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SPATIAL VARIABILITY OF PHYSICAL PROPERTIES OF A TROPICAL SOIL: II SOIL WATER RETENTION CURVES AND HYDRAULIC CONDUCTIVITY

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F O R E W O R D

In November 1970 the PANEL ON THE USE OF NUCLEAR TECHNIQUES IN SOIL PHYSICS AND IRRIGATION STUDIES recommended that the Joint FAO/IAEA Division of Atomic Energy in Food and Agriculture initietes a coordinated research programme on "The Use of Isotope and Radiation Techniques in Studies of Soil Water Regimes". This programme starded in 1972, involving twelve countries well distributed around the world. Brazil participates in that programme through the Research Agreement nº 1280 between Joint FAO/IAEA Division and Center for Nuclear Techniques in Agriculture. Progress in this research has been reported in annual reports. The present technical report envolves part of the work carried out in Piracicaba, Brazil, during the period 1975/1976.

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SPATIAL VARIABILITY OF PHYSICAL PROPERTIES OF A TROPICAL SOIL: II SOIL WATER RETENTION CURVES AND HYDRAULIC CONDUCTIVITY*

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1. INTRODUCTION

The objective of this paper is to characterize a field site from the soil physics point of view, to improve the use of soil and water resources under a particular condition and, to develop means for controlling the dynamics of soil-water movement.

Spetial emphasis is given to the variability in space and in time of physical soil properties, in order to make it possible to define representative means which will indeed describe the movement of water and of salt in the entire field.

The field site chosen has an area of aproximately 10 ha and belongs to the Campus of the Escola Superior de Agricultura "Luiz de Queiroz", University of Piracicaba, being located 580 meters above see level at a latitude of 22° 42' 30" S and a longitude of 47° 38' 00" W. Other details about the field site can be found in *REICHARDT et al. (1976)*.

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* Contribution from Center for Nuclear Energy in Agriculture (CENA), Piracicaba, SP., Brazil and "Instituto Agronômico de Campinas" (CIA), Campinas, SP., Brazil.

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Experimental Design and Procedures

In order to determine the variability of some important physical characteristics of the field, three plots of 5 x 5 m were randomly chosen over the 10 ha site at ESALQ Field Station. Mercury tensiometers were placed in each plot at depths of 15, 45, 75, 105, 135 and 165 cm to follow metric soil water potential changes during the internal drainage process in the profile. In the center of each plot, three neutron acess tubes, one meter appart from each other, were installed down to the depth of 2 m. Soil water content measurements were made at the same depths at which tensiometers were placed. The plots were leveled and enclosed for water ponding. During steady state infiltration of water its rate was measured at three different locations in each plot by pushing into the soil 30 cm diameter plastic cilinders and following the decrease of ponding head per unit time. At the end of the infiltration period, plots were covered with plastic and the internal drainage observed through tensiometer and neutron probe readings, for periods of approximately two months. This procedure was followed in 1975 and after staying fallow for approximately six months, the procedure was repeated in 1976.

One meter appart from each plot a pit was made down to the depth of 2 m. Soil samples were taken in triplicate at depths of 15, 45, 75, 105, 135 and 165 cm for measurements of 1) soil bulk density, 2) soil particle density, 3) texture, 4) pH in water, 5) organic carbon, 6) cations (Ca^{2+}, Mg^{2+} and K^{1+}), 7) salt content in a 1:1 soil water extract, and 8) specific surface. This data are presented in *REICHARDT et al. (1976)*.

Soil water characteristics curves were obtained from simultaneous tensiometer and neutron probe readings made at different time intervals during the internal drainage process.

At any given time the amount of water stored S_L (cm) from

the soil surface to any depth L, was obtained integrating θ (cm^3 of water/cm^3 of soil) over the soil profile, according to:

$$S_L(t_i) = \int_0^L \theta\,(z,t_i)\,dz \quad \ldots\ldots\ldots\ldots \quad (1)$$

the function $\theta\,(z,t_i)$ being approximated by a step function, considering a neutron probe reading as an average value over a depth of 30 cm.

Soil water fluxes q_L ($cm \cdot day^{-1}$) during drainage were obtained from storage differences at time t_i and t_j.

