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THE ROLE OF MACROPORES IN THE HYDROLOGY OF FIELD SOILS

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ABSTRACT

The aim of this report is to give a review of present knowledge of water flow in soil macropores. A large variety of research papers deal with this subject and their results are considered in the context of hydrological evidence. Macropores are defined as voids in which the water is not subject to capillary forces. Their minimum width has to be larger than about 4 mm.
A classification of these voids is proposed and a model of water-flow in a combined micropore/macropore system is presented.
A drainage experiment carried out on a large undisturbed soil sample produced some of the parameters required by the model.

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1. INTRODUCTION

There is a trend in current theoretical hydrology to model the processes of catchment hydrology more and more completely. As a result of this process gaps of knowledge become app rent where information from different branches of hydrology does not fit together well. One such domain of limited knowledge is the hydrological role of water flow in large pores in the soil. Many observations in soil hydrology have shown that during or shortly after heavy rainstorms or intensive irrigation, water may move much faster through upper soil horizons than infiltration equations based on the flow equat. on of RICHARDS (1931) would predict. As one example, GERMANN (1976) has reported that following a total precipitation of 30 to 40 mm in two days, tensiometers in a loess soil show a very rapid response down to depths of 40 to 60 cm. These observations suggest that an important part of the infiltrated water flows through a system of large pores that allow fast flow velocities and bypass the soil matrix. This system will be called the macropore system of the soil. The importance of macropores was recognized at least as early as the middle of the last century and has been periodically mentioned in the literature in a qualitative way.

According to the classical basics of pedology, a soil is defined as a product of parent rock, climate, topography, organisms, human influence and time (see for example FITZPATRICK, 1971). Considerable success in understanding the behaviour of soil water has been achieved by conceptualising the soil as a continuous porous medium. The Richards equation is based on this approach. However, pedalogical considerations suggest that discontinuous components, such as macropores due to soil fauna or soil structure, should be expected in field soils. It is suggested here that these discontinuities may be important hydrologically, but cannot be adequately represented by simple porous media concepts.

The most important macropores have a width between 1 mm and 1 cm. The water in smaller voids is subject to capillary forces, and in larger cavities the maximum possible flow rate is so large that it would not normally be exceeded by normal rainfall rates. Most of the channels formed by earthworms and the main roots of annual crops lie within this range. Although macropores may consistitute only a small portion of the total porosity of a soil, and may flow for only restricted periods of time, they may have a far greater importance in terms of the volumes of water and solutes moved. This has important implications for the movement of pollutants and the use of field soils to filter waste water.

The development of research on water flow in macropores may have been restricted by the recent predominance of the Richards equation in analysing water flow in soils, and the tendency to treat all rapid water discharge from hillslopes or catchments as overland flow. The Richards equation works well in describing the flow of water in laboratory columns of homogeneous soil, but has been less successful in applications to field soils, particularly at forested sites. In agronomy, where there is greater emphasis on soil water storage for crop production than on soil water flow, the possibility of rapid flows in large soil voids is reflected in the persistence of terms such as field capacity in describing the characteristics of field soils (see for example VEIHMEYER and HENDRICKSON, 1949).

2. DESCRIPTION OF SOIL MACROPORES

2.1 When is it a macropore?

The capillary tension at which water is held in a cylindrical pore of radius r may be approximated by

$$\psi = - \frac{2\sigma}{r \cdot \rho \cdot g} \qquad 2.1$$

where σ is the surface tension at the air-water interface, ρ is the density of water, g is the gravitational acceleration and ψ is in units of length. As pore radius increases, the capillary tension decreases to a point at which ψ is small relative to gravity forces on the water in the pore. Pore sizes above this size may be called macropores. However, the definition of this point is purely arbitrary. Indeed, many papers refer to macropores or non-capillary pores without defining the size range involved. McDONALD (1967) defines macropores as 'those which are drained by an applied tension equivalent to a column of water 60 cm height'. REEVES (1980) in a study of percolation through fissures in chalk, refers to microfissures with an opening of 1-200 µm, macro-fissures 0.2 - 2.0 mm and enlarged macrofissures in the range from 2-10 mm.

For the purposes of this report an arbitrary value of ψ = -1.0 cm will be used to define the boundary between micropores and macropores. The equivalent mean pore radius calculated from equation 2.1 is about 1.5 mm. It should therefore be safe to define pores in the range of 2 mm and above as macropores.

2.2 The occurence of soil macropores.

The larger macropores have the advantage of being readily visible, and are often mentioned in the qualitative description of soil profiles. More quantitative information is rarer. WILLIAMS and ALLMAN (1969) found macropores in a loess profile down to a depth of 10 m. The pores were cylindrical with a diameter of 5 to 10 mm and ranged in frequency from about 100 per m^2 close to the surface down to about 50 per m^2 at a depth of about 8 m. They were not sure whether the pores were formed by plants or by animals. GREEN and ASKEW (1965) reported on ant and earthworm activities producing macropores of 1 to 10 mm diameter to a depth of at least 1 m. They considered that the macropore system could have been established for several hundred years. Large voids of biotic origin such as mole runs may also be of considerable age and may not be apparent from the surface (MELLANBY, 1971). EHLERS (1975) gives numbers of earthworm channels per unit area for different depths within a loess profile (see Table 2.1) and in a destructive but exhaustive study of a block of forest soil with the dimensions of 2 x 2 x 6 m, AUBERTIN (1971) catalogued the occurrence of macropores of biotic origin (Table 2.2). The A2-horizon of this soil was found to contain 36% of macropores by volume, the bulk of them loosely packed with organic material. At a depth of about 1 m in the C-horizon, there still remained 2.1% macropores by volume.

TABLE 2.1 (after EHLERS, 1975)

Number of earthworm channels per m^2 in tilled and untilled soil

Depth (cm.)	Tilled				Untilled			
	2-5	5-8	8-11	Total	2-5	5-8	8-11	Total
	(diameter in mm)				(diameter in mm)			
2	21	5	1	27	75	40	2	117
20	60	18	1	79	99	41	1	141
30	124	58	5	187	200	91	5	305
60	174	165	9	348	183	172	8	363

Soil structure may also lead to the occurrence of interconnected pore space at the macropore scale in between soil aggregates or peds. Soils with a high clay content that are liable to cracking due to dessication may be particularly prone to such structural macropores. This type of macropore may be expected to change form as the soil wets and dries. Such cracks may not close completely even after prolonged wetting (see for instance BEVEN, 1980) and may be used by roots and soil fauna. However, macropores formed by biotic agents may also be completely unrelated to the soil structure and earthworm channels passing through peds of compact clay soil without reference to the interpedal pore space have been observed at the site described by BEVEN (1980).

Macropores that are visible in the soil profile need not be hydrologically active. In part this may be due to the lack of a sufficient supply of water (see discussion in section 3.1) but it may also be a result of a lack of connectivity of the macropores, some of which may be dead-ended. Infiltration of dyed water or impregnation with dyed resin have been used to demonstrate the effect of water flow through macropores (see for example AUBERTIN, 1971; BOUMA *et al.*, 1977, 1978, 1979; EHLERS, 1975; KISSEL *et al.*, 1973) and BOUMA *et al.* (1977) have been able to distinguish hydrologically effective macropores from those isolated from the flow system on the basis of dye colouring after an infiltration experiment (see table 2.3).

2.3 A simple classification of soil macropores

From a morphological point of view the following groups of macropores may be distinguished:

(a) Cracks or planar voids mainly appear in fine textured soils due to shrinkage during periods of drying or after frost expansion. It is expected that once a crack has been established it will reappear under conditions similar to those that caused it unless the soil is subject to disturbance.

A simple crack may be assumed to consist of two planar, more or less parallel walls with a width and a length depending on the

TABLE 2.2 (after AUBERTIN, 1971)

Surface-area percentages of biotic macropores exposed in the front face, as related to horizon, depth, pH, and bulk density

Horizon	Depth	Total biotic macropores	"Distrubed" area[1]	Open root channels	Open animal passageways	pH	Bulk density
	Inches	Pct.	Pct.	Pct.	Pct.		G/cc.
A1	0.1	-	-	-	-	5.0	0.94
A2	1.6	36.0	28.9	2.9	4.3	4.5	1.26
B1	6.9	11.4	5.8	3.8	1.8	4.5	1.49
B21t	9.16	16.6	13.3	2.8	.5	4.6	1.61
B22t	16.23	14.4	12.6	1.8	0	4.6	1.59
IIB3	23.31	10.8	9.8	1.0	0	4.6	1.34
IIIBC	31.36	5.8	4.8	1.0	0	4.7	1.63
IIIC1	36.42	2.1	1.6	.5	0	4.6	1.72
IIIC2	42.47	2.3	2.1	.2	0	4.6	1.82
IIIC3	47.56	.8	.6	.2	0	4.6	1.78
IIIC4	56.68	.1	0	.1	0	4.6	1.79
IIIC5	(60.63)	0	0	0	0	4.3	1.81
IIIC6	68.74	< .1	0	< .1	0	4.3	1.80

[1] "Disturbed" area refers to those portions of the soil profile that were dissimilar in colour, texture, and structure, from the bulk of the surrounding soil mass. These areas were always darker in colour and highly porous. Frequently they contained bark linings.

TABLE 2.3 (after BOUMA *et al.* (1977))

Average total number of channels and vughs and length of planes for each of the four samples per horizontal cross-section of 70 cm^2 (area thin section) and the average number and length of coloured macropores (also expressed as a percentage of the total number).

	Number of channels								
	Total	Coloured	%	Total	Coloured	%	Total	Coloured	%
Sample I	266	125	47	94	47	50	666	393	59
Sample II	797	167	21	210	65	31	404	186	46
Sample III	560	34	6	113	17	14	267	91	34
Sample IV	647	13	2	156	3	2	790	71	9

soil structure and the degree of shrinkage. Complex crack systems can, to a first approximation for a single plane, be resolved into a number of simple crack systems (DOLEZAL, 1976a; SCOTTER, 1978). Crack walls are often covered with films of clay, silt or organic matter (cutans) which is most commonly translocated from upper horizons. It must be expected that crack systems will be subject to important seasonal variations.

(b) Cylindrical holes or channels are mainly formed by animals or plant roots. In non-disturbed soils, such as forest soils, they may remain for a long time and do not depend on typical seasonal variations. They may be loosely packed with organic or mineral debris. The orientation of such channels may have a strong lateral as well as a vertical component.

(c) Vughs are voids that are intermediate in shape between cracks and channels (BOUMA *et al.*, 1977). Note that non-vertical cylindrical pores may take on the appearance of a vugh when seen in a horizontal plane (an exposed surface or thin section) through a sample.

(d) Soil pipes are large macropores (diameter of 4 cm and more) that carry channeled turbulent flow laterally down hillslopes. Such pipes have been described in the peaty topsoil of upland podsols in Britain (JONES, 1971; NEWSON, 1976), where they may be the result of preferential water movement down lateral cracks initiated by dessication (GILMAN and NEWSON, in press). They are distinguished morphologically by their size and preferred lateral orientation. They are included to show that this simple classification may be incomplete.

3. A BRIEF REVIEW OF SOIL MACROPORES IN THE LITERATURE

3.1 Macropores and catchment hydrology

The importance of the role of macropores in catchment hydrology has been the subject of considerable debate but of very little quantitative study, except perhaps the research of BURGER, which is mentioned below. The idea that such readily visible channels in the soil might dominate the flow of water through the soil is an attractive one and has recurred in the literature for a considerable time. Thus perhaps opening the discussion

> 'the permeability of a soil during infiltration is mainly controlled by big pores, in which the water is not under consideration of capillary forces' (SCHUMACHER, 1864).

Similarly

> 'The drainage water from a soil may thus be of two kinds: it may consist (1) of rainwater which has passed with but little alteration in composition, down the open channels of the soil; or (2) of the water discharged from the pores of a saturated soil. The latter water, the true drainage of the soil, will itself escape to a greater or less extent through the channels already mentioned. The respective proportions of direct and general drainage will vary much in different soils, and under different circumstances. In a light soil, of naturally free drainage, channels can play but an insignificant part, the rain being at once absorbed by the main body of the soil, and freely discharged again from its pores when saturated. In heavy soils, on the other hand, both the absorption and the discharge of water can take place but slowly, and the part which natural channels play in freeing the soil of water is more considerable. In a heavy soil channel-drainage will most cases precede general drainage; a portion of the water escaping by the open channels before the body of the soil has become saturated; this will especially be the case if the rain fell rapidly, and water accumulates on the surface.' (LAWES, *et al.* 1882).

> 'In considering flow through natural upper soil horizons, the formulae of soil mechanics do not generally apply. Here porosity is not a factor of individual soil particle size but rather of structure determined by particle aggregates which form a three-dimensional lattice pattern. This structure is permeated throughout by biological channels which in themselves also function as natural hydraulic pathways. A single dead root channel, worm hole or insect burrow may govern both the drainage of water and the escape of air through a considerable block of soil.' (HURSH, 1944).

