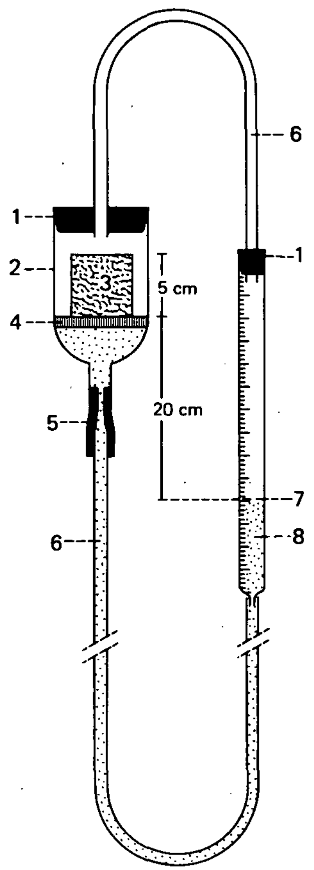
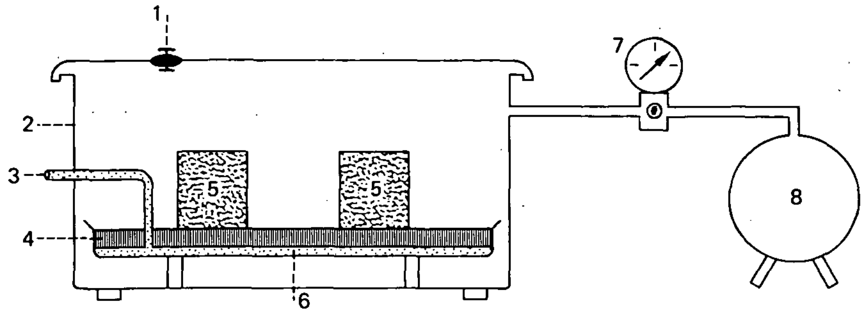


Fig. 10. Schematic diagram of apparatus for moisture retention measurements at high pressure potential (after BAKER et al., 1974).

An attractive procedure for obtaining a moisture retention curve, uses in situ tensiometry and moisture content measurements by neutron probe or otherwise. Values thus obtained were compared with those determined on cores in the laboratory (BAKER et al., 1974). Results were only satisfactory when large cores were used for the laboratory determinations. These data are important and encouraging because they relate to swelling soils. In situ measured moisture potentials include the overburden potential, which will not be very relevant for the laboratory samples. More studies of this kind are needed to define retentivity curves for field soils.

Moisture contents at pressures lower than about —500 cm (pF 2.7) must be determined in a different manner because application of negative pressure is not easily possible due to air coming out of solution at the reduced pressure. Pressure devices are then used to obtain water contents. Rather than "pull" the water out, it is "pushed" out. Air pressure is applied to samples, which are placed on a very fine porous ceramic plate, the pores of which are filled with water. A water-filled space below the plate is connected with the outside atmosphere (Fig. 11). Pressures are expressed in terms of the equivalent height of a vertical water column (cm) (see also previous sections 3.2.3 and 3.2.5).

The differential water capacity of a soil (C_w or C_θ) is the rate of change of w (or



1 = safety pressure release valve
2 = extractor
3 = outflow tube
4 = porous ceramic plate
5 = soil sample
6 = neoprene diaphragm
7 = pressure regulator
8 = compressor

Fig. 11. Schematic laboratory setup of a pressure-plate extractor for moisture retention measurements in the range 0.1 - 15 bar.

θ) with the soil matric potential, expressed per unit volume (p^m), to be specified in Pa^{-1} or $mbar^{-1}$ (ISSS, 1976). This definition implies that there is no pneumatic potential due to compressed soil air; otherwise the water capacity would have been defined in terms of the pressure potential. The overburden potential is included but is absent in non-swelling soils and should be specified for swelling soils. If the latter is not done, this will imply that the retentivity curve was obtained at zero envelope pressure. Then the differential water capacity is a function of the wetness potential (see previous section 3.2.5). Simplification is obtained by considering the water capacity as the slope of the retentivity curve: $C_\theta = d\theta/dh$, (dimension: m^{-1}) (see section 4.1).

An example follows (Table 2), which illustrates the measurement of a moisture retention curve. Data were obtained using large clods taken in the C horizon (clay) of a Dutch Alluvial Soil (Ooivaaggrond according to the Dutch system of soil classification, Fluvaquentic Eutrochrept according to the American system). The field-moist clods were coated with a SARAN solution in aceton, which leaves (after evaporation of the acetone) a thin flexible skin around the clod permeable to vapour but not to liquid (GROSSMAN et al., 1968). The skin expands with the clod upon wetting, which occurs through one flattened side of the clod from which the skin has been removed, to allow good contact with water. The skin contracts when the clod shrinks as water is removed by moisture extraction, for example on an apparatus shown in Figure 10. Weighing under water allows a determination of clod volume, which can be related to clod weight

Table 2. Basic data for calculation of two points of the moisture retention curve using the Saran method.

Fieldweight (g)	Fieldweight plus plastic	Weight at saturation	Vol. (cm^3)	Weight at pF 2.0	Vol. (cm^3)	Weight 105° C	Vol. (cm^3)
565.2	580.5	656.5	367	626.7	363	446.0	243

(corrected for the weight of the plastic) to obtain a pF curve. Weighing under water requires that the skin is re-applied to the flattened side of the clod, to be removed again afterwards before the clod is equilibrated at another (lower) pressure potential. Use of the SARAN method is more satisfactory for swelling soils with clay than methods using sampling cylinders with a fixed volume. The example is presented for one clod and for only two moisture contents, corresponding with saturation and a moisture potential of —100 cm (pF 2.0) reached by desorbing the saturated clod (Fig. 10).

The calculation first determines the weight of the plastic, which is 580,5 — 565,2 = 15,3 g. Weight at 105° C = 446,0 — 15,3 = 450,7 g. The measured volumes of the soil are obtained by subtracting the volume of the SARAN from the volume of the SARAN coated clod (density SARAN = 1.4 g/cm³). Here 15.3 g = 10.9 cm³. The volume of the stove-dry soil is 243.0 cm³ — 10.9 cm³ = 232.1 cm³. At saturation and at pF 2.0 volumes are 356.1 cm³ and 352.1 cm³ respectively. Three "bulk densities" can be calculated, based on the one weight and the three volumes. These values are 1.95 g/cm³, 1.26 g/cm³ and 1.28 g/cm³ respectively. The first value is not a realistic field value, but the other two values differ due to differing volumes. Such differences are ignored if one sample with a fixed volume (of usually 100 cm³) in a rigid cylinder is subjected to saturation and desorption. The moisture content (w) at saturation = (641.2 — 450.7) / 4.507 = 42.3 %. The moisture content θ : 42.3 x 1.26 = 53.3 % by volume. For pF 2.0 these two values are 35.6 % and 45.6 % respectively.

One problem using the SARAN method should be recognized. Samples, even when taken in small cylinders, can be considered to be sampled at random, whereas sampling of natural clods, whatever their size, may represent a somewhat biased sample, in that interaggregate pores are underestimated.

Results obtained with the SARAN method can be used to calculate Coefficients Of Linear Extensibility (COLE) as follows (GROSSMAN et al., 1968).

The COLE is defined as: $\dfrac{L_m - L_d}{L_d} = \dfrac{L_m}{L_d} - 1.$

where: L_m = length of the moist (1/3 bar = pF 2.5) samples and L_d = length of the stove-dry sample.

Assuming that swelling is identical in all directions, we find for irregularly shaped clods:

$$\text{COLE} = \sqrt[3]{\frac{L^3_m}{L^3_d}} - 1 = \sqrt[3]{\frac{V_m}{V_d}} - 1.$$

where: V_m = volume of moist sample and V_d = volume of stove-dry sample. It is practical to define Linear Extensibility values (LE) also for saturation and other pressure potentials than 1/3 bar (pF 2.5). However, to avoid confusion, this type of use should always be clearly identified. For example: $LE_{(sat)}$; $LE_{(pF2.0)}$ etc.

3.4 Hysteresis

Unfortunately, the moisture content is not a single function of the pressure potential. The moisture content corresponding with a certain pressure is higher when it has been reached by desorption of an initially wetter sample as compared with the moisture content reached by wetting (adsorption) of an initially drier sample. One could say

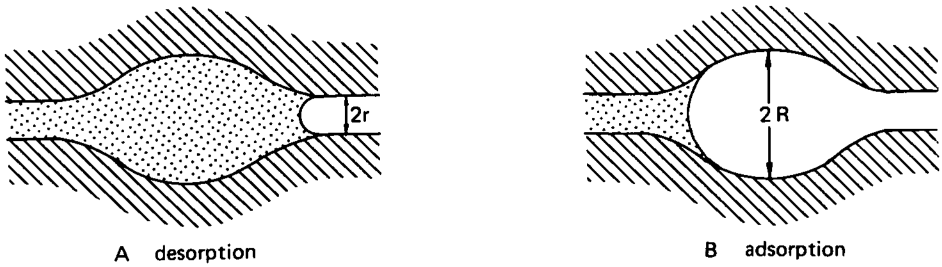


Fig. 12. Cross-section through an idealized void illustrating the hysteresis phenomenon (after ROSE, 1966).

that it takes more energy (it is more difficult) to get water out of the soil once it is in, than to get it back in, once it is out. The phenomenon of hysteresis can be illustrated by considering a schematic void partly filled with water (Fig. 12). The water-filled idealized void (A) will drain in the course of a desorption proces if the negative pressure exceeds the relatively large capillary force corresponding with the *smallest* pore diameter (2r) in the system. The idealized air-filled void (B) will *fill* with water in the course of an adsorption process as soon as the relatively small capillary force, corresponding with the largest pore diameter (2R) is sufficiently strong to pull the water in. This comparison shows that the water content of a soil at a given moisture tension will be greater following desorption than following adsorption (see Fig. 13 for an example in sand). The moisture retention curve is usually determined by a desorption process; starting with saturated soil. Values thus obtained may not apply to water contents and moisture potentials occurring in the field when an initially dry soil is wetting up, for example in the rainy season. A physically correct characterization of the hysteresis phenomenon is very complicated (GILLHAM et al., 1976).

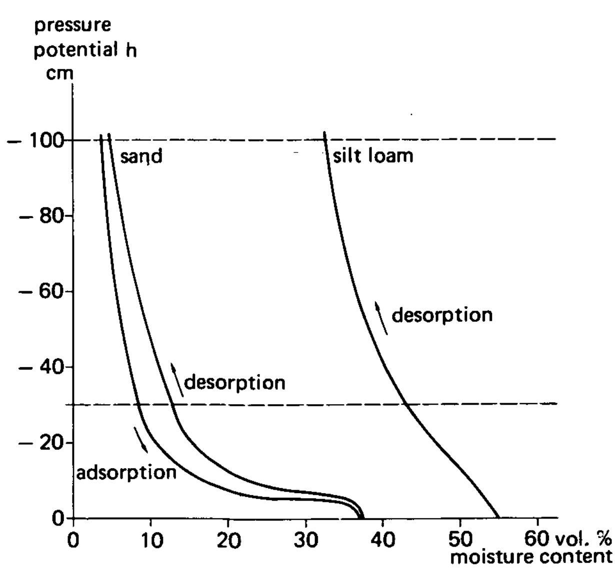


Fig. 13. Adsorption and desorption curves for a coarse sand and a desorption curve for a silt loam.

3.5 The equivalent pore-size distribution

If a decrease in the soil water pressure from h_1 to h_2 cm results in the withdrawal of a volume V_{12} cm^3 of water, then we may say that the volume V_{12} was contained in capillary pores with an "effective" radius between r_1 and r_2, where h_1 and r_1 and h_2 and r_2 are related by equation 3.1 (Fig. 5). The real pores are, of course, not tubular capillaries, but the release of a certain volume of water within a pressure range can be "translated" into a release from a recognizable type of pore which has a known relationship between size and corresponding pressure. The concept of the equivalent pore-size distribution is often used in the soil science literature. The procedure will be illustrated using Figure 13 where two moisture retention curves are given. The volumes of water released in both media between pressures of —30 and —100 cm are 8 % and 12 % for the sand and the silt loam respectively. Translated into equivalent pore sizes, we can say that pore sizes (2r) between 93 and 28 micron constitute 8 vol % and 12 vol % respectively in both media. These values compare favourably with those obtained by micromorphological analysis of thin sections using point counts (BOUMA and DENNING, 1974). *This, however, is only true for sands.* A pF curve for a coarse sand was derived from morphometric point counts using equation 3.1 to "translate" measured pore sizes into corresponding pressure potentials, and was compared with a measured curve (Fig. 14). Good agreement was found between the morphometric curve and the physical curve, *but only if the latter was determined by adsorption* (see section 3.4). Agreement with the physical curve, determined by desorption, was not as good.

Poor agreement between morphometric pore-size distributions (whether measured by point counts or with automatic equipment: the Quantimet: ISMAIL, 1975) and those derived from desorption curves is due to the effects of "pore necks" in the porous medium. More specifically, the relatively good agreement obtained when the comparison is made with *adsorption* data, points to the importance of the hysteresis phenomenon. Real pore sizes, which are made visible and which are accurately measured with morphometric techniques, can only be estimated in sandy materials by means of physical techniques if adsorption curves are used. Use of desorption curves results in erroneous estimates of real pore-size distributions in all soils.

Some additional problems in determining "equivalent" pore-size distributions using physical data, will be discussed in the following three points:

(i) Clayey samples may lose water even though no pores become filled with air. The latter may be due to shrinkage of the sample, whereby the volume of the released water is equal to the reduction in soil volume. The model of equivalent pore size obviously does not apply here. This aspect is particularly relevant for unripened soils.

(ii) Structured clayey soils usually have planar voids, which have a different relationship between size and pressure. The capillary model is then irrelevant and the term "equivalent" has much less meaning than if applied to a sand or sandy soil with pores that look more like capillaries.

(iii) Large mistakes can be made when large pore sizes are related to low suctions. Assume that a negative pressure of 10 cm (pF 1.0) is applied to a 5 cm high sample. The pressure on top of the sample is then —15 cm (equivalent pore diameter: 186 μm) and at the bottom —5 cm (diam. 560 μm). Which is now the equivalent pore diameter to be derived from the overall sample which will give one "average" pressure of —10

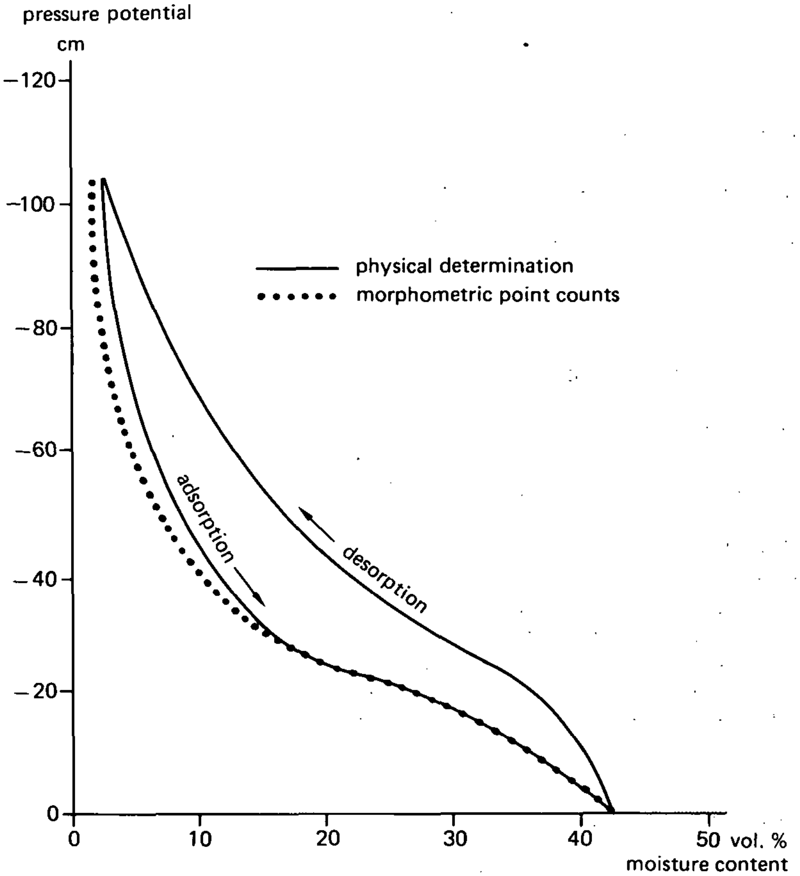


Fig. 14. Moisture retention curves for a sand, determined with physical techniques, by adsorption and desorption, and calculated with morphometric point counts (after BOUMA and DENNING, 1974).

cm (pore diameter: 280 μm)? The relative error in pressure is 50 % and in pore diameter 33 %. The error is much less at lower pressures and smaller equivalent pore sizes. The relative error should always be estimated before equivalent pore-size distributions are determined and reported.

This review clearly indicates the many problems associated with estimating a pore-size distribution from moisture retention data. In any case it is important to refer always to *"equivalent* pore-size distributions" when such calculations are based on moisture retention data. References to "pore-size distributions" are clearly unrealistic. These can only be determined by morphometric techniques. However, observations are then restricted to pores with a diameter of more than 30 μm (ISMAIL, 1975).

4. WATER MOVEMENT IN SOIL

4.1 Flow theory: Darcy's law and some flow equations

There is a considerable amount of experimental evidence to show that, whatever the state of saturation, the volume flux is proportional to the gradient of the hydraulic potential in the direction of the flux. This, basically, is Darcy's law:

$$v = -K \cdot \frac{dH}{dx} \tag{4.1}$$

where: v = flux, which is the volume of water, crossing unit area per unit time ($m^3.m^{-2}.s^{-1} = m.s^{-1}$) (negative because it is in the direction of decreasing potential) dH = difference in hydraulic potential and dx = difference in distance over which the difference in hydraulic potential is measured. K is called the hydraulic conductivity, which is the flux resulting from unit gradient in hydraulic potential (dH/dx = 1), and which has the dimension of m/s (often reported in cm/day).

Knowing the flux of water is often of great practical interest, whether dealing with downward infiltration or capillary rise. Equation 4.1 shows that the hydraulic conductivity (K) is a physically well defined flux. In other words, many fluxes may exist in the soil at any given time with varying gradients of the hydraulic head. However, there is only one K value at any particular pressure potential. This aspect, which defines it as a characteristic soil constant, can be more clearly illustrated by splitting the hydraulic head into its two component potentials: the pressure potential h and the gravitational potential z.

It follows for horizontal flow: $v = -K \cdot \dfrac{dh}{dx}$

for vertical flow: $v = -K \cdot \dfrac{d(z+h)}{dz}$

$$\text{or: } v = -K \left(1 + \frac{dh}{dz}\right). \tag{4.2}$$

This important expression will be referred to frequently in the following sub-chapters. Its usefulness is somewhat limited to conditions of steady state, that is, conditions where the flow rate does not change with time or changes very slowly, so that different consecutive semi-stable phases can be distinguished. The significance of the hydraulic conductivity as a characteristic hydraulic constant is illustrated in Figure 15 which presents six hypothetical examples of both horizontal and vertical steady flow conditions at constant gradients. Fluxes differ at the centre point of each of the six lines, because of different hydraulic gradients. But the K value at these centre points is the same for all examples.

Use of negative and positive signs may be confusing, even more so since use is not consistent in the literature. A simple procedure can be helpful, particularly for vertical steady flows (Fig. 15). The flux v is determined by the hydraulic or pressure gradient since K = 1 cm/day at h = —30 cm. Arrows can be used to indicate the direction of movement from high to low potential. This is very simple for horizontal flow (z =

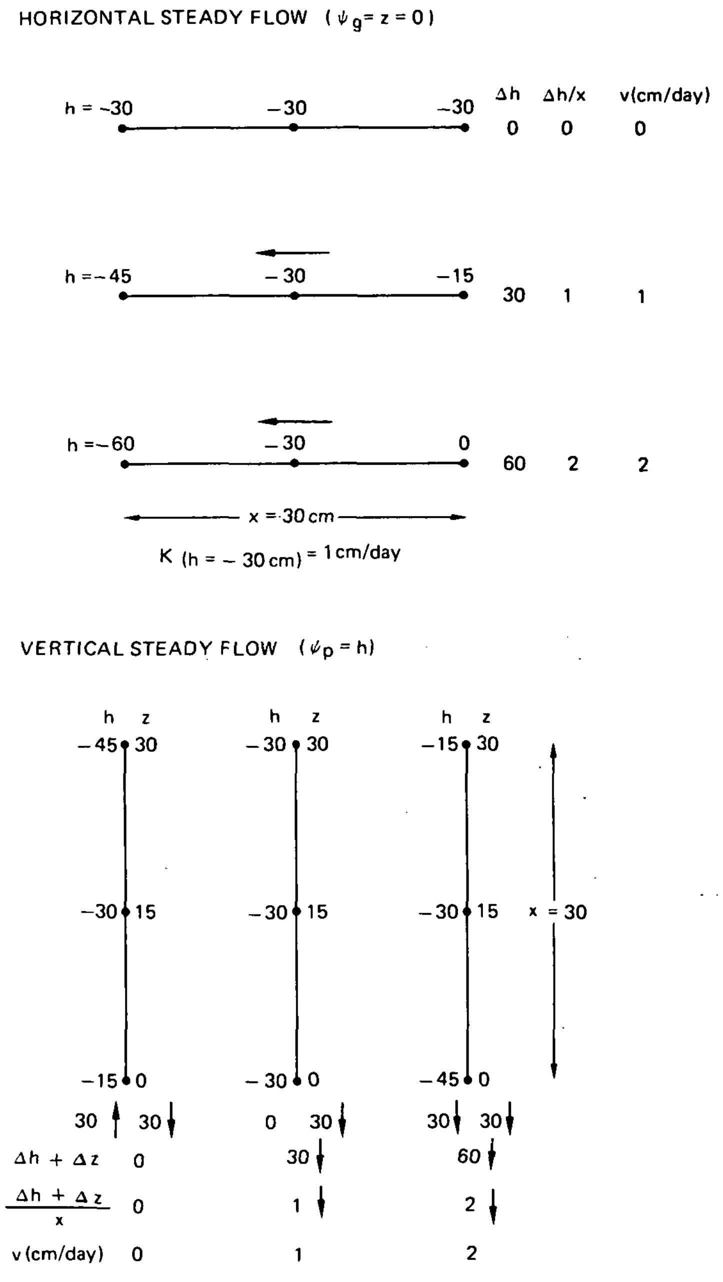


Fig. 15. Schematic diagram illustrating the importance of the hydraulic conductivity K as a hydraulic constant. The pressure potential at the center point of each of the six lines is identical at h = —30 cm, but the flux v is different due to different potential gradients, even though K (1 cm/day) is constant at a pressure potential of —30 cm.

0). For vertical flow, h (negative) and z (positive) their differences over 30 cm and their direction from high to low can be noted separately. If both arrows point the same way, Δh and Δz values can be added. If not, they must be subtracted and size and direction of the remaining hydraulic gradient should be obvious (Fig. 15).

Other mathematical expressions have been developed that are useful for characterizing moisture systems that do change with time. Of course, every soil moisture system in nature belongs to that category. The additional expressions are presented for the sake of completeness, the mathematically inclined reader is referred to the literature (Rose, 1966; Childs, 1969).

Modification 1. To obtain an expression for a changing water content, the $\delta h/\delta z$ quotient is multiplied by $\delta\theta/\delta\theta$, where θ = moisture content by volume. This is a trick, which allows the definition of the characteristic D (soil water diffusivity). This diffusivity is needed to describe problems where water contents are changing.

$$v = -K\left(1 + \frac{\delta h}{\delta z}\right) = -K\left(1 + \frac{\delta h}{\delta z} \cdot \frac{\delta \theta}{\delta \theta}\right) \tag{4.3}$$

$$= -K - D\frac{\delta\theta}{\delta z}$$

where: $D = K\dfrac{\delta h}{\delta\theta}$ ($m^2.s^{-1}$).

The magnitude of $\delta h/\delta\theta$ can be determined as the slope of the moisture retention curve. The water capacity C was defined as $\delta\theta/\delta h$ (section 3.3) so it follows that $D = K/C$.

Modification 2. Often what is measured experimentally is not a flux but changes in volumetric water content (θ) as a function of time at various depths. An expression for $\delta\theta/\delta t$ can be obtained by simple differentiation of equation 4.3 with respect to z as follows:

$$\frac{\delta v}{\delta z} = -\frac{\delta K}{\delta z} - \frac{\delta}{\delta z}\left(D \cdot \frac{\delta\theta}{\delta z}\right). \tag{4.4}$$

The change from $\delta v/\delta x$ to the required $\delta\theta/\delta z$ can be derived from moisture conservation considerations. The change of water content of any elementary volume of soil must be equal to the net flow across the boundaries of this volume. This requirement can be expressed as follows:

$$\frac{\delta\theta}{\delta t} = -\frac{\delta v}{\delta z}.$$

Substitution in 4.4 results in:

$$\frac{\delta\theta}{\delta t} = \frac{\delta}{\delta z}\left(D\frac{\delta\theta}{\delta z}\right) + \frac{\delta K}{\delta z}. \tag{4.5}$$

Equations 4.1 through 4.5 are needed to describe water movement through soils in physical terms. The derivation of these expressions is not complicated. However, solving them is quite another matter. Only few analytical solutions are available. Generally, numerical methods must be used to solve the equations for specific boundary conditions. Lately, computer methods used in simulation have been very successfully applied (e.g. DE LAAT et al., 1975; WIND and VAN DOORNE, 1975; DE WIT and VAN KEULEN, 1972). These equations are included in this text without further elaboration to allow their recognition in articles in the literature which describe water movement in soil, also those using simulation techniques.

4.2 A further discussion of the hydraulic conductivity

The abstract, theoretical discussion of the hydraulic conductivity in section 4.1 must now be translated into a format that can be related to water movement as it is recognized by the soil scientist in the field. Flow of water obviously occurs through the voids in the soil. An analysis of relationships, if any, between the flow rate and soil pore geometry is therefore relevant as an attempt to relate the flow rate to visible soil features.

Physical equations have been developed for certain types of pores to relate pore sizes to flow rates at a given hydraulic-head gradient (CHILDS, 1969). For a cylindrical pore of radius r, we find

$$Q/t = \frac{\pi g \varrho r^4}{8\eta} \text{ . grad H.} \tag{4.6}$$

For a plane slit of width D, and unit length,

$$Q/t = \frac{g \varrho D^3}{12\eta} \text{ . grad H.} \tag{4.7}$$

where: Q/t = volume of liquid (m^3) flowing through the tube per second ($m^3.s^{-1}$), ϱ = density of water ($kg.m^{-3}$); g = gravitational constant ($m.s^{-2}$); η = viscosity ($N.m^{-1}$); grad H = hydraulic gradient ($m.m^{-1}$).

These equations are graphically expressed in Figure 16, demonstrating the significant effect of pore size on flow rates. For example, these graphs show that a tubular (cylindrical) pore with a diameter of 100 μm will conduct about 2 cm^3/day at a gradient of 1 cm/cm (or 4 x 2 = 8 cm^3/day at a gradient of 4 cm/cm), since v is equal to K at a gradient of 1 cm/cm. A plane slit with a width of 100 μm (and unit length) will conduct 700 cm^3/day. A plane slit with a length of 4 cm will conduct 8400 cm^3/day if the gradient is 3 cm/cm (12 x 700 cm^3/day). The determination of hydraulic conductivity would be a very simple matter if soil pores were continuous cylindrical tubes or planar slits with known dimensions. Real soil pores are, of course, very irregular in shape (see Chapter 2) and schematization of soil pores in terms of tubes and plane slits represents a very drastic approach. Nevertheless, such representations may be useful to illustrate flow phenomena and the approach will be used here with that purpose in mind.