Hydraulic conductivity K_L ($cm \cdot day^{-1}$) at depth L was measured as suggested by *DAVIDSON et al. (1969)* and used successfuly by *HILLEL et al. (1972)*, *NIELSEN et al. (1972)* and *NIELSEN et al. (1973)*:

$$K_L(\theta) = \frac{S_L(t_i) - S_L(t_j)}{(t_j - t_i) \cdot \partial\Psi/\partial z} \quad \ldots\ldots\ldots \quad (2)$$

where Ψ (cm) is the total soil water potential head, assumed to be sum of the soil water pressure head h (cm) and the gravitational head z (cm).

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2. RESULTS

Figures 1 to 6 show soil water retention curves for the depths of 15, 45, 75, 105, 135 and 165 cm. *NIELSEN et al. (1973)* have shown that values of soil water content, measured in a particular field of 150 ha, follow a normal probability function. Assuming the same bahavior for the present field, a mean soil water retention curve was calculated to represent the whole field over all depths. This curve is presented in Figure 7.

Table 1 shows data of steady infiltration rates assumed to be equal to the saturated hydraulic conductivity. Again following *NIELSEN et al. (1973)* these values were assumed to be log-normaly distributed. Figure 8 shows the calculated frequency distribution using table 1 data, which follows equation:

$$df(x) = \frac{N}{x\,\sigma\,\sqrt{2\pi}} \exp - \left[(\ln x - m)^2 / 2\sigma^2\right] dx \quad . \; . \; . \; . \; . \; (3)$$

Where f is a frequency, N the total number of events, x the variable studied (hydraulic conductivity values), m the mean of ln x and σ the standard deviation of m. For this distribution, the mode is equal to exp $(m - \sigma^2)$ the median is exp (m) and the mean exp $(m + 0.5\,\sigma^2)$.

Figures 9 to 14 show $K(\theta)$ relationship for the same depth intervals described above. Assuming K log-normaly distributed for each θ, a mean $K(\theta)$ relation was calculated to represent the whole field over all depths. This is shown in Figure 15.

3. DISCUSSION AND CONCLUSIONS

Soil Water Retention Curves

Data show that curves of different plots differ considerably for the depths of 15 and 45 cm. This is due to the greater variability of all physical properties *(REICHARDT et al., 1976)* of this two layers which are more affected by environmental factors. In general the variability of the soil water retention curves is greater for higher potentials (between 0 and - 75 cm of water) and also greater between plots than between curves obtained at different times, (1975 and 1976).

The soil studied in this experiment is very well structured so that its retention properties near saturation are extremely low. This looks evident when analysing total porosity values presented in *REICHARDT et al. (1976)* which are on the average around 0.521 $cm^3.cm^{-3}$, the smallest value found being 0.408 $cm^3.cm^{-3}$. During steady infiltration, soil water contents of the order of 0.368 $cm^3.cm^{-3}$ where measured. This is due to the high structure of the soil which assures high hydraulic conductivity values near saturation and due to the depth of the water table which is located several meters below soil surface. Under field conditions it is virtually impossible to saturate this soil. Analysis of soil atmosphere composition show a very well aerated profile *(VICTORIA et al., 1976)* even during the rainy season.

Although tensiometer readings were close to zero during steady infiltration, it is important to mention that under field conditions it is very difficult to make precise measurements of low negative pressure heads (around - 10 cm of water) due to small difference in the exact depth at which the tensiometer cup is placed, the hight of the mercury level above soil surface which is irregular and the difficulty of reading an exact height of a

mercury column in a manometer.

Data here presented show the importance of the gravitational potential head in the water retention properties of the profile. During the internal drainage procedure it was observed that soil water fluxes became "negligible" at an average soil water content of 0.310 $cm^3.cm^{-3}$. Subtracting this value from the average soil water content during steady infiltration, it can be seen that 0.058 $cm^3.cm^{-3}$ can be lost due to gravity in very short periods of time (less than 2 days).

Figure 7 shows the average soil water retention curve for all depths and plots. The shaded area shows the interval $m \pm \sigma$, which gives an idea of the limitations of its use when studying the dynamics of water over a large area of the soil here analysed.