'Most of the observed storm flow seeped through root holes, cracks, decayed root channels and earthworm holes. It was obvious that these openings were interconnected and that many were open to the soil surface.' (WHIPKEY, 1967).

'Thus it is postulated that a temporary perched water table occurs soon after the onset of rain and subsurface flow rapidly develops in the top 20 cm recharged by pipe flow infiltration due to innumerable macropores.' (BONELL and GILMOUR, 1978).

The fact, that water can flow through macropores is not in dispute. What generates considerable discussion is the relative importance of macropore flow to the hydrological response of the soil and of complete hillslopes and catchments. Quantitative work on the movement of water through the soil has been predominantly based on the Richards equation which assumes that the velocity of flow is proportional to the hydraulic gradient in the pores of the soil matrix as expressed in Darcy's law. Such flow will respond to gravity, capillary and osmotic forces in the pores of the soil matrix. This type of law derives from experiments on saturated laboratory samples under low hydraulic gradients (DARCY, 1856) and the linear relationship assumed in Darcy's Law may not always hold (SCHEIDEGGER, 1957; DE VRIES, 1979). By allowing the constant of proportionality (hydraulic conductivity) to have a non-linear and possibly hysteretic relationship to the degree of saturation the Richards equation can be extended to the case of unsaturated soils (KLUTE, 1952), and GARDNER (1967) stated quite plainly, that 'modern infiltration theory is based on the theory of water movement in unsaturated soils', when it may be expected that all macropores in the soil will be dry.

If the working definition of macropores outlined in section 2.1 is realistic, i.e. that water in macropores is not subject to significant capillary forces, then it is reasonable to assume that under unsaturated conditions pressures within the macropores are essentially atmospheric. Under saturated conditions hydrostatic pressures and the effects of entrapped air may also be important. An analysis of the relationship between water flow in the macropores and seepage in the adjacent soil matrix in terms of Darcy-type flow in the micropores would suggest that

1. Flow will only be maintained in the macropores if the immediate soil micropores are saturated, and

2. Flow will only take place from the micropores to the macropores if the matrix is saturated.

This has led some workers to question the importance of macropores on the basis that flow in macropores requires the maintenance of saturated conditions and fully saturated conditions occur only rarely in the field (see for instance HIBBERT, 1967). Indeed many of the studies that have tried to demonstrate the importance of macropores have been based on artificial surface inputs to the soil either by sprikling or ponded infiltration, at rates that must increase the probability of flow in macropores open to an essentially saturated surface (e.g. McDONALD, 1968; WILLIAMS and ALLMAN, 1969; CHAMBERLIN, 1972; RITCHIE *et al.*, 1972; DE VRIES and CHOW, 1978). However, there is also evidence that macropore flow can be maintained through otherwise unsaturated soils (WHIPKEY, 1967; PILGRIM *et al.*, 1978). In this case it is important that there should be a supply of water to the macropores and that the rate of supply should be sufficient to overcome the infiltration rate from the sides of the macropores into the soil matrix. Sources of supply might be surface flow due to a rainfall intensity greater than the infiltration capacity of the micropores, or concentrated flow reaching the soil surface due to stemflow or concentrated drip from the vegetation cover (WHIPKEY, 1967). Such sources of supply may be expected to vary both within and between individual storms. Once water enters a macropore it needs only to wet the adjacent matrix for flow to be maintained, and the depth of penetration may be enhanced by the hydrophobic nature of some soils when dry (PIERCE, 1967). This process has been demonstrated on a large scale for water flow in natural soil pipes down a hillside (GILMAN and NEWSON, in press), and for a clay matrix of low permeability where even at the end of a winter season the soil inside the structural units was unsaturated and flow was shown by means of dyed water to take place predominantly in the interpedal cracks and other macropores (BEVEN, 1980). REYNOLDS (1966) used powdered dye to trace the movement of natural rainwater and showed that the deepest penetration was related to the presence of macropores. It is clear that the presence of macropores can increase the surface available for infiltration by a significant factor, spreading the input of water over some depth in the profile.

It is difficult, however, to draw together the essentially qualitative small scale evidence on the role of macropores in water transport in field soils into a realistic assessment of importance at the hillslope or catchment scale. It would appear that macropores may have a dual role to play in that by increasing the infiltrating surface in the soil, they might decrease the amount of storm runoff (surface or subsurface) reaching a stream channel, while by providing rapid flow routes within the soil they might increase the amount of subsurface storm runoff. An analysis of subsurface flow by FREEZE (1972) based on Richards' equation has suggested that only in soils of very high conductivities can subsurface flow make a significant contribution to storm runoff and that saturation excess overland flow on a variable contributing area as described by DUNNE and BLACK (1970) is a more likely mechanism of storm runoff generation for most soil types. It seems likely that, under some conditions the presence of structural and biogenetical macropores in the soil may serve to increase the range of soils for which subsurface stormflow as described by HURSH (1944) and WHIPKEY (1967) is a valid alternative mechanism. Macropores could provide both very rapid flow rates within saturated zones and a fast pathway for water to reach a saturated zone (BONELL and GILMOUR, 1978). CHILDS *et al.* (1957) for example reported clay soils with saturated conductivities in the field equivalent to gravelly soils. Under some conditions, such as flow through a litter layer (EAGAN, 1967) subsurface storm flow may grade into overland flow with similar velocities. The fact that the postulated mechanisms of storm runoff response are not mutually exclusive has been stressed by BEVEN (1975) and PILGRIM *et al.* (1978). A view of subsurface flow in which macropores conduct flow, even if only under saturated conditions, is one of non-uniformity of processes in both space and time (WEYMAN, 1970; WHIPKEY, 1967; AUBERTIN, 1971; PILGRIM *et al.*, 1978). The difficulties of analysing flows of this nature are demonstrated by ARNETT (1974), who shows that flow patterns on a hillslope in Claydale, North Yorkshire, were poorly correlated with slope angle, slope length, surface roughness and soil texture, but were apparently closely related to seasonal cracking of the topsoil and the lateral distribution of living and dead Bracken rhizomes. There is a growing body of evidence that macropores play an important role in the recharge and movement of groundwater aquifers; as a result of tracing experiments (ATKINSON and SMITH, 1974; RODDA *et al.*, 1976) and the interpretation of flow, well and water quality (see WILLIAMS and ALLMAN, 1969; REEVES, 1980; SMITH, 1979; RUSHTON and RATHOD, 1979). Further evidence that macropores may be important at the catchment scale can be obtained indirectly from short term water balance studies. A well-documented flood flow in the Rietholzbach catchment, Switzerland, can be used to illustrate this. The catchment has a surface area of 3.2 km^2 and is situated in the Swiss Prealps between 600 and 900 m above sea level. Land use in the catchment is 20% forest and 80% grassland. Cumulative precipitation and runoff for the heavy storms of 7 and 8 August, 1978 are shown in figure 3.1. The difference between these curves is a measure of cumulative water storage in the catchment, which is at least twice as high as the measured increase of soil water content (as determined at 17 sites with a Wallingford neutron probe from 0.1 m to a depth of 1.4 to 1.8 m). Soil moisture content in the lower horizons of most of the soil profiles did not change significantly. Thus there is a large volume of water that was neither observed percolating through the soil nor running off at the weir. (The interception loss of the short grass vegetation is estimated to be about 5 mm). A single overland flow measurement site did not register any surface runoff and a later infiltration experiment demonstrated, that a rainfall intensity at least twice as high as the highest observed one would be necessary to generate any surface runoff. All the evidence suggests that the 'lost' water was transmitted rapidly through the soil, via macropores constituting a very small portion of the pore space (perhaps within the measurement accuracy of the neutron probe), to deeper layers. During the two weeks after the event, discharge from the Rietholzbach catchment remained higher than normal baseflow levels, indicating the discharge of deeper lying water storage (see JAHRESBERICHT, 1978).

3.2 Field observations of water-flow in soil macropores

One of the most comprehensive studies on soil macropores has been carried out by BURGER between 1922 and 1940. The original aim of this research was to demonstrate the influence of different types of land use on the hydrological behaviour of soils. In particular several different forest treatments were studied. Starting from ENGLER's comparative observations in the catchments Sperbelgraben (55.8 ha, 99% forested) and Rappengraben (69.7 ha, 31% forested, 5% arable land and meadows, 64% pasture land) within the same geological formation (ENGLER, 1919), he proceeded to explain differences of infiltration capacities by differences in soil air capacity. The latter expression

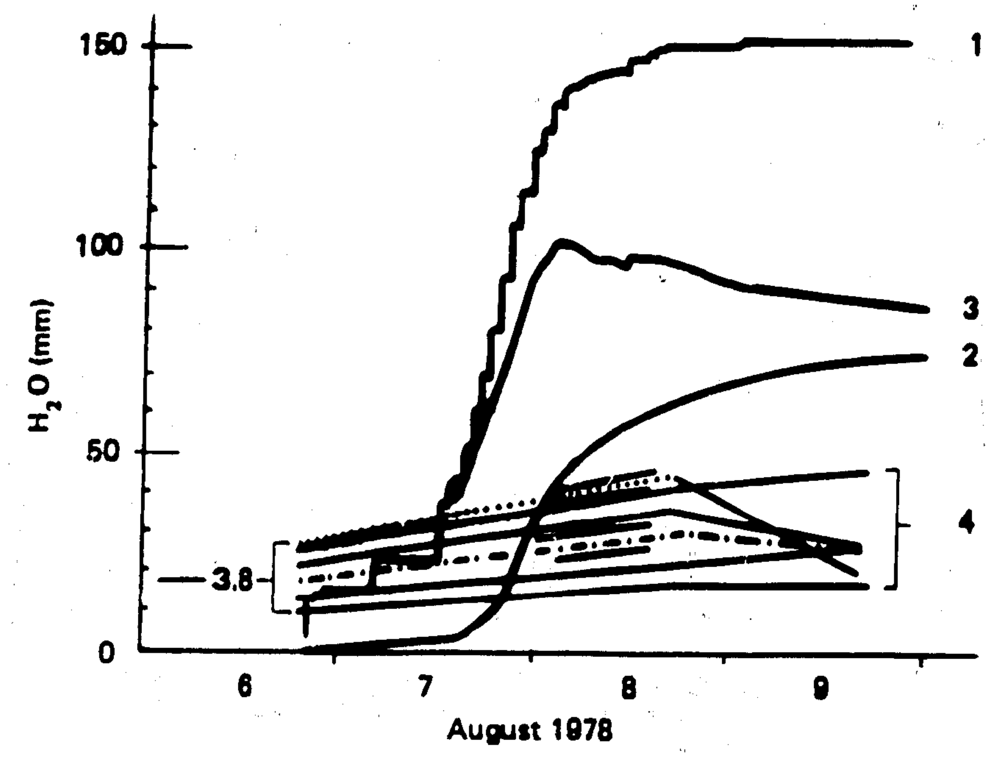

Figure 3.1 **Cumulative precipitation, run off and storage within the catchment, compared with the soil water storage during the storm of August 7 and 8, 1978 (data from Rietholzbach catchment, Switzerland. See 'Jahresbericht 1978 der VAW, 1979)**

(1): cumulative precipitation
(2): cumulative run off
(3): (1) – (2) cumulative water storage within the catchment
(4): soil water storage, determined with a Wallingford soil moisture probe (0 to 1.5 m) measured August 3, 8 and 9 at 10 different sites

was introduced in soil physics by KOPECKY (1914) and was interpreted by BURGER (1922) 'to be independent of the total pore volume of a soil, but obviously dependent on the volume of pores in which the water is no longer held by capillary forces'. This soil air capacity may be taken as a measure of the amount of macropores, according to the definition used in this report. The soil air capacity was determined on undisturbed soil cores, taken with iron cylinders of a height of 10 cm and a cross sectional area of 100 cm^2 (still called the "Burger-cylinder"). The core, remaining in the cylinder, was fully saturated to its top over a period of 24 hours. Afterwards it was gravitationally drained on a layer of gravels for 1 hour. Then the sample was weighed, dryed at 120°C, weighed again and the volume of the soil material determined. The soil air capacity is then the difference of the total volume of the sample minus the volume of the soil material minus the water volume after drainage, expressed as a percentage of the total volume. The infiltration capacity was the time needed for a column of 10 cm of water to infiltrate into the soil. For this purpose another of these cylinders was put completely into the soil with a watertight extension tube above. The top of the soil was protected with a fine sieve. Then one litre of water was poured into the extension and the time measured for its infiltration. BURGER also mentioned that a second, immediately following, infiltration of the same quantity of water always needed more time that the previous one. So he used only the first measurements. As it will be pointed out below, this kind of infiltration measurement is not a very satisfactory measurement, especially carried out with relatively small cylinders. Therefore this infiltration capacity will be treated like a soil structure test rather than an absolute value. BURGER also indicated the huge variability of this parameter within a soil unit. Nevertheless, with these two parameters he was able to demonstrate the influence of different management practises on the hydrological behaviour of the soil as well as the destruction of the macropore system due to camping activities within a forest (BURGER, 1922, 1927, 1929, 1932, 1937, 1940). The immense collection of data on this subject would be worth a re-examination from more recent points of view. Here only two conclusions are reproduced. From fig. 3.2 it is clear that an increase of air capacity increases the infiltration capacity of of different soils. Figures 3.3a and 3.3b show that the infiltration capacity may decrease due to an increase in initial water content.