The dominant effect of pore sizes on permeability is evident when comparing K values of a soil material that are measured at different degrees of saturation. Unsaturated soil below an infiltrating surface may have different causes, such as the occurrence of

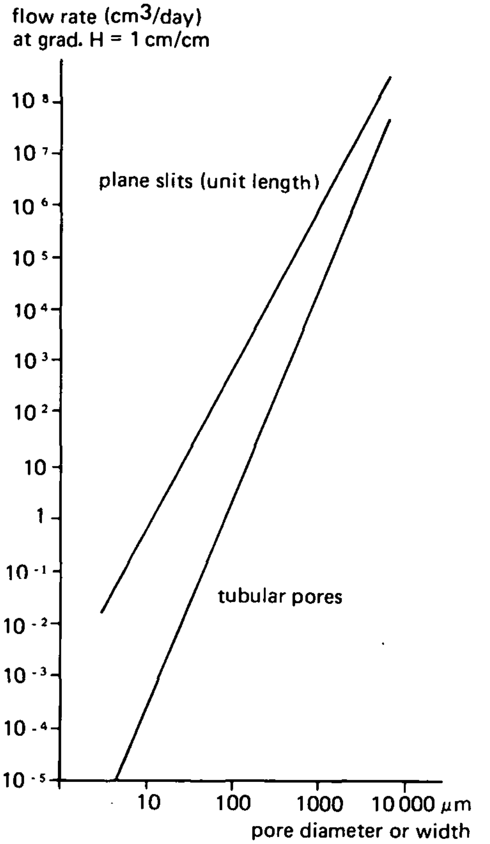


Fig. 16. Flow rates through tubular or planar voids as a function of pore size at a hydraulic gradient of 1 cm/cm.

a physical barrier to flow at the surface of infiltration or an application rate which is lower than the saturated hydraulic conductivity. We may assume three different soil materials, with pore-size distributions schematically represented in Figures 17 and 18. The model assumes that the pores are very long. The uppermost "soil" is coarse porous (like a sand) and the lowest one is fine porous (like a clay). Without any physical barrier (a "crust") on the soil surface and with a sufficient supply of water, all pores are filled and each will conduct water downward as a result of the potential gradient of 1 cm/cm, due to gravity. The larger pores will conduct much more water than the smaller ones (see Fig. 16). Suppose a weak crust forms over the tops of the tubes (Fig. 17). Pores will fill with water only if the capillary force they can exercise is strong enough to "pull" the water through the crust. The larger the pore, the smaller the capillary force that can be exercised (see section 3.2.3). Therefore, larger pores will empty first at increasing crust resistance, creating unsaturated soil and negative soil moisture potentials, which, in turn, lead to a strong reduction in the hydraulic conductivity of the soil.

With no crusts present, similar processes can occur when the rate of application of water to the capillary system is reduced (Fig. 18). With abundant supply, all pores are filled. As this supply (which is supposed to be divided evenly over the infiltrating

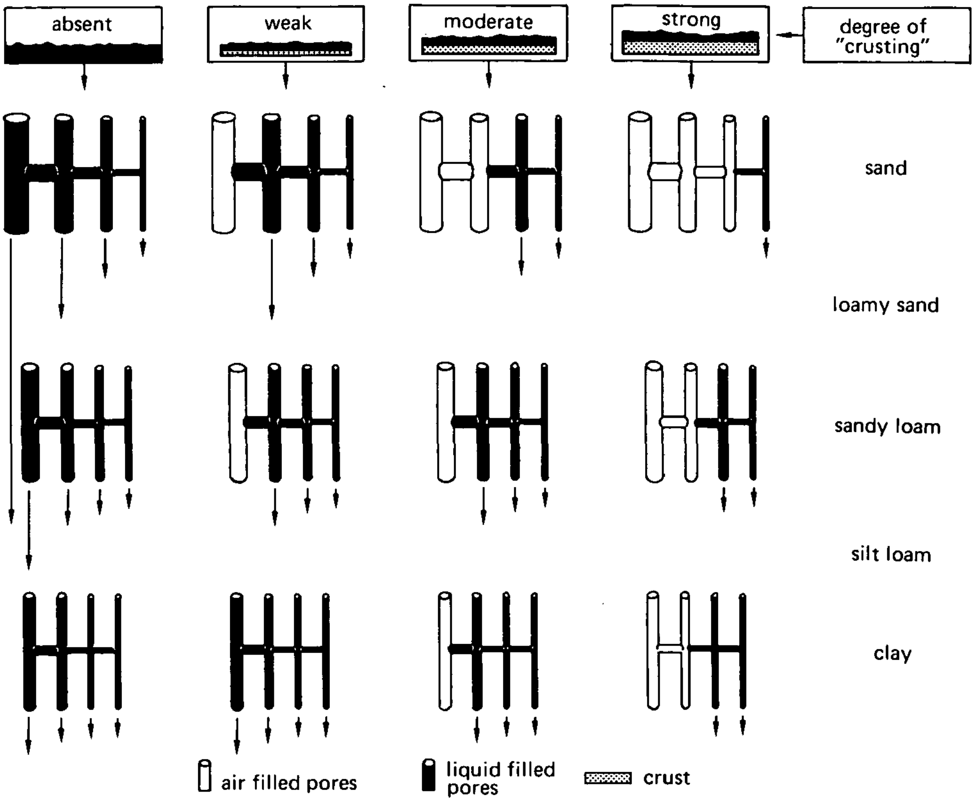


Fig. 17. Schematic diagram showing the effect of increasing the degree of ￼crusting on the rate of percolation through three "soil materials".

pore system) is decreased, there is not enough water to keep all pores filled during the downward movement of the water. Larger pores will empty first, as they conduct most liquid, while at the same time they exercise only relatively small capillary forces. *In this system a certain size of pore can be filled with water only if smaller pores have an insufficient capacity to conduct away the applied water.*

The degree of reduction in K upon desaturation and increasing soil moisture tension is thus characteristic for the pore-size distribution. Coarse porous soils have a relatively high saturated hydraulic conductivity (K_{sat}), but K drops strongly with decreasing potential. Fine porous soils have a relatively low K_{sat}, but K decreases more slowly upon decreasing potential. Experimental curves, determined in the field with the crust test (section 4.4) show such patterns for natural soil. Figure 19 shows curves for the sand C horizon of the Plainfield loamy sand, the sandy loam IIC horizon and the silt loam B2 horizon of the Batavia silt loam, and the clay B2 horizon of the Hibbing loam. The curves for the pedal silt loam and clay horizons demonstrate the physical effect of the occurrence of relatively large cracks and root and worm channels. Soil structure inside the peds is very fine porous and these fine pores hardly contribute to flow. The large pores between peds and root and worm channels give relatively high K_{sat} values

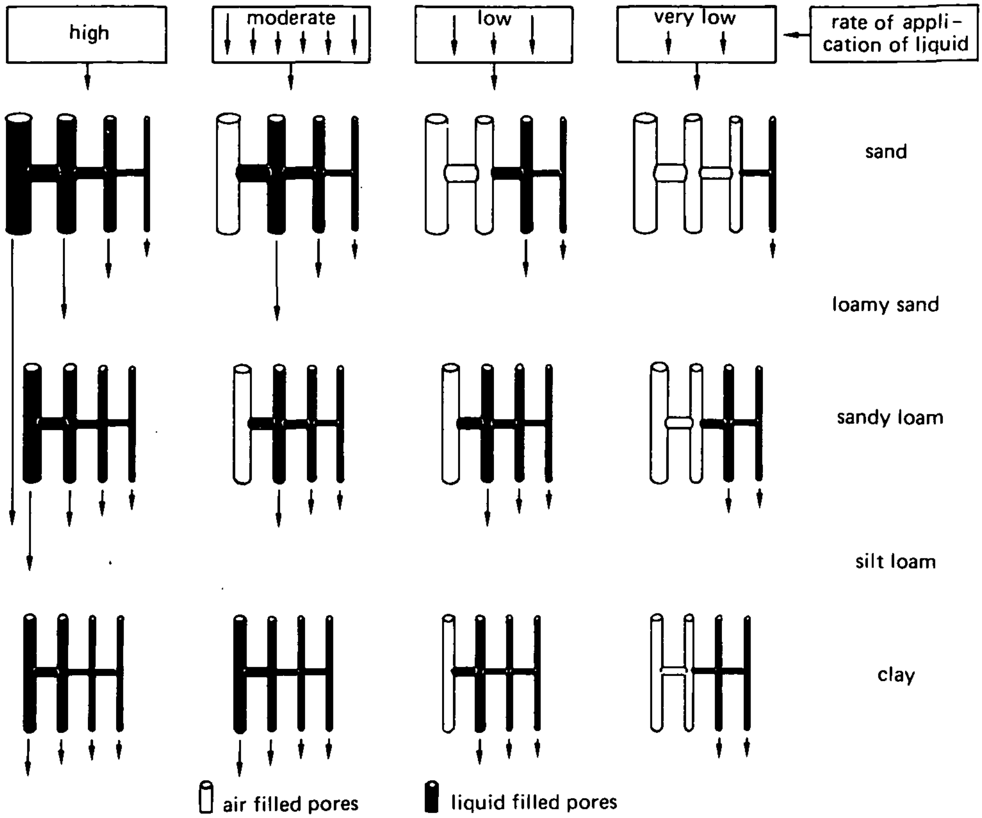


Fig. 18. Schematic diagram showing the effect of decreasing the rate of application of liquid on the rate of percolation through three "soil materials".

(140 cm/day for the silt loam), but these pores are not filled with water at low tensions and K values for these pedal soils drop therefore very strongly between saturation and 20 cm tension (1.5 cm/day for the silt loam).

What is shown in Figure 19 as measured hydraulic conductivity curves is reflected in the schematic diagram of Figures 17 and 18. Going from left to right, the moisture potential decreases (the tension increases), as well as the flow rate through the schematic pores. This corresponds with the trend in Figure 19, as discussed. Such K curves can be shown as graphs and graphical techniques can be used to derive specific K values and hydraulic gradients. However, many attempts have also been made to develop equations that describe the K curves. Such equations are useful for calculations of moisture flow. One equation used is as follows (RIJTEMA, 1965):

$$K = K_0 \, e^{-\alpha \cdot h}$$

where: K_0 = K at saturation, α = characteristic soil constant, h = pressure potential. This equation applies to pressures of approximately —200 cm water. RIJTEMA (1965)

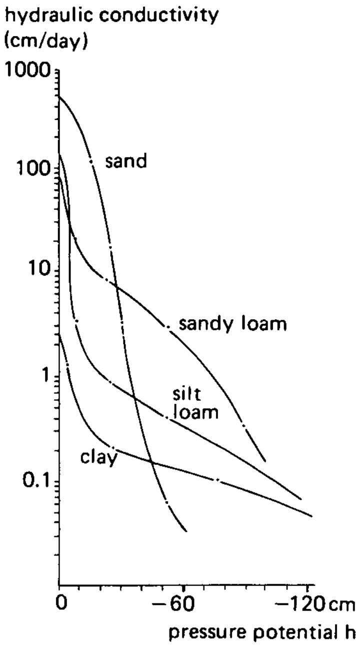


Fig. 19. Hydraulic conductivity (K) as a function of the pressure potential measured *in situ* with the crust-test procedure.

discussed literature data and found that K values in the range from pF 2.3 to the wilting point (pF 4.2.) obeyed a simpler relationship, as follows:

$$K = a.h^{-n}$$

where: n = 1.4 for a very wide range of soils and a = constant.
BAKER and BOUMA (1976) presented an equation for silty clay loam soil horizons for the potential range of 0 to —150 cm, as follows:

$$K = a+b.h^{-c}$$

where a, b and c are characteristic constants.

The ability to predict hydraulic conductivities following an analysis of soil-pore systems, has long been pursued by soil scientists. Efforts have only partly been successful and reflect a clear overlap between the disciplines of soil physics and soil morphology. The topic will therefore be discussed separately in the following chapter.

4.3 Predicting K by means of soil-pore analysis

4.3.1 Calculations based on the capillary "pore-neck" model

The sizes of soil pores can be directly measured by means of morphometric analysis (sections 2.1, 2.3 and 3.5) or can be derived as an equivalent pore-size distribution from moisture retention data (section 3.5). However, flow through pores in a three-dimensional soil volume is governed by the smallest pores in the system (the "bottle-necks"). So whether or not the used pore-size distribution is correct, the main problem will always be the definition of the size of the "necks" which will determine the overall permeability. MARSHALL (1958), following earlier work in England, hypothesized that the size of the "neck" (r_t) in a pore system could be expressed as follows:

$$r_t = \sqrt{e.n.^{-2} \left[r_1^2 + 3r_2^2 + 5r_3^2 + \ldots. (2n\text{-}1)\, r_n^2 \right]}$$

where: e = soil porosity ($m^3.m^{-3}$) measured physically, n = the number of small equal classes in which the total pore volume is divided $r_1 > r_2 > r_3 \ldots > r_n$.

This expression, developed for tubular pores, emphasizes the importance of the smaller pore-size classes in defining the circular "neck-size". The expression has been used by most researchers, with some modification, to derive K values from pore-size distributions which, in turn, were always derived from moisture retention data (GREEN and COREY, 1971). This procedure explains the occurrence of one potential problem: equivalent pore-size distributions derived from desorption characteristics are generally *not* a very good measure for the pore-size distribution due to the hysteresis phenomenon (section 3.5).

Be that as it may, K values are calculated with the following equation by MARSHALL (1958):

$$K_i = \frac{e^2.n^{-2}}{8} \left[r_1^2 + 3r_2^2 + 5r_3^2 + \ldots. (2n\text{-}1)r_n^2 \right].$$

The calculated permeability is known as the intrinsic permeability K_i which can easily be transformed into K, as defined in this text. This transformation is not further pursued here, because it is not essential to the discussion.

This expression will yield K not only for saturated but also for unsaturated conditions. Assume a moisture content and a corresponding moisture potential at which the largest pore size (r_1), becomes filled with air. Then K corresponding with those conditions, is:

$$K_i = \frac{e_1^2.n_1^{-2}}{8} \left[r_2^2 + 3r_3^2 + 5r_4^2 + \ldots. (2n\text{-}1)r_n^2 \right].$$

Here, e_1 is the water-filled porosity when pores with size r_1 are empty and n_1 is n-1. This procedure of elimination can be continued until a series of points is obtained, each one corresponding with a particular moisture content. The equation, as written, uses r values and can be directly applied to morphometric data (BOUMA and ANDERSON, 1973; BOUMA and DENNING, 1974). However, use of desorption data requires a transformation from r to a corresponding pressure h as discussed in section 3.2.3 (equations 3.1 and 3.2 and Fig. 5). The complicated equation, resulting from this transformation, has been used by many authors to calculate K (see reviews by GREEN and COREY, 1971 and KLUTE, 1972):

$$K\ (\theta)_i = \left(\frac{K_s}{K_{sc}}\right) \cdot \left(\frac{30\delta^2}{\varrho g \eta}\right) \cdot \left(\frac{e^p}{n^2}\right) \sum_{j=1}^{m} (2j + 1 - 2i)h_j^{-2}$$

$$i = 1,2 \ldots. m$$

where: $K(\theta)_i$ = the calculated conductivity for a specified water content, $(\theta)_i$ = last water content class on the wet end: i = 1 = pore class corresponding with θ_{sat}, i = m = pore class with lowest water content for which K is calculated, p = constant, δ = surface tension of the liquid, ϱ, g, e, n and h were defined previously.

Of particular interest is the matching factor K_s/K_{sc} which is the ratio between the measured hydraulic conductivity at saturation and the calculated one. The direct application of the Marshall equation yielded good results for sands and was intended for that purpose. Later, the application was extended to other soils as well and calculated K values differed very significantly from the measured ones. The factor p was changed by MILLINGTON and QUIRK (1964) and KUNZE et al. (1968) but only the introduction of matching factors succeeded in producing "good" agreement between measured and calculated values. KUNZE et al. (1968) needed matching factors for different soils varying between 0.5 and 0.004. The latter value indicates that the measured K_{sat} was 250 times smaller than the calculated one: a clear indication that something in the method did not work. Also, DENNING et al. (1974) reported very small matching factors. These authors described another feature of the method, when applied to clayey, pedal soil materials. They produced a wide variety of curves for one soil horizon when varying n, which represents the number of pore classes in which the moisture retention curve is divided for the calculation of K. This is theoretically unacceptable and also indicates the inadequacy of the underlying theory. Varying n did not affect the results of calculations for sands, loamy sands and sandy loams. For those apedal soils, fair results were obtained when using the matching factor. One final comment must be made regarding the presentation of results in the literature. The underlying pore-neck hypothesis was based on r values. These were translated into h values according to the capillary pore model. The output of the calculation yields K as a function of h. But results are always presented as K-θ curves, which require yet another transformation using the moisture retention curve (h-θ relation). Agreement between calculated and measured values often looks reasonable when shown as a K-θ plot, with a compressed θ scale. However, such plots may be misleading.

The calculation of the K-h relationship is not very promising on the basis of the used pore interaction model despite all proposed modifications. Use for sandy, apedal soil materials is feasible when matching factors are used, but the latter are just empirical values without any physical meaning except for their ability to make things "fit". Direct application of the "pore-neck" concept to morphometric measurements was reported by BOUMA and ANDERSON (1973) and BOUMA and DENNING (1974). Results were good for sands but poor for soils containing clay. Large matching factors were needed also for these calculations, which were difficult to execute because of low total porosities obtained with the morphometric point count method (see section 2.3, Table 1).

LALIBERTE et al. (1968) presented another method for calculation of the K-h curve (see review by KLUTE, 1972). This approach uses an empirical functional relationship between the relative conductivity K/K_s and the pressure potential h. The air-entry value is of essential importance in this approach and this value is often very difficult to determine in natural soils with relatively large root or worm-channels and cracks. Satisfactory use of this method can only be expected when good techniques can be developed for the determination of the air-entry value (Bloemen, pers. comm.). As yet this method is not operational for all soils. Due to its empirical nature, the method is not very interesting in the context of this discussion on relations between pore size and hydraulic conductivity.

Summarizing, calculation of hydraulic conductivity on the basis of the capillary pore interaction model as defined by Marshall and later modified by others, are successful for sandy apedal soils, but even then matching factors are needed. Physical methods (using an equivalent pore-size distribution derived from the desorption curve) and morphometric methods are equally successful. The physical methods are quicker and cheaper and should be preferred. Application of the capillary pore-interaction model to clayey, often pedal soil has not been successfull for either physical or morphometric methods. A number of possible reasons was discussed in sections 2.2 and 3.5. They concern low total pore volumes determined with morphometric counts, hysteresis, volume decrease upon water extraction in swelling soils without creation of air filled voids, and the occurrence of non-tubular pore types such as planar voids. Direct measurement of hydraulic conductivity may therefore be the only practical possibility left. However, before such methods are briefly reviewed, one example will be presented to illustrate specific use of morphometric data for calculating the hydraulic conductivity of pedal silt loam soils with well developed peds and planar voids.

4.3.2 Calculations based on the plane-slit model for clayey, pedal soil

Morphometric methods for determining the pore-size distribution were not useful when calculations of the hydraulic conductivity were based on the capillary pore-interaction model. The physical procedure, using equivalent pore-size distributions, was more attractive in sandy soils and results in clayey, pedal soils were equally poor for both procedures. An understanding of soil structure, as discussed in section 2.1, allows a further evaluation of the problem. Pedal soils (Fig. 1, right side) consist of peds with fine simple packing voids, which are separated from adjoining peds by relatively large cracks (planar voids). Also for these soils, it is very important to define pore-necks in the porous system, which will determine the overall permeability. A "neck" based on an at random connection of capillary tubes (MARSHALL, 1958) does not represent a viable model for pedal soils. A new pore interaction model must therefore be developed.

As an introduction we consider horizontal sections through abstract models of "soil" (Fig. 20). The upper figure shows a surface of 100 cm^2 occupied by square blocks ("peds"), mutually separated by a distance d. The blocks are impermeable and water moves only along the plane. The lower figure shows the same surface with one tubular pore (channel). Water can only move through this pore. K_{sat} values were calculated for different sizes of peds and pores, using equations 4.6 and 4.7 presented in section 4.2 (see Fig. 16). Porosities were also calculated for these models. Results are reported in Table 3. Some obvious conclusions can be drawn:

(i) Few small pores can conduct large quantities of water. This is due to the fact that the capacity for transmitting liquid is proportional with the factor r^4 for channels (r = radius) and d^3 for planar voids (d = width). Small differences in measured pore size

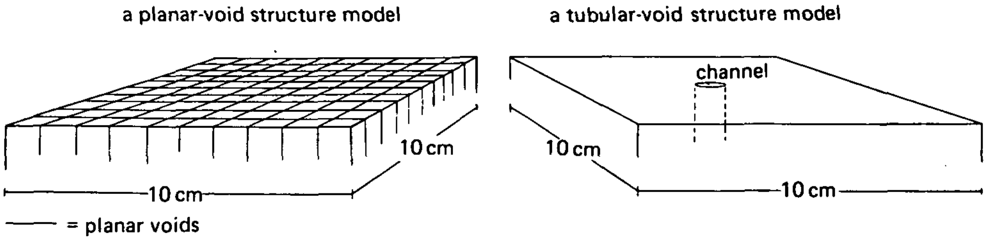


Fig. 20. Horizontal sections through a tubular and a planar void model of soil structure.

Table 3. Hydraulic conductivities and porosities of planar and tubular pore models of soil structure (Fig. 20).

Planar void model

Size of blocks	Width (d) of planar voids in μm	Porosity in %	K in cm/day
1 cm^2	10	0.2	1.5
	50	1.0	180
	100	2.0	1 440
4 cm^2	10	0.1	0.7
	50	0.5	90
	100	1.0	720
	1 000	10	719 712

Tubular pore model

Diameter of channel (2r) in μm	Porosity in %	K in cm/day
100	0.8×10^{-4}	0.02
200	0.3×10^{-3}	0.34
500	0.2×10^{-2}	14.6
1 000	0.8×10^{-2}	211.2
4 000	0.13	54 130.5

values will therefore have a large effect on the calculated permeability. This is not a very good starting point for any method that tries to calculate K from morphometric pore data.

(ii) Occurrence of only a few pores may result in high permeabilities, even though their contribution to pore volume is very small. For example, a structure with blocks of 4 cm^2 separated from adjoining blocks by 100 μm wide planes, has a high K of 720 cm/day (Table 3). However, they contribute only 1 vol % to porosity, which is well within the experimental error when porosities of natural soils (let alone equivalent pore-size distributions) are determined by physical methods. These phenomena are even more expressed for tubular voids (Table 3). Physical methods are totally inadequate as a tool to characterize size distributions of voids, such as planes and channels, which are essential to flow in pedal soil. The latter was proved by means of dye-studies which clearly demonstrated exclusive flow during saturated conditions along these larger voids (ANDERSON and BOUMA, 1973).

Measuring the widths of planar voids or channels by morphometric techniques is then the only viable alternative, presenting a specific and exclusive application of these techniques. However, the "necks" in the system of larger voids still remain to be defined. The following procedure was devised. A summary will be presented; for more details and examples the reader is referred to the publications by BOUMA and ANDERSON (1973) and ANDERSON and BOUMA (1973).

First a width (d) distribution is obtained for planar voids. Thin sections can be used but also soil peels. A width distribution must be corrected for effects of swelling if values are obtained from air-dry samples. SARAN data can be used for this purpose (section 3.3). Assume that there were n size classes of planar pores in the air-dry soil. Correction for swelling resulted in a reduction of the number of open size classes from n to n—x, where x is the largest pore class *closed* by swelling (all classes between 1 and x are also closed). The average width of each class is denoted by d. The length of each size class of open planes (l_i) in a saturated soil is then estimated as follows:

$l_i = L.\ (P_i/P_t)$

where: L = total length of all planes (traced on a photograph of the soil sample), P_i = number of planes in a certain size class i, as determined by a ribbon-count procedure, and P_t = *total* number of planes counted in the ribbon-count. The ribbon count involves observation of the soil sample in a series of narrow parallel ribbons. The number and sizes of planes which occur in these sampling ribbons are noted. Average ped height (v) is determined from either field investigations or morphometric analysis. The key part of the method is the prediction of the probability that planar voids of a certain size are continuous throughout a sample, which consists of a certain number of layers, each with thickness v (the height of the peds). *The hypothesis is made that interconnected open planes govern the flow process.* This is schematically shown in Figure 21 for two conditions A and B. The overall permeability of sample A is governed by the width of the planar void in layer 2; for sample B it is layer 4. The possibility that certain sizes of planar voids do occur in the flow system is a function of:

(i) Their number. In other words, if one size class occurs very dominantly (P_i/P_t is large) the probability is high that planes with this size will occur.

(ii) The length of the sample. The longer the sample, the higher the probability that a small plane will occur in the flow system, even if they do occur sporadically (P_i/P_t small). This is illustrated in Figure 21B. A sample consisting of three layers of peds would have been much more permeable, than the one pictured with four layers, due to the small plane in layer 4.

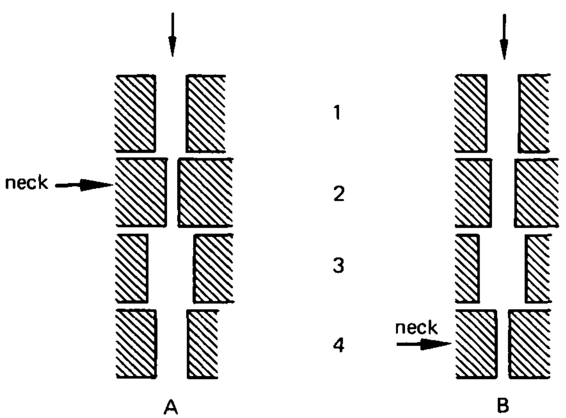


Fig. 21. Schematic diagram illustrating the occurrence of planar pore "necks".

Chances of a size class of planes to be vertically continuous can be estimated as they are proportional to the l_i values for each open class considering a structure model with, for instance, five layers of peds. For example: open planes, interconnected through the layers of peds are of size class (n — x + 1) or larger (size class n — x was the largest plane closed by swelling as the reader may recall). What we are interested in now is to find out whether or not larger size planes are likely to be continuous throughout as well. Considering the contact between the first and second layer of peds, the probability of a plane of size class (n — x + 2) to be connected *with a pore similar or larger than itself* is proportional to the l_i values involved: $(l_t - l_n - x + 1)/l_t$, where l_t = total length of planes which are still open after swelling ($l_t < L$). A similar expression can be derived for the first three layers of peds etc. until we arrive at the probability for 5 layers of peds (our example) as follows:

$$\left[(l_t - l_n - x + 1)/l_t\right]^{5-1}$$

If the probability of a plane being continuous into a pore similar or larger than itself is lower than 5 %, we ignore the possibility and assume that the plane is connected with planes with a smaller width.