Hydraulic Conductivity

Figure 8 shows the theoretical distribution of saturated hydraulic conductivity values. As it can be seen the values for the mode, median and mean are 5.3, 10.0 and 13.7 $cm.hr^{-1}$ respectively. These values together with those presented in table 1 confirm the low retention properties of the profile.

For infiltration studies over a larger area of this soil, due to this great variability of hydraulic conductivity data, the question remains of which the mode or the mean represent better the field. Although the mean represents better the whole range of the variability, the mode is the most frequent value. Hence, over a large area, a much greater surface will be characterized by hydraulic conductivity values around the mode.

The $K(\theta)$ relations show again a greater variability for the top layers (15 and 45 cm). Also differences between plots

seem greater than differences between the relations obtained at different times.

The mean K(θ) relation shown in figure 15 shows also the great variability of the profile. Using this mean values of K, errors of ± 100 orders of magnitude can be made. This fact makes it very difficult to use these relations for the estimation of soil water fluxes under field conditions. For the deeper layers (135 and 165 cm) however, much better estimatives can be obtained so that fairly good calculations of drainage fluxes below the root zone can be obtained.

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Table 1 - Saturated hydraulic conductivity in $cm \cdot hr^{-1}$, measured at soil surface.

	Plot I	Plot II	Plot III
	5.5	5.1	30.0
1.975	4.1	5.7	26.5
	5.8	12.9	34.1
	5.3	3.2	8.7
1.976	27.6	6.7	17.2
	18.0	4.5	20.6

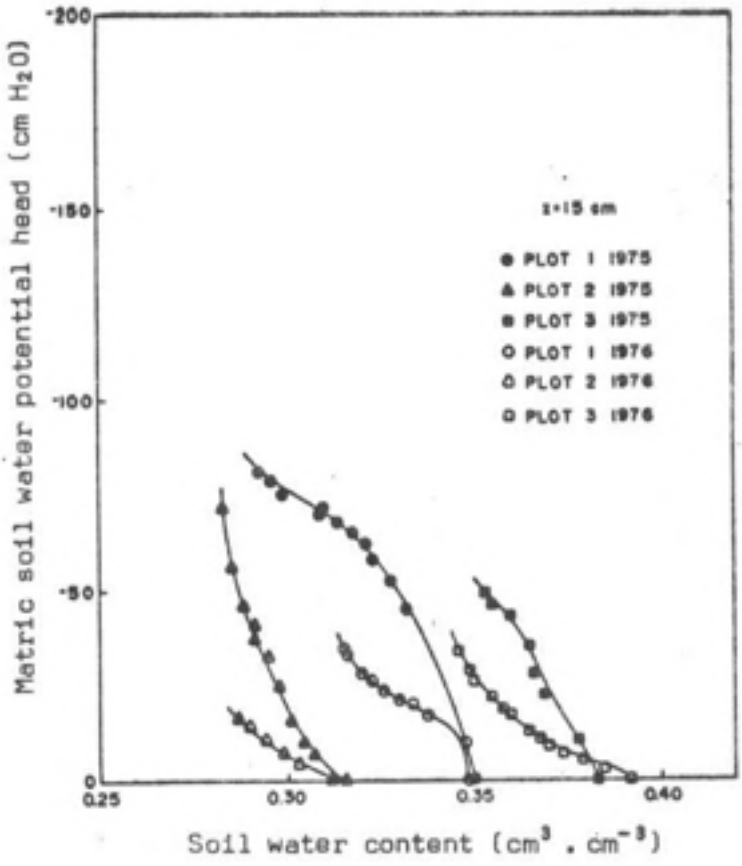

Figure 1 - Soil water retention curve for z = 15 cm.