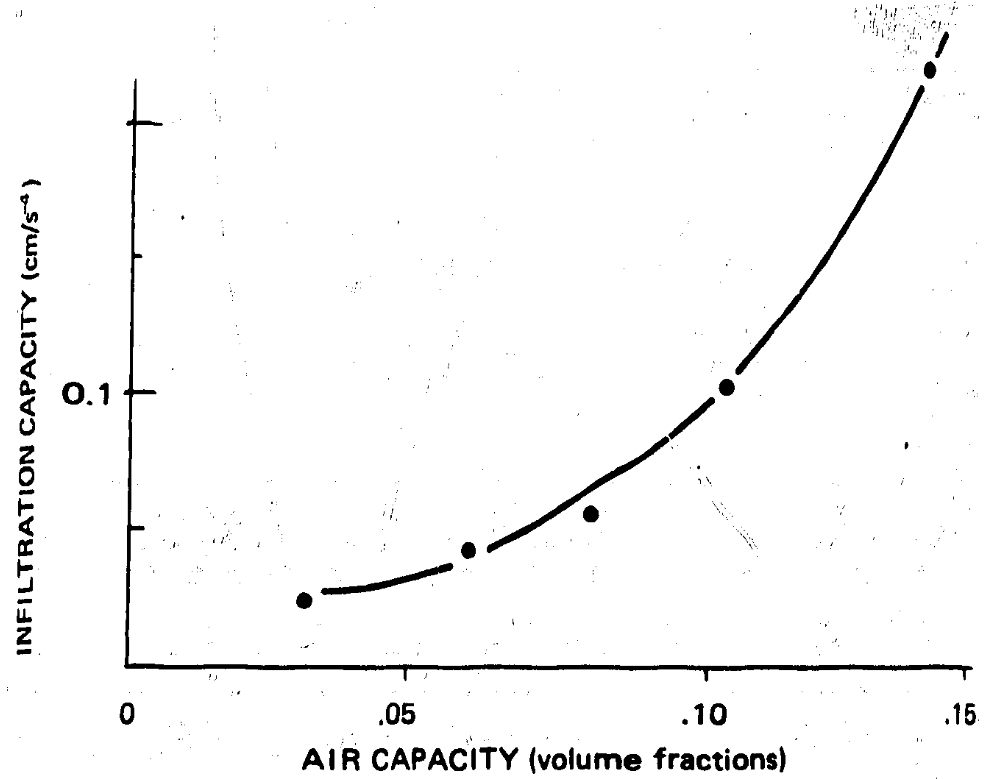

Figure 3.2 Infiltration capacity as a function of air capacity, which is a measure for the total volume of soil macropores, according to Burger (1922)

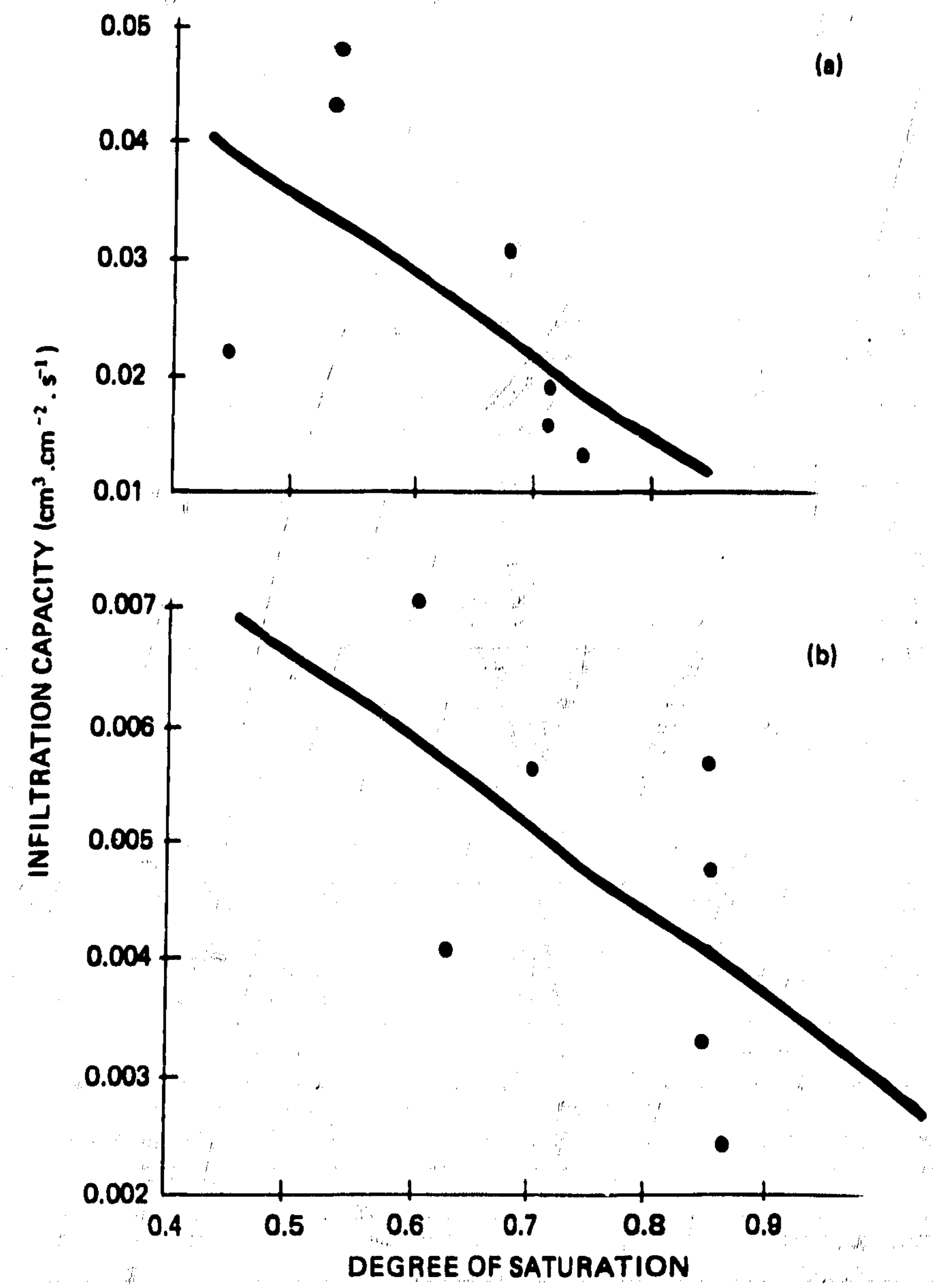

Figure 3.3 Infiltration capacity as a function of soil saturation at the beginning of the infiltration experiment (Burgher, 1922)

(a) Soil with relative high air capacity of 0.14 (vf)
(b) Soil with relative low air capacity of 0.06 (vf)

The infiltration capacities of single worm holes of different diameters have been measured by EHLERS (1975). He found by regression analysis

$$y = 32.2\ x^{3.8} \qquad 3.1$$

where y is the infiltration capacity ($cm^3 \cdot s^{-1}$) and x the radius (cm) of the worm hole. He infiltrated dyed water for a period of 5 min. The value of the exponent is of particular interest: according to the theory of Poisseuile it should be the fourth power of the radius. PETERSON and DIXON (1971) observed during an infiltration experiment, that the opening of a pencil-sized pore increased the pore space of the infiltration area of 1.35 m^2 by only 0.002% but increased the infiltration rate from 6.0 to 10.5 $cm.h^{-1}$. The flow into this pore was estimated to be about 30 $m.min^{-1}$. This velocity corresponds to a Reynold's number of 3334, which would suggest turbulent flow.

There are many observations and measurements on water flow in macropores at the plot scale. Some remarks have to precede the literature review on this item. Reports only assuming a macropore system are not mentioned further. Measurements with infiltrometers are also not metioned further, except where they are included in experiments explicitly describing the effect of macropores. BERDANIER and HANNA (1970) reported great variability in infiltration capacity measured in this way even within the same soil cartographic unit and the report of HILLS (1970) on the use of cylindrical infiltrometer is also not very optimistic on the interpretation of the results. Water may either percolate laterally away and indicating too high an infiltration capacity, or be stopped by entrapped air giving too low an infiltration capacity.

DE VRIES and CHOW (1978) reported from sprinkler experiments in a forest soil, that the infiltrating water partially bypassed the A-horizon and very strange water potential distributions in the vertical direction occurred. AUBERTIN (1971) concluded from his infiltration experiments, also carried out on a forest soil, that the water flow capacity of a field soil is several hundred times greater than the hydraulic conductivity of the soil matrix. RITCHIE *et al.*, (1972) compared the infiltration capacity of a 10 m^2 plot, a 2.5 m^2 plot, big undisturbed soil samples (diameter 0.7 m length 0.55 m) and disturbed samples of the same soil. The infiltration capacity decreased with 2 to 4 $cm.d^{-1}$ from the two plots, to 0.5 $cm.d^{-1}$ for the sieved sample. Even the large undisturbed soil sample used by these authors did not seem to be big enough to represent the field soil. KISSEL *et al.* (1973) found from infiltration experiments on a ploughed soil with Cl^--solution under steady state conditions that the chlorine from was 8 to 10 times more advanced than piston flow in the soil matrix would have predicted. DIXON and PETERSON (1971) proposed six types of macropore structure to describe infiltration. Their macropore-channel concepts link infiltration and surface roughness. They were able to change the macropore structure from one type to another by treating the surface and thereby change the infiltration capacity of the soil in question (see also PETERSON and DIXON, 1971). Entrapped air also diminishes the infiltration capacity of a soil, as has been shown by DIXON and LINDEN (1972), LINDEN and DIXON (1973, 1975, 1976) under border irrigation conditions. They found the infiltration capacity to be reduced by a factor of ten when the soil air pressure increased only by 5 mbars. This amount of pressure increase seems to be quite common shortly after infiltration has started. However, the infiltration of 40 $cm.h^{-1}$ resulting from border irrigation is extremely high.

3.3 Laboratory experiments on water flow in macropores

In a number of experiments on saturated laboratory soil columns breakthrough curves of solutes have been measured. The nature of the flow within the column can be interpreted from the form of such a breakthrough curve. For example, when the solute reaches the bottom of the column in a sharply defined front at the same time as a volume of water equal to one pore volume of the soil sample has drained out, then a piston-like displacement flow can be inferred (see curve 1 in figure 3.4). When 50% of the change of concentration is reached in the outflow before 50% of the pore volume has drained out, then some kind of bypass system can be inferred. In this case there are at least two possibilities: first, that a significant portion of the flow takes place in a macropore system; second, that the flow takes place in the micropores but that a significant portion of the micropores constitute dead space, that is pore space not interacting with the flow of solute. The interpretation of a breakthrough curve of this type would then also depend on the velocity of solute flow through the sample. RITCHIE *et al* (1972) and KISSEL *et al* (1973) describe breakthrough experiments in large undisturbed samples of swelling clay soil that suggest that about 60% of the soil water was inactive, while for a disturbed sample (sieved material from the same soil) this fell to 40%. They inferred that sieving and repacking the soil destroyed the structure that was responsible for the different behaviour of the undisturbed soil.