Calculations are also made for the larger pore classes n — x + 2, n — x + 3 etc. Assume that class n — x + s represents the last class to have a probability of 5 % of being continuous into a similarly sized or larger pore class within the defined thickness of the sample. Now only a fraction of open pore length contributes to conductivity. To be more exact: $(l_n - x + 1) + (l_n - x + 2) + \ldots. (l_n - x + s)$. Total open pore length was l_t cm. That leaves:

$$l_t - [\,(l_n - x + 1) + (l_n - x + 2) + \ldots. (l_n - x + s)\,].$$

According to the model, flow into planes of size class n — x + s + 1 or larger will have to pass planes of size class n — x + 1 (the smallest open plane) or larger. The probability that flow occurs through size class n — x + 1 is generally highest. Therefore l_i values of all pores larger than class n — x + s are assigned to class n — x + 1. The K calculation, which is based on the equation for planar flow (section 4.2, equation 4.7), is therefore composed of two parts: the first part relates to flow through pores that are likely to be continuous throughout (either through pores of its own size or larger) and the second part relates to flow through pores that are unlikely to be continuous in that manner. The appropriate l_i values belonging to these (large) pore classes are attributed to the smallest open class. By calculating flow through plane slits of varying length and width, a total flow (cm^3) is obtained for the area of investigation (a soil peel or a large thin section). To obtain K, the flow rate must be divided by this area (S). The following equation is used:

$$K = \left[\sum_{i\,=\,n-x+1}^{i\,=\,n-x+5} \left(\frac{\varrho g d_i^3}{12\eta}\right) \times \frac{l_i}{S}\right] + \frac{\varrho g d^3_{n-x+1}}{12\eta} \times \frac{l_t - [\,(l_n - x+1) + (l_n - x+2) + \ldots. (l_n - x+5)\,].}{S}$$

For explanation of the symbols the reader is referred to section 4.2.

BOUMA and ANDERSON (1973) and ANDERSON and BOUMA (1973) applied this method successfully to a number of pedal soils. However, the method is laborious and the assumptions are rather drastic. For example, horizontal flow is ignored, only flow through planar voids is considered and it is assumed that flow occurs through interconnected open pores after swelling. The latter, in particular, is improbable since planes closed by swelling will frequently occur in the flow system as well. In addition, recent dye experiments have indicated that flow patterns may be different than hypothesized. Moreover, only K_{sat} is determined and no information is obtained about conductivities for unsaturated soil. The method seems therefore more interesting from the viewpoint of research than as a practical procedure for calculating K_{sat}. Direct measurement is faster and cheaper. However, the procedure allows a better understanding of flow through pedal soils and this justifies its discussion. A particular illustration of this aspect will be presented in the following section.

4.3.3 Representative sizes of soil samples for physical analyses

The measurement of hydraulic conductivity in small cores has offered many practical problems, particularly in clayey, pedal soil. Large variability in measuring results has been approached in vain by statistical techniques. Our plane-interaction model allows an estimate of the effect of sample height on measured K_{sat} values. The probability of plane continuity is a function of sample height, as discussed. Conductivities were

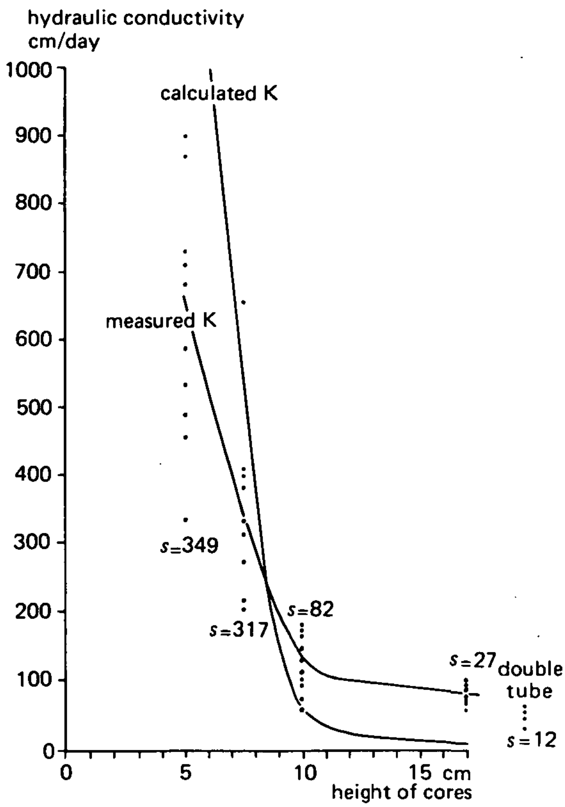


Fig. 22. Measured and calculated hydraulic conductivities and standard deviations (s) for series of cores with different heights, K values measured with the double-tube method are included as a reference (Batavia silt loam) (after ANDERSON and BOUMA, 1973).

calculated for the B2 horizon of a Batavia silt loam (Fig. 22) following criteria of the method just discussed. Then, series of cores with different heights were measured with standard laboratory techniques. The results (Fig. 22) demonstrate the clear relationship between sample height and measured K_{sat}. The trend predicted by the plane-interaction theory is quite correct, even though there are some differences with the predicted values. In other words, *any* K_{sat} *can be measured when relatively small cores are used in the laboratory*. The standard core size of 100 cm³ (5 cm high) is very inadequate for characterizing a pedal soil horizon of perhaps 30 cm thick. Soil morphological analyses are very important for determining optimal sizes of soil samples to be subjected to physical measurements. It is useful to refer to the size of the elementary unit of structure. A standard 100 cm³ sample of a coarse sand will contain at least (very approximately) 65 000 individual grains. Such samples are realistically considered to be representative of entire soil horizons. However, a soil horizon consisting of 1 cm subangular blocky peds, would have to be characterized (when applying the same relative standard) by a sample of at least 65 000 cm³ (or 65 litres). Perhaps this size is unrealistically large but more effort should be made to define minimal sizes of representative samples for different soil horizons by determining variability as a function of sample size.

4.4 Methods for the determination of hydraulic conductivity

4.4.1 Introduction

Many methods, both for field and laboratory use, exist for measuring hydraulic conductivity. An excellent and complete review has been presented by KLUTE (1972). This Soil Survey Paper is not intended to be a methods manual. In any case, no written account can ever be a satisfactory substitute for a practical demonstration. A description will be presented of methods that have been used by the author. Emphasis will be on practical aspects of limitations and applicability. The reader is referred to detailed publications for more specific information, if needed.

The development and evaluation of methods for measuring hydraulic conductivity is a very active area of research, at this time. Data to be presented in section 4.4 is therefore bound to be subject to change and modification.

4.4.2 Field methods

The crust test will be discussed in some detail and operational aspects of the instantaneous-profile method will be reviewed.

4.4.2.1 The crust test

The crust-test procedure is used to measure hydraulic conductivities (K values) of unsaturated soil in situ (HILLEL and GARDNER, 1969). The method consists of a series of infiltration runs, each of which yields one point on the hydraulic conductivity curve (K-h). The natural groundwater should be sufficiently deep so as not to allow interference with the measurement procedure. This sometimes means that measurements can only be made in summer or the dry season. Curves can cover K values in a moisture pressure range between 0 and —150 cm, but the test should be limited to pressures above about —50 cm for practical and technical reasons to be discussed in this section.

Each run is made in a soil pit, using an excavated cylindrical column with a height and diameter of approximately 30 cm. The equilibrium infiltration rate ($cm^3.cm^{-2}.s^{-1}$) through a gypsum-sand crust applied on top of the column in a cylinder infiltrometer is measured, as is the corresponding sub-crust negative pressure with a tensiometer. Separate runs are made through three or four gypsum-sand crusts (with different gypsum-sand ratio's), following a sequence from high to low crust resistances. Crusts with a high resistance yield a low infiltration rate (which corresponds with a K value when the hydraulic gradient below the crust is 1 cm/cm) at a relatively low moisture content and a relatively high soil moisture suction. The method was discussed by BOUMA et al. (1971), BOUMA and DENNING (1972) and BOUMA et al. (1974a).

A profile description is made before applying the test to determine characteristic depth ranges, usually corresponding with major horizon boundaries, within which separate columns should be excavated.

A horizontal plane is prepared at the desired level at the test site by using a putty knife and a carpenter's level. A cylindrical column of soil, at least 25 cm high, with a diameter of 30 cm, is carved out from the test level downward, taking care to chip or pick the soil away from the column as the desired form is approached, so as to prevent undue disturbance of the column itself. A ring infiltrometer, 30 cm in diameter and 10 cm high, with a 2.5 cm wide brim at the top, is fitted onto the column. The level of the soil surface inside the infiltrometer should be approximately 1.5 cm below its brim to avoid an excessively large space between the soil surface and the infiltrometer cover, which is applied later. The sides of the column are then sealed with aluminum foil to avoid evaporation losses and soil is packed around it. Complete sealing is not necessary because water will not flow from the column under unsaturated conditions. A half-inch thick acrylic plastic cover with a diameter of 35 cm and with sponge-rubber glued to it is bolted to the top of the infiltrometer with ring nuts. An intake port and air-bleeder valve are provided in the cover. The inner surface of the cover is cone-shaped towards the air-bleeder to facilitate air removal during filling. The cover is removed after the cylinder has been positioned, to allow the crusts to be placed. Thin pencil-size, mercurytype tensiometers attached to 3 mm thick plastic tubing are implanted from the sides to a point 1 cm below the crust in the column (for details see BOUMA et al., 1974a).

In the first experiments with the crust-test procedure, various puddled soil materials were used for crusts (BOUMA et al., 1971). Additional field experience, however, showed that some of these crusts (in particular the ones with a relatively low resistance) were rather unstable and easily disturbed, due to continuous swelling of the clay particles. A different procedure was therefore developed in later experiments, using dry gypsum powder thoroughly mixed with varying quantities of a medium sand. After sufficient wetting and continuous mixing, a thick paste is obtained. This material is then quickly transferred to the prepared soil surface in the column and applied to the top with a putty knife as a continuous crust to the wall of the cylinder to avoid boundary flow. Crusts of this type harden within about 30 minutes, thereby providing a stable porous medium with a fixed conductivity value.

Crust resistance can be varied by changing the relative quantities of gypsum and sand. Crusts composed of gypsum only have the highest resistance. For example, a subcrust moisture pressure of —52 cm was induced in a sand column capped with a 5 mm thick gypsum crust with 3 mm water on top. Another crust, formed from a pre-wetted mixture composed of 14 % gypsum and 86 % sand by volume, as measured in the field using a graduated cylinder, induced a sub-crust of only —11 cm.

A series of crusts is applied to the same column for succeeding runs. A measurement of the moisture pressure should be obtained before crusts are applied, to determine the possible range of K values that can be measured. For example, if natural moisture pressures are only —20 cm, it would be useless to apply crusts that would induce lower pressures. No flow would occur. The best sequence is to start with a heavy crust and to follow with crusts of progressively decreasing gypsum contents. Each infiltration run through a particular crust yields one point of a curve of hydraulic conductivity versus pressure potential h (—cm) (see Fig. 23).

One important addition has been made to the crust test recently (BAKER and BOUMA, 1976). After measurement of the K value, corresponding with the lightest crust, this crust is removed and a measurement of K_{sat} is made in a column without a crust. Removal of the crust generally results in removal of some of the underlying soil which has the benefit of producing an open, natural soil surface. This is essential to obtain a reliable measurement for saturated conditions. To avoid lateral flow from the saturated column, a gypsum collar is poured around the cylinder after removal of the aluminum foil. Two types of measurements are made: (i) with the column attached to the subsoil and (ii) with the column detached (Figs. 23 and 24). This is easily achieved by slightly

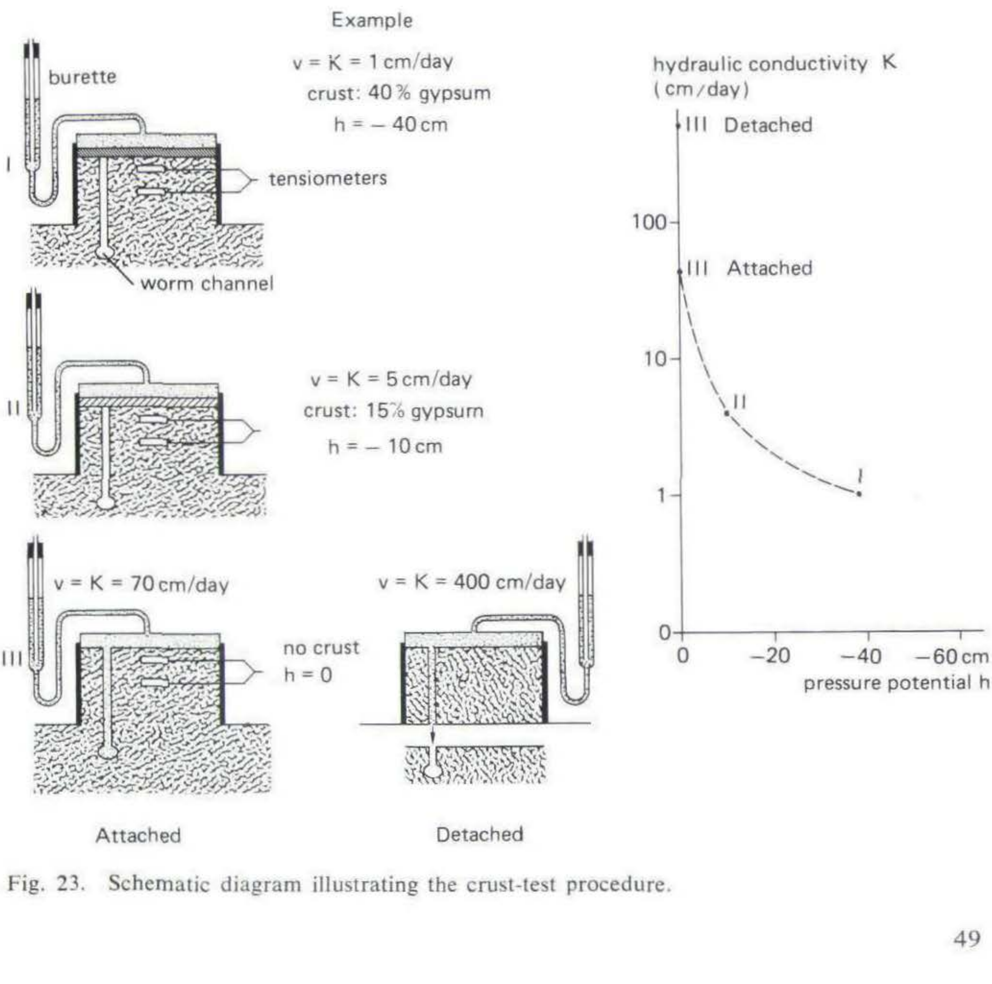


Fig. 23. Schematic diagram illustrating the crust-test procedure.

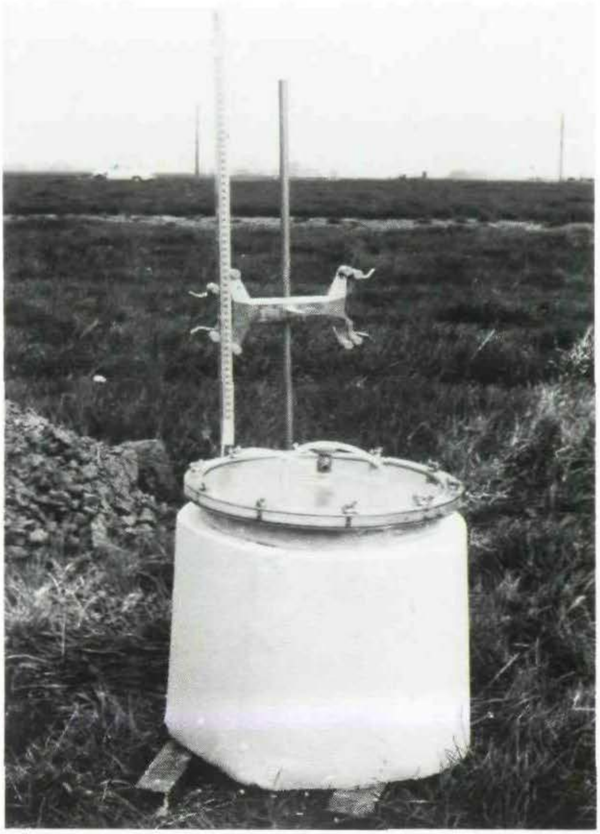


Fig. 24. Measurement of K_{sat} on a large undisturbed soil core which was carved out *in situ*, set in gypsum and removed from the profile.

lifting the gypsum collar with a spade. This creates, in fact, a very large core with a diameter of 30 cm and a height of 25 cm. Such sizes are needed for reliable measurements in pedal soils (see section 4.3.2.). The differences between the "attached" and "detached" measurements are interesting because they are a measure for the continuity of large pores, such as channels and planar voids. These voids may be continuous, even through a 25 cm high sample. However, in the attached column such voids will most probably *not* be continuous. Then, they will just fill with water which will slowly drain through surrounding voids, whereas they would greatly increase the saturated conductivity if continuous throughout the sample (Chapter 3).

The crust test procedure is initiated by ponding water on top of the crust to a level slightly higher than the top level of the cylinder. Water applied this way does not have to be applied through the burette and this procedure saves time. The cover is then bolted to the cylinder. To avoid leakage the sponge-rubber gasket between cover and cylinder should be flat and should cover clean surfaces before the cover is applied. A piece of semi-rigid tubing, attached to a water-filled burette, is then attached to the intake port in the cover. The tubing connecting burette and cylinder should be completely full of water to avoid irregular flow rates. Air can be removed from the tubing by turning the unattached burette upside down after closing it at the top and by patting against the tubing.

The air-bleeder valve should be open during filling until the small space between the crust and the cover of the cylinder is full of water. A Mariotte device, in a burette, maintains a constant pressure of about 3 mm water over the crust (BOUMA et al., 1974a). The infiltration rate into the soil, corresponding to the rate of movement of the water level in the burette, is recorded as soon as the tensiometers show that equilibrium has been reached. Soil moisture potentials are measured in the columns and are derived from the mercury rise in plastic tubes along calibrated scales or from transducer readings (Fig. 25).

The infiltration rate, when constant for a period of at least one hour, is taken to be the unsaturated K value at the subcrust pressure when the pressure gradient is zero (this means only gravitational flow occurs: $v/i = 1$). In some cases, a pressure gradient remains at steady-state conditions. Hydraulic conductivity is then calculated according to $K = v/i$, where v = infiltration rate and i = hydraulic gradient below the crust (in such a case $\neq 1$). However, the hydraulic gradient is usually close to unity and use of one tensiometer at 3 cm depth below the crust generally suffices to obtain a representative pressure corresponding with unit hydraulic gradient.

Discussion: The crust test is a rather laborious technique which is most useful to be applied at low suctions. Use of 100 % gypsum crusts may yield K values that correspond with pressures below —100 cm water except for sands. Flow rates are then often near 1 mm/day or lower and relatively long times are needed to obtain equilibrium. The method is not very practical for such conditions. Other methods, for example

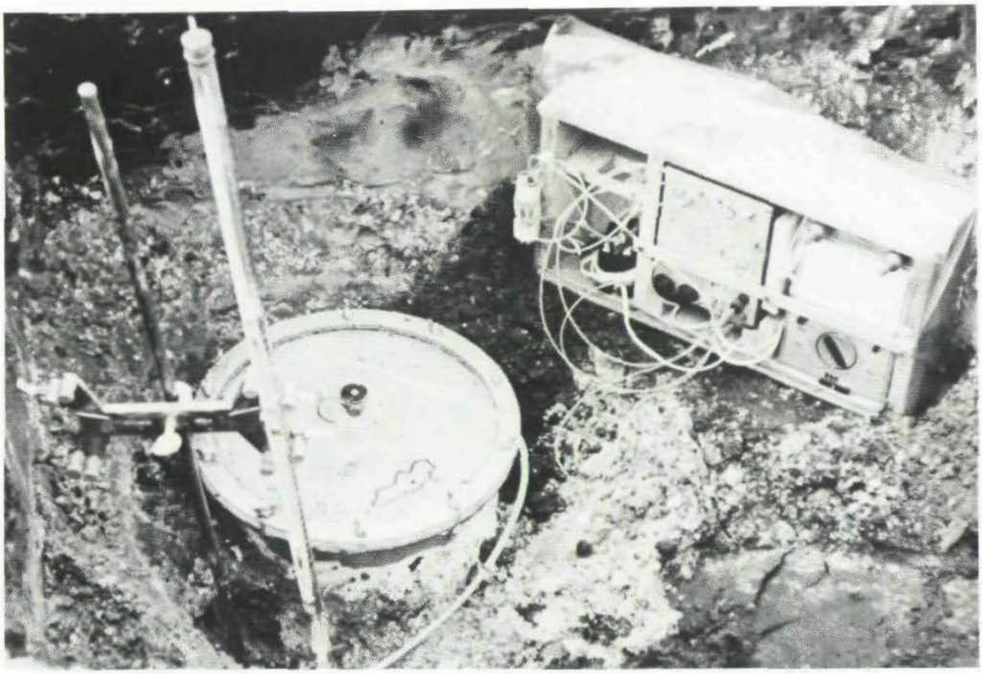


Fig. 25. Measurement of the pressure potential below the crust, using an electronic transducer system (right) developed by J. W. Bakker (Institute for Land and Water Management Research, Wageningen).

the evaporation method, to be discussed later, are more suitable. Measurements *at* saturation without a crust are of great importance. There really is no other satisfactory method for measuring K_{sat} of soil well above the groundwater. Much experience has been gained with the Bouwer double tube in Wisconsin (BOUMA et al., 1974a) but this method is even more cumbersome than the crust test, and more important, it has some serious flaws. Vertically anisotropic soils can not be characterized adequately with this method. Summarizing, the crust test is seen as a very useful method for obtaining K values in soil above the groundwater in the range from saturation to a moisture pressure of, perhaps, —50 cm or so. Recent experience suggests that only light crusts should be used to avoid the small flow rates and low pressures that are associated with heavier crusts with more gypsum.

4.4.2.2 The instantaneous profile method

The continuity equation may be applied to one-dimensional flow and integrated to yield:

$$V_{(z_2,t)} = V_{(z_1,t)} - \int_{z_1}^{z_2} \frac{\delta\theta}{\delta_t} \, . \, dz$$

where: $V_{(z2,t)}$ and $V_{(z1,t)}$ are the fluxes at positions z_2 and z_1, and time t. If the water-content distribution $\theta_{(z,t)}$ is known, the integral can be evaluated and if either $V_{(z_2,t)}$ or $V_{(z_1,t)}$ is known, the other flux can be calculated. If the hydraulic-head distribution is also known, $H_{(z,t)}$, which is the hydraulic gradient at a given position and time, can be evaluated. The ratio of the flux and hydraulic gradient at the given position and time is then the hydraulic conductivity at the water-content pressure head found at that position. KLUTE (1972) has reviewed different variations of this method. Detailed examples were given by HILLEL et al. (1972) and Baker (in BOUMA et al., 1974a). The discussion in this section will therefore be limited to practical aspects.

In this method a fallow plot of soil is artificially wetted to saturation. The plot is then covered to prevent evapotranspiration. Internal drainage occurs when the groundwater is sufficiently deep. By the use of tensiometers, placed at depths corresponding to the lower part of major horizons, pressure potentials are measured. (These include over-burden potentials in swelling soils and pneumatic potentials, see section 3.2.5). Moisture contents at these depths are also determined simultaneously. These two characteristics are measured frequently, starting at saturation, which represents time zero.

A nearly level site is selected for the experiment so that water may be ponded more or less uniformly over the entire area. To retain the water, a simple earth dike can be constructed surrounding the plot. A water depth of about 2 or 3 cm over all portions of the pond is desired, but may be difficult to achieve on sloping land. On slopes of low grade, uniform distribution of water is no problem, but on steeper slopes uneven infiltration and sub-surface flow may occur. Water infiltrating over the entire area of the plot moves downward under the influence of gravity. Because the soil surrounding the plot is drier than the soil directly under the pond, lateral movement of water can be expected to occur. This effect is greatest at the boundaries of the area and decreases toward the centre. Flow is nearly vertical below the centre of the plot. Here, approximate one-dimensional flow is achieved and measurements are made there.

Problems may arise due to flow of external moisture from upslope through the soil of the plot. This directional flow is most significant at saturation when downslope movement due to gravity is the dominant force involved. Saturation may occur in the form of a real water table or of a "perched" water table where water is temporarily ponded above slowly permeable superficial horizons or strata. Data obtained under such conditions, which are quite common due to the occurrence of illuvial B-horizons, can be extremely difficult to interpret. But also in perfectly level soils, problems may occur. In homogeneous media, such as some sands, an almost uniform gradient is found during drainage. In multi-layered soils where horizons have different hydraulic conductivities, a non-uniform gradient develops. Impeding horizons such as plough layers, argillic horizons or pans near the surface will prevent free drainage from occurring within the experimental volume of soil. Soil moisture pressures above such impeding layers may approach or reach zero while pressures beneath the layer can remain much lower. Such impeding horizons behave as a "crust", preventing saturation from occurring in horizons below them. Therefore, part of a given multi-layered soil may never reach saturation, and K values close to saturation cannot be determined for these horizons by a free-drainage method such as this. This offers no problems when natural drainage is being characterized, because these horizons will never be saturated either under natural conditions. However, use of soil for sub-surface seepage fields, will, for example, require the entire K-curve. This method is then inadequate. The method also requires a rather large area. A circular plot of a diameter of 3 metres has been used successfully (DAVIDSON et al., 1969), but under extreme conditions of drying in the surrounding soil, this may be an inadequate volume to allow undisturbed internal drainage at the centre of the site (VEPRASKAS et al., 1974). In such a case, the diameter of the plot may be expanded to ensure unaffected drainage at the centre. After the profile has been wetted, it must be isolated from such factors as precipitation and evapotranspiration. These can be controlled with two sheets of, preferably black, plastic sheeting large enough to cover the plot easily.

Other forms of plot preparation are possible. One involves more digging but provides better control over environmental factors. This requires the digging of a ring-like trench around a 1 metre diameter undisturbed column of soil. The trench is made to a depth somewhat below the deepest horizon of which the conductivity is to be determined. Following this, a detailed profile description can be made from the wall of the column, providing accurate horizon boundaries for location of the tensiometers. The sides of the column must be covered with plastic sheeting or aluminum foil to prevent evaporation. The surface is prepared as described previously, and a ring or dike is constructed to retain the ponded water on the top for wetting. This arrangement has the distinct advantage that tensiometers can be implanted from the sides which eliminates many of the problems of vertical placing, such as the need to seal the cavity around the tensiometer shaft. For this application, small 6 mm diameter tensiometers, as described for the crust-test method, can be used. The neutron moisture probe is a convenient tool for moisture-content measurement here.

One of the main advantages of the column method is that one-dimensional flow is maintained. There is no lateral interference with drainage or downslope moisture flow. Internal drainage proceeds uninterrupted. Effects of neighbouring vegetation are eliminated and problems associated with slope are greatly reduced. However, the procedure is very elaborate and costly. Accurate measurements of soil-moisture tension and soil-moisture content are necessary for the implementation of this technique. The careful placing of tensiometers in specific horizons of the profile, the sealing of these tensiometers in place and methods of moisture-content determination were discussed by BAKER et al. (1974) and BOUMA et al (1974a).

Two methods can be used to determine soil moisture contents: (i) Values can be derived from a moisture retention curve of the horizon, "translating" measured pressure potentials (h) into θ values. Use of a desorption curve is required to simulate the drainage process. (ii) Values can be measured directly *in situ*, for example with the neutron-probe, or — less attractive — by periodic gravimetric sampling. A cross-check of data is possible when both procedures are followed. BAKER et al. (1974) reported such a comparison and found good agreement between results obtained by both methods. Moisture retention curves were derived from *in situ* neutron-probe measurements of the moisture content *and* in situ tensiometer readings, made at the same time. These moisture retention curves compared favourably with those determined with the conventional technique of using (large) samples that were saturated and desorbed in the laboratory. Detailed calculations of K using basic data obtained with the instantaneous profile method, were reported by HILLEL et al. (1972) and by Baker (in BOUMA et al., 1974a).