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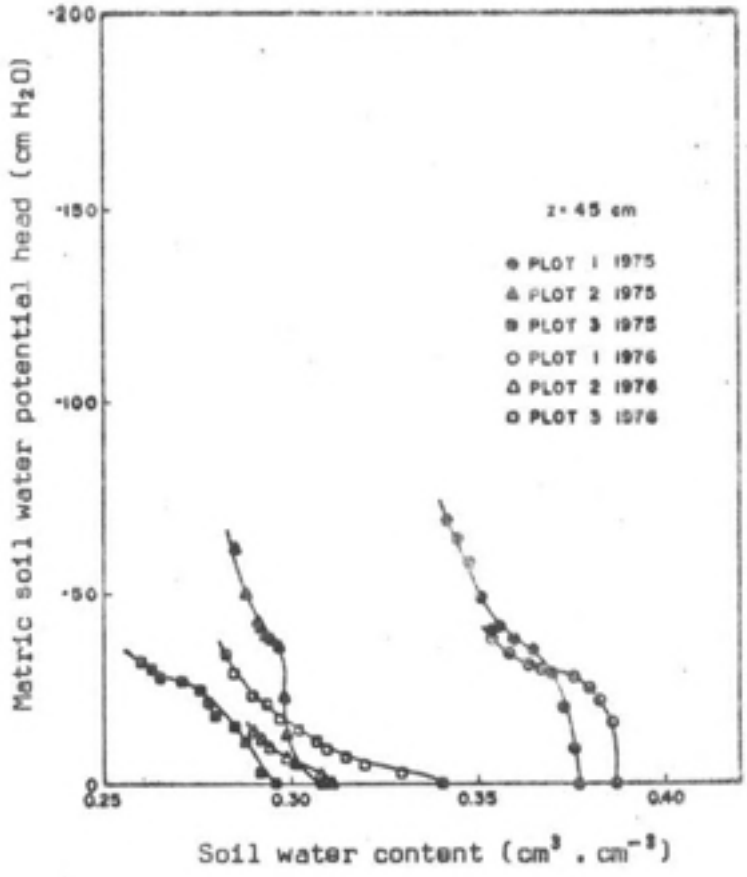

Figure 2 - Soil water retention curve for z = 45 cm.

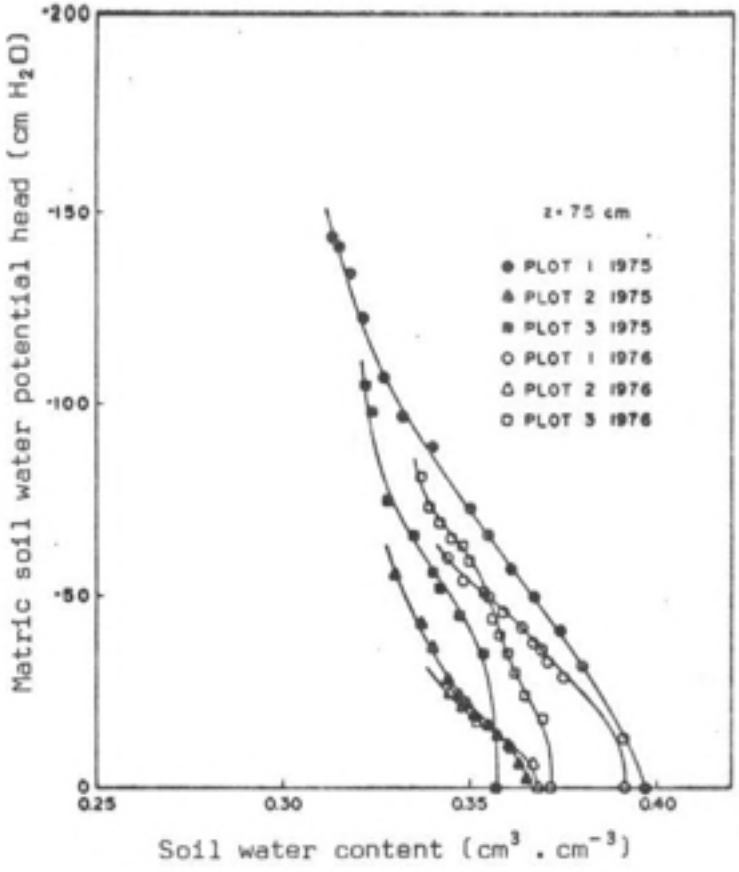

Figure 3 - Soil water retention curve for z = 75 cm.