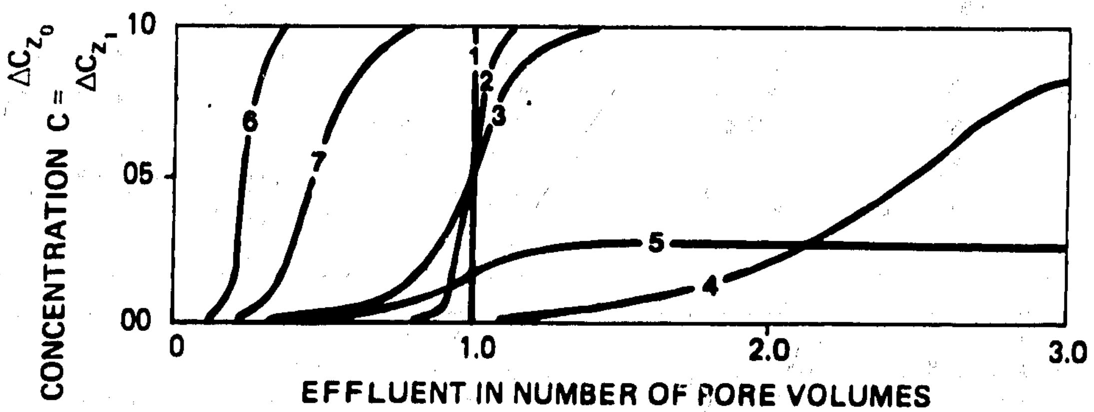

Figure 3.4 Different types of breakthrough curves
(1) piston flow
(2), (3) solute without adsorbance at the soil particles (Cl^-, e.g.)
(4) solute with adsorbance at the soil particles (Mg^{2+}, e.g.)
(5) solute with partial degradation during flow (NO_3^-, e.g.)
(6), (7) water flow through macropores is suggested curves
(1) to (5) after Flühler (1975)

ANDERSON and BOUMA (1977a,b) described breakthrough experiments with chlorine solution in fully and partially water saturated soil columns of differing structure. When the supply of water was unlimited, the water moved mainly in the cracks and voids of the soil (interpedal pore space according to the authors). However, applying a gypsum crust to the top of the sample, thereby limiting the flow rate to about 1 cm/day, changed the shape of the breakthrough curves dramatically. Much more solute was exchanged by the slowly flowing water. BOUMA and WÖSTEN (1979) obtained similar results from the same type of experiment for two differently structured clay soil columns. They also compared the pore size distributions for the soil of the columns as calculated from breakthrough curves, water retention curves and morphometric measurements. The latter procedure suggested a bimodal distribution with some macropores and a lot of very fine pores (diameters from 5 to 0.5 μm), whereas the other techniques suggested a much smoother distribution. However, by making some simple assumptions about the nature of the macropores, the authors predicted the start of the breakthrough quite precisely.

SCOTTER (1978) proposed a model for the breakthrough of chlorine and phosphate solutions as influenced by macropores, based on the laboratory measurements of KANCHANUSUT *et al* (1978). The experiments were carried out in small columns 5 cm long, in which the soil was packed artificially and voids of differing geometry had been manipulated. These authors proposed that the threshold between macropores and micropores should be at 0.1 mm for planar voids and at 0.2 mm for cylindrical channels.

3.4 Water flow in macropores of heavy clay soils

It is well known that crack systems in clay soils may be expected to vary in form seasonally, as the soil wets and dries. Consequently, seasonal changes in the hydrological properties of the macropore systems of such soils is also expected. PHILIP (1969) suggested a theory in which an additional overburden potential is introduced to describe the effect of swelling on the hydrostatic water properties of heavy clay soils. These ideas have been used for example by SPOSITO (1975), SMILES (1976), and SPOSITO and GIRALDEZ (1976). The experimental determination of soil water properties during swelling is quite difficult. A method for the determination of diffusivity for swelling soils is given by SMILES and HARVEY (1973). Similar methods were also used by DOLEZAL (1976b) to determine the hydrostatic and hydrodynamic properties of the micropore system in a swelling soil.

However, the evidence reviewed in the previous sections would suggest that even if there are only a few macropores they will be much more efficient at conducting water than the micropores of a clay soil. It has already been noted that there may be an effective system of fissures even at the end of a wetting period (BEVEN, 1980). DOLEZAL (1976a) has presented a theory of the 'hydraulic efficiency' of macropores which is equivalent to an infiltration capacity. Together with methods for describing the macropore system of swelling soils (eg. NOVAK, 1976 a,b) this approach could lead to a practical way of modelling water flow in the macropores of heavy clay soils. KUTILEK and NOVAK (1976) describe a model that combines water flow within the cracks of a shrinking soil with infiltration into the micropore system.

3.5 Summary

The literature reviewed in chapter 3 can be broadly summarised in terms of three approaches or levels of gaining information about soil macropores.

(a) The microscopic level is concerned with the detailed description of the size and structure of macropores and macropore systems. The experimental techniques used at this level are generally destructive and time consuming. Thus, measurements cannot be repeated on the same sample and changes over time cannot be investigated. The results are more useful in improving understanding of how water flows in macropores, rather than how important it will be in hydrological terms under particular conditions. A number of research centres are developing methods at the microscopic level, and a lot of information should be forthcoming in the near future.

(b) The mesoscale level is represented by the results of experiments on large soil samples and columns. Here the nature of the water flow through the sample is the focus of attention. Results from experiments at the mesoscale level are expected to lead to greater knowledge of the relationship between macropore structure and water flows, as well as the conditions under which flow in the macropores takes place.

(c) The macroscopic level is represented by field experiments at the plot scale in undisturbed soils. Hypotheses derived from experiments at the first two levels should be applied and tested at the plot scale. In the future it will be necessary to extrapolate from the plot scale to larger soil units, with the aim of assessing the hydrological importance of macropores at the catchment scale.

4. CONSIDERATIONS IN MODELLING WATER FLOW THROUGH MACROPORES

From the foregoing discussion it is clear that models of soil water flow based on Darcy's Law may fail to predict the behaviour of some soils in the field where macropores affect the infiltration and redistribution of water. It is recognized that some very specific conditions are required to generate flow in macropores. However, it is suggested that these are not 'special' conditions but may occur widely and although periods of macropore flow may be small in relation to the time scale of interest, the effect on soil water profiles may be expected to be of much greater significance.

Any approach to modelling water flow in field soils cannot reject approaches based on Darcy's Law since such models have been well proven in the past, (see for example HAVERKAMP *et al.*, 1977; ZARADNY, 1978; BRESLER *et al.*, 1979). Thus any model of a combined micropore/macropore system should degrade to a Darcy-type model in the case where there are no macropores at all. This suggests the introduction of a domain concept in modelling a combined system with the micropores as one domain, that conforms to hydraulic principles based on Darcy's Law, and the macropores as a second domain with interactions between the two domains in some physically realistic manner. There is some justification in the use of such a domain concept from experimental measurements on undisturbed cores of field soils. For example data from BRÜLHART (1969) can be interpreted in terms of three flow zones (see figure 4.1): Zone 1 is that of completely saturated soil where, assuming a fixed soil structure, the saturated hydraulic conductivity is the greatest possible conductivity of the soil. Zone 2 can be assumed to be flow in both macropores and micropores. The lower boundary of zone 2 has been placed at a potential of −1 cm (which we have used as an approximate definition of the lower bound of macropores) and suggests a macropore volume in the order of 1% of the sample. In zone 2 the micropores will remain essentially saturated but small changes in the macropore water content causes changes in the apparent hydraulic conductivity of two orders of magnitude. The term 'apparent hydraulic conductivity' is used since if, by definition, water moves under gravity forces alone; the hydraulic gradient in the unsaturated macropores cannot be properly defined. From hereon we shall use the term 'specific discharge' to describe free water flow in the macropores, as it does not imply a definition in conjunction with a hydraulic gradient. In zone 3 the macropores are essentially drained in so far they no longer make a significant contribution to water seepage and the micropores begin to drain as well.

The use of a domain concept to model water flow in a combined micropore/macropore system was suggested at least as long ago as MUSKAT (1946) in a study of saturated flow in fissured limestone. Further suggestions have been made by DE QUERVAIN (1972) and WAKAHAMA (1974) in relation to water flow through snow, and EDWARDS *et al.* (1978) with specific reference to flow through soil. In addition SCOTTER (1978) has produced a model for solute flows in a saturated macropore system. The model of EDWARDS *et al.* comes closest to the approach adopted in the present work. They restricted

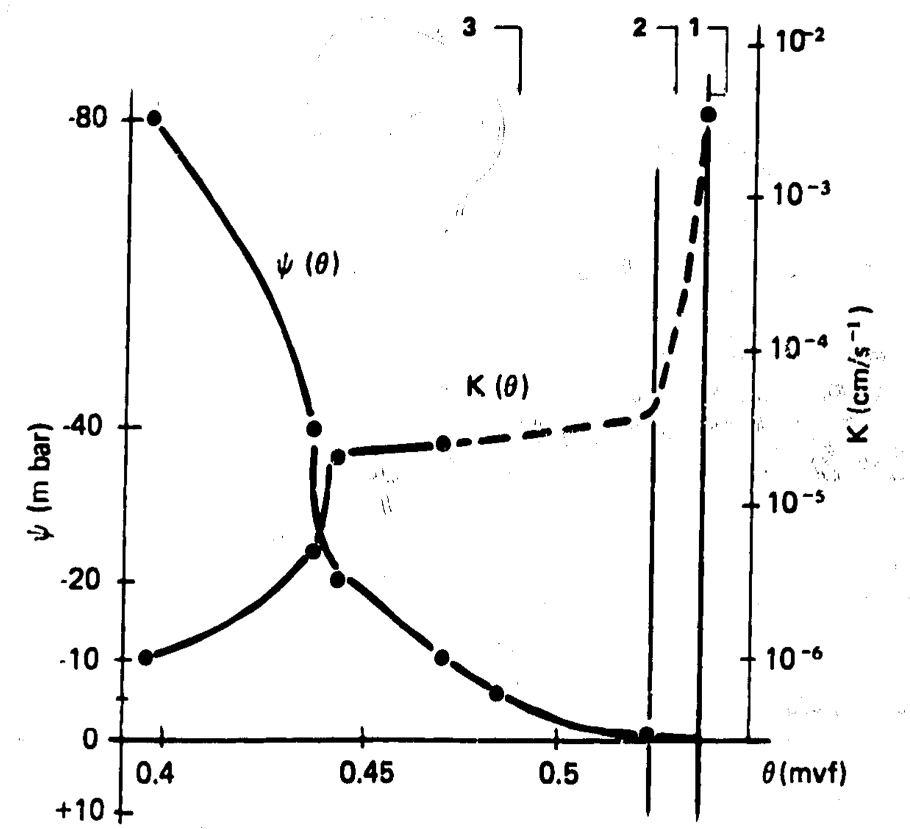

Figure 4.1 The characteristic curves and zones of flow in an undisturbed core of a field soil

(1) saturated zone
(2) macropore zone
(3) unsaturated zone

● data after Brulhart (1969)
--- function assumed

their analysis to flow in a cylindrical column around a single pore. Flow in the micropores is treated in two dimensions (vertically and radially away from the macropore) using a solution of the Richards' equation. Flow in the macropores is modelled using a simple accounting procedure for given depth increments with input at the surface and infiltration losses to the micropores at the side of the macropore. Velocities of flow in the macropore are assumed to be nonlimiting and surface runoff is assumed to occur only after the macropores are full. Numerical experiments with this model clearly demonstrate the importance of macropores on infiltration and runoff, at least under conditions of heavy rainfall.

However, a more general model is required to predict the influence of macropores at larger scales than single pores. The results of such a model should integrate the effects of a realistic distribution of pores at given levels in the profile of a field soil. Ideally such a model should also take account of the effects of 'necks' in the macropore system (BOUMA *et al.*, 1977) and of air pressure effects on flow rates (LINDEN and DIXON, 1976). A first attempt at such a model is described in the following section.

5. A MODEL OF BULK FLOW IN A COMBINED MICROPORE/MACROPORE SYSTEM

In keeping with the discussion of the previous section the micropores and macropores will be treated as separate flow domains, allowing exchange of water to take place between them. Both must satisfy a continuity equation. Thus in one dimension:

$$\frac{\partial \theta_{mi}}{\partial t} = -\frac{\partial}{\partial z}(Q_{mi}) + S \qquad 5.1$$

$$\frac{\partial \theta_{ma}}{\partial t} = -\frac{\partial}{\partial z}(Q_{ma}) - S \qquad 5.2$$

where θ_{mi} and θ_{ma} are volumetric water contents in the micropores and macropores respectively, Q_{mi} and Q_{ma} are flow rates (specific discharge) in the two domains, S is an interaction term treated as a source/sink term, z is the flow direction and t is time. Equations 5.1 and 5.2 imply that both domains are being treated in a spatially averaged sense. This is the usual formulation of soil water flow models, based on a Darcy-type flow law which will be adopted here for the micropore phase. In this case the representative volume over which a suitable average can be taken should be relatively small for broadly homogeneous soils. This volume will usually be within the scale of the sampling range of a neutron probe, for instance. For the macropore system, because of the greater spacing between pores, a suitable representative volume will be much larger (perhaps 1 to 10 m^2 in area with a depth corresponding to the distribution of the macropores in the soil profile) to obtain a spatial sample that is statistically characteristic of the soil. It will therefore be much more difficult to measure the characteristics and nature of the response of the macropore system than of the micropore system, particularly as the macropores will generally constitute only a small volume of the soil (perhaps in the range of 1 to 5%) and also may not be obviously linked to typical soil horizons. If a model of the combined system is to be practically useful, its predictions should be applicable at scales larger than the representative volume for the macropore system. This is some justification for the averaging approach adopted here.