Discussion: KLUTE (1972) considers the instantaneous profile method to be the best for measuring K values in the field. One major disadvantage, so far not mentioned, is the very long duration of the procedure. Many weeks may be needed to reach pressure potentials near —100 cm in soils in which natural drainage from saturation occurs without evaporation. A limited range of K values is thus obtained at an immense investment of time. ARYA et al. (1975) have proposed a modified method in which evaporation is allowed to occur as well. This results in a shorter experimental period in which data for a larger range of potentials is obtained. This procedure could be more attractive. Neither method provides reliable data for saturated conditions or for moisture potentials near saturation. The crust test method, or any other method that proves useful, can play an important role if data near saturation are required. Measurement *in situ* constitutes one very major advantage of the instantaneous-profile method. Data derived does apply to a natural undisturbed volume of soil as it occurs in the field with all its characteristics in terms of variability, internal irregularities etc. Field testing of these types of methods should receive high priority in soil physical research. Soil survey input is essential to allow testing to be meaningful. As already discussed, the occurrence of pedogenic soil horizons may impose severe limitations on the method, and knowledge of profile characteristics can explain what might otherwise appear to be odd results.

4.4.3 Laboratory methods

Two methods will be reviewed briefly, emphasizing practical aspects of application rather than technical details. This selection is quite arbitrary (see KLUTE, 1972) but represents experience of the author. WIND (1966) published a method which can be used to calculate K from evaporation data. This method is of the "instantaneous profile" type (KLUTE, 1972, and section 4.4.2.2.) and uses undisturbed, large cylinders of soil. The initially saturated column evaporates at the top only. Moisture potentials are measured at different depths in the column during the experiment, yielding gradients of the pressure potential. The flux for different depths can be calculated from the changes in moisture content and total weight. The quotient of flux and hydraulic gradient yields hydraulic conductivity (K) values according to:

$$V = -K\left(\frac{dh}{dz} + 1\right).$$

WIND (1966) provides very detailed examples and the reader is referred to his publication. Four observations made by Wind are of general operational significance:

(i) For vertical *upward* flow the pressure gradient has to be *decreased* by 1 to compensate for the increase in potential due to gravity. Values of dh/dz are small in the low tension range in the lower part of the column and cannot be measured accurately. But small differences have a large effect on the calculated K value. For example, assume V has been calculated as 1 mm/day. Small gradients of 1.6 or 1.1 cm/cm translate into K values of 1.7 and 10 mm/day respectively. This procedure will therefore be unreliable in the "wet" range of moisture potentials.

(ii) The flow velocities at a certain depth equal the evaporation rate (weight loss of the entire column) minus the moisture lost by the soil *above* that depth. The errors in the flow velocities become larger as more calculated "moisture losses" for increasingly deeper layers are subtracted. In other words, velocities for the bottom layers of the column, which are wet and have negative potentials near saturation, are the least reliable.

(iii) The determination of hydraulic conductivities of soils up to moisture pressures corresponding with pF 4.2 requires the use of gypsum blocks rather than tensiometers since the latter cannot record moisture pressures lower than appr. —800 cm at the most (section 3.2.4). But gypsum blocks do not yield very reliable results when pressures are higher than —100 cm.

(iv) Very long experimental periods are required when K values at low pressures are to be derived. The same problem does exist when the instantaneous profile field method (section 4.4.2.2) is applied.

The first three points suggest that the method is not very reliable for obtaining K values in the "wet" range of soil moisture conditions. The crust test (section 4.4.2.1) is the most suitable then. The fourth point is operational and implies simply that much time and effort is needed to obtain such data.

Application of a newly developed method by ARYA et al. (1975) can overcome practical limitations expressed in the fourth point. In this quick method, an undisturbed small column of soil in a cylinder (diameter 5 cm, height 9 cm) is saturated. Then, hot (130° C) air is blown towards the exposed upper surface of the otherwise entirely closed core. This happens while the cylinder is standing on a balance, which allows very frequent measurements of total weight. The hot air is turned off after approximately 20-25 minutes. The decrease of weight during the measurement is plotted on a graph to see if a plot of weight versus $\sqrt{t}$ forms a straight line. This is generally true after a short irregular initial period, and this observation allows the application of the Boltzmann substitution which allows the mathematical solution of the flow equation (KLUTE, 1972). As soon as the hot air is turned off, the core is gently pushed from the cylinder and is cut in small, 5 mm thick fragments from which the gravimetric moisture content is determined. This results in a graph that expresses the moisture content (θ) as a function of depth in the core (x). Then, D (diffusivity) can be calculated as follows:

$$D(\theta) = \frac{1}{2t} \cdot \frac{dx}{d\theta} \int_{\theta_x}^{\theta_i} x.d.\theta \, .$$

This procedure can only be applied when the porous medium acts as "semi-infinite".

In other words, only when the moisture content at the bottom of the core does *not* change during the experiment. Empirically, an experimental time of about 20-25 minutes has been found to satisfy this requirement, whereas required low moisture contents at the top of the core were also obtained during this period. As discussed, K(h) relations can be derived from D(θ) as follows:

$$K\ (h) = D(\theta) \cdot \frac{d\theta}{dh}$$

where the latter value is the moisture capacity C, the slope of the moisture retention curve (section 3.3). This method is being tested at present at our laboratory. ARYA et al. (1975) reported good results; also in comparison with other methods. Some major advantages of the method are obvious: It is quick, cheap and simple. More testing is necessary before a final judgement can be made, but the first results are encouraging. However, also with this method results at negative pressures near saturation are not very reliable due to difficulties in determining slopes of rear vertical dθ/dx lines in the lower part of the core (see also WIND, 1966). Again, the crust test may be used for that moisture range.

4.5 Some examples of flow problems

4.5.1 Introduction

Two types of mathematical expressions describing the movement of water in soil, are found in the literature. One type is derived empirically from experimental observations in the field in a particular area of watershed, with only minor concern for the physical principles involved (for example, HOLTAN et al., 1975). Results relate to that particular area alone and cannot be easily extrapolated to other and different, areas. The other type is derived from flow equations and gives some insight into the manner in which various soil physical properties, such as hydraulic conductivity and the moisture characteristic, influence water movement. The latter approach can follow two procedures: (i) Partial non-linear differential equations which describe moisture flow in the soil (see Chapter 4) are solved analytically (for example: PHILIP, 1969; HILLEL, 1971, BAVER et al., 1972). These equations are difficult to solve even for simple boundary conditions. Considerable knowledge of mathematics, physics and computer science is required to understand the principles and solutions of flow problems. (ii) The partial differential equations are solved by numerical techniques using a computer (WIND and VAN DOORNE, 1975). Special computing systems have been developed to handle problems of numerical integration. Simulation languages are available that do not require advanced mathematical or computer programming knowledge (for example, DE WIT and VAN KEULEN, 1972; VAN DER PLOEG, 1974). Numerical methods, which use the computer, have also been applied to simulate soil moisture regimes for fields or entire regions, rather than for single soils only (for example, DUFFY et al., 1975; DE LAAT et al., 1975). The basic limitations of the empirical approach and the complexity and restricted applicability of the analytical techniques put an increasing emphasis on numerical methods as a practical means for predicting field soil moisture regimes. A discussion of these techniques and methods is beyond the scope of this text and the reader is referred to the literature for examples and details. Of interest here is the general observation that both analytical and numerical methods need basic physical data in terms of hydraulic conductivity (K) as a function of θ or h and the moisture characteristic, preferably in terms of adsorption and desorption. Any discussion of these characteristics, their methods of determination and, particularly, relationships with soil textural and structural features (to be discussed in Chapter 5) has therefore practical significance. Computers are, of course, not always available to solve flow problems. There remains a need for simple, often approximate methods that can be used by the field soil scientist to predict certain aspects of hydrological behaviour BOUMA, 1973). Unsteady flow systems where flow rates and hydraulic gradients vary with time, generally cannot be simulated satisfactorily by approximate methods. However, this can be done in systems where a steady flow condition exists. Two examples will be given of simple procedures which are of interest for field soil scientists. The first concerns steady upward and downward flow through unsaturated soil away from and towards a water-table, and the second concerns steady infiltration through surface crusts.

4.5.2 Steady up- and downward flows in unsaturated soil

The Darcy flow equation for vertical flow, as discussed in section 4.1, can be written as:

$$v = -K \left(1 + \frac{dh}{dz}\right)$$

where: $\frac{dh}{dz}$ = gradient of the pressure potential (cm cm^{-1}), K = hydraulic conductivity (cm day^{-1}) and v is flux (cm^3 cm^{-2} day^{-1}).

Integration yields:

$$\int dz = -\int \frac{K}{v+K}\, dh.$$

K varies with h. Therefore the term $\frac{K}{v + K}$ is no constant. So:

$$z_n = -\int_{h_o}^{h_n} \frac{dh}{1 + \frac{v}{K}}.$$

This equation can be integrated by steps by computer (see DE LAAT et al., 1975) but a rather simple graphical procedure can be used as well (BYBORDI, 1968; CHILDS, 1969 p. 224). Z_n = height above a reference layer (a horizon boundary, water-table etc.) at which the pressure h_n is experienced. By choosing a steady flux v (+v for upward flow and —v for downward flow!) and by reading appropriate "average" K values from a K curve (which should be available as a graph) a complete profile of z versus h may be plotted if a sufficiently high number of potential increments is used. This procedure is well suited for calculating h→z curves for downward and upward steady flows in soils consisting of many layers or horizons. The graphical method will be illustrated with one specific example, describing upward and downward flow in a hypothetical profile.

Example calculation. The hypothetical profile consists of a 60 cm silt loam layer on top of sandy loam. Hydraulic conductivity curves (BOUMA, 1975) are presented in Figure 26. Groundwater is (unrealistically) supposed to remain at 1 metre below the soil surface, and the steady upward and downward flow rate has been chosen as 1 cm/day. Calculations are first made for steady *downward* flow and are started at the groundwater level, where h = 0. First, the height Z above this level is determined where an h of, say, —10 cm is experienced (△h = —10 cm). More exact results are obtained when smaller values of △h are used and this is necessary when K changes strongly with increasing h, for example in a sand with low moisture contents. Some representative K value must now be defined for the △h selected. WIND and VAN DOORNE (1975) analysed this problem and concluded that it is rather dangerous to use the arithmetic mean or any other mean of the two relevant conductivity values. when hydraulic gradients are high. Realistic and representative K values can, of course, only be obtained by integration. But for the approximate method discussed here, which applies to steady state flow systems with low hydraulic gradients, the average K value (Fig. 26) has been used for the selected △h. It follows:

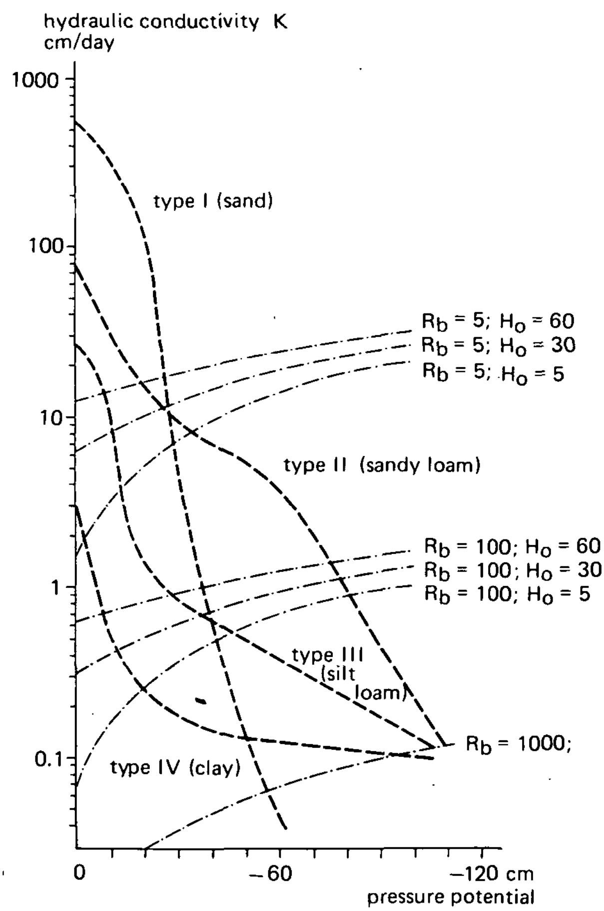


Fig. 26. Hydraulic conductivity curves for four major types of soil and curves expressing the hydraulic effects of impeding barriers of different resistances (see text).

$$Z_{h=-10} = -\frac{1}{1+\frac{v}{K}} \cdot \int_{h=0}^{h=-10} dh = -\left[\frac{-h}{1+\frac{v}{K}}\right]_{h=0}^{h=-10}$$

$$= -\left[\frac{-10}{1-\frac{1}{57}}\right] = 10.18 \text{ cm.}$$

where: $K = 57$ cm day^{-1} is the average K value in the h range considered. In other words, for $v = -1$ cm day^{-1}, a pressure of —10 cm is experienced at 10.18 cm above the water-table. This is close to static equilibrium. Other heights are similarly calculated:

$$Z_{h=-20} = -\frac{1}{1+\frac{v}{K}} \cdot \int_{h=-10}^{h=-20} dh = -\left[\frac{-h}{1+\frac{v}{K}}\right]_{h=-10}^{h=-20}$$

$$= -\left[\frac{-10}{1-\frac{1}{26}}\right] = 10.4 \text{ cm.}$$

A pressure of —20 cm is thus reached at 20.58 cm above the water-table etc. Note that the selection of $\triangle h = -10$ cm, involves averaging very different K values. For example, from zero to —10 cm, the average K = (80 + 35)/2 = 57 cm/day. But these differences are hardly expressed in the calculated Z values. Substituting K = 80 cm/day, rather than 57 cm/day, results in $Z_{10} = 10.12$ cm. Similarly K = 35 cm/day corresponds with $Z_{10} = 10.29$ cm. Such differences are insignificant. If the entire profile were composed of sandy loam, a pressure of —72 cm would be found at the soil surface (Fig. 27). The hydraulic conductivity of 1 cm/day corresponds with an equilibrium pressure of —80 cm (see K curve sandy loam in Fig. 26). The calculated curve does not reach this value but approaches this line (Fig. 27). However, a different soil material is found at 40 cm above the groundwater. A pressure of —38 cm is found at the interface as follows from the graph (Fig. 27). Now the integration procedure is continued *using the K curve for the silt loam,* realizing that in this soil material a K of 1 cm/day corresponds with a pressure of —20 cm (Fig. 26). The integration must therefore proceed from h = —38 cm to h = —30 cm.

$$Z_{h=-30} = -\frac{1}{1+\frac{v}{K}} \cdot \int_{h=-38}^{h=-30} dh = -\left[\frac{+h}{1+\frac{v}{K}}\right]_{h=-38}^{h=-30}$$

$$= -\left[\frac{8}{1-\frac{1}{0.68}}\right] = 17 \text{ cm.}$$

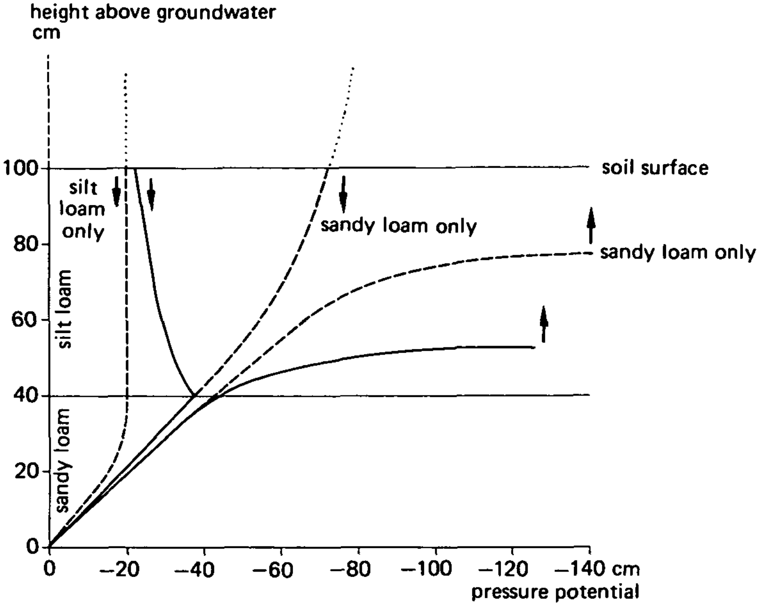


Fig. 27. Pressure potentials in a hypothetical two-layer soil profile during steady up- and downward flow. Hypothetical curves are included for a profile entirely consisting of sandy loam.

In other words, a pressure of —30 cm is found at 17 cm above the interface (Fig. 27). The remainder of the curve is calculated, as discussed. The pressure at the soil surface is —22 cm, because the equilibrium pressure-line at —20 cm is not reached due to insufficient thickness of the silt loam layer.

Calculation of steady *upward* flow involves the introduction of a positive flow rate. Calculations are again started at the groundwater level for the 1 cm/day rate and results are reported in Figure 27. The chosen flow rate could be maintained to a higher level in the sandy loam as compared with the silt loam. If the entire soil had had a sandy loam texture, a capillary rise would have occurred at a rate of 1 cm/day to a height of 80 cm above the groundwater. However, occurrence of the silt loam reduces this height to about 55 cm.

Numerous calculations can be made, varying thicknesses and hydraulic characteristics of the layers used. Moisture retention data can be used to translate Z → h curves into Z → θ curves, allowing an estimate of water- and air contents. Examples, in which this method was applied, were presented by Bybordi (1968), Bouma (1973), Bouma et al. (1975b), Giesel et al. (1973) and Renger et al. (1975).

4.5.3 *Steady infiltration through surface crusts*

A special case of the two-layer flow system occurs when very thin layers with different hydraulic properties occur in a flow system. An example will be discussed concerning thin barriers (crusts) on top of infiltrative surfaces (Hillel, 1971; Falayi and Bouma, 1975).

Assuming steady infiltration (as will be occurring during the crust-test measurement, discussed in section 4.4), the flux through the crust (v_b) should be equal to the flux in the sub-crust soil (v_s).

$$v_b = v_s \text{ or } K_b \left(\frac{dH}{dz} \right)_b = K_s \left(\frac{dH}{dz} \right)_s \tag{4.8}$$

where K_b and K_s are hydraulic conductivities of the barrier and the underlying soil with dH/dz the hydraulic head gradient in both materials. The hydraulic head gradient will be approximately unity in the soil at steady infiltration (BAVER et al., 1972). Assuming flow in the soil thus to result only from gravitational forces:

$$v = K_{s(M)} = K_b \cdot \left(\frac{H_o + M + Z_b}{Z_b} \right) \text{ or}$$

$$\frac{K_{s(M)}}{H_o + M + Z_b} = \frac{K_b}{Z_b} = \frac{1}{R_b} \tag{4.9}$$

where: $K_{s(M)}$ is the unsaturated K value of the soil at a pressure of —M cm, H_o is the positive hydraulic head on top of the barrier by ponded liquid, Z_b is the thickness of the barrier and R_b = hydraulic resistance of the barrier. R_b can be determined from Darcy's law as applied to the barrier:

$$v = K_b \cdot \frac{\Delta H}{Z_b} \quad \text{or} \quad v = \frac{K_b}{Z_b} \cdot \Delta H = \frac{\Delta H}{R_b} \tag{4.10}$$

ΔH can be determined as: $H_o + M + Z_b$. When the flux v is known, from the measured tensiometer pressure and the K curve, R_b can be determined from 4.10. K_b can be calculated if barrier thickness (Z_b) is known; otherwise the use of R_b is the most appropriate.

The hydraulic effects of barriers can thus be predicted using equation 4.9 when K curves are available for the soils below the barriers and when R_b values are known for the barriers themselves.

This is illustrated in Figure 26 where K curves, measured in situ with the crust test are shown for a sand, a sandy loam, a silt loam and a clay soil. The other curves were derived from equation 4.9 assuming different R_b and H_o values, and a value of 2 cm for Z_b. The curves are composed of all points where the relationship between $K_{s(M)}$ and M, as expressed by equation 4.9, is valid for the assumptions made. Curves were drawn in Figure 26 for R_b = 5, 100 and 1,000 days, combined with H = 5, 30 and 60 cm (for R_b = 1,000 only H_o = 5). Points of intersection of both types of curves represent the only possible hydraulic conditions, in terms of pressures below barriers and flow rates, at the specified H_o, Z_b and R_b values. Some conclusions of practical interest can be drawn from Figure 26: (i) Infiltration rates decrease and pressures below the barrier decrease as the resistance of the barrier increases. The effects are a function of the capillary properties of the underlying soil, as expressed by the K curve (section 4.2). (ii) Identical barriers induce different moisture pressures in different soils because their hydraulic effect is not only dependent on their own resistance but also on the capillary properties of the underlying porous medium (see equation 4.9). For example, a crust with R_b = 100 days, H_o = 5 cm induces pressures of —80 (sandy loam), —45 (silt loam), —40 (sand) and —20 cm (clay). Associated flow rates are 0.9, 0.55, 0.5 and 0.25 cm/day, respectively. The statement: "The ultimate rate of acceptance of the soil is identical to the rate of acceptance of the barrier", which is frequently quoted is physically incorrect. (iii) Increasing the hydraulic head on top of a barrier with fixed R_b increases the flow rate and increases pressures in the soil, but effects are generally minor. For example, a barrier with R_b = 100 days induces flow rates of 0.5 cm/day (H_0 = 5), 0.7 cm/day (H_0 = 30) and 1 cm/day (H_0 = 60) in sand. Corresponding tensions are 42, 40 and 38 cm, respectively. The effect of increasing the head is a function of the capillary properties of the porous medium and thus, the shape of the K curve. (iv) Barriers with a small resistance (R_b = 5 days) will *not* affect the clay soil (except for H_0 = 5 where a pressure of 3 cm and a flow rate of 1.8 cm/day is induced). In fact, the $R_b \rightarrow H_0$ curves reflect hydraulic conditions *imposed* by the crust and the K curves those *allowed* by the soils. The most limiting of the two determines conditions if the curves do not cross.

4.6 Flow through pedal soil: hydrodynamic dispersion

4.6.1 Introduction

The flow velocity of water defined in section 4.1 as part of the Darcy equation is a bulkflow velocity which is derived by observing the rate of movement of a water surface. In other words, if a water surface on top of, say, a soil core moves down at a rate of 1 cm per day, we define v as 1 cm/day. This is also the case with the crust-test measurement where the volume of liquid that is supplied by the burette is measured (section 4.4.1.1). This volume is translated into a flux by dividing the volume by the total area of infiltration. *So, in fact, a hypothetical velocity — which the water would have if flowing through the given cross-section unobstructed by solid particles — is derived.* The real flow velocity in the soil pores is higher because at least 40% of the soil is composed of solid particles. This is of practical importance for all soils but particularly for the finer textured soils where flow occurs along the ped faces and through other larger voids which may contribute very little to the total pore volume (section 4.3).

If the soil were composed of simple capillary tubes of specific sizes, calculations of the real flow velocity in those pores would be easy. However, pores vary in shape, width, and direction, and the actual flow velocity in the soil pores is variable. At the best, therefore, one can refer to some "average" velocity (v′) that can be calculated on the basis of the water-filled porosity at each tension.

$$v' = \frac{v}{\theta}$$

where: θ is the water-filled porosity ($cm^3.cm^{-3}$) as derived from the moisture retention curve. At unit hydraulic gradient, we find:

$$v' = \frac{K}{\theta}.$$

The "average" velocity v′ is often still a rather useless characteristic. Flow velocities vary considerably within irregular, natural porous systems (see section 4.3).

It is much more interesting to know the maximum and the minimum velocities for a given flow regime which occur simultaneously during flow through large and fine pores rather than the "average" velocity. Use of tracing techniques can provide data that can be used to analyse such heterogeneous flow patterns. The percolating water can be replaced by a solution of $CaCl_2$, which is used as a tracer, and the chloride concentration in the column effluent can be monitored as a function of time to obtain a measure for the range of flow velocities within the sample. Chlorides are used because they are not adsorbed by the soil particles. If the maximum and minimum velocities are both close to the average velocity (for example in a sand) we will find a relatively long period in which water (without chlorides) that was initially present in the column, is replaced ("pushed out") by water with chlorides. There is a relatively well defined, sharp boundary that moves down the column, separating the two liquids. As soon as all the water is replaced, the column effluent will contain chlorides and the effluent concentration will be identical to the concentration of the applied chloride solution. This type of displacement is called: "piston-flow".

Maximum and minimum velocities may differ considerably in, for example, fine textured soil materials with channels and planar voids. Then, chlorides may be found very soon in the column effluent because fast movement occurs along the larger voids during saturated flow, whereas very slow movement is found simultaneously through the fine porous peds. The concentrations of chlorides in the column effluent will therefore show quite a different pattern in time as compared with conditions where the maximum and minimum velocities are close together. The first chlorides will appear long before all the untraced initially present water is displaced from the column. The column effluent will reach the chloride concentration of the applied chloride solution after displacement of *more* than one pore volume because the chloride solution (which flows rapidly through the larger voids) mixes in the column effluent with untraced water which is slowly displaced from the fine porous aggregates at the same time. There is no well-defined boundary moving down the column, separating the two liquids. These flow regimes which are schematically represented in Figure 29 can *sometimes* be physically characterized in terms of the apparent dispersion coefficient (D), if breakthrough curves follow an "ideal" pattern.

The phenomenon of uneven displacement of one liquid by another is called "Hydrodynamic Dispersion" (see for example BRENNER, 1962). This process is of great practical interest for certain problems relating to physical soil behaviour. The rate of vertical movement of liquid waste or chemical fertilizer (particularly nitrates) through a soil may be highly underestimated if a hypothetical vertical flow rate is estimated by $v' = K/\theta$ as discussed, where θ refers to the total water-filled porosity. Few planar voids or channels occupy a much smaller porosity but may conduct large quantities of liquid in a short time (section 4.2). This can be further illustrated by considering saturated flow through a hypothetical 30 cm high soil core with a diameter of 11 cm and a porosity of 50 %. Assume that the measured flux is 15 cm/day. If water is only shallowly ponded on top of the core we find that $v = K$. The "average" flow velocity in the pores (v') is $100/50 \times 15 = 30$ cm/day.

This implies that an estimate can be made of the time ("travel-time") needed for a particle of liquid to pass through the soil core. This would, obviously, be $30/30 = 1$ day. However, assume that the core consists of a very slowly permeable clay with one continuous channel with a diameter of 500 micron. This channel contributes only 2×10^{-3} vol % to total porosity but accounts fully for the measured permeability of 15 cm/day (section 4.2). We now can calculate the real v in the channel as $100/(2 \times 10^{-3}) \times 15$ cm/day $= 750\,000$ cm/day and a "travel time" of only 3.5 seconds rather than 1 day. This estimate is more realistic. Fortunately, such pores are not always continuous in larger natural soil samples, but still the example illustrates the problems involved.

A specific example relating to infiltration into well structured soil was presented by Magdoff (in BOUMA et al., 1974b). He calculated the theoretical pattern of infiltration, using a measured K-curve and moisture retention data defined in terms of desorption *and* adsorption. The calculation method, which assumes that the soil is isotropic, uses a numerical computer analysis (HANKS et al., 1969). Results (Fig. 28) indicate that the model predicts a much stronger initial wetting than is measured in the field, where a high infiltration rate of 3.2 cm effluent in 20 min can only be explained by flow along larger vertical pores, bypassing peds. Tensiometric measurements are governed by conditions inside peds where the contact between soil and cup is made. This explains why the tensiometer reacts slowly and why the pressure potential does not reach values higher than —45 cm; even though the model predicts saturation (quite erroneously). Unexpected rapid vertical movement of liquid waste or fertilizer offers obvious practical problems. Both processes may remove nutrients beyond rooting depth and may also result in groundwater contamination. Hydrodynamic dispersion creates another, different problem. Irrigation practices or leaching of soil with pure water to remove salts or pollutants, both require efficient use of usually scarce water. Preferential vertical movement along larger pores during saturated flow results in a relatively high water use before complete leaching is achieved in the slowly permeable soil between these pores. The ability to predict such flow patterns for different soils would be of significant practical interest. Use of K data alone cannot provide answers, but the theory of hydrodynamic dispersion may be very useful, as will be shown in the following sections.