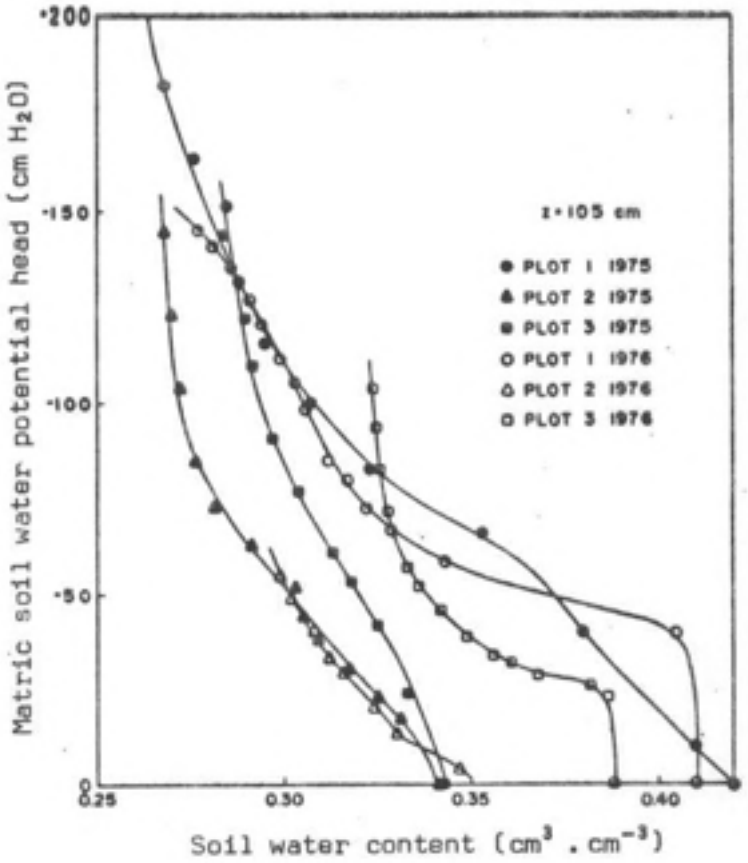

Figure 4 - Soil water retention curve for z = 105 cm.

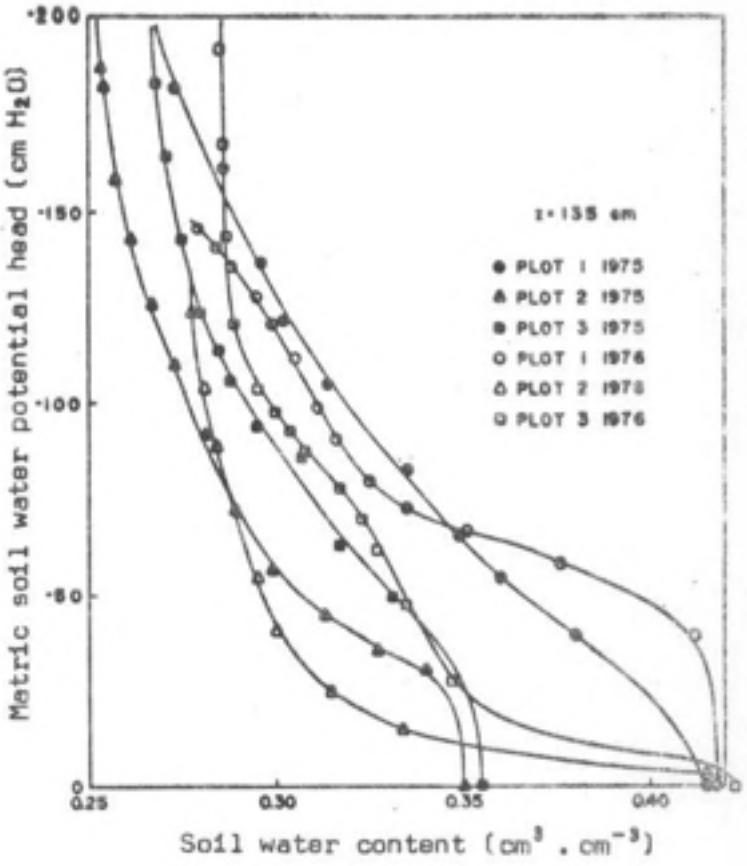

Figure 5 - Soil water retention curve for z = 135 cm.