The flow model for the micropore system is completed by a Darcy-type flow law:

$$Q_{mi} = -K(\Theta_{mi})\frac{\partial \phi}{\partial z} \qquad 5.3$$

where K is the hydraulic conductivity function and ϕ is the total hydraulic potential $\phi = \psi + z$ where ψ is the capillary tension in the micropores. Substituting into equation 5.1

$$\frac{\partial \Theta_{mi}}{\partial t} = \frac{\partial}{\partial z}\left[K(\Theta_{mi})\left\{\frac{\partial \psi}{\partial z} + 1\right\}\right] + S$$

or

$$\frac{\partial \Theta_{mi}}{\partial t} = \frac{\partial}{\partial z}\left[D(\Theta_{mi})\frac{\partial \Theta_{mi}}{\partial z}\right] + \frac{\partial K(\theta_{mi})}{\partial z} + S \qquad 5.4$$

where D is the diffusivity function. Equation 5.4 is solved by an implicit finite difference scheme of the form

$$
\begin{aligned}
(\Theta_j^{t+1} - \Theta_j^t/\Delta t = \Big[& \alpha(D_{j+\frac{1}{2}}^{t+1}\{\Theta_{j+1}^{t+1} - \Theta_j^{t+1}\}/\Delta z - D_{j-\frac{1}{2}}^{t+1}\{\Theta_j^{t+1} - \Theta_{j-1}^{t+1}\}/\Delta z) \\
& + (1-\alpha)(D_{j+\frac{1}{2}}^{t}\{\Theta_{j+1}^{t} - \Theta_j^{t}\}/\Delta z - D_{j-\frac{1}{2}}^{t}\{\Theta_j^{t} - \Theta_{j-1}^{t}\}/\Delta z)\Big]/\Delta z \\
& + \alpha(\{K_{j+\frac{1}{2}}^{t+1} - K_{j-\frac{1}{2}}^{t+1}\}/\Delta z + S^{t+1}) \qquad 5.5 \\
& + (1-\alpha)(\{K_{j+\frac{1}{2}}^{t} - K_{j-\frac{1}{2}}^{t}\}/\Delta z + S^{t})
\end{aligned}
$$

where Θ refers to Θ_{mi}, j is a node index, t is a time index, α is a time weighing parameter, and Δz and Δt are depth and time increments, respectively. Single valued (non-hysteretic) relationships between θ_{mi}, ψ_{mi} and $K(\Theta_{mi})$ are assumed in the model.

In the macropore system, it is assumed that unless all the pores are saturated the ambient pressures are atmospheric throughout. For this one dimensional representation, below the level of saturation in the macropore it is assumed that pressures are hydrostatic above the lower limit of the macropores. This allows a simple separation of the macropore flow model into unsaturated and saturated zones but implies that all the macropores are sufficiently interconnected to minimise the effects of restrictions on flow. No attempt is made in this first model formulation to specifically take account of the effects of 'necking' or air pressure effects noted above.

There is little information available on the nature of flows in macropores, but it seems sensible to assume that in general

$$Q_{ma} = Q_{ma}(\Theta_{ma}) \qquad (5.6)$$

The complexities of the macropore network are such that it is impossible to specify the relationship on theoretical grounds alone. The approach taken here is to make some assumptions about the nature of flow in the macropores in order to specify the f o r m of equation 5.6. Given that $Q_{ma}(\Theta_{ma} = 0) = 0$ a measurement of fully saturated flow through the macropore system can than be used to provide a scaling factor to complete the specification, subject to those assumptions remaining reasonable in the case of fully saturated flow. It is expected, however, that the precise form of equation 5.6 will not be essential to building a useful model of the joint macropore/micropore system since, assuming that flow rates in the macropores will be maintained only for short periods of time, it will be more important to model the supply of water to the macropore system correctly.

As an initial approach the following is suggested. It is assumed first that an approximate pore size distribution can be specified for the macropores. This distribution may be made up of a collection of N circular pores each of average radius R_i, $i = 1 \ldots N$ and M longitudinal cracks of average width D_j, $j = 1 \ldots M$ and length L_j, $j = 1 \ldots M$. In general there may be N class sizes of pores, each containing n_i, $i = 1 \ldots N$ individual pores, and similarly M class sizes of cracks. It is necessary to make a further assumption about the way in which the distribution of pores and cracks wets and dries in order to relate average flow through the macropores to average water content, θ_{ma}. Several different assumptions could be made but the simplest is to assume that as the macropores wet and dry there is always an equal depth of water attached to the walls of the pores over all the size classes, until during wetting, the smaller pores become and stay fully saturated. The process continues until it is the largest pores that saturate last.

For flow in cylindrical pores and straight sided cracks the case of fully saturated laminar flow has been treated by CHILDS (1969). Maintaining the assumption that flow remains laminar in the macropores we can follow a similar development for partially saturated flow. For the case of a cylindrical pore of radius R we consider flow in a vertical annular cylinder with a thickness of (R-r) (figure 5.1).

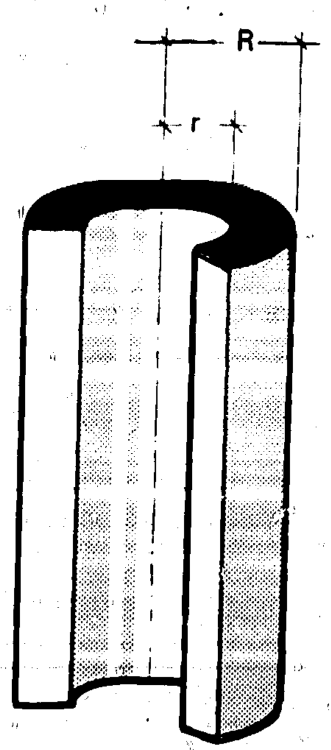

Figure 5.1 Definition diagram for film flow in a cylindrical pore
R radius of the pore
r radius of the air filled cylinder

Then for steady flow, the downwards force due to the weight of water must be balanced by a retarding force F or

$$g\rho\pi\,\ell(R^2 - r^2) = -F \qquad 5.7$$

where in general F is expected to be a function of the depth and velocity of flow, the surface area of the pore walls and the skin and form resistance of the pore. For laminar flow, we can assume that F results from viscosity effects alone such that in an inner annular cylinder of thickness (y - r)

$$-g\rho\pi\,\ell(y^2 - r^2) = 2\pi y\,\ell\mu\,\frac{dv}{dy} \qquad 5.8$$

where v is the velocity of flow at radius y, and μ is the dynamic viscosity of the water. Thus

$$\frac{dv}{dy} = -\frac{g\rho(y^2 - r^2)}{2\mu y}$$

$$= -C(y - \frac{r^2}{y}) \qquad 5.9$$

where $C = \frac{g\rho}{2\mu}$

Integrating, noting that v = O at y = R

$$v = \int_R^r -C(y - \frac{r^2}{y})\,dy = \frac{C}{2}(R^2 - y^2) + C\,r^2(\ln y - \ln R) \qquad 5.10$$

To obtain the discharge of the flow in the pore, q_p, we consider the incremental discharge due to a small part of the flow at radius y such that

$$d(q_p) = 2\pi yv\,dy \qquad 5.11$$

and $$q_p = \int_{y=r}^{y=R} 2\pi y(\frac{C}{2}(R^2 - y^2) + C\,r^2(\ln y - \ln R)\,dy$$

$$= \pi C(\frac{R^4}{4} + \frac{3}{4}r^4 - r^2R^2 + r^4\ln R - r^4\ln r) \qquad 5.12$$

which reduces to Child's formula for flow in a saturated tube for r = O.

For flow in a partially saturated crack of width D with a symmetrical depth of flow, d, on each side (figure 5.2) a balance of forces per unit width at a depth y in a laminar flow yields, for one side only,

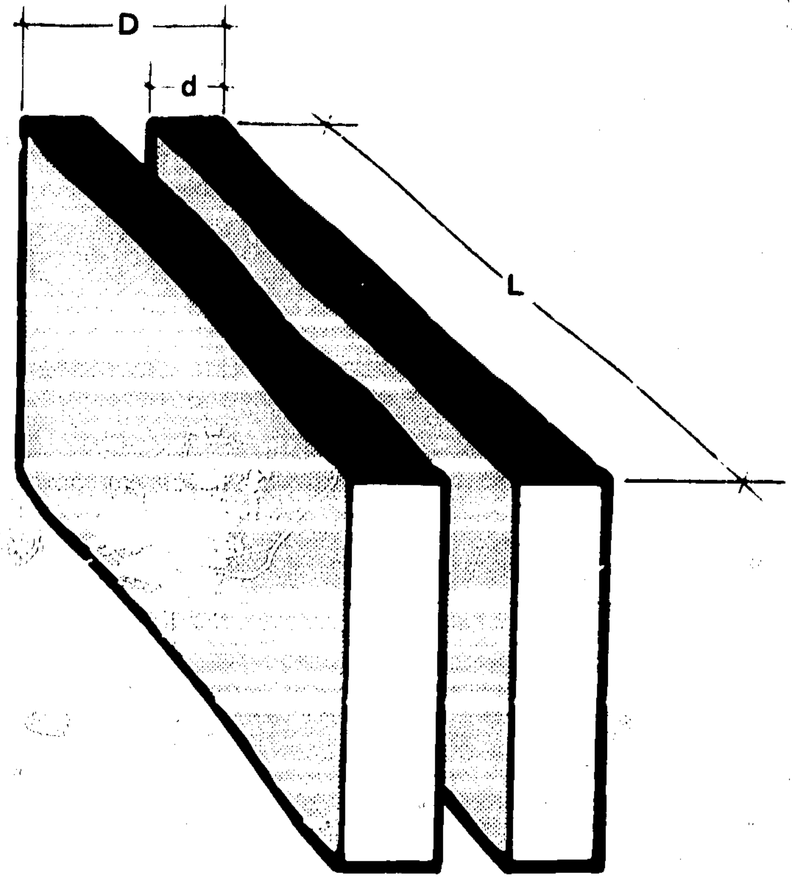

Fig. 5.2 Definition diagram of film flow in a planar crack
D width of the crack
L length of the crack
d thickness of the water film

$$g\rho\ell(d-y) = \ell\mu \frac{dv}{dy} \qquad 5.13$$

or

$$\frac{dv}{dy} = \frac{g\rho(d-y)}{\mu}$$

Integrating and noting that $v = 0$ at $y = 0$ gives

$$v = \int_0^d \frac{g\rho(d-y)}{\mu} dy = \frac{g\rho}{\mu} d\,y - \frac{g\rho y^2}{2\mu} \qquad 5.14$$

$$= C(2\,d\,y - y^2)$$

Then considering the incremental discharge in the crack of length L due to flow at a depth y

$$d(q_c) = 2\,L\,v\,dy \qquad 5.15$$

and

$$q_c = 2L \int_{y=0}^{y=d} C(2dy - y^2)dy$$

$$= \frac{4}{3}L\,Cd^3 \qquad 5.16$$

which reduces to Child's formula for flow in a fully saturated crack for $d = \frac{D}{2}$.

Given equation 5.12 and 5.16 and assuming that an approximate pore size distribution for the macropores can be specified, a final relationship between q_{ma} and θ_{ma} may be defined as follows.

$$q_{ma} = \frac{1}{A}\left[\sum_{i=1}^{N}\sum_{j=1}^{n_i} q_p + \sum_{i=1}^{M}\sum_{j=1}^{m_i} q_c\right] \qquad 5.17$$

$$\theta_{ma} = \frac{1}{A}\left[\sum_{i=1}^{N}\sum_{j=1}^{n_i} \pi\,(R_j^2 - r_j^2) + \sum_{i=1}^{M}\sum_{j=1}^{m_i} 2\,L_j d_j\right] \qquad 5.18$$

remembering that once pores of a given size are saturated (ie $r_j = 0$, or $d_j = D_j/2$) they make no further contribution to q_{ma} or θ_{ma}.