4.6.2 Theory of hydrodynamic dispersion

The theory of hydrodynamic dispersion and its application to soil science has been discussed by many authors (for example: BRENNER, 1962, ROSE and PASSIOURA, 1971; CASSEL et al., 1974). The process is complicated and involves subtle interaction between

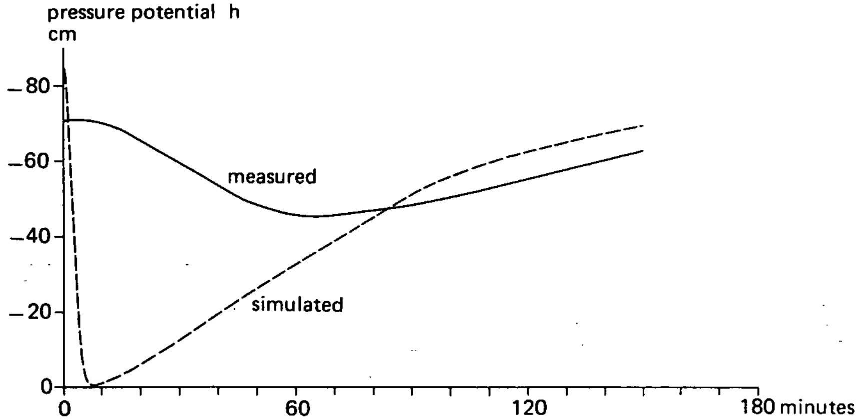


Fig. 28. Infiltration and redistribution of a 3.2 cm dose of effluent on a silty clay-loam soil horizon, comparing measured and simulated curves for a depth of 5 cm below infiltrating surface.

convection and molecular diffusion. It seems, however, well established that dispersion can be formally described by a one-dimensional diffusion equation in which the coefficient of diffusion is replaced by one of dispersion. Most of the experimental work on dispersion has consisted of following the break-through curve of one solution as it displaces another from a column, as discussed earlier. To explain this, we must first briefly review the underlying dispersion theory.

The following equation is assumed to describe hydrodynamic dispersion during steady flow:

$$\frac{\delta C}{\delta T} + v' \frac{\delta C}{\delta X} = D \cdot \frac{\delta^2 C}{\delta X^2} \tag{4.11}$$

where: T is the time from the commencement of the displacement, X is the distance from the point of introduction of the displacing fluid, C = solute concentration, D = dispersion coefficient and v′ is the "average" velocity of the fluid.

Dimensionless variables can be introduced to allow a general application of the equations based on relative dimensions:

$$c_e = \frac{C_e - C_f}{C_o - C_f}, \quad x = \frac{X}{L} \quad \text{and} \quad t = \frac{v'T}{L}$$

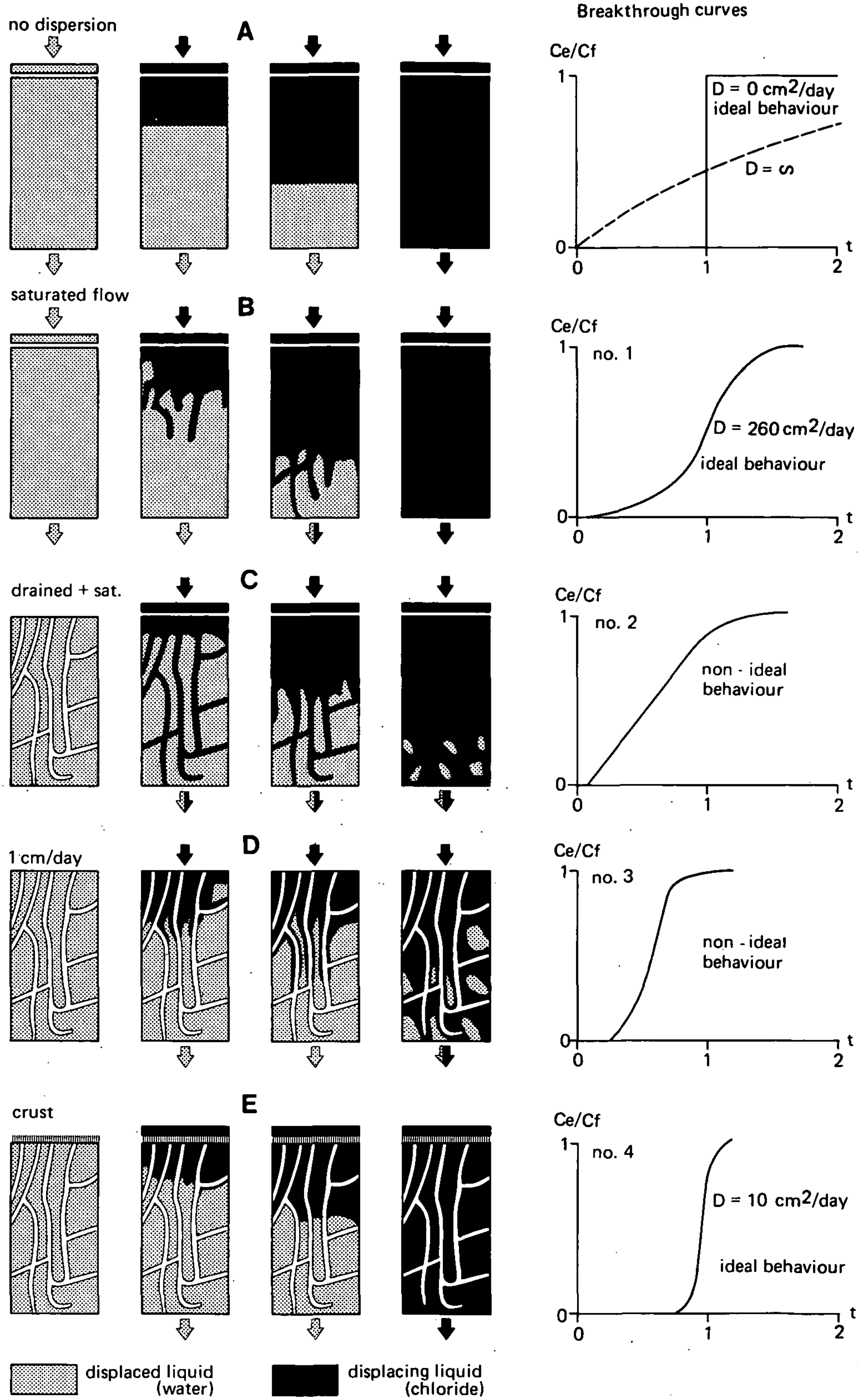
where: C_e = exit concentration; C_o = initial concentration of the liquid in the column and C_f = concentration of incoming liquid, L = length of the column. The dimensionless time t corresponds physically to the number of water-filled pore displacements introduced into the medium since the start of the experiment. Equation 4.11 can now be transformed to:

$$\frac{\delta c}{\delta t} + \frac{\delta c}{\delta x} = \frac{1}{4P} \cdot \frac{\delta^2 c}{\delta x^2} \tag{4.12}$$

where: P (= v′L/4D) is the Péclet number.

This equation can be solved for particular boundary conditions that were discussed by BRENNER (1962) and ROSE and PASSIOURA (1971). The reader is referred to these and other publications for further details. Of particular interest for this review is the manner of presentation of data. Brenner published tables and graphs which present the solutions of equation 4.12 in a very accessible form. One such table is reproduced as Table 4 and shows the dimensionless exit concentration c_e as a function of Péclet number P (=v′L/ 4D) and t (=v′T/ L). Use of the tables or figures can be illustrated by analysing break-through data (Fig. 29). The concentration of column effluent can be expressed as c_e (as discussed) or as 1 — c_e which produces visually more attractive curves. The 1 — c_e value is equal to C_e/C_f when C_o = 0. This dimensionless ratio is *zero* as long as no displacing liquid leaves the column and *one* when the column effluent has the concentration of the incoming displacing liquid. Then, all initially present liquid has been removed. Breakthrough curves always show the exit concentration as a function of the number of water-filled pore volumes passed. Or better, expressed as t = v′T/L. Unfortunately, different procedures are followed in reporting data. Sometimes 1 — c_e or c_e is used, with or without logarithmic scales. Also t is presented both ways. This may be confusing.

A breakthrough curve yields values of c_e (which are measured by determining the chloride concentrations in the column effluent as a function of time) and t (calculated



no dispersion
A
Breakthrough curves
Ce/Cf
D = 0 cm2/day
ideal behaviour
D = ∞
saturated flow
B
no. 1
D = 260 cm2/day
ideal behaviour
drained + sat.
C
no. 2
non - ideal
behaviour
1 cm/day
D
no. 3
non - ideal
behaviour
crust
E
no. 4
D = 10 cm2/day
ideal behaviour
t
0
1
2
displaced liquid
(water)
displacing liquid
(chloride)

Table 4. Table from BRENNER (1962) with dimensionless exit concentrations (c_e) as a function of the Péclet number (*P*) and dimensionless time (*t*). Such tables can be used to determine dispersion coefficients if the breakthrough curve indicates "ideal" behaviour.

t	*P* = 1	*P* = 2	*P* = 4	*P* = 6	*P* = 8
0.00	1.000 (0)	1.0000 (0)	1.0000 (0)	1.0000 (0)	1.0000 (0)
0.10	0.9999	1.0000	1.0000	1.0000	1.0000
0.20	0.9951	0.9997	0.9999	1.0000	1.0000
0.30	0.9635	0.9943	0.9998	1.0000	1.0000
0.40	0.8985	0.9666	0.9958	0.9994	0.9999
0.50	0.8121	0.9059	0.9716	0.9915	0.9972
0.60	0.7177	0.8170	0.9126	0.9559	0.9768
0.70	0.6245	0.7127	0.8142	0.8733	0.9110
0.75	0.5802	0.6588	0.7541	0.8140	0.8556
0.80	0.5378	0.6054	0.6898	0.7455	0.7865
0.85	0.4975	0.5535	0.6236	0.6710	0.7071
0.90	0.4597	0.5037	0.5578	0.5941	0.6218
0.95	0.4241	0.4566	0.4941	0.5178	0.5354
1.00	0.3909	0.4125	0.4338	0.4448	0.4517
1.05	0.3600	0.3715	0.3778	0.3769	0.3731
1.10	0.3313	0.3336	0.3267	0.3157	0.3038
1.15	0.3017	0.2989	0.2806	0.2615	0.2435
1.20	0.2801	0.2673	0.2396	0.2144	0.1923
1.25	0.2573	0.2386	0.2035	0.1741	0.1493
1.30	0.2363	0.2126	0.1721	0.1403	0.1153
1.35	0.2170	0.1891	0.1448	0.1122	0.8775 (—1)
1.40	0.1992	0.1681	0.1214	0.8905 (—1)	0.6612
1.45	0.1828	0.1492	0.1014	0.7026	0.4936
1.50	0.1677	0.1323	0.8446 (—1)	0.5512	0.3654
1.60	0.1112	0.1037	0.5807	0.3341	0.1958
1.70	0.1188	0.8110 (—1)	0.3954	0.1990	0.1023
1.80	0.9987 (—1)	0.6327	0.2670	0.1168	0.5224 (—2)
1.90	0.8397	0.4926	0.1791	0.6769 (—2)	0.2620
2.00	0.7058	0.3830	0.1195	0.3882	0.1294
2.20	0.4986	0.2308	0.5242 (—2)	0.1244	0.3033 (—3)
2.40	0.3522	0.1386	0.2268	0.3881 (—3)	0.6827 (—4)
2.60	0.2487	0.8308 (—2)	0.9714 (—3)	0.1186	0.1490
2.80	0.1756	0.4972	0.4128	0.3571 (—4)	0.3175 (—5)
3.00	0.1240	0.2973	0.1744	0.1062	0.6512 (—6)
3.50	0.5195 (—2)	0.8204 (—3)	0.1991 (—4)	0.4938 (—6)	0.1253 (—7)
4.00	0.2176	0.2260	0.2241 (—5)	0.2222 (—7)	0.2238 (—9)
4.50	0.9116 (—3)	0.6224 (—4)	0.2504 (—6)	0.9790 (—9)	0.3861 (—11)
5.00	0.3819	0.1714	0.2787 (—7)	0.4267 (—10)	0.6531 (—13)

←

Fig. 29. Schematic diagrams illustrating chloride displacement patterns in pedal soil as a function of the flow regimes and measured breakthrough curves for columns of Table 5. The upper diagram represents lack of dispersion and the adjacent graph also shows a breakthrough curve for infinite dispersion (not shown in a diagram).

Table 5. Travel-times of liquid to first breakthrough and to total displacement (300 ppm in effluent) for four flow regimes in ten columns with two types of natural structure (numbers between parentheses refer to cumulative outflows). Each number is the average of five measurements. Schematic diagrams of the displacement patterns in columns 1 through 4 are represented in Figure 29.

Flow regime	First trace	300 ppm	Initially water-filled	D cm^2/day
Column type subangular blocky				
First series:				
Saturated flow (9 cm/day)	18 hrs (550)	5 days (3240)	1680 cm^3	260
Second series:				
Drained + saturated flow	1 hr (84)	2 days (1920)	1525 cm^3	—
Third series:				
Dose: 1 cm/day	3 days (273)	24 days (1470)	1525 cm^3	—
Fourth series:				
Crust: 1.1 cm/day	8.6 days (750)	21 days (2160)	1525 cm^3	10
Column type prismatic				
First series:				
Saturated flow (1.6 cm/day)	21 days (1390)	38 days (2780)	1770 cm^3	2
Second series:				
Drained + saturated flow	15 days (181)	42 days (2055)	1645 cm^3	260
Third series:				
Dose: 1 cm/day	10 days (680)	32 days (2014)	1645 cm^3	10
Fourth series:				
Crust: 0.22 cm/day	49 days (820)	104 days (1650)	1645 cm^3	0

from $t = v'T/L$). The Brenner table lists a number of combinations of c_e and t that belong together according to dispersion theory, with its specific boundary conditions, for one particular P value.

Assume that a measured curve represents combinations of c_e and t that correspond approximately with the listing given by Brenner for $P = 4$. The observed dispersion process can thus be described by $P = 4$. The apparent dispersion coefficient can be derived from $D = v'L/4P$, because, v' and L are either known or defined. The two extremes of dispersion are shown in Figure 29A. Lack of dispersion ($D = 0$, $P = \infty$) is shown by a straight vertical line at $t = 1$. The column effluent has the concentration of the incoming liquid as soon as all liquid initially present is "pushed out". Total dispersion ($D = \infty$; $P = 0$) is found when incoming liquid leaves the column as soon as it is introduced (this breakthrough curve is shown only in Fig. 29A). Then, it takes a very long time before the concentration of the column effluent is exactly equal to that of the incoming liquid. This is due to slow diffusion of untraced water from soil that was initially bypassed. Real conditions fall between these two extremes if the dispersion process can be described by equation 4.11, but dispersion theory describes "ideal" behaviour, which corresponds with results obtained by the solution of a mathematical equation. Most natural soils do not have "ideal" behaviour either because of their particular structure or because of the applied flow regime. Descriptions of such systems, which either show "tailing" (PASSIOURA, 1971) or rapid total displacement (ANDERSON and BOUMA, 1977), are theoretically more complex and will not be discussed here (see section 4.6.4).

4.6.3 Practical applications

Hydrodynamic dispersion is discussed because hydraulic conductivity and moisture retention data are often inadequate for predicting the real velocity of liquid in soil. The apparent dispersion coefficient (D) can be used, if available, to predict the maximum flow velocity and the point of breakthrough of a newly applied liquid if "ideal" behaviour of the flow system is expected. In addition, D also will predict the moment of total displacement. The procedure to be followed will now be discussed for a specific example. Assume a 50 cm high soil core with a porosity of 50 % and a measured K_{sat} of 10 cm/day. D is known to be 125 $cm^2\ day^{-1}$. We first estimate v': 100/50 x 10 = 20 cm day^{-1}. Then: P = v'L/4D = 2.0. Now we can use the appropriate Brenner table for determining t by input of P = 2 and the desired (dimensionless) exit concentration. Assuming an influent concentration of 300 ppm, we can calculate the time of appearance of the first trace of chloride (say 5 ppm) by estimating c_e as $\frac{5 - 300}{0 - 300} = 0.9833$ (liquid in column has $C_0 = 0$). This corresponds, according to the Brenner table (Table 4) with approx. t = 0.02. It follows from t = v'T/L that T = (0.2 x 50) / 20 = 0.5 day. Identical calculations can be performed for any other exit concentration. An estimate of the time when total displacement occurs is of interest. Then $c_e = \frac{299 - 300}{0 - 300} = 0.0033$ or 0.33×10^{-2}. We find from the table that t = 2.93 . So : T = (2.93 x 50) / 20 = 7.325 days. It takes about a week before all initial liquid is replaced from the column. Without any known, or assumed D value, only estimates of breakthrough and displacement can be made assuming piston-flow and associated lack of dispersion. This would have resulted in a prediction of a travel time T of 50/20 = 2.5 days. Or, in other words: after 60 hours the column effluent contained the first chlorides and the concentration of column influent (300 ppm) was reached very quickly after that. This estimate would have been a very poor one; and a dangerous one if the newly applied liquid had not contained just chlorides but pathogenic viruses of bacteria as well. A key aspect of this example calculation was the estimate of D. Direct measurement is always the best and most reliable procedure. However, once breakthrough characteristics are measured it does not make sense to turn around and calculate them on the basis of such a measured breakthrough curve and a derived D value. Procedures are therefore needed to estimate D values. Recent research results may have pointed to one possible procedure, although much testing remains to be done. Detailed results of these studies will be presented elsewhere (ANDERSON and BOUMA, 1977), but a summary will be presented here.

Apparent dispersion coefficients were determined with a chloride tracer in 55 cm long undisturbed columns from four pedal soil horizons of three pedons. The dispersion behaviour of five medium subangular blocky structures was compared with five coarse prismatic structures (SOIL SURVEY STAFF, 1951) all with a silty clay loam texture. Undisturbed columns were sampled from the subangular blocky B2 horizons of Morley and Batavia silt loams (Typic Hapludalf and Mollic Hapludalf, respectively, according to the U.S. Soil Classification system, SOIL SURVEY STAFF, 1975), and from the prismatic Almena silt loam B2 and Batavia silt loam B3 horizon (the Almena silt loam is classified as Aeric Glossaqualf). The bulk density of the subangular blocky and prismatic structures was 1.50 g/cm^3 and 1.55 g/cm^3 respectively.

Four series of experiments were made determining breakthrough curves for different flow regimes. Results for both types of structure are reported in Table 5. Schematic drawings are shown in Figure 29B -E, conceptually illustrating the observed patterns for the subangular blocky structure only.

The first series of experiments determined breakthrough curves for saturated flow. Breakthrough curves for the subangular blocky structure corresponded with "ideal" behaviour in terms of dispersion and the average D was 260 cm^2 day^{-1} (Table 5). Quite different behaviour was observed for the five prismatic soil horizons. The average D was 2 cm^2 day^{-1} indicating much less dispersion. The two populations of soil structure, which can be routinely distinguished by field soil scientists, behaved significantly different. This offers perhaps the potential to estimate D values based on a structure description and a texture determination. Research will be continued to further develop such relationships that may allow reasonable estimates of D, which is crucial to allow estimates of the degree of hydrodynamic dispersion. This phenomenon is not restricted to saturated flow. Dispersion may be very pronounced during intermittent flows and may be very low during steady state unsaturated flows when only small pores conduct liquid. Some experiments were made with the same ten soil columns to test and demonstrate these effects and results will be discussed because they have practical significance and may act to further illustrate the dispersion phenomenon and the calculation of D values.

In the second series of experiments, an unlimited quantity of chloride-traced water was supplied to drained columns, which were at hydraulic equilibrium. This implies that larger pores were filled with air in the upper part of the cores because moisture potentials at equilibrium corresponded (in cm) with the height above the bottom of the cores. *Dispersion in all soil colums was very pronounced,* because the traced water moved almost instantly into the air-filled voids reaching the bottom of the cores quickly, *thereby bypassing the peds* (Table 5 and Fig. 29C). This effect was most pronounced, once again, for the subangular blocky structures where more vertical voids apparently offered more opportunities for vertical pore-continuity. These flow processes can, strictly speaking, not be considered in terms of hydrodynamic dispersion because more is involved than displacement of liquid. The breakthrough curve could not be described by dispersion theory (Fig. 29C). Extremely high initial dispersion ($D = 3000$ cm^2/day) should have resulted in total displacement at $t = \pm 4$ days, according to the Brenner tables. In other words it should have taken a very long time to "wash" all untraced water from the interiors of the peds. However, total displacement as evidenced by a column effluent concentration of 300 ppm, was already observed after two days, and the corresponding t value, in turn, was associated with a D of only 1.2 cm^2/day. This changing dispersion pattern may partly be due to some swelling of the drained column after saturation, following the liquid application. Certain pathways that were initially open, may have closed after a few hours as a result of the swelling of peds. A better explanation may be found by recollecting that total displacement is *supposed to have occurred* as soon as column effluent reaches the 300 ppm chloride concentration. *But this condition does not necessarily always imply total displacement of all untraced water from the soil.* Pockets of untraced water within peds may diffuse so slowly downwards and laterally to *pathways where the tracer moves down relatively quickly,* that the dilution of the tracer solution in the column effluent becomes immeasurably small when standard techniques are used.

The theoretical analysis of this problem, considering "mobile" and "immobile" parts of the soil water is complicated and has been the subject of much recent physical research (section 4.6.4). This second experiment yielded results in the prismatic soil horizons that *could* be expressed with some difficulty by the dispersion theory (Table 5). D values averaged 260 cm^2/day, a very strong increase from the 2 cm^2/day measured during steady saturated flow. Fewer vertical larger voids in the prismatic soil may have resulted in the observed difference between the two structures. This second series of experiments was of more practical relevance than the first. Intermittent applications of liquid to soil are common whether we are dealing with natural rainfall, irrigation or liquid waste application practices. Then we must know how fast and how deep liquid will penetrate the soil, and how effective it is in displacing the liquid that was initially present.

The second series of experiments clearly demonstrated extreme "short circuiting" of liquid when applied to drained pedal soil. Methods are needed for reducing possible associated practical problems, such as groundwater pollution or inefficient leaching. Since pore patterns in the soil cannot readily be changed, other methods have to be explored to reduce dispersion. There are two obvious possibilities: (i) Formation of a barrier on top of the surface of infiltration which will not allow flow into larger soil voids and (ii) Reduction of the loading rate to a level where flow will occur *through* rather than *around* the peds.

A third and fourth series of experiments was made to investigate and demonstrate these effects. The same columns were repeatedly used after flushing the chlorides from the previous experiment with water. Results can be summarized as follows: A daily application of 1 cm and a 50 % gypsum-sand crust were chosen to represent these two effects. Use of a 1 cm/day dose strongly reduced dispersion (Table 5, Fig. 29D). The initial breakthrough in the subangular blocky structures corresponded with a D of 100 cm^2/day rather than 300 cm^2/day when the dose was unlimited. Corresponding times were 3 days and 1 hour respectively. However, again the columns did not show "ideal" behaviour at the reduced daily dosing rate for the same reasons as discussed for the unlimited dose. A concentration of 300 ppm in the column effluent was measured at $t = 1$ (rather than $t = 1.2$ when the dose was unlimited). This interesting difference is due to the repeated drainage of the larger pores during the daily 1 cm dose, whereas those pores remained full of water after the unlimited dose was applied (comparing Figs. 29B and D). *The repeated daily drainage results in a more pronounced bypassing of the peds by the intermittently applied liquid.* The effect of applying a 1 cm/day dose to the prismatic columns was quite obvious, reducing D from 260 cm^2/day tot 10 cm^2/day (Table 5). Application of the crust results in an uninterrupted 24-hours a day flow of liquid into the crust-topped soil, *excluding the larger pores* (see section 4.5.2) (Fig. 29E). The effects were of particular interest for the subangular blocky soils, in which the flow rate through the crust averaged 1 cm/day. The crusted columns accepted therefore the same daily quantity of liquid as when the instant daily dose of 1 cm was applied. D for the crusted subangular blocky columns, which showed "ideal" behaviour, averaged a low 10 cm^2/day (Table 5). D values for the crusted prismatic columns were close to zero. The examples discussed apply to only a few case-studies and more work is needed to provide a broader base for soil structure interpretations in terms of potential dispersion behaviour. The reported experiments and a further analysis of the dispersion phenomenon can be broadly summarized as follows.

Whether or not newly applied liquid will disperse as it flows through a soil depends on a number of factors which are often inter-related: (i) the initial moisture condition of the soil which determines lateral gradients, (ii) the rate and procedure of application of liquid, (iii) the lateral hydraulic conductivity of the soil in the peds, (iv) the number and patterns of occurrence of relatively large vertical pores. *The applied liquid will flow downward through relatively large pores in unsaturated soil only if the capacity of the small pores inside the peds to conduct liquid is inadequate (section 4.2).* This can be because of flow barriers at the ped surfaces (clay skins etc.) or a low K and/or a low lateral hydraulic gradient in "wet" peds.

The depth of direct flow through the larger pores will depend on: (i) the vertical continuity of the larger pores, (ii) the lateral hydraulic gradients which "pull" the liquid from the larger voids, and (iii) the quantity of applied liquid. The larger this quantity, the larger the probability that direct flow will extend to a greater depth, provided that this is allowed by the geometry of the larger pores in the flow system. The aspect of vertical pore continuity is of particular interest and will be a function of sample height (see section 4.3.2). Whether D is a linear function of height ($D = v'L/4P$) in pedal soils is still an open question. Attempts by PASSIOURA (1971) to describe dispersion in aggregated porous media, may not always be relevant for natural pedal systems because of "unnatural" very small aggregates used in his test.

Finally, consideration of dispersion phenomena is essential from a practical point of view when soil suitability for liquid waste disposal is determined. Data by GREEN and CLIVER (1974) and ZIEBELL et al. (1975) clearly demonstrate that the transient retention time ("travel time") of a liquid waste within a soil is an important characteristic for purification. In fact, the longer pollutants are in contact with the soil, the higher the probability that they will be removed ("that the liquid waste is being purified") by processes of adsorption and filtration. Saturated flow through a 60 cm column filled with coarse sand did not adequately remove pathogenic viruses but the flow regime of 5 cm/day was very effective in removing 10^6 infectious viral units per ml (GREEN and CLIVER, 1974) (Fig. 30). Fecal indicators passed through 55 cm long columns of a prismatic silty clay loam soil (which was also used for the dispersion experiments discussed in this section) if the fecal solution was added to drained columns at a rate of 10 mm/day. Removal was complete when added to saturated columns (not shown in Figure 30) or to columns to which a 3 mm daily load was applied and the difference can be explained by considering the longer retention times associated with the latter two flow regimes (Table 5). Of particular interest is the difference between the 3 mm and the 10 mm daily loading rate (Fig. 31). The latter rate results in rapid vertical movement along larger natural pores, not allowing adequate contact with the soil and thus resulting in poor purification. The former rate is apparently slow enough to

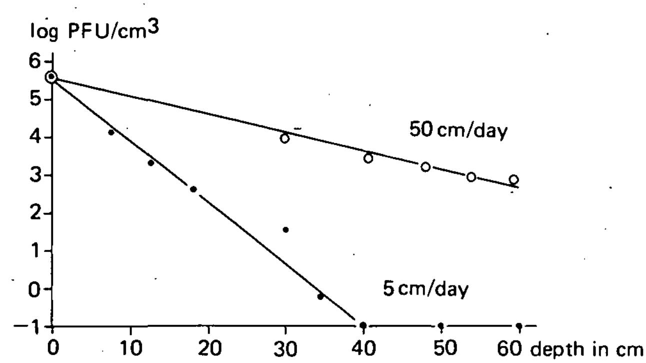


Fig. 30. Penetration of poliovirus type 1 (strain CHAT) into 60 cm long sand columns at room temperature comparing the degree of removal as a function of the flow regime. Input concentrations are extremely high. PFU = Plague Forming Unit (after GREEN and CLIVER, 1974).