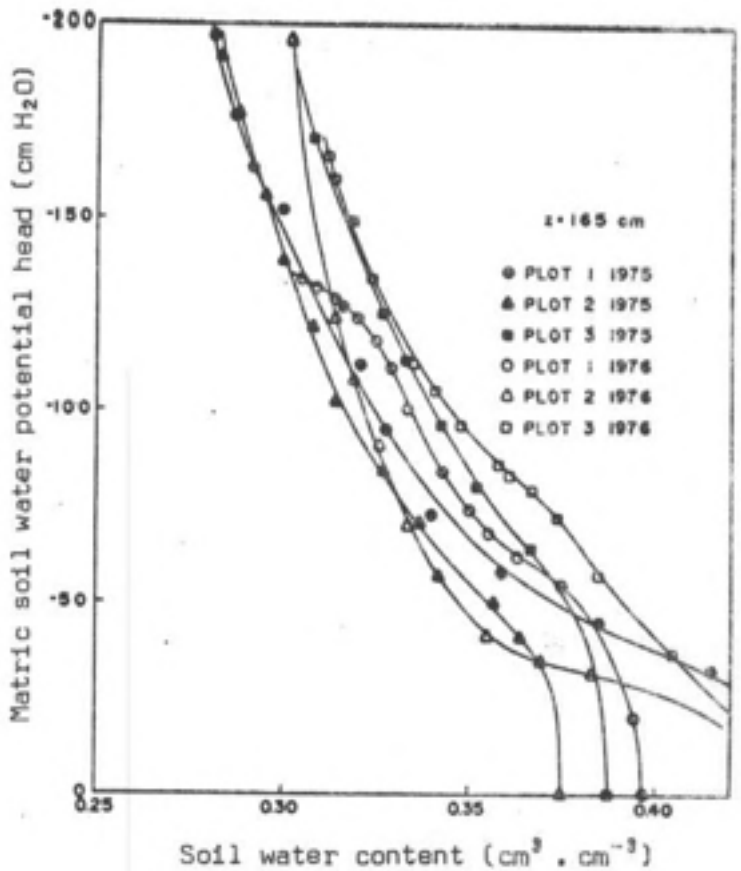

Figure 6 - Soil water retention curve for z = 165 cm.