As an example we shall consider a hypothetical porous medium with microporosity of 0.49 and a macroporosity of 0.01. We assume that the micropores are just saturated so that the addition of further water will cause the water in the macropores to flow. For the present, interaction between the two systems is ignored and a unit hydraulic gradient is assumed in the micropores so that a steady flow rate is maintained equal to the saturated hydraulic conductivity, say 0.01 cm/min. Figure 5.3 compares discharge for various arrangements of macropores over an area of 1 m^2 and different values of Θ_{ma}, from equations 5.12 and 5.16 alone with no scaling as suggested above. On the basis of these results a further generalisation has been introduced into the model by approximating the curves of Figure 5.3 by a function of the form

$$q = a(\Theta_{ma})^b$$

$$Q_{ma} = K_{ma} \cdot q/q_{sat}$$

where K_{ma} is the measured hydraulic conductivity at saturation and q_{sat} is the flow at full saturation, predicted from 5.17. Thus

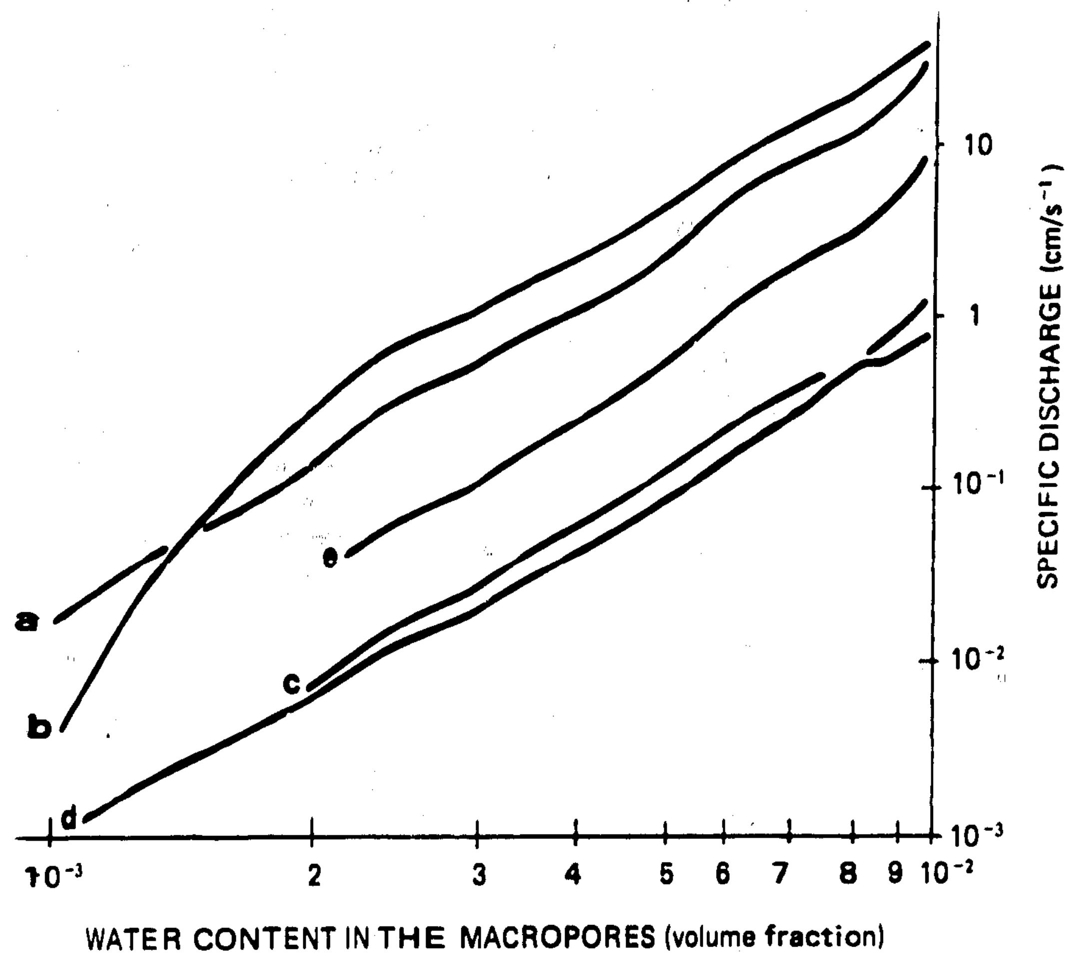

Fig. 5.3. Specific discharge of different types of macropores versus their water content (volume of the macropores: 0.01 fraction of volume)
- (a) cylindrical pores, diameter 1.0 cm
- (b) cracks 0.5 x 5 cm
- (c) cylindrical pores, diameter 0.2 cm
- (d) cracks 0.1 x 1 cm
- (e) mixture of pores (diameters from 0.2 to 1.0 cm) and cracks (lengths from 1.0 to 5.0 cm, widths from 0.1 to 0.5 cm)

$$Q_{ma} = K_{ma} \cdot a' (\Theta_{ma})^{b} \qquad 5.19$$

where $a' = a/q_{sat}$. Introducing this into Equation 5.2 leads to

$$\frac{\partial\theta_{ma}}{\partial t} = - \frac{\partial}{\partial z} \left[K_{ma} \cdot a' \cdot (\Theta_{ma})^{b}\right] - S \qquad 5.20$$

or

$$\frac{\partial Q_{ma}}{\partial t} = -c\left[\frac{\partial Q_{ma}}{\partial z} + S\right] \qquad 5.21$$

where $c = \frac{\partial Q_{ma}}{\partial \Theta_{ma}}$. This has the form of a kinematic wave equation (LIGHTHILL and WHITHAM, 1955) with $c = b\left[a'K_{ma}\right]^{1/b} \theta_{ma}^{(1-1/b)}$ representing the non-linear kinematic wave velocity. Equation 5.21 may be solved by an implicit finite difference scheme of the form

$$(Q_j^{t+1} - Q_j^{t})/\Delta t = - \alpha(\beta . c_j^{t+1} - (1-\beta)c_{j-1}^{t+1})(\{Q_j^{t+1} - Q_{j-1}^{t+1}\}/\Delta z + S^{t+1})$$
$$- (1-\alpha)(\beta c_j^{t} - (1-\beta)c_{j-1}^{t})(\{Q_j^{t} - Q_{j-1}^{t}\}/\Delta z + S^{t}) \qquad 5.22$$

where α is a time weighting parameter and β is a depth weighting parameter. Because of the shock front nature of the infiltration of a wetting front into the macropores, the time step to maintain reasonable accuracy of the solution will necessarily be restricted under such conditions. It has been found that satisfactory solutions are obtained when the time step is limited by the criterion

$$\Delta t \simeq \Delta z/c_{max} \qquad 5.23$$

corresponding to the fastest rate of wave propagation across one depth increment.

Equation 5.22 is used for every node in the unsaturated part of the macropores domain. In the saturated part a simple accounting procedure is used at each time step to calculate the rise or the fall of the surface of saturation depending on the rate of inflow from the unsaturated part above, the rate of loss to the micropore system and the current macropore water content above the surface of saturation.

To complete the combined model it is necessary to consider the S term of equations 5.1 and 5.2. This must also be treated as a macroscopic average and it is convenient to consider that the S term is controlled within the micropore system with the form of a Darcy-type flow in a horizontal direction, such that

$$S = - K(\theta_{mi}) \frac{\Delta\psi}{\Delta x} \qquad 5.24$$

where $\Delta\psi/\Delta x$ is a representative hydraulic gradient and Δx is a characteristic length, which will be related to the average spacing between macropores. The hydraulic gradient may be taken as

$$0 \quad , \quad \theta_{ma} = 0,\ \theta_{mi} < \theta_{mi,sat} \qquad 5.25a$$

$$(\psi(\theta_{mi})-0)/\Delta x \quad , \quad 0 < \theta_{ma} < \theta_{ma,sat} \qquad 5.25b$$

$$(\psi(\theta_{mi})-d_w)/\Delta x \quad , \quad \theta_{ma} = \theta_{ma,sat} \qquad 5.25c$$

where d_w is depth below a water table in the macropores.

To determine the characteristic length, Δx, the following procedure was adopted. It was assumed that the averaged θ_{mi} values predicted by the model would be characteristic of finite volumes within the micropores equidistant from neighbouring macropores. For the specified macropore distribution at a given level let there be N^* cylindrical pores and a total length, L^*, of cracks. In addition let the area at that level be subdivided into an area W_p nearer a pore rather than a crack and W_c nearer to a crack rather than a pore. Thus on average the area surrounding a pore will be

$$\frac{W_p}{N^*} = \pi(\Delta x_p)^2 \qquad 5.26$$

where Δx_p is a length characteristic of the spacing between pores.

For the area W_c we assume that the length of cracks L^* is arranged in patterns of regular hexagons such that the average area of a hexagon is

$$\frac{W_c}{L^*} = \frac{2}{\sqrt{3}} \quad (\Delta x_c) \qquad 5.27$$

where Δx_c is a length characteristic of the spacing between cracks. The final characteristic length is given by the weighted average.

$$\Delta x = (W_p \Delta x_p + W_c \Delta x_c)/A$$

where A is the total area ($A = W_p + W_c$).

An example of the use of the model is shown in figure 5.4. This simulation used soil characteristics for the micropore domain similar to those of the Yolo Light Clay published by PHILIP (1957). A run of the micropore solution alone for the case of continuous infiltration into a dry soil gave results that were a satisfactory approximation to results published by PHILIP and KNIGHT (1974). However the combined model assumed a macropore system extending to a depth of 1m with constant characteristics. Figure 5.4a shows the short, intense rainfall rate applied at the upper boundary in the simulation together with the calculated cumulative infiltration and rainfall excess for model runs with and without macropores, and the rise and fall of the water table in the macropores for the former case. Figure 5.4b shows the evolution of the moisture profiles in the macropores and micropores. Figure 5.4c shows the results of the run without macropores. The effect of the macropores on infiltration into the profile as a whole is well demonstrated by comparison of figures 5.4b and c. The micropore moisture profiles of 5.4b demonstrate the effects of positive pressures beneath the water table in the macropores on losses to the unsaturated micropores.

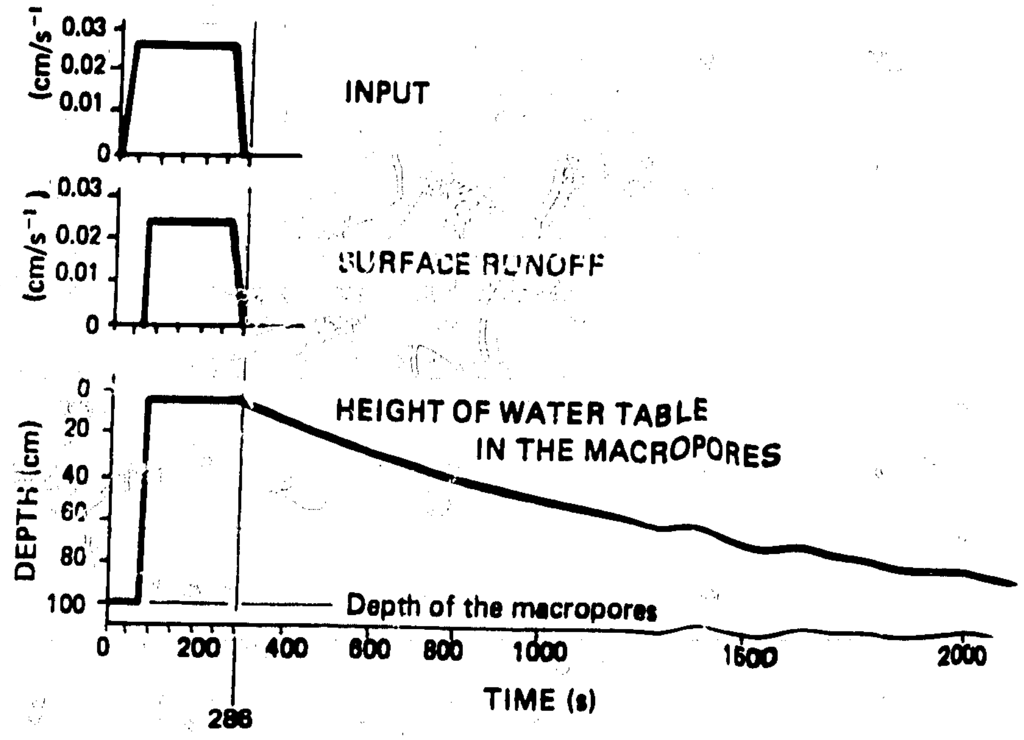

Figure 5.4a Input, surface runoff and height of the water table in the macropores used and produced by a combined micropore/macropore model