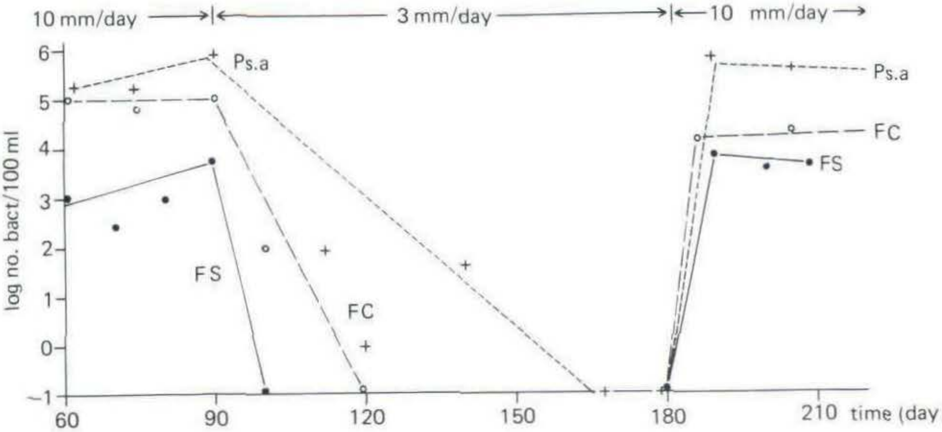


Fig. 31. Concentration of pathogenic bacteria as a function of the flow regime in large 55 cm long undisturbed columns of a prismatic silt loam. FS = Fecal Streptococcus, FC = Fecal Coliform, Ps.a = Pseudomonas aeruginosa. The lower dosing rate induces flow through rather than around the peds and results in better purification (Modified from ZIEBELL et al., 1975).

allow the peds to absorb the liquid laterally and to transport it very slowly downwards, thereby ensuring good contact and purification.

Travel times (determined by a chloride tracer) cannot be compared with those associated with the movement of biological materials or adsorbing chemical compounds. The chloride anions are *not* adsorbed by the negatively charged clay particles and their rate of movement may be considered representative for water. However, other compounds may be adsorbed or oxidized during travel through the soil. Precipitation reactions may also occur. The estimation of the travel times of such compounds is therefore very difficult, requiring additional mathematical equations and is the subject of detailed investigations in several laboratories. Consideration of these effects is far beyond the scope of this text.

4.6.4 Model experiments: the presence of "immobile water"

4.6.4.1 A description of some model experiments

Explanations of dispersion phenomena in natural soils, as previously presented, are of a hypothetical nature because pore patterns are generally heterogeneous and irregular and cannot be known exactly even after detailed two-dimensional micromorphological analysis. Model experiments may therefore be essential for testing certain assumptions (see section 5.3.2 for another example of this approach). Model studies, if properly designed, allow an independent evaluation of the effect of single, well defined factors. Detailed results of these experiments will be published elsewhere (BOUMA and ANDERSON, 1977) but major aspects will be summarized here.

Model columns (10 cm diameter and 30 cm high) were prepared from mixtures of silty clay loam and coarse sand to form clay loam and sandy loam textures. The mixtures were thoroughly mixed when dry. They were then moistened and kneaded until a thick paste was formed which was shaped into a mould to the required dimensions. The columns were allowed to air-dry slowly for two weeks. Following the drying, the columns were slowly remoistened before being permanently placed in tightly fitting plastic cylinders. This procedure induces the formation of a very homogeneous and stable microstructure, representing an ideal "textural porosity" (BOUMA, 1969). Three series of two columns each were prepared from both soil materials. The first series had no large pores. The sandy loam and clay loam columns had saturated hydraulic conductivities of 19 cm/day and 1 cm/day respectively. The second series had one vertical 5 mm cylindrical channel, drilled through the centre, and the third series had three such continuous channels in a triangular pattern (see Fig. 32). The saturated conductivities of these perforated cores were very high and of no interest for the experiments to be reported. Three moisture regimes were compared: saturated flow, daily application of 5 mm liquid and infiltration through a gypsum-sand crust (50 % by volume).

A chloride solution was again used to determine breakthrough curves. Columns were flushed with de-ionized water between measurements. Some major results can be summarized as follows (see Table 6 and schematic drawings in Fig. 33).

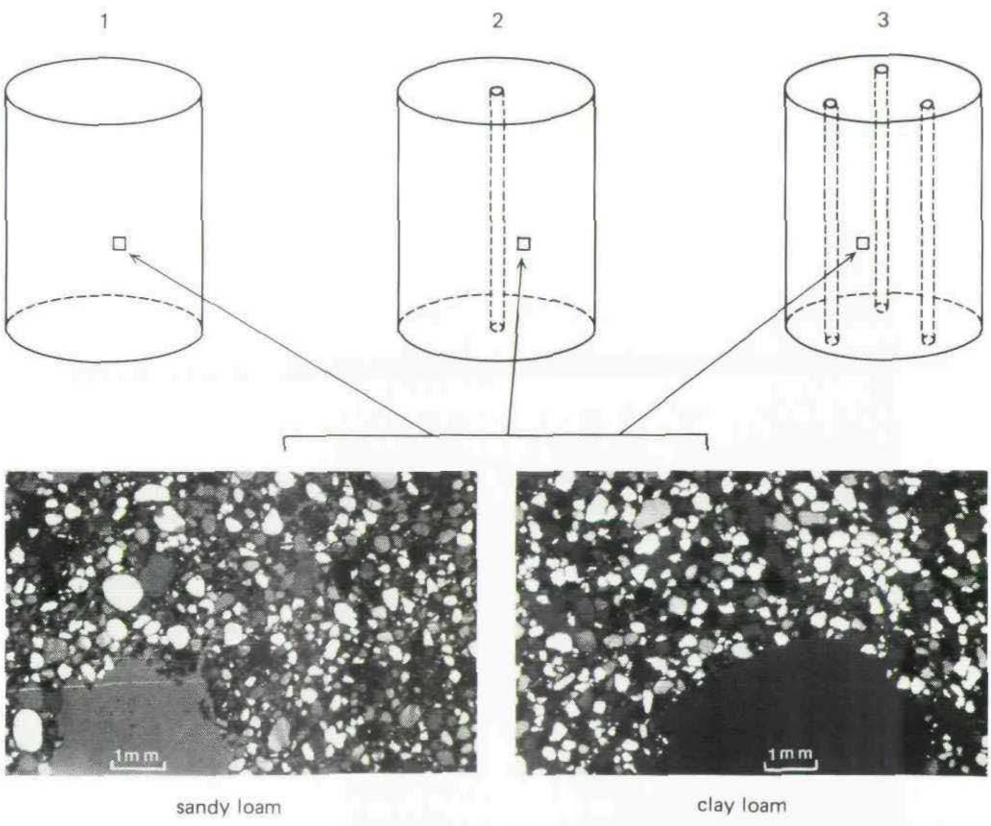


Fig. 32. Three types of artificial cores used for model experiments on dispersion. For these types two characteristic microstructures, as shown on the photo's, have been investigated.

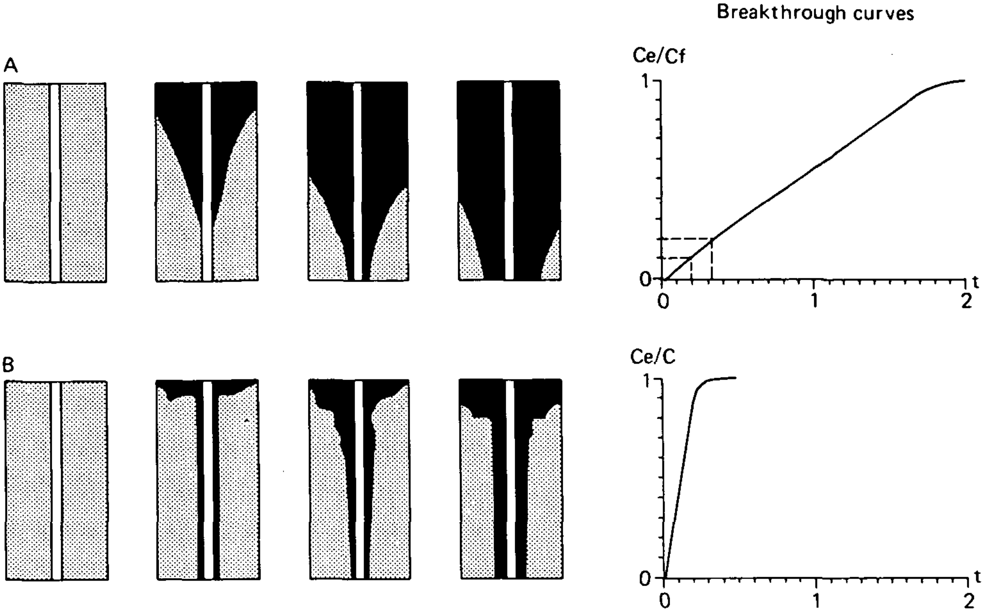


Fig. 33. Schematic diagrams illustrating chloride displacement patterns in artificial sandy loam (A) and clay loam (B) columns with one artificial vertical channel, at a loading rate of 5 mm/day. Corresponding measured breakthrough curves are shown in the adjacent graphs and illustrate "non-ideal" behaviour.

Table 6. Travel-times of liquid to first breakthrough and to total displacement (300 ppm in effluent) for three flow regimes in six soil columns with simulated structures (numbers between parentheses refer to cumulative outflows).

Column type	Flow regime	First trace	300 ppm	Initially water-filled (cc)	D cm^2/day
Sandy loam	Sat. (19 cm/day)	7 hrs (400)	22 hrs (1300)	765	60
Sandy loam-1 channel	Dose: 5 mm/day	12 min (24)	25 days (980)	566	—
Sandy loam-3 channels	Dose: 5 mm/day	40 min (16)	31 days (1250)	520	—
Sandy loam-50 % crust	Crust: 4 mm/day	9 days (250)	38 days (1150)	550 (av.)	2.5
Clay loam	Sat. (1 cm/day)	6 days (570)	19 days (1670)	940	3.8
Clay loam-1 channel	Dose: 5 mm/day	5 min (13)	3 days (130)	809	—
Clay loam-3 channels	Dose: 5 mm/day	7 min (12)	5 days (200)	764	—
Clay loam-50 % crust	Crust: 2.5 mm/day	22 days (849)	64 days (1460)	810 (av.)	0.7

(i) Incomplete displacement of the initially present untraced water was observed for the 5 mm a day liquid applications in all clay loam columns at the time when column effluents reached the concentration of the added chloride solution ($C_e = C_f$).

For example, only 130 cc of liquid had left the column, with one channel, whereas 809 cc of untraced liquid was initially present in the column. This must imply that pockets with untraced water ("immobile" water) remain inside the columns. The daily applied chloride solution will flow down the large vertical pore or pores but will also be pulled into the surrounding soil by (low) lateral gradients of the pressure potential thereby displacing untraced water. Due to the low permeability of the clay loam, the process occurs very slowly and is apparently effective only in a relatively small zone around the vertical channels. The effluent concentration is therefore equal to the influent concentration long before all untraced water is displaced. An estimated 94 % (one vertical channel) and 92 % (3 vertical channels) of the initially present untraced water is still present in the cores at the time when $C_e/C_f = 1$ (see calculation procedure in section 4.6.4.2.) (Fig. 33).

This phenomenon was also found in natural soils with aggregates of several centimetres size, and contrasts with data on "tailing" by PASSIOURA (1971) who discussed dispersion in "aggregated" soils with artificial aggregates of 1-2 mm. Then, distances of diffusion are much smaller and results of dispersion experiments (at low flow rates) may not necessarily be representative for many natural pedal soils. Flow processes in the more permeable sandy loam columns are quite different. The volume of liquid that left the column at the time when $C_e = C_f$, is higher than the volume of untraced water which was initially present in the column (Table 6). However, this does not imply that all initially present untraced water has been displaced from the column. The liquid that leaves the column is composed of a chloride solution flowing (intermittently) through the larger pores *and* of untraced water which is slowly displaced from the soil. An exact quantitative separation between these two quantities of "mobile" versus "immobile" water is complex (VAN GENUCHTEN and WIERENGA, 1976). An estimated 28 % (one vertical channel) and 36 % (three vertical channels) of the initially present untraced water is still present in the cores at the time when $C_e/C_f = 1$ (see calculation procedure in section 4.6.4.2) (Fig. 33). But even without these estimates, the experiments show that *identical large pores have a different hydraulic function in different types of soil materials, due to the different hydraulic behaviour of the surrounding soil matrix.* This conclusion is based on an analysis of simulated small intermittent flows, occurring in reality as rainfall, waste application or irrigation, and does not, of course, apply to saturated flow where large vertically continuous pores will conduct very high quantities of liquid in all soil materials.

(ii) Dispersion patterns are identical in each of the two soil materials when infiltration occurs through a crust. Results for *all* crusted columns are therefore reported together in Table 6 for each soil material. Whether there are three, only one or no vertical channels does not make a difference because these large pores do *not* conduct liquid (section 4.1). Breakthrough characteristics for the sandy loam columns are significantly different from those for the clay loam columns, even though the crusts are identical. The difference is due to the different pore size distributions (section 4.1). Complete displacement is achieved in all columns at $C_e/C_f = 1$. This conclusion is derived from the fact that calculations of D values, as discussed earlier, are possible since breakthrough curves represent "ideal" behaviour (BRENNER, 1962). These D values are very low, indicating that flow processes approach "pistontype" flow, as discussed earlier. Crusts are thus very effective in improving leaching efficiency.

Differences in dispersion between these crusted columns and columns of the same soil material without a crust (during saturated flow) are considerable. This is particularly true for the sandy loam texture (Table 6). In both cases flow occurs through "the fine-pores"; either because flow through large pores (which are present) is not possible because of the crust, or because large pores are not present. Note that the microfabrics themselves are, of course, identical. The observed differences in dispersion during flow through "the fine pores" in each of the two soil materials are of interest. Conceptual distinctions between "large" structural pores and "fine" textural pores sometimes seem to suggest, unintentionally perhaps, two rather distinct levels of soil physical behaviour (KLUTE, 1973). Similar separations are made in micromorphology (BREWER, 1964; BOUMA and ANDERSON, 1973). Our observations indicate that such distinctions may have less physical meaning than is often implied.

(iii) Breakthrough data for columns with one or three vertical channels were not very different for each of the two soil materials. Very soon all have a trace of chloride in the column effluent (Table 6). One important phenomenon is shown by the data in this table. Columns with three holes allow a somewhat better lateral absorption of the chloride solution than those with one hole. This is evident from the slightly *longer* times until the appearance of the first trace of chlorides and from the larger cumulative outflows at the time when $C_e = C_f$. In other words, in the latter case chloride solution flowing (intermittently) along the larger channels is diluted somewhat more efficiently by untraced water being displaced from the column, when three channels are present. However, these differences are minor when compared with those obtained for the different soil materials. This can be specifically illustrated by comparing the estimated volumes of "immobile" water at $C_e/C_f = 1$ (see point ii). The values for each soil material, which express the effect of the number of vertical channels, are close together (94 % and 92 % for the clay loam, and 28 % and 36 % for the sandy loam for one and three channels, respectively). These values are, however far apart when the different soil materials are compared ignoring the effect of the channels (93 % versus 32 %). The number of vertical channels simulates the number of natural vertical voids, such as worm or rootchannels and open planar voids between peds. Structures with small peds will have more planar voids per unit horizontal soil surface than those with large peds. Of course, channels in this model were continuous throughout the sample, whereas larger pores in natural samples are generally not (BOUMA and ANDERSON, 1973).

4.6.4.2 Estimation of volumes of "immobile" water

The calculation and prediction of volumes of "immobile" water is complicated (see VAN GENUCHTEN and WIERENGA, 1976 and references). An estimate can however be made with a simple procedure using a graphical step-integration technique for the breakthrough curve. The breakthrough curves in Figure 33 can be used to illustrate the principle. The C_e/C_f values range from 0 (no chloride exiting) to 1 (exit concentration equal to influent concentration). A small interval of C_e/C_f can be chosen, for example for C_e/C_f between 0.1 and 0.2. Corresponding t values can be read from the breakthrough curve (t = 0.15 and C_e/C_f = 0.35 respectively for the upper curve in Fig. 33).

Conditions at $C_e/C_f = 0.15$ are considered representative of the interval. This implies that a volume of column effluent of (0.35 — 0.15).P cm^3 was composed of 85 %

displaced water and 15 % undiluted chloride solution. Here, P = total volume of water-filled pores (t = 1). The volume of displaced water in the interval considered is therefore (85/100).(0.35—0.15).P cm^3. Such values are calculated for other C_e/C_i intervals as well, until the entire breakthrough curve has been characterized. Then, all calculated volumes of displaced water are added up and then subtracted from the total volume of initially present water (P), to yield an estimate of the volume of "immobile" water. Of course, results will be most accurate when very small intervals are used. Breakthrough curves can also be used to determine "equivalent pore-size distributions" of pores contributing to flow (KLINKENBERG, 1957).

5. USING SOIL SURVEY DATA TO PREDICT ASPECTS OF SOIL HYDROLOGICAL BEHAVIOUR

5.1 Introduction

Basic soil physical characteristics were discussed in the previous chapters and methods of determination were reviewed. Since flow of water can only occur through pores, attempts were made to relate morphological observations of pore systems to physical behaviour in terms of moisture retention and hydraulic conductivity. (BOUMA and ANDERSON, 1973). Generally, research in soil physics is focussed on the application of flow equations which produce the most satisfactory results in isotropic non-swelling porous media. Attempts to work with natural, heterogeneous soil have been few in number (although there is a marked increase lately) and have been hampered by what appeared to be a lack of knowledge regarding the constitution of natural, porous systems. On the other hand, morphologists have concentrated on the development of techniques and classification systems for voids, rather than on a physical interpretation of the observed patterns. The remainder of this chapter discusses some contributions by soil survey to the general understanding, characterization and prediction of water movement in unsaturated soils. A brief review of basic principles of soil survey and classification may facilitate assessment to readers who are not familiar with activities in this particular field of soil science.

5.2 A brief review of relationships between soil survey, classification and interpretation

Soil survey practices and systems of soil classification in different countries have been described in manuals (e.g. SOIL SURVEY STAFF 1951, 1975; EHWALD et al., 1966; DE BAKKER and SCHELLING, 1966). This review will not focus on the details of such systems but will attempt to summarize some basic principles and relationships. For a more thorough analysis, the reader is referred to BUOL et al. (1973), SCHELLING (1970) and DE BAKKER (1970) who also cite many references.

Soil genesis is the study of the development of soils on the land surface of the earth in the upper portion of geological materials which have been affected by what are considered the "state-factors" of soil formation, such as climate, organisms, topography, parent material and time. These factors refer to external conditions, which often interact. They induce soil forming processes of a very complex nature that result in a "soil" with characteristic horizons which represents a significant change from the initial geological material (the "parent" material). This change is so profound that soil science could be established as a separate discipline and not just as a sub-science of geology. Discussions as to which "state factors" are expected to have been most active in producing certain types of soil are often inconclusive because only the end result can be seen, and unsatisfactory. Fortunately, the modern tendency is to describe and characterize a given soil by quantitative methods and to base classification and interpretations on such data, rather than on subjective considerations alone. A very important distinction must be made between soil classification (giving a name to a given soil) and the map legend (describing which soils are to be found within an area delineated on a soil map). SCHELLING (1970) points out that this distinction is often rather obscure in soil publications and may thus result in confusion among users since they are not quite clear as to what can and what cannot be expected. Some elaboration of these two aspects is therefore needed.

Soil classification represents an effort by the human mind to create logical order in what would otherwise be an overwhelming quantity of "soils". In fact, every new observation of the vertical sequence of soil horizons at any given location could represent the distinction of a new "type" of soil since no two soils are exactly alike. In a normal survey environment such a procedure would offer serious practical problems when applying soil survey data on a regional basis. The distinction of different "types" of soil that fit into a comprehensive soil classification system, each having a defined limited range of properties, has the purpose of restricting the "infinite" number of soils to a more manageable quantity. For that purpose all classification systems use different levels with more classes as the "level" becomes lower.

This will be further illustrated in section 5.4. Defining criteria to establish the levels and the classes requires observation and manipulation of basic soil data which were gathered during on-site investigations, or which were determined in the laboratory by analyzing samples taken in situ. These basic data form the backbone of each classification system. The application of classification systems implies by necessity that not all gathered data are used in the determination of the "soil type" according to the classification system used. Unused data will be on field sheets in some file and it has been argued that soil classification is an unnecessary screen imposed by the limitations of the human mind, and that it would be preferable to store all basic data and the location of the spot of observation on magnetic cards or tapes to be recalled at will. Be that as it may, soil classification is a reality and needs to be discussed. The need to distinguish different "types" of soil requires criteria and concepts to allow a rational discussion. What, for example, is the minimum size of a "soil"? Some definitions will be cited from SCHELLING (1970). An "individual" has been defined as "the smallest natural body that can be defined as a thing complete in itself". A "natural individual" is an individual that is discrete and independent of the observer. An "artificial individual" on the contrary, is a human construction. The *pedon* is an artificial individual of arbitrary size varying between 1 and 10 m^2 and is a function of the lateral variations in soil characteristics used for the classification (SOIL SURVEY STAFF, 1975). The pedon is thus the smallest "soil", that can provide basic data needed for classification and mapping purposes. The *polypedon* is a unit for which the boundary criteria of soil classification units (e.g. soil series) are used. Different pedons may occur within the outer boundaries of the polypedon. Whether or not a pedon is considered to be "different" depends, of course, on the classification system. In any case, different pedons (the "inclusions") are not part of the population of identical pedons in the polypedon. The latter will therefore be a "body with holes" (SCHELLING, 1970). The entire landscape is composed of polypedons with well defined boundaries which are imposed by the classification system. The latter is thus of great importance. The development of a soil classification system requires the establishment of certain classes, which are abstract fields formed by the concept of that class, and defined with simple or complex characteristics which form the basis for membership of the class. Generally, soil classification systems are developed to be used for making soil maps which are intended to be used for many years and for many different purposes. Therefore, only those characteristics are used for classifications that are permanent or that have a variable but *constant* cyclic character.

The distinctions between permanent and variable is somewhat flexible, but the intention is to avoid a changing soil classification following a forest fire, a single ploughing chemical fertilization or differences in soil management which may be reflected in structural differences. As discussed, characteristics used for classification should preferably be obtained by measurements rather than estimates and definitions should be very specific to allow identical judgments by different soil scientists.

A discussion of the procedures which are followed in developing a soil classification system is beyond the scope of this text. The reader is referred to SCHELLING (1970) for a discussion of the procedures and to various publications that present soil classification systems for different countries.

This all implies that a fair amount of subjectivity is bound to be part of the development of a soil classification system. So far, this review has emphasized soil classification and occurrence in nature of pedons and polypedons, the latter as objective, geographic bodies of soil. The delineation of all polypedons within a landscape may be possible, but would involve an immense investment of time and money since many observation points would be needed. Practical soil survey does not generally allow that degree of detail if only for economic reasons. A soil map consists of delineated areas which represent soil bodies. A "delineated soil body" is a geographic body of soil that corresponds to a delineation on a soil map (SCHELLING, 1970). The soil surveyor in the field, draws those lines not only on the basis of soil observations in a limited number of borings and pits, but also based on his practical experience, which has taught him that there are areas in the field within which the genetic processes "apparently" varied within fairly narrow limits. Soil classification systems could therefore be developed in which polypedons have similar boundaries as soil landscape bodies, which can be recognized and delineated because directly visible field characteristics exist, such as landscape position and vegetational cover, which provide indications as to the boundaries of the soil landscape bodies.

The number of soil borings that can be made by the soil surveyor, to verify his assumptions, should depend only on the intricacy and complexity of the soil pattern in the landscape, but a general rule of thumb requires at least three to six borings per 1 cm^2 of the soil map to be made (BIE, 1970). For a soil map of scale 1 : 50 000, this implies one boring per 4 to 8 ha. For a 1 : 20 000 survey this is reduced to one boring per 2 to 3 ha, densities that are obviously inadequate to allow a detailed evaluation of the occurrence of different types of polypedons. The experience of the soil surveyor is thus heavily taxed. Surprisingly little research has been conducted to verify the accuracy of the soil mapping process. The few published studies are not very encouraging (BIE and BECKETT, 1973; WEBSTER and BUTLER, 1976; AMOS and WHITESIDE, 1975). However, it is possible that certain criteria, used for classification, are not very relevant for the particular aspect of soil behaviour which is of prime interest to a particular user. In other words, what may appear as a "hole" in the polypedon (an "inclusion") according to the soil classification system, may *act* quite comparably to the polypedon proper. On the other hand, soils that must be considered part of the defined polypedon according to the soil classification system, may *act* quite differently from other, identically named polypedons. For example, virgin and cultivated soils may have identical pedological classifications but may act differently in terms of their capacity for transmitting water (BOUMA and HOLE, 1970; BOUMA et al., 1975b).

The aspect of soil behaviour, introduced here for the first time in this section is of prime importance for practical users of soil survey, and its prediction forms the main justification for the soil survey programme. Users of soil surveys are, in fact, often only interested in knowing where different soils occur and how their suitability or limitations for a particular use (be it agricultural or otherwise) can be predicted as clearly as possible. In addition, there is an increasing demand to be informed about methods that can be used to *overcome* restricted suitability or limitations, thereby fully realizing whatever potential a soil has for a given use (McCORMACK, 1970; BOUMA, 1973, 1974). Soil maps are used to delineate areas for which a certain interpretation applies, which is defined in terms of expected soil behaviour. Such areas may include individual soil mapping units or combinations of several mapping units which are expected to have a comparable behavioural pattern. Soil survey interpretation procedures add a new dimension to the concepts used in defining procedures of soil classification and mapping. The latter two are essentially based on an analysis of the long range effects of processes of soil genesis that produce more or less permanent soil characteristics. Those characteristics, in turn, are indicators for the processes that caused their formation, but only, of course, if they are still active. If so, such processes may govern current soil behaviour. For example, periodic saturation and reduction of soil may result in gleying (section 5.3). Thus, periodic saturation and reduction of the soil is assumed to occur when gleying is observed in a soil. Such a soil may therefore not be suitable for a variety of uses and this is expressed by the soil survey interpretation. Cultivation of soil by man introduces a major and often unpredictable, short-range variability which may result in a wide range of behavioural patterns. Soils at different locations, which are of the same type according to a pedogenic soil classification system, may behave quite differently in terms of, for example, permeability or structural stability, following different soil management in its recent history (BOUMA, 1969). There are often no clues whatever to be derived from soil morphological or genetic features that would explain why those differences did occur. Soil morphology, and in particular soil structure descriptions which provide data on pore-size distributions, can sometimes *explain* why physical behaviour is different, but only "afterwards". Moreover, direct physical measurement is often more attractive to substantiate the occurrence of such differences. This single observation of soil conditions is of much less significance than one that relates to permanent soil characteristics such as soil texture and the occurrence of genetic soil horizons, which are used in soil classification. What is observed now in a tilled soil, for example in terms of structure, may be quite different in another season or year with a different management. Common questions asked by soil users as to the most advisable soil management to be used to overcome limitations can, therefore, hardly be derived from general soil survey data. Special studies by experts in soil tillage and management are needed to achieve that purpose. However, the systematic gathering of the general experience of farmers, and others, in a certain area, in addition to information gained directly from soil profile characteristics, may allow a reasonable assessment of practical soil behaviour for a wide variety of uses, with the constraint that conclusions relate to current technology and that they are therefore bound to become obsolete before too long. *In any case, soil survey interpretations are based on the hypothesis that soil behavioural patterns do correspond with patterns that are shown on a soil map.*

More specific testing of this hypothesis is very much needed, but its practical usefulness has been demonstrated (for example: HAANS and WESTERVELD, 1970). To allow a more specific focus on some of the problems raised, soil suitability for agricultural production has been defined as "the degree of success with which a crop or range of crops can be regularly grown on a certain soil, within the existing type of farming, under good management and under good conditions of parcellation and accessibility" (VINK and VAN ZUILEN, 1974). Soil limitations for non-agricultural soil survey interpretations have also been defined in terms of the degree of success with which conventional technology can be applied to achieve satisfactory soil performance. Slight and moderate limitations describe conditions where conventional technology (not specified in the interpretation scheme) can be successfully used, whereas severe and very severe limitations do not allow use of conventional technology and require (again, unspecified) new technology (SOIL SURVEY STAFF, 1975; BOUMA, 1974). A case study dealing with soil disposal of septic tank effluent will be presented in section 5.4 to demonstrate the use of soil survey information for predicting not only soil limitations for a *specified* conventional technology but also, more importantly, for developing, testing and applying specific new technology.