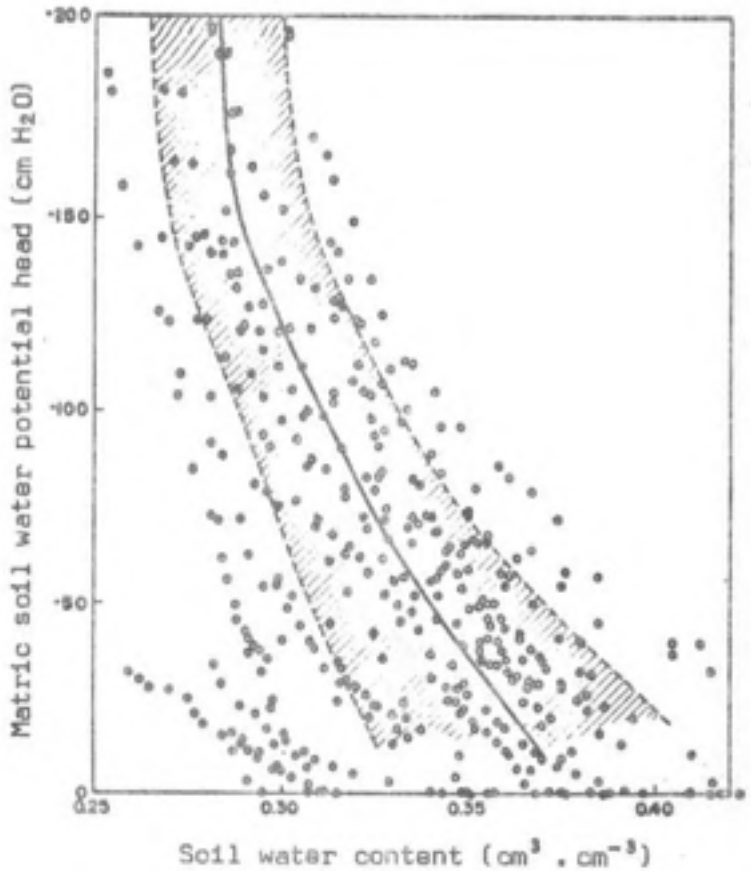

Figure 7 - Soil water retention curve for the entire field over all depths. Points are experimental. The solid line represents an average curve obtained averaging soil water contents at fixed values of matric soil water potential head. The shaded area represents the interval $m \pm \sigma$.

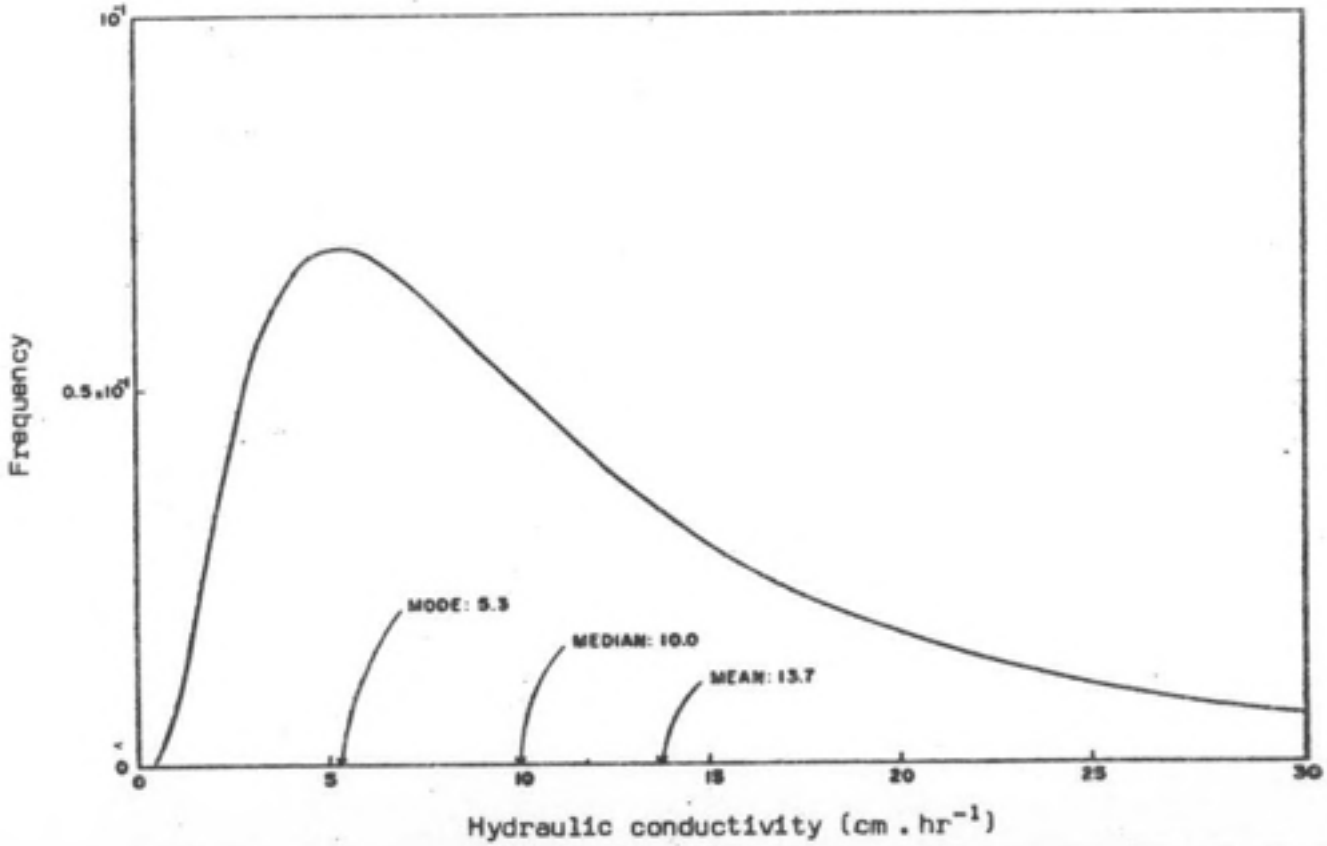

Figure 8 - Theoretical frequency distribution of hydraulic conductivity data, according to equation (3).