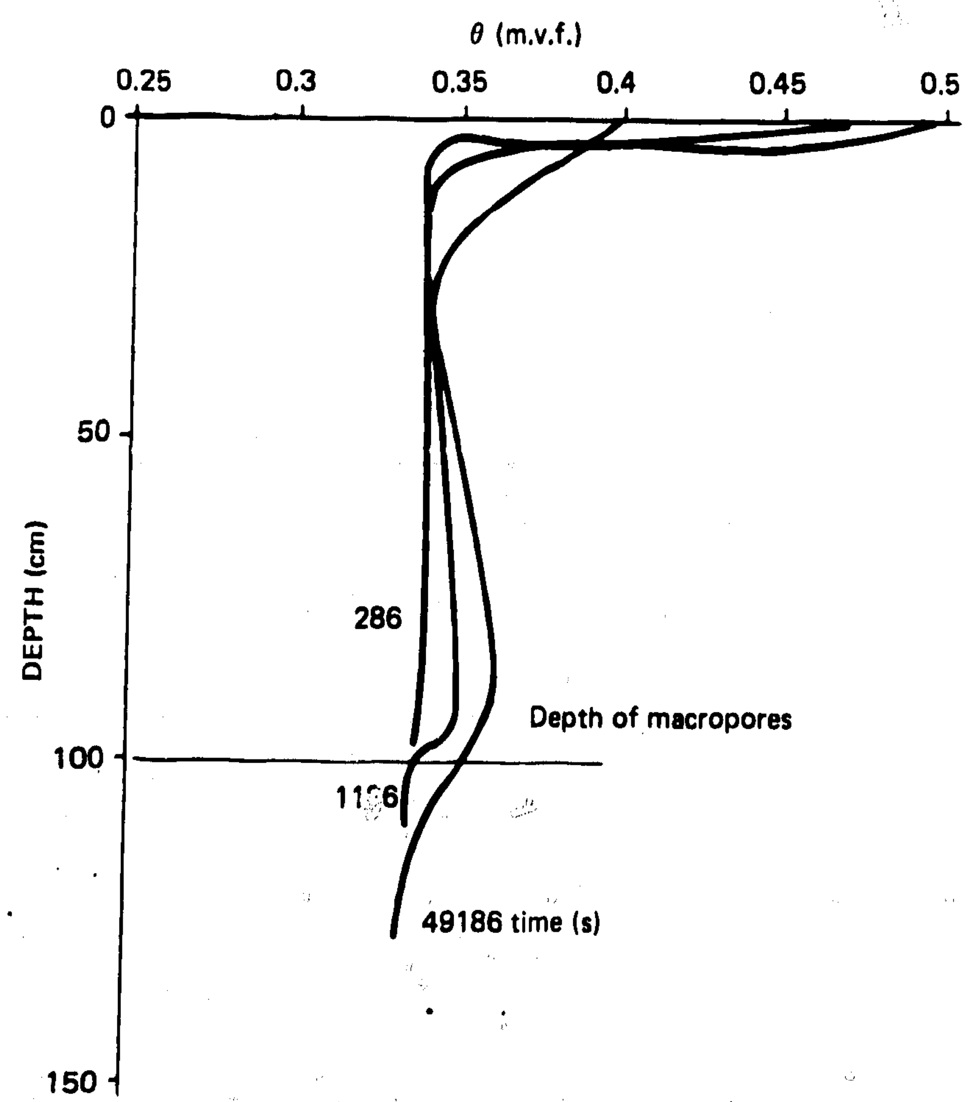

Figure 5.4 b Soil moisture profiles at different times during and after infiltration into a soil with macropores (0.01 fractions of volume)
Soil: Yolo light clay (data after Philip, 1957)

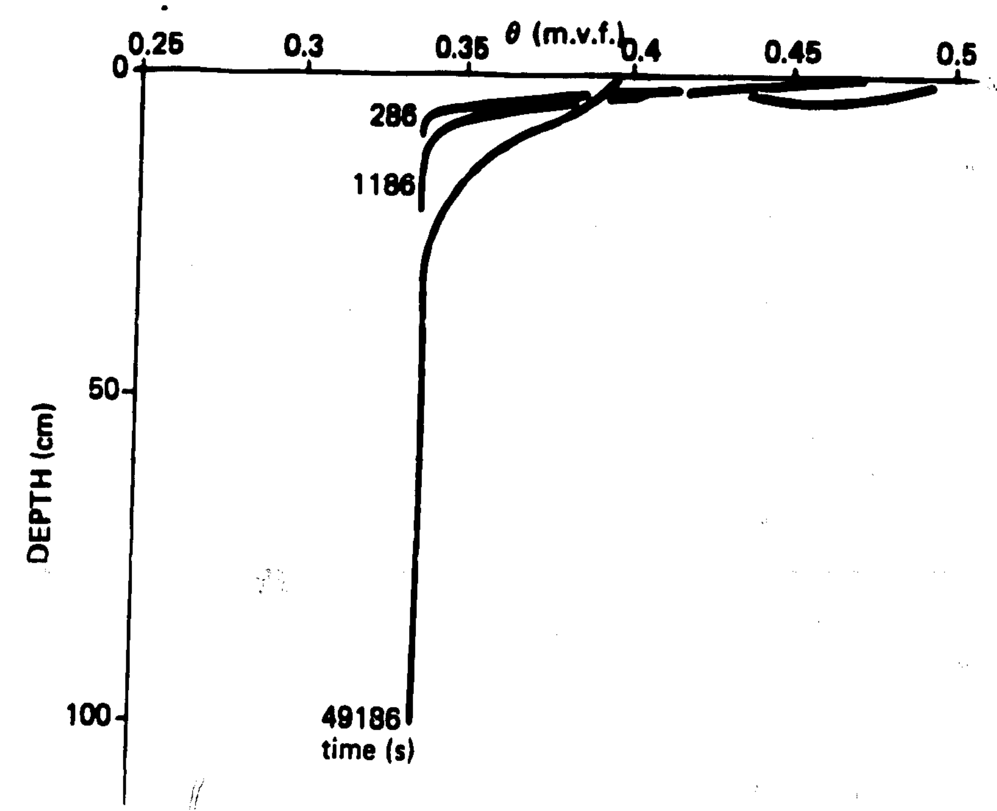

Figure 5.4 c Soil moisture profiles at different times during and after infiltration into a soil without macropores
(Same soil as in Figure 5.4 b)

6. EXPERIMENTAL APPROACH TO THE WATER FLOW IN THE MACROPORES OF A LARGE SOIL SAMPLE.

6.1 Introduction

The model of bulk water flow in a combined macropore-micropore system outlined above requires the following information:

a. Hydraulic conductivity K as a function of the soil water suction ψ or as a function of the volumetric water content θ.

b. Soil water suction ψ as a function of the volumetric water content θ.

c. The parameters K_{ma}, a and b in equation 5.19 that describe the specific discharge in the macropores Q_{ma} as a function of the volumetric water content in the macropores θ_{ma}.

The soil water properties of an unsaturated sample can be determined using an outflow method, as described by BENECKE *et al* (1976) and modified by GERMANN *et al* (1978). Discharge through the sample at full saturation can either be measured by a throughflow experiment, or at the start of drainage using the same outflow method. The total volume of the macropore system can also be determined during a drainage experiment, starting at full saturation and using tensiometers to indicate the threshold between macropores and micropores, as described by BOUMA *et al* (1977 a,b). The advantage of using an outflow method lies not only in getting all the information required with one experiment on one sample but also in avoiding internal erosion and disturbance of the sample. AUBERTIN (1971) for example describes a dramatic reduction in apparent hydraulic conductivity during a throughflow experiment.

6.2 Description of the method

A large undisturbed soil sample, sealed with polyester resin is sealed onto a porous ceramic plate, through which the sample can be saturated or drained by altering the level of the outflow (figure 6.2). Tensiometers connected to water filled manometers, are used to indicate the soil water potential at the upper and lower boundaries of the sample. The potentials P_u and P_ℓ of the upper and lower tensiometers, and the total amount of drainage Q are recorded over time. During the drainage experiment the level of the outflow should be lowered in such a way that a hydraulic gradient of about unity is maintained within the sample. The hydraulic properties of the soil micropore system can be calculated for relatively short time steps, Δt_i, from equations 6.1 to 6.3 (see also figure 6.1).

$$\bar{K}_i = \frac{\Delta Q_i \cdot \Delta z}{A \cdot \Delta t_i \, \Delta\phi_i} \qquad 6.1$$

where $\bar{K}_i$ is the average hydraulic conductivity (cm.s^{-1}) during the period

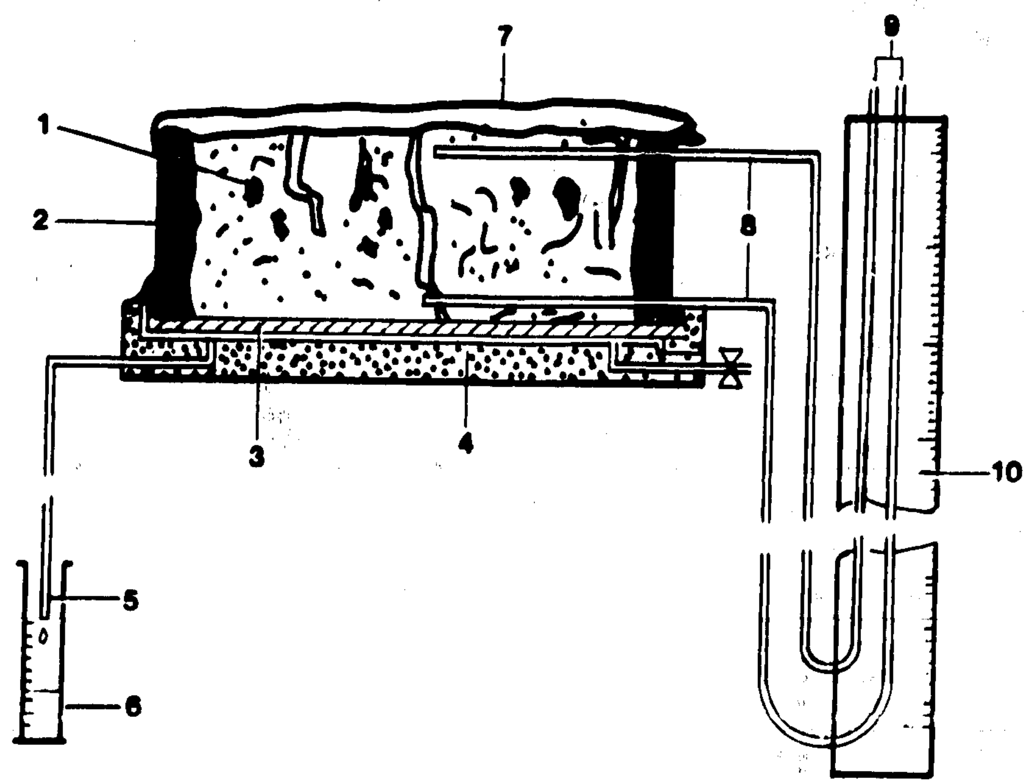

Figure 6.1 Design of the outflow method

(1) large undisturbed soil sample
(2) wall of polyester resin and glass fibre
(3) porous ceramic plate
(4) support ring in aluminium
(5) outflow
(6) burette
level can be varied (stepwise or continuously)
(7) polyethylene sheet (preventing evaporation)
(8) upper and lower tensiometers
(9) manometer tubes (glass)
(10) manometer scale, adjusted to the bottom of the sample z_u and z_l level of the upper and the lower tensiometer, respectively

$\Delta t_i = t_{i+1} - t_i$ (s) (t_{i+1} and t_i are times of readings)

$\Delta Q_i = Q_{i+1} - Q_i$ (cm^3) and (Q_{i+1} and Q_i are amounts of total outflow from the beginning of the drainage experiment until t_{i+1} and t_i respectively).

A cross sectional area of the bottom of the sample (cm^2)

$\Delta z = z_u - z_\ell$ (cm) (z_u and z_ℓ height of the upper and lower tensiometer above a datum, for example the zero point of the scale).

$\Delta\phi_i = (P_{u,i+1} + P_{yi} - P_{\ell,i+1} - P_{\ell,i})/2$ (cm) $P_{u,i}$ and $P_{\ell,i}$ are the potentials at the upper and lower tensiometer respectively at time t_i.

$$\bar{\psi}_i = (\psi_{u,i+1} + \psi_{u,i} + \psi_{\ell,i+1} + \psi_{\ell,i})/4 \qquad 6.2$$

where $\bar{\psi}_i$ average suction within the sample during the period (cm)
$\psi_{\ell,i} = P_{\ell,i} - z_\ell$; $\psi_{u,i} = P_{u,i} - z_u$

$$\bar{\theta}_i = \theta_E + \left[Q_{tot} - (Q_{i+1} + Q_i)/2 \right]/V \qquad 6.3$$

where $\bar{\theta}_i$ average water content in the sample during the period Δ_{ti} ($cm^3 \cdot cm^{-3}$)

θ_E water content of the sample at the end of the drainage experiment (gravimentrical determination by drying the sample at 105°C)

Q_{tot} total amount of outflow at the end of the drainage experiment (cm^3)

V volume of the sample (cm^3)

$\bar{K}_i$, $\bar{\psi}_i$ and $\bar{\theta}_i$ are corresponding points on the soil water property curves $K(\psi)$, $\psi(\theta)$ and $K(\theta)$

6.3 The experimental procedure

(a) Preparation of a large undisturbed soil sample

From the horizon in question a lump of soil is carefully dug out by taking away the surrounding soil. For the present apparatus the diameter should not exceed 26 cm and the length should be at least 20 cm to have enough soil volume for preparing a final length of about 10 to 15 cm. The sides of the sample are first sealed with polyester resin B and then with two layers of glass fibre and polyester resin A. When this covering is hard, the sample is dug out and transported to the laboratory.

(b) Description of the outflow apparatus

According to figure 6.1 the equipment consists of a highflow porous ceramic plate (air entry value 1.5 bar, hydraulic conductivity 5.10^{-6} $cm.s^{-1}$, diameter 27.5 cm, thickness 0.9 cm) and its support in aluminium; a pair of tensiometers (length of the cups: 8.4 cm, diameter 0.6 cm, air entry value 1 bar, hydraulic conductivity 10^{-6} $cm.s^{-1}$), which are connected with a water column manometer; and an outflow tube which conducts water draining from the sample to a burette to record its amount. The level of the outflow tube and the burette can be moved vertically either stepwise or continuously. Covering the sample with a plastic sheet avoids evaporation.