5.3 Use of soil morphological data

5.3.1 Introduction

Soil morphological data are used to predict certain aspects of soil hydrological behaviour. Two approaches can be distinguished. The first approach is a direct one and relates for example to the occurrence of coatings on natural soil aggregates which may be indicative of flow channels in the soil (JONGERIUS, 1970) or to the occurrence of mottling which may be indicative of the soil moisture regime (SOIL SURVEY STAFF, 1951, 1975). These soil properties refer to features that can be directly observed and interpreted by a trained pedologist who, in fact, believes that he knows the art of "reading" a soil profile. This interpretation is very attractive because it is much cheaper than performing elaborate soil physical measurements. However, one must always hope that the pedologist does not overestimate his reading skills and that he frequently tests and refines his visual skills by making physical measurements or observations. The second approach uses soil morphological data, which has no direct physical interpretive meaning by itself, to either *calculate* physical characteristics or *correlate* them with measured physical characteristics. Several examples of this approach are discussed in this publication: (i) calculation of K in pedal soils (section 4.3.2), (ii) the correlation of apparent dispersion coefficients with soil structure descriptions (section 4.6) and (iii) the correlation of the K-h curve with soil morphometric data (section 5.4). The remainder of section 5.3 will deal with a discussion of the physical interpretation of soil mottling as an example of the *first* approach. This example is presented as a specific case study, with results that do not necessarily apply to soils elsewhere. An attempt was made to avoid using complex soil morphological terminology in the discussion. For more details the reader is referred to VEPRASKAS et al. (1974); VENEMAN et al. (1976) and VEPRASKAS and BOUMA (1976).

5.3.2 The physical interpretation of soil mottling phenomena: a case study

Soil mottling is described as: "marked with spots of different colours of grey and red" (SOIL SURVEY STAFF, 1975). Saturated soil *may* induce low redox potentials that can reduce ("dissolve") iron (and manganese) compounds but only in the presence of an energy source such as organic matter (VAN SCHUYLENBORGH, 1973). Reduced iron (and manganese) are soluble and can move with the water through the soil until they oxidize again when redox potentials reach a critical higher value (VAN SCHUYLENBORGH, 1973). A brown soil often derives its colour from very fine, well distributed iron particles. Reduction of iron, which is relatively easy because of the small size of the particles, and removal by flowing water may permanently change that colour to grey. The part of the soil from which the iron is removed is then "gleyed", and may have chromas of two or less according to the Munsell soil colour notation (SOIL SURVEY STAFF, 1975). Intermittent saturation and drainage of the soil, resulting in seasonal reduction and oxidation, may have the effect of dissolving, transporting and precipitating iron (and manganese) in patterns that can be indicative for the moisture regime of the soil. Such precipitates of iron (and maganese) are often relatively large and easily visible and do not dissolve as well as much finer particles when reducing conditions occur again.

Periodic saturation of soil is the common process which causes reduction. A logical consequence of this observation is to consider soil mottling to be indicative of periodic saturation. SOIL SURVEY STAFF (1975) uses the occurrence of chroma's of two or less to be indicative of saturation during some period of the year. Soil morphology is thus used to predict one aspect of soil behaviour. This reasoning represents the reverse of the one used in soil classification where state factors determine, through processes of soil genesis, the morphology of the soil which, in turn, determines its classification (section 5.2). In interpretation, as discussed, soil morphology is used to hypothesize about the processes of soil genesis (reduction and oxidation) and the state factor that caused them to occur (groundwater fluctuation, or in a broader sense, the climate). Some potential problems are readily evident: (i) Saturation of soil does not always coincide with the occurrence of reducing conditions, due to the possible lack of an energy source (VAN SCHUYLENBORGH, 1973). So mottling may indicate saturation, but lack of mottling does not necessarily imply lack of it. (ii) What is visible now in the soil may not result from current processes but may be associated with climates of the past. Many geological deposits have been subject to pedological processes for thousands of years. A more recent and typical example relates to drainage of very wet soil. Soil mottles in such soils are indicative of a soil moisture regime which was drastically changed by man. (iii) Even when mottling of soil can be correctly associated with periodic saturation, it is still quite important to know *how long* a soil is saturated. Criteria currently used do not specifically define the duration of this period, although mottling itself is characterized in terms of having "chromas of two or less" (SOIL SURVEY STAFF, 1975).

Practical use of soil mottling criteria and results of field and laboratory research (which was inspired by observed inadequacies in the underlying concepts) will now be presented as a case study. This study was initiated because of legal problems following a routine application of mottling criteria. Soil suitability for the disposal of septic tank effluent in sub-surface seepage beds is judged by several factors such as soil permeability, the level of bedrock, the slope and the groundwater level (BOUMA, 1974). The latter factor is of interest here. If the highest level of the groundwater occurs within 90 cm (three feet) below the bottom of the sub-surface disposal system (which is usually 1.50 m (5 feet) below the soil surface) then the soil is considered unsuitable for on-site disposal of septic tank effluent. This implies that, in many areas of the United States, new houses cannot be built according to the health code or local laws. Such decisions are far-reaching and are bound to offend the spirit of free enterprise. Groundwater levels are generally not observed very frequently and indirect criteria such as mottles are indispensable for estimating the highest groundwater level during the year.

Our case study (VEPRASKAS et al., 1974) involved a parcel of land with a Grays silt loam soil (Typic Hapludalf, fine silty, mixed, mesic). This soil formed in about 90 cm of loess (silty clay loam) covering glacial outwash sand with a water-table at 20 cm depth below the soil surface. The lower 30 cm of the loess cover was mottled with chromas of 2 in the soil matrix. This implies the occurrence of "saturation during some part of the year" (SOIL SURVEY STAFF, 1975). This analysis was made by a soil scientist during an on-site investigation, and since the depth of mottling was well within the critical depth as defined by the health code, the site was turned down to be used for construction of a home. The owner of the plot did not accept this conclusion and placed a groundwater tube to a depth of 80 cm, well within the mottled layer. He never observed free water in the pipe, even during the wet season. Our investigation involved the placing of tensiometers in the soil and running the instantaneous profile method (section 4.4). Detailed results, reported by VEPRASKAS et al. (1974), showed that the lower part of the loess layer became very wet but was *never* really saturated. The wetness resulted in reduction processes within peds, which mobilized the manganese and some of the iron. Movement towards larger air-filled voids, such as channels and planar voids, resulted in the oxidation of these compounds along these voids. Technically speaking, there was no water-table in the silt cover but only in the underlying sand at a considerable depth. This would allow for adequate purification during downward percolation and the site was therefore approved for sub-surface waste disposal. This decision shifted the burden of proof from the house-owner to the soil survey personnel. What used to be a straightforward criterion was now subject to doubt. Clear indications were obviously needed to reliably estimate the occurrence or non-occurrence of a water-table, and preferably, its duration. The studies were therefore expanded to other comparable sites where a 10 month monitoring programme was started and to the laboratory where model experiments were made. The monitoring programme involved in situ measurements of soil moisture pressure and a detailed description of soil morphology in selected soil profiles which were located in a topo-sequence with soil moisture conditions ranging from relatively dry to very wet. The detailed results are presented elsewhere (VENEMAN et al., 1976) but the basic conclusions can be shown schematically (Fig. 34) and can be summarized as follows:

(i) Occurrence of high chroma colours inside peds (chromas higher than two according to the Munsell Soil Color Charts) and some black manganese coatings on peds were associated with wet conditions and very short periods of saturation of one day or less, *not* predicted by current criteria which associate saturation of unspecified length with chromas equal to or smaller than 2 (type 2 in Fig. 34).

(ii) Occurrence of chromas of two inside the peds, few black manganese but common red iron coatings on peds, were associated with short periods of saturation not exceeding a few days. However, high pressure potentials (low tensions) near saturation occurred for several months (type 3 in Fig. 34).

(iii) Occurrence of chromas of one inside peds (giving the soil a real grey, "gleyed" appearance), very few black manganese coatings or concretions but clear bleached ("gleyed") zones around larger tubular or planar voids were associated with periods of continuous saturation for several months. The gleyed zones around the pores (called "neoalbans" by the authors) are often separated from the surrounding soil by a concentric concentration of iron or manganese (a "quasicutan" in BREWER's nomenclature (1964) (type 4 in Fig. 34).

The observed moisture regimes, measured in what fortunately turned out to be an average year in terms of the weather, are schematically presented in Figure 34. The reader is referred to the original publication for the complete graphs (VENEMAN et al., 1976). The different behaviour of manganese and iron upon reduction and oxidation can be explained by considering that manganese is reduced more easily than iron, whereas iron is more readily oxidized. The critical redox potentials involved are a complex function of the type of mineral and of the pH (VAN SCHUYLENBORGH, 1973).

A simplified diagram was constructed (Fig. 34) to demonstrate that certain soils never

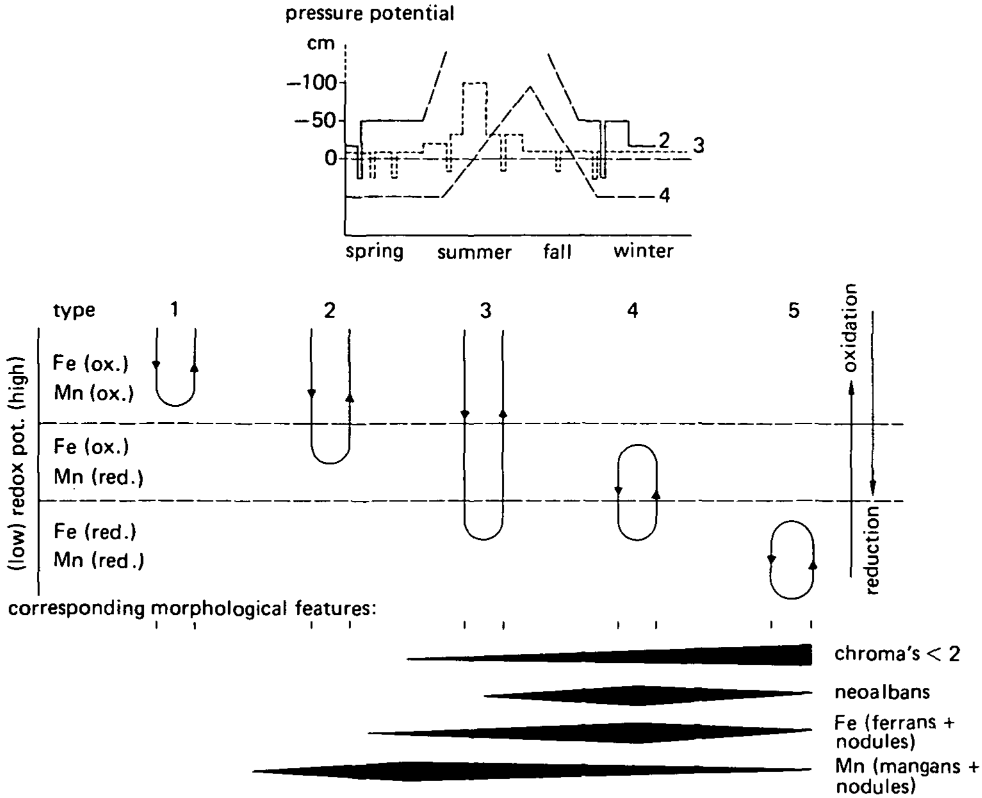


Fig. 34. Schematic diagram showing relationship between measured soil moisture regimes, associated Redox potentials and morphological features (see text).

get wet enough to even reduce the manganese (type 1) whereas others mobilize only the manganese (type 2) or also iron (type 3). In all three cases oxidation occurs in the dry season. Type 4 reduces both iron and manganese but only iron is oxidized. The manganese remains soluble and may be removed from the soil by leaching with the groundwater. Type 5 finally, is an extreme version of type 4, in which permanently reduced iron and manganese may be leached from the profile leaving a completely gleyed soil. These very schematic drawings may be helpful in explaining the morphology observed, although the reader should always be conscious of the simplifications involved. One aspect remains to be discussed. A consideration of the level of redox potentials (which are assumed to be among other factors, a function of the duration of saturation) can explain whether or not manganese and/or iron concentrations, and gleying occur. The flow regime in the soil is also very important, because this determines *where* oxidized compounds will accumulate. For example, reduction of iron may not result in gleying if the permeability of the soil is so low that it does not allow removal of the iron. Also, drainage of soil may occur irregularly when water leaves the larger pores first allowing precipitation of iron and manganese along the air-filled voids. This pattern is common in soils that are seasonally wet but not significantly saturated (types 2 and 3, Fig. 34). The processes are very different in soils that *are* at some time saturated. Then, larger pores are filled with water by precipitation or a rising water-table and water will move from those pores *into* the surrounding unsaturated soil, possibly reducing the soil adjacent to the pore. Precipitation of the reduced compounds may occur as the liquid penetrates the air-filled ped. Such precipitates often form a concentric red or black circle at some distance from the void. Complete saturation of the soil follows for several months and the interiors of the peds may be entirely reduced (Fig. 34, type 4 or 5). In situ measurements of redox potentials confirmed this hypothesis (VEPRASKAS and BOUMA, 1976). The precipitates of iron or manganese cannot be reduced easily because of their relatively large size.

To summarize, this research allowed refinement of soil mottle interpretation for certain Wisconsin soils. Even so, some major hypotheses were needed to present a somewhat coherent theoretical framework. Model experiments were therefore designed to specifically test some of the assumptions made. This is unusual for soil morphological studies, which is unfortunate because, without those, the pedologist might overestimate his capacity to interpret soil morphological features. The model experiments (VEPRASKAS and BOUMA, 1976) tested the hypotheses underlying the genesis of mottling types 2 and 4 (Fig. 34). First, model soil cores were made with a silty clay loam texture which had a brown colour. The colour condition at time zero was therefore known; a fact which is unusual for field studies. One series of cores was penetrated with small vertical holes and the core was subjected to a constant pressure of —7 cm water in a porous cup while evaporation occurred at the top. The water used was a nutrient solution to allow reduction processes. The moisture regime thus established resembled that of mottling type 2. After five months, clear manganese coatings had formed at the upper surface of the core and redox measurements, made in situ, showed that reduction of Mn could have occurred. In other words, the hypothetical link between: (i) an unsaturated but wet moisture regime, (ii) reduction of manganese and (iii) formation of manganese cutans along larger pores, was experimentally supported. Another series of identical cores with holes was subjected to slow saturation from the bottom up, for three weeks followed by drainage for three weeks. This cycle was repeated five times. Afterwards, bleached, gleyed zones were observed along the vertical artificial tubes through the cores and Mn and Fe had precipitated in circular patterns around it. Measured redox potentials indicated possible reduction of both iron and manganese. This morphology corresponded with type 4 and even though the time cycle in this experiment was quite different from the field cycle, it supported the hypotheses made to correlate field morphology with the duration of saturation.

This case study on soil mottling is relevant not only because of the importance of this phenomenon in soil survey, but also as a demonstration of an approach in soil morphological research which could be applied equally well to other morphological aspects of soil. The approach, as discussed, involves: (i) an evaluation of the indicative value of current morphological criteria by applying modern analytical techniques (in our case study: in situ measurement of soil moisture and redox potentials), (ii) a modification of morphological criteria following such experiments, by making them more specific or by broadening their meaning (in our case study: introduction of the detailed Brewer terminology and distinction of four rather than two moisture regimes) and (iii) an independent test of the newly developed criteria by model experiments in which relevant factors are well defined and controlled by experimental conditions (in our case study: model experiments on mottle formation). Another example is to be found in section 4.6.4.

Studies of seasonal fluctuations of the water-table made in the Netherlands have followed an essentially similar but broader approach due to the more regional character of the studies (VAN HEESEN, 1970; VAN DER MOLEN, 1970; VAN WALLENBURG, 1973). Morphological criteria could *not* be defined uniformly for the entire country, so regional relationships, existing between the depth of the water-table and profile and field characteristics, were used successfully to locally extrapolate measured values in observation tubes.

5.4 Use of soil classification and soil survey

5.4.1 Introduction

Computer simulation models have been developed in recent years to predict soil water movement. The models may describe hydrological conditions in a single soil (VAN KEULEN and DE WIT, 1970; VAN DER PLOEG, 1974; WIND and VAN DOORNE, 1975) or those in a region (DE LAAT et al., 1975). The models need input data in terms of precipitation and evapotranspiration if they describe natural soil moisture regimes. But some may also be used to predict the hydraulic effects of the application by man of, for example, liquid waste or irrigation water. Then, application rates are needed as input data. In addition to these data, which define the boundary conditions of the flow system, characteristic soil hydraulic constants have to be available. These are the liquid retentivity curve (discussed in section 3.3) and the hydraulic conductivity (discussed in Chapter 4). The availability of ever faster and cheaper computers and the degree of success with which simulation models have been used for a wide variety of applications, make it probable that simulation techniques will become increasingly important in years to come. Application to certain regions in particular will become more relevant as questions on land use issues receive more emphasis. The latter type of use of simulation models requires an estimate or measurement of hydraulic constants, as discussed, for soils in the region. At present, soil maps are often the only readily available source of information as to where different types of soil (as defined by the map legend) occur. The distinction of those different types of soil according to pedogenic criteria of a soil classification system, does not, of course, necessarily imply that all areas on a soil map which are named for a particular type of soil have identical soil hydraulic constants as well. Soil mapping units may have inclusions of other soil types which may have different physical characteristics (section 5.2). Soils which are considered identical from a pedological point of view, may be different when judged by soil physical criteria and identical physical behaviour may occur in soils which are different from a pedological point of view. Very little quantitative data are available at present on the physical relevance of pedological criteria used by soil surveyors for predicting soil moisture regimes. This item will be further discussed in section 5.4.2 by reporting results of field measurements of hydraulic conductivity, which made use of soil maps to locate experimental sites.

Use of simulation techniques may have important consequences for soil survey interpretation. Current schemes, whether they relate to agricultural or non-agricultural use of the soil often emphasize soil limitations (or soil suitability) when conventional technology is used (section 5.2). Use of new technology can often successfully overcome limitations or increase soil suitability for a given use. Practical users of soil survey are generally more interested in knowing whether certain limitations can be overcome, and if so, how. This may be more urgent than knowing what those limitations are supposed to be. Simulation techniques or application of simple physical calculations are of essential importance for predicting the hydraulic effects which may be associated with applications of such new technology. A relevant case study, relating to on-site soil disposal of liquid waste, will be discussed in section 5.4.3.

5.4.2 *Estimating hydraulic conductivity from soil survey data*

Computer simulation models, which attempt to describe water movement in natural soils, can only produce significant results if basic soil physical data, as discussed, are available for the soils that occur (DE LAAT et al., 1975). Gathering such data requires the application of elaborate and costly techniques by trained personnel and technical and financial facilities are often inadequate to be able to allow this. But even if many measurements could be made, their number would have to be limited and it would still remain necessary to extrapolate data from measured to unmeasured soils that were somehow characterized as being comparable in terms of relevant soil physical characteristics. The extrapolation does, of course, not necessarily require soil survey data and may follow interpolation techniques (WEBSTER and BURROUGH, 1972). The remainder of this chapter will deal exclusively with the important soil hydraulic conductivity (K). Attention will be confined to the use of soil survey data for extrapolating purposes. Initially, there is no substitute for the measurement of K (Chapter 4). The purpose of the procedures to be discussed here, would be to allow prediction of K for unmeasured soil when a large amount of experimental data is available. The relevant question is therefore as follows: which soil survey data, if any, can be used by the field soil scientist to predict K, and how should this be done? An important distinction must be made between data to be gathered by the soil surveyor on the site itself and data to be derived from a soil map without on-site work. One very popular scheme used in the Netherlands, which was based on a literature review, defines K for twenty soils as a function of soil texture and soil moisture retention (RIJTEMA, 1971). Soil texture can be estimated reasonably well by many soil scientists on the site to be tested. The classification of the soil, used on large scale maps generally allows some estimate of soil texture. This implies that sometimes on-site investigations may not always be necessary. However, some soils with identical textures have quite different K curves due to differences in soil structure and associated drainage class. Then, a prediction based on texture alone is inadequate and can be improved by including structure and drainage class as diagnostic criteria. The latter two characteristics are exclusively provided by soil survey data, as expressed by soil classification. Two case studies will be presented here as examples for the practical feasibility of this approach.

Case study 1

Hydraulic conductivities of unmottled well to moderately well drained soils developed in silt loam deposits, covering sandy outwash in the southern part of Wisconsin were compared with those of poorly drained, highly mottled soils developed in silt loam deposits, covering glacial till in the central part of this state. Measurements were made in situ in B horizons of the silt loam deposits with the crust test (section 4.4.2). Examples of soil series of the first group are the Batavia and Plano silt loams and of the second group the Almena and Withee silt loam (for descriptions see BOUMA, 1973) (see Fig. 36.) Hydraulic conductivity curves were significantly different as shown in Figure 35. Use of texture alone as a diagnostic tool for predicting K would not have allowed this important distinction. Soil classification refers the reader on the soil

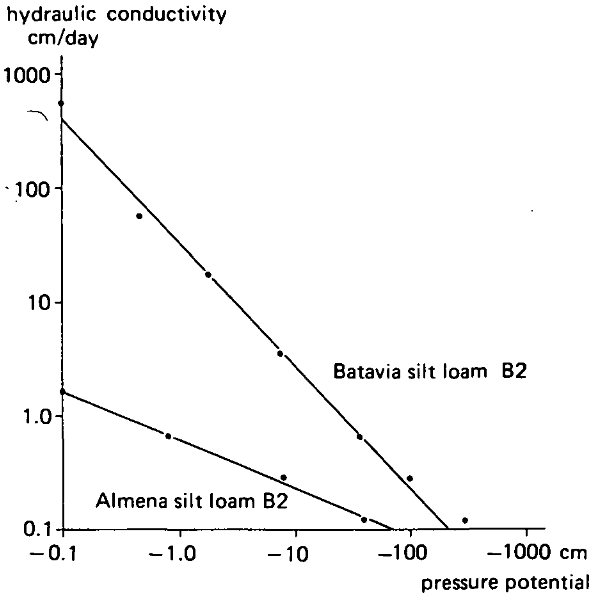


Fig. 35. Hydraulic conductivity curves for B2 horizons of the Almena and Batavia silt loams which have identical textures but different structures.

series level to a "standard" description of the central concept of the classification unit and this contains information on structure, drainage class and other relevant features. Soil classification can thus be a useful tool in predicting K data but not necessarily always as will be illustrated in case study two.

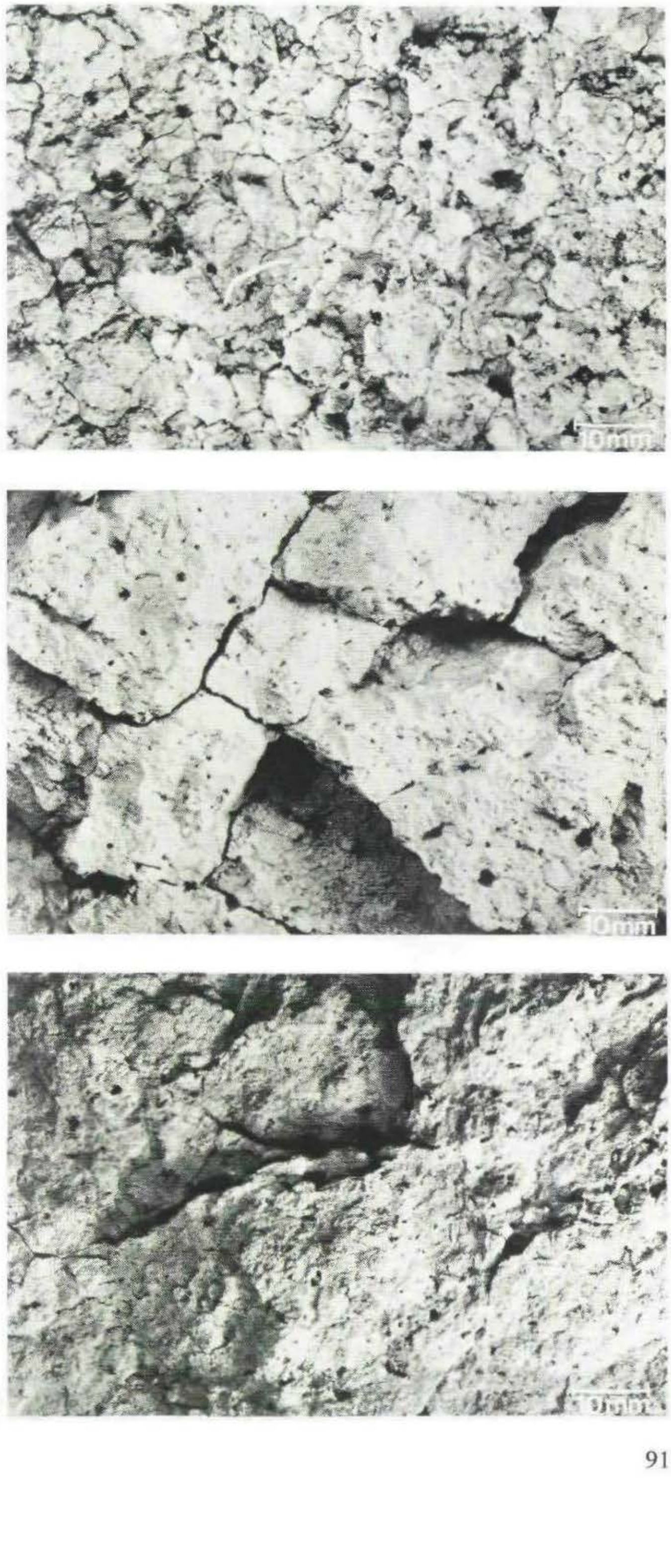
Case study 2

Measured hydraulic conductivities of subangular blocky and prismatic soil horizons in the Batavia and Plano silt loam (Fig. 36) were compared to evaluate the diagnostic value, in terms of estimating K, of the pedogenic distinction between these two soil series. This distinction is essentially based on the presence of a dark topsoil in the Plano (prairie vegetation) and a lighter coloured topsoil in the Batavia (forest vegetation). The subangular blocky structures were found in the B2t horizons of seven Plano silt loams and five Batavia silt loams and the prismatic structures were found in the B3 horizons of these soils. All soil horizons had a silty clay loam texture.