(c) Mounting the sample on the apparatus

The bottom of the sample should be flat. Care has to be taken not to block the macropores when cutting the sample. Fine sand gives a good contact between the sample and the ceramic plate, which has previously been fully saturated with deaired water. After putting the sample on the ceramic plate, the polyester cover is sealed with silicon rubber against the support ring in an air-tight manner. The tensiometers are installed and connected with the manometer The sample is slowly saturated by lifting the level of the outflow tube. The tensiometers indicate saturation when the manometer level reaches the level of the outflow tube (capillary rise of the water in the manometer tubes has to be taken into account).

(d) Drainage procedure

Two types of run are carried out at present:

- Saturation to the top of the sample and drainage to an equilibrium suection profile above an outflow level at the bottom of the sample.

- After saturating the sample again to its top, the outflow tube is lowered to a certain depth below the bottom of the sample.

Care has to be taken to maintain a hydraulic gradient of about unity during the different experiments and changes over time should be fairly slow. Fulfilling these conditions will yield an approximately linear distribution of suction and water content in the vertical direction, which is assumed in computations based on equations 6.1 to 6.3. The single free variable in these experiments is the vertical position of the outflow tube over time. The suction profile depends on the hydraulic conductivity within the sample and its development over time, and the suction at the bottom of the sample. This suction depends on the hydraulic conductivity of the porous plate, which remains constant during the experiments, and the level of the outflow. Taking these factors into consideration allows the ideal position of the outflow to be estimated.

6.4 The results of two drainage experiments

From a dark brown A-horizon of a brown earth soil a sample was taken in the manner described above. The height of the sample was 12.8 cm; the cross-sectional area of its bottom was 565 cm^2, and the tensiometer cups were installed at 1.9 and 10.7 cm above the bottom. According to figure 6.2, the outflow tube in run 1 was at the bottom of the sample (level 0). This provoked a decrease of the potential within the sample and the outflow stopped at about 0.01 moisture volume fraction (mvf). This amount is taken as the total volume of the macropores in this sample. The average suction at this point is therefore at about -6.4 cm in this case and not as proposed in Section 2.1 at -1.0 cm. Run 2 started at the same conditions of full saturation as run 1, but the outflow was put at -66 cm. With this run the transition from saturated to unsaturated conditions was clearly demonstrated. Between time 0 and time t_1 the sample remained fully saturated: for $t_1 \leq t \leq t_2$ both saturated and

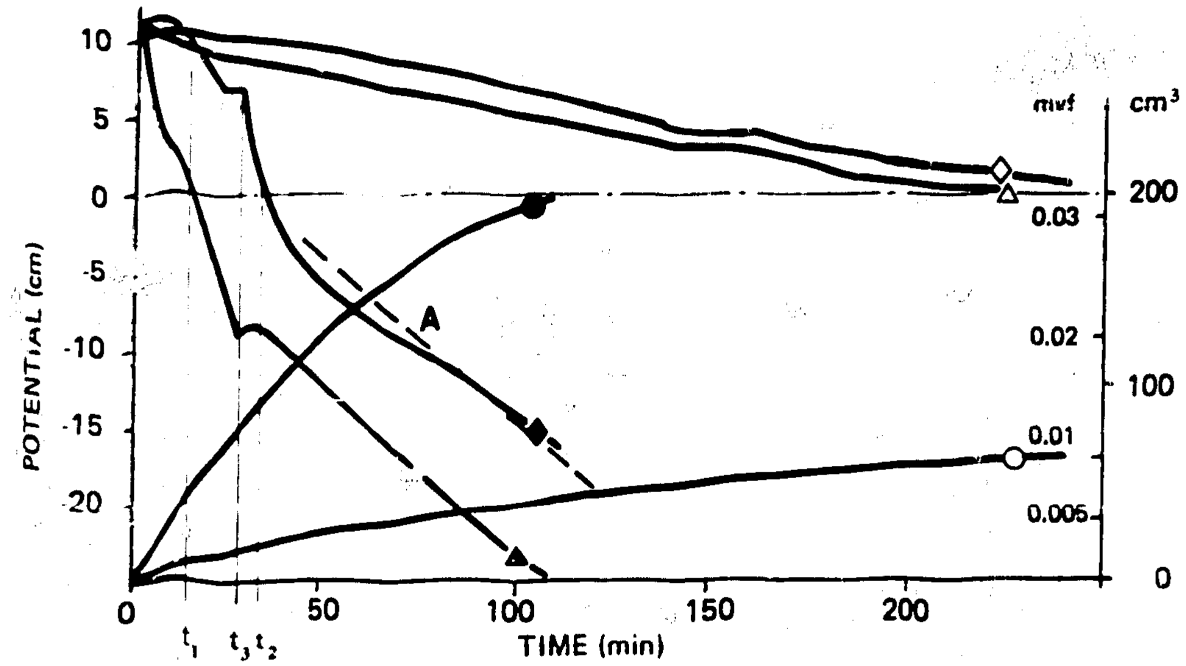

Figure 6.2 Potentials and cumulative outflow from two drainage experiments

— — (A): gradient of unit
for the period $t_1 < t < t_2$ saturated and non saturated parts within the soil sample are expected, demonstrated by particular development of the potentials shortly before and after time t_3

◆◇ potentials at the upper tensiometer (cm)
▲△ potentials at the lower tensiometer (cm)
●○ cumulative outflow (cm³ and vf)
◇△○ run (1)
◆▲● run (2)

unsaturated conditions are expected and for $t > t_2$ unsaturated conditions are established. The irregularities at time t_3 indicate that the macropores may empty discontinuously in surges, suggesting that tortuosity and imperfect connectivity are factors that in general should not be neglected. Similar, but much smoother changes are indicated during run 1 between 150 and 200 minutes.

If a linear potential gradient is assumed through the sample, then unsaturated conditions in run 2 may be assumed to start at $t = (t_1 + t_2)/2$. The amount of cumulative outflow at this time in run 2 was about 10% higher than at the end of run 1. Given the response time of the tensiometers (about 1 minute) and the irregularities observed within the sample, this result is quite satisfactory. During the unsaturated part of the experiment, the requirement of maintaining a hydraulic gradient of about 1 was reasonably fulfilled, compared with line A in figure 6.2.

Computing the specific discharge for the periods during which the water is assumed to flow predominantly in the macropores (run 1, run 2 for $t < t_1$)

Figure 6.3 shows the computed specific discharge for the periods during which water is assumed to flow predominantly in the macropores (run 1, run 2 $t > t_1$), the unsaturated hydraulic conductivity $\overline{K}_i$ (calculated from

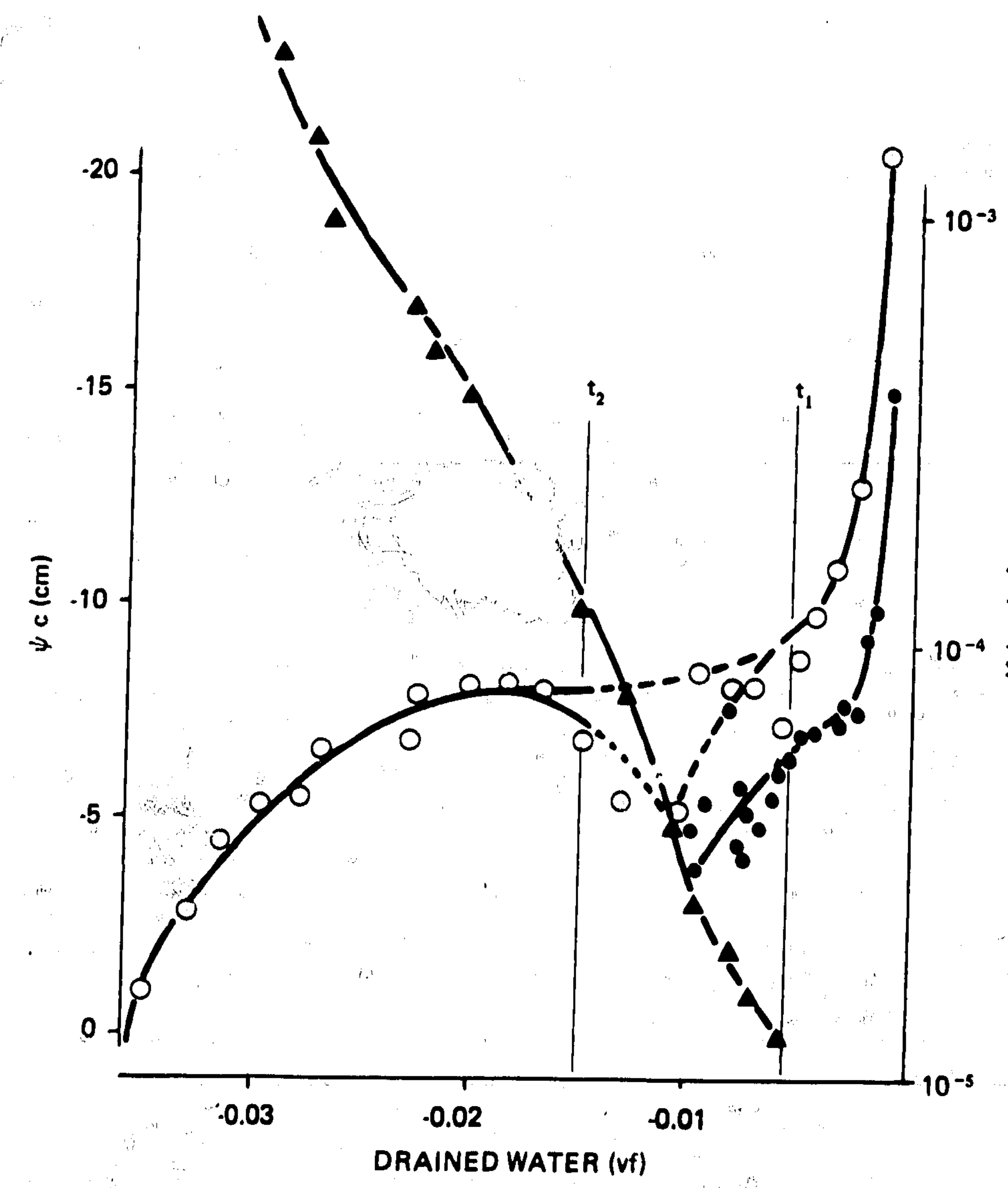

Figure 6.3 Soil water properties, deduced from run 1 and run 2, versus the drained amount of water

● hydraulic conductivity from run 1
○ hydraulic conductivity from run 2
▲ retention curve from run 2
times t_1 and t_2 correspond with those in Figure 6.2

equation 6.1 for run 2 t > t_2) and the corresponding water contents $\overline{\theta}_i$. It is clear that the flow irregularities mentioned above affect the results down to a drained water volume of about 0.02 mvf. Between saturation and this point the hydraulic gradient within the sample is not properly defined.

The accuracy of the results depends on the sensitivity of the measurements relative to the volume of flow. Typical values for the volume of the sample are about 5000 cm^3, for the cross-sectional area of the sample 500 cm^2, and for the height of the sample, 10 cm. The amount of outflow can be easily measured with an accuracy of 0.1 cm^3 and the soil water potential to 0.1 cm. The sensitivity of the tensiometers is quite high due to their design (hydraulic conductivity 10^{-6} cm s^{-1}, surface area 15 cm^2, and inner diameter of water manometer 0.1 cm). A volume change of only 0.008 cm^3 of water is need to indicate a change of potential of 1 cm. If the sample is carefully mounted and aritificial voids are avoided, the average change of water content can be computed to an accuracy of 2.10^{-5} cm^3 cm^{-3}. Assuming that the volume of macropores will be in the range 1 to 5.10^{-2} volume fraction of soil, the accuracy of measurements should be adequate for the problem.

7. CONCLUSIONS

This report has explored several aspects of the hydrological properties of soil macropores. Further research effort in this direction must overcome considerable problems, primarily due to the temporal and spatial variability of macropores, but on the evidence presented would appear to be worthwhile. In particular it would be valuable to link the 'static' properties of macropores with established morphological properties of the soil, and to develop field measurement techniques to characterise the dynamics of both macropores and micropores at a scale characteristic of the bulk macropore system. In time, it may be possible to make a further link between the 'static' and dynamic properties of macropores, at least at the plot scale.

However, the ultimate aim and test of macropore concepts in soil hydrology should be applications at the catchment scale. Thus, it will be necessary to extrapolate from the plot scale to larger soil units. The relative ease of mapping the morphological properties of soils at this scale may allow extension of the scale of study of the dynamics of macropore systems, at least to a first approximation. We make no claim for the universal applicability of macropore concepts, only that an assessment of the role of macropores in governing the volume and velocity of soil water and solute flows is worth making.

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