Measurements were made in situ by the crust test and provided K-data down to pressure potentials of about —150 cm. Results, reported in detail by BAKER and BOUMA (1976) are summarized in Figure 37 which shows that curves for the subangular blocky structures were not significantly different from those for the prismatic structures in both soil series. In other words, the distinction of the two series and the difference in soil structure did not constitute significant diagnostic features for predicting K. These distinctions may, of course, be significant as a diagnostic feature for other soil properties.

Case study 2 illustrates that hydraulic conductivity data for certain types of soil should be represented not as a single curve obtained after a single measurement, but as a band which expresses the effect of natural variability by using a series of measure-

Fig. 36. Horizontal soil peels through the B2 horizon of the Batavia silt loam (A), the B3 horizon of the Batavia silt loam (B) and the B2 horizon of the Almena silt loam (C).



A

B

C

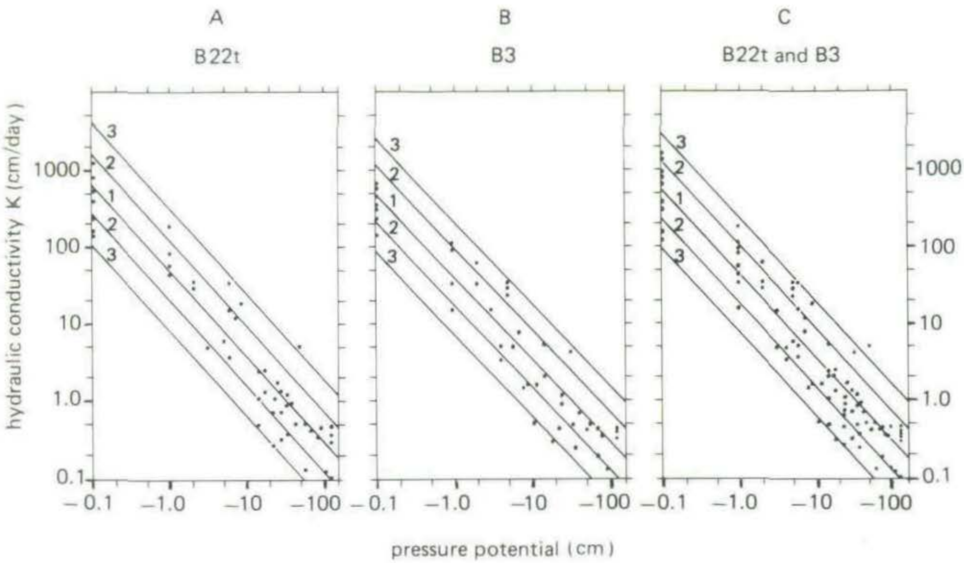


Fig. 37. Hydraulic conductivity bands for four soil horizons in two different soil series (Batavia and Plano silt loam). Regression curves (1) and 68 % confidence (2) and 95 % confidence (3) intervals are included (statistical analysis based on untransformed values). (After BAKER and BOUMA, 1976).

ments at different locations. Such "bands" (Fig. 37) allow a realistic statistical treatment of K data in terms of predicting the probability that K will be *below* or *above* certain values at any given pressure potential. There is some analogy here with data on precipitation which not only defines the amount of rain to be expected on a given day of the year, but also the more interesting probability that precipitation will be above or below a certain value for any given day (RIJTEMA, 1971).

5.4.3 Use of soil for on-site liquid waste disposal: a case study on soil survey interpretation

5.4.3.1 Introduction to the problem of waste disposal

Soil survey interpretations in the United States include an evaluation of soil limitations for on site disposal of septic tank effluent (USDA-SOIL CONSERVATION SERVICE). This evaluation, is based on the use of conventional technology with septic tanks and subsurface drainage systems, which is strictly regulated in health codes (BOUMA, 1974). Technical aspects of waste disposal in soil are not relevant for this publication, but the function of soil survey information in applying both conventional and newly developed modern techniques can be reviewed as a case study to demonstrate how valuable this information can be and what type of research may be needed to improve the procedures followed. Soil survey interpretations for on-site disposal of septic tank effluent follow a general procedure of defining different degrees of limitations for soils relating to absorption and purification of effluent in terms of "slight", "moderate", "severe" and "very severe". This range describes a sequence of increasing difficulties to be encountered when existing septic tank technology is applied in the construction of on-site disposal systems. Construction in soils having slight or moderate limitations is considered feasible, but soils with severe or very severe limitations cannot generally be used. More specifically, the "severe" limitation implies that major soil reclamation, special design or intensive maintenance would be required to construct a satisfactory system. The soil survey approach follows some attractive, realistic assumptions: (i) satisfactory on-site disposal and treatment of septic tank effluent is possible by means of soil absorption in soils with slight limitations. However, problems may occur even in these soils due to construction practices or other causes not directly related to the soil. Limitations are therefore *always* present although they can generally be overcome with present technology; (ii) satisfactory on-site disposal is not necessarily impossible on soils with severe or very severe limitations, but even a professional application of available technology will generally not result in a satisfactory system. The absolute impossibility of on-site disposal is thus never implied, but a practical limit is suggested at least for the present time.

This system of interpretation is based on practical experience but also on regulations in Health Codes which define suitable soil and site conditions very specifically as follows: (i) the hydraulic capacity of soil to transmit liquid is generally expressed by the "percolation rate" which should be faster than 60 cm/day (the expression in the Code is min/inch); (ii) at least 90 cm of unsaturated soil should be present between the bottom of the seepage system and a high groundwater table or bedrock. Bedrock can be observed at all times, but the groundwater level may only be high in wet periods of the year. A seepage system has to function at all times and the highest groundwater level is therefore most relevant. Many Health Codes allow observation of the highest level of soil mottling to be used as an indicator for the seasonally highest water level (section 5.3.2); (iii) limiting site characteristics such as excessive slopes and location in floodplains are also considered.

Following these criteria, unsuitable soils can be defined as having severe or very severe limitations in soil survey interpretations. Interpretive maps can be prepared based on the soil maps, showing areas occupied by these soils and maps of this type can be very useful for land-use planning purposes, based on current health regulations. However, many practical questions are currently being raised regarding the availability of alternative waste disposal procedures for soils with severe or very severe limitations. Recent research, based on new test procedures and a more strict analysis of liquid-waste disposal problems, has indicated that such alternatives may be available. This development may present the opportunity to expand the current interpretation concept by not only describing soil limitations for current technology for certain soil types but by also defining specific alternative "construction and management packages". When applied these would overcome these limitations, so as to make satisfactory on-site disposal a practical reality in specific areas that can be delineated on soil maps. This approach is not necessarily limited to soils that are now considered to have severe or very severe limitations. The very definition of "slight" limitations implies that problems can arise if mistakes are made in designing, constructing or applying effluent to a system in a soil with a "slight" limitation. A (small) "package" can be defined for these soils to assure construction of a good system. Generally, soils with moderate limitations could be expected to need a more elaborate "package". Basically, however, modifications or refinements of current practices should be adequate for soils with slight or moderate limitations, whereas drastically new approaches would be needed for soils now classified as having severe or very severe limitations. This approach to interpretation offers a major advantage in that derived maps could be more useful by not only showing limitiations for use of current technology but, more important, by also showing potential implications of applying innovative technology. Rather than just describing the limitations, emphasis would be shifted to defining soil potential and to defining means for realizing it. The difference in approach by emphasizing the potential of a soil for a given use rather than the limitations (using current technology) is important for a very wide range of soil uses. A conceptual percentage of "successful operation" of a particular waste disposal system, varies in different types of soil. This experience is translated into classes of limitations as discussed. Development of new technology can often overcome limitations ("technically, anything can be done anywhere") but risk of failing increases as site conditions become more difficult. This technical approach does not consider economic aspects, which are often quite relevant for small scale on-site liquid waste disposal. Practical aspects will now be discussed in terms of the role of soil survey in developing innovative technology "packages". The discussion in the following section relates only to liquid waste disposal. Conclusions to be reached do, of course, not necessarily always apply to other soil survey interpretations.

5.4.3.2 The role of soil survey in extrapolating results of interdisciplinary research

The development of a satisfactory on-site disposal and treatment system for septic tank effluent requires the integration of the know-how of a large group of specialists, because biological, chemical and physical processes are involved. Pedologists should play a role in this interdisciplinary team by: (i) selecting representative experimental sites with the use of soil maps, (ii) interpreting soil morphology in terms of expected long range physical behaviour which may differ from results from, by necessity, short term monitoring programmes. An example is the interpretation of soil mottling (section 5.3.2), (iii) emphasizing that model experiments in soil columns or simulation programmes should always be based on conditions found in nature. There is, unfortunately, a need for this watchdog function. For example, column experiments using small sieved aggregates which are supposed to simulate a well structured soil with large natural peds may not be representative even though bulk densities are identical (CASSEL et al., 1974). Use of undisturbed, large soil columns is essential for obtaining reliable results (ZIEBELL et al., 1975). Sophisticated simulation programmes do not yield meaningful practical results if basic assumptions are not realistic (as exemplified by Magdoff, in BOUMA et al., 1974b).

These general requirements can be illustrated by reporting experience obtained in the development of a new on-site liquid waste disposal system in the interdisciplinary Small Scale Waste Management Project in Wisconsin, USA (BOUMA et al., 1975a). Slowly permeable soils with seasonally high groundwater-tables were considered unsuitable for the on-site disposal of septic tank effluent according to the State Health Code. With no alternatives available, construction of new homes in unsewered areas was therefore halted and this caused many problems. The new mound system (BOUMA et al., 1975a) was developed as an alternative, and this system is now accepted by the health authorities. First, some major soil series were selected as being representative of these particular problem soils. Use of soil maps and practical experience of field soil survey personnel are essential in this selection process. Not only the total acreage occupied in the State, but also the area of occurrence was important for this selection, since pressure to build new homes was most evident in the southern most populated part of the State. Once soil series were selected, experimental sites were chosen using soil maps, and on-site investigations were made to check the soil map. A series of column experiments was designed and K curves were measured in situ to determine soil potential for infiltration and purification of waste (DANIEL and BOUMA, 1974, MAGDOFF et al., 1974). They indicated that the hydraulic conductivity (K) was adequate to allow infiltration of the expected daily volume of waste within not too large an area. But a drainage system would be needed to remove the liquid after infiltration from sub-surface trenches because soil mottling in these soils indicated water levels up to 30 cm below the soil surface (BOUMA et al., 1975a; BOUMA, 1975). However, associated bacteriological experiments, made in large undisturbed soil columns, indicated fast movement of fecal bacteria in this well structured soil along natural larger voids (ZIEBELL et al., 1975). So rather than remove excess water from the soil, it was concluded that the system should move away from the water by using a covered mound of sand on top of the original soil surface in which a seepage bed was built (BOUMA et al., 1975a). A number of systems were built and monitored and official permission for their use was obtained. All too often scientific endeavours end at this point, whereas procedures that should lead to the practical application of the new technology, start here. So even though the three functions of pedologists mentioned earlier are important in the design and testing stage, their role in "extrapolating" the new designs to "identical" sites elsewhere may be even more important. Three procedures of such extrapolation can be used (BOUMA, 1974) as follows (see Fig. 38).

Application of *procedure one* requires detailed on site investigations, to determine key soil properties and variability in an area of perhaps 100 m^2 in which a seepage system is to be built. Key properties for on-site liquid waste disposal are: hydraulic conductivity and site characteristics such as slope and landscape location and the occurrence of groundwater and bedrock. The measured properties can be compared with those for the sites where experimental systems operate successfully. If agreement is considered acceptable, extrapolation will occur. This procedure is, of course, time consuming and is confined to isolated, single sites.

Application of *extrapolation procedure two* (Fig. 38) assumes that taxonomic classifications, based on pedological criteria, adequately reflect key soil properties for liquid waste disposal. This is realistic, within limits, for groundwater or bedrock levels, since soil series descriptions list a range of observed diagnostic features. However, hydraulic conductivities of pedons within a series, or phases of series, may vary. Moreover, textural differences may occur in the same soil series at depths exceeding 1.50 m which may not be reflected in separate phases, and this may strongly affect liquid movement from a sub-surface seepage trench. For example, soils in glaciated areas may have formed in a relatively homogeneous loess cover of 1.50 m over heterogeneous sandy and clayey glacial deposits. Diagnostic surface- and sub-surface horizons may then be identical in identical taxonomic units at different locations, even though rates of movement through the subsoil may be quite different, so much so that several experimental designs for seepage systems may be required. A routine application of the second extrapolation procedure would make available increasingly detailed data on magnitude and variability of key properties in soil series, or phases of series, assuming that on-site measurements needed for extrapolation procedure one are made as well. This could mean that on-site measurements would not be needed any more for well-characterized taxonomic soil series with proved low variability at some future time. The required procedure would then be to: (i) classify a soil at any prospective new site following taxonomic schemes of the Cooperative Soil Survey; (ii) check the variability of key properties measured elsewhere in this series, and (iii) determine whether this variability is sufficiently low to allow direct interpretation of site suitability or whether variability has been so high that on-site determination of key

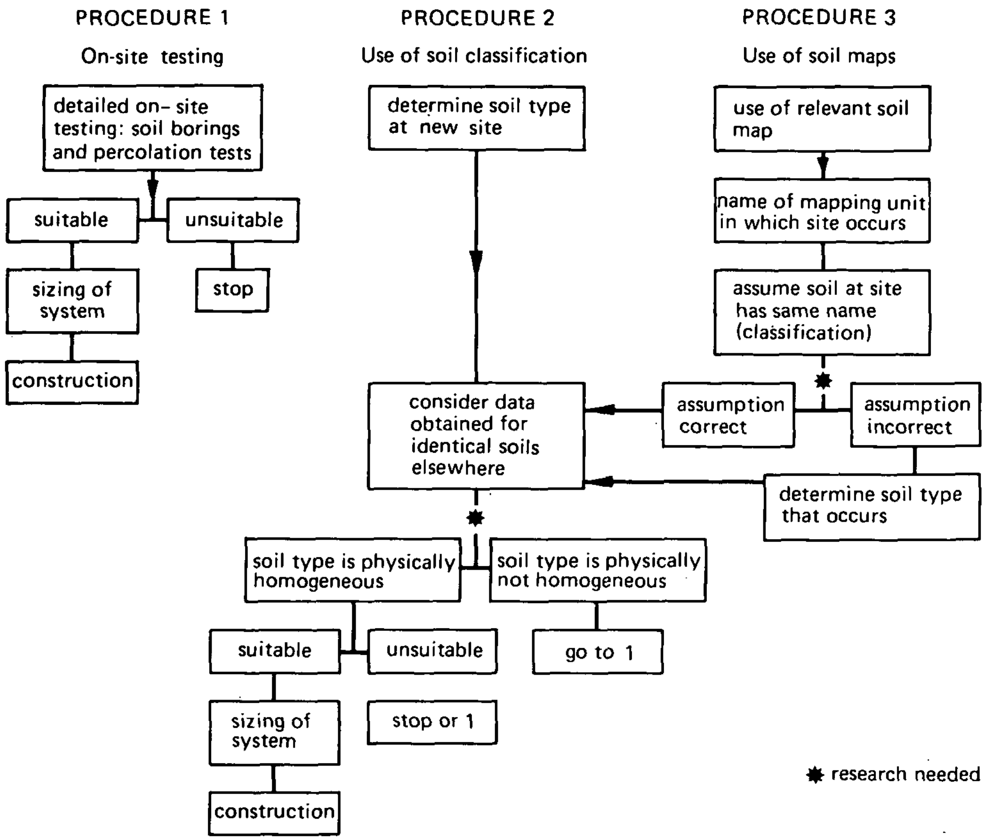


Fig. 38. Schematic diagram illustrating different procedures of data-extrapolation using soil survey procedures. This diagram is based on studies of soil disposal of septic tank effluent and may be less relevant to other soil uses.

properties is still necessary. Soil series with proven low variability of key properties could be directly associated with certain types of innovative systems.

The *extrapolation procedure three* (Fig. 38) is more difficult to accept from a conceptual point of view. Soil mapping units are cartographic units which delineate areas within landscapes and which are normally named for the dominant series within the mapping unit. This unit would contain the named series, other series that do not significantly differ from the named series with respect to "use and management" and some percentage of contrasting inclusions (15-20 % often quoted). Use of this procedure can only be satisfactory if mapping units named for a soil series have few inclusions and if the soil series itself has a low variability of key properties, as discussed in the context of the second extrapolation procedure. Another problem is one of scale posed by the use of detailed soil maps which do not allow showing separate areas if smaller than approximately 1 ha (2 acres). However, a seepage area of 90 m^2 (approx. 0.01 ha) is considered large and this difference shows that this third procedure of extrapolation is not acceptable, except when applied to very homogeneous soil mapping units. However, few data are available on the variability of soil mapping units and the third extrapolation procedure can only become viable if more data of this type are generated. The financial risks involved in using extrapolation procedure three may be much larger than the cost of on-site investigations following procedures two and one. Building a house of $ 40,000,— obviously justifies the cost of on-site testing. Other applications of soil survey may be more appropriate for use of procedures three and, perhaps, two. The economic value of soil survey in relation to different applications has been discussed by BIE and ULPH (1972) and BIE et al. (1973).

Limitations of extrapolation procedures two and three should not lead to the conclusion that the first procedure is the only viable one because the other procedures have some specific and attractive advantages: (i) on-site K measurements and observations needed for extrapolation procedure one, are relatively costly and time-consuming, whereas using limited on-site observations needed for soil taxonomy (extrapolation procedure 2) would be much more economical; (ii) soil maps can be used, assuming that soil mapping units are reasonably homogeneous, to show the potential future impact of innovative technology on land use patterns before any development has taken place (BEATTY and BOUMA, 1973). Many soils not suitable for on-site disposal now may be used in the future and this has implications which depend on how large an area will potentially be affected and where such areas occur. If this analysis indicates the potential for major change in large areas, attempts may be started early to create new zoning laws if such developments are considered undesirable. If, on the other hand, this analysis indicates that future effects will be minimal, concern can be channelled in time to more worthy causes.

Procedure two is currently being tested in Wisconsin and preliminary results are encouraging (procedure three has only been accepted for one soil series).

5.4.3.3 A generalized scheme for soil survey interpretation

The relevance of this case study should go beyond technical aspects of on-site disposal of septic tank effluent which is, incidentally, considered to be a very major area of concern in the United States. In addition, the study illustrates an approach which may be useful for other types of soil survey interpretations. The approach is schematically represented in Figure 39 which compares the conventional procedure (left side) with this innovative one. The following points are essential for this comparison:

(i) The conventional procedure uses practical, empirical experience, which is often also reflected in health codes or other laws or regulations, to define soil survey interpretations in terms of limitations or suitability. As discussed, this useful procedure is confined to applying conventional technology and can often not be used to answer the common question as to what can be done to overcome the problems encountered.

(ii) Interdisciplinary research, on a scientific basis, is needed to analyze the problems that occur and to develop and test new technology. Field soil scientists should play a role in the research, as discussed earlier. Work at this level requires that soil data be more specific; so measurements rather than estimates are needed.

(iii) The interdisciplinary research should be focussed on the production of technology and management "packages" which specifically define *what* has been developed, *how* it can be used, and *where*. The Wisconsin studies on liquid waste disposal resulted in "packages", which included a list of suitable soil series. Such "packages" are ideal vehicles to define soil *potential* for a given use (MCCORMACK, 1974).

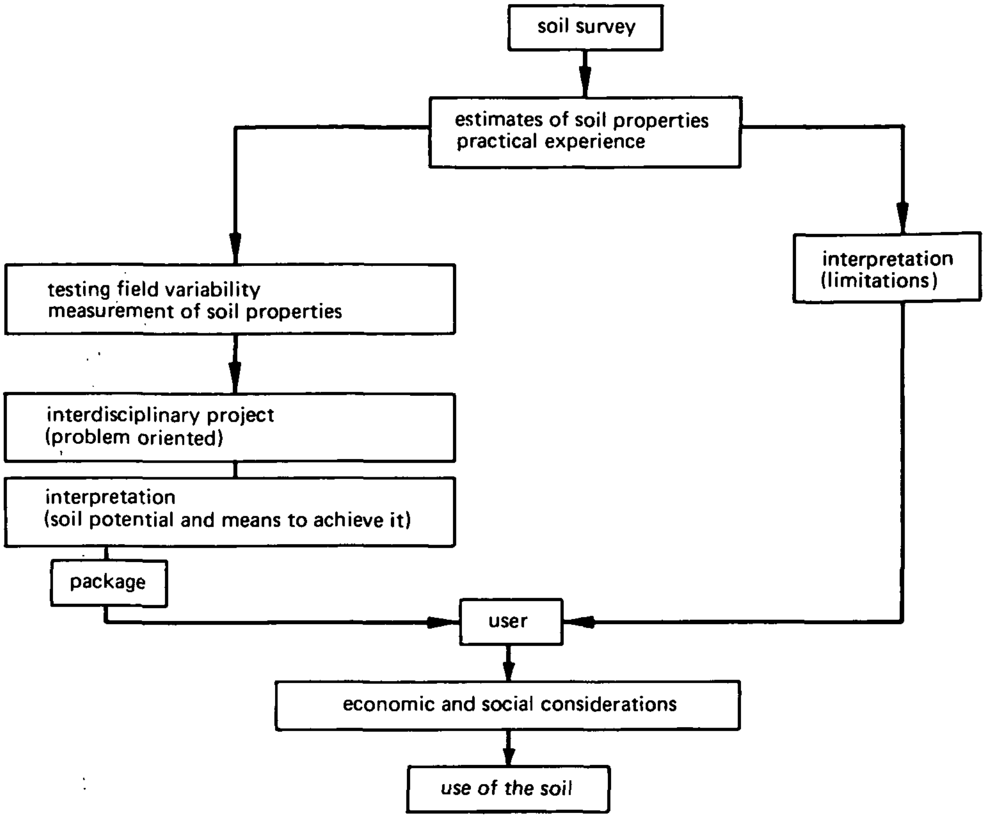


Fig. 39. Flow diagram illustrating procedures which may define soil limitations and soil potential for a given use.

(iv) Soil survey data should be better characterized to allow extrapolation of research data from measured to identical, unmeasured soils. Special emphasis should be given to: (a) the variability of selected physical characteristics within given taxonomic units of the soil classification system (BAKER and BOUMA, 1976; VENEMAN et al., 1976) and (b) the "purity" of major soil mapping units. However, other extrapolation procedures may prove to be useful as well (WEBSTER and BURROUGH, 1973).

6. GENERAL SUMMARIZING CONCLUSIONS

Data presented in the previous chapters and associated reviews and discussions can be summarized in some general conclusions:

1. Morphological studies of soil pores are most significant if specific advantages of morphological techniques are utilized rather than if just an attempt is made to reproduce results which could have been more easily obtained with physical methods. These specific advantages are the ability to distinguish different types of pores and their dimensions and relative volumes. Of particular interest is, in addition, the number of tubular and the length of planar voids within a given cross-section of soil, as these values are essential for calculations of hydraulic conductivity. In addition, morphological data are also essential to define optimal sample sizes for different soil horizons to be used for soil physical determinations.

2. The derivation of "equivalent pore-size distributions" from moisture retention data, to be used for K calculations, appears to be most meaningful for sandy soils. Extending the application of the pore interaction model of Collis-George and Marshall from sands (where it applies quite well, as was shown with our morphological data) to all other soils may have been unwise. The introduction of matching factors (which have no physical meaning) has created a questionable illusion of success. Also, real pore-size distributions in sandy soils, as measured in thin sections, can best be derived from physical adsorption rather than from the usual desorption curves. This conclusion followed from our measurements and from an analysis of the hysteresis phenomenon.

3. The *in-situ* crust test procedure appears to be most suitable for obtaining K values *at*, but particularly *near* saturation. All other discussed methods tend to be insensitive at such moisture contents. In addition, different variations of the instantaneous-profile method can be used to obtain K (or D) values at lower pressure potentials. But these methods are very time consuming and require advanced instrumentation. The simple, cheap and rapid hot-air method of Arya should be extensively tested because this method, if found acceptable, would finally allow the determination of K and D values on the necessary large scale.

4. A characterization of the soil moisture regime in more quantitative terms, based on the well established physical flow theory, is necessary for improved soil survey interpretations which emphasize soil potential rather than soil limitations for a given use. However, this characterization can only be realized if field soil scientists and surveyors are familiar with the concepts involved. An attempt was made in this publication to present an essentially non-mathematical explanation of some key aspects of the flow theory.

5. Soil morphological features, such as iron and manganese mottles as discussed in this publication, can be very useful for predicting certain aspects of the dynamics of the soil moisture regime. However, physical interpretations of morphological features should be checked by means of exact physical measurements and should be independently confirmed, if feasible, by model experiments. If this is done, there are two advantages: (i) correct physical interpretations of morphological soil features can save

much expensive physical work by allowing extrapolation of physical data on the basis of a morphological description, and (ii) the physical interpretation of morphological features is thus continuously extended and refined, increasing its usefulness.

6. The definitions of "representative" hydraulic conductivity and moisture retention data for soils represented on soil maps require multiple measurements in each unit and a definition in terms of probabilities that values will be higher or lower than a given value, rather than a definition in terms of a single "average" value. There is a clear analogy with the presentation of climatic data.

7. Soil survey thinking has traditionally been centred around pedogenesis. Now, however, soil classification systems have been developed (on the basis of pedogenesis), and emphasis is increasingly placed on interpretations. What may be "identical" from a pedological point of view, may *act* differently from some physical point of view. Also, what may be "different" pedologically, may act "identical" in some physical sense. However, one basic underlying assumption of the traditional soil survey discipline is the expectation that the different "products" of a given combination of state factors of soil genesis (which are logically categorized in a classification scheme) will also "behave" in distinctly different patterns. This expectation implies that data, measured in a particular type of soil, can be extrapolated to unmeasured soils with identical classifications. Much testing remains to be done to justify this procedure. Specifically, the percentage inclusions within major mapping units and the range of physical properties of major soil types should be defined. This approach elaborates on and refines concepts that have traditionally been used in soil surveying. However, alternative procedures are also being developed now which use statistical interpolation techniques and computer technology rather than soil classification to extrapolate data from measured to unmeasured soils.

8. Flow through aggregated ("pedal") soils with planar voids (cracks) and worm and root channels cannot be well described with the traditional flow theory which requires the presence of an isotropic and homogeneous porous medium. Chloride breakthrough curves can be determined to define the minimum travel-time of newly applied liquid through a given volume of soil at a given flow regime *and* the time required for total displacement of the initially present liquid. These data are of great practical significance and can not be derived from K data assuming piston-type flow, as occurring in sands. Breakthrough curves, which can sometimes be expressed mathematically in terms of apparent coefficients of hydrodynamic dispersion, provide an attractive and simple measure for the vertical pore continuity in large, natural soil samples. Examples presented in this publication demonstrate that: (i) the behaviour of a given soil can be manipulated by (in these studies) varying liquid application techniques. Such data are essential for interpretive purposes which increasingly require a positive definition of required soil management resulting in a fulfilment of whatever potential a soil has for a given use, and (ii) different structures, now routinely described in the field but hardly used in interpretive work, behaved significantly different in terms of dispersion, thereby suggesting their possible usefulness as an indicator of this particular aspect of physical behaviour. Emphasis in this type of structure description should be on macroscopic features and on simplicity to avoid unreproducible results which may be associated with the use of very detailed descriptive schemes.

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