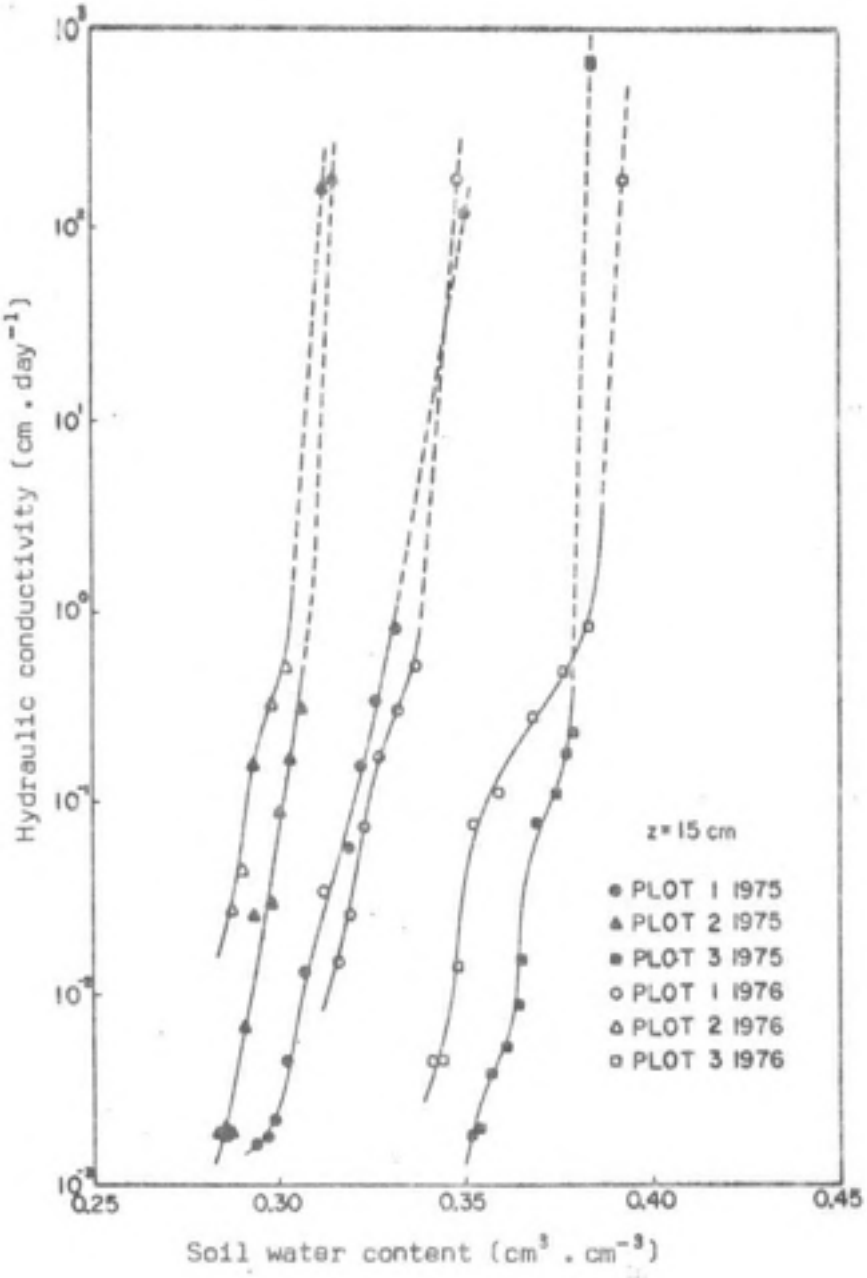

Figure 9 - Hydraulic conductivity as a function of soil water content for z = 15 cm.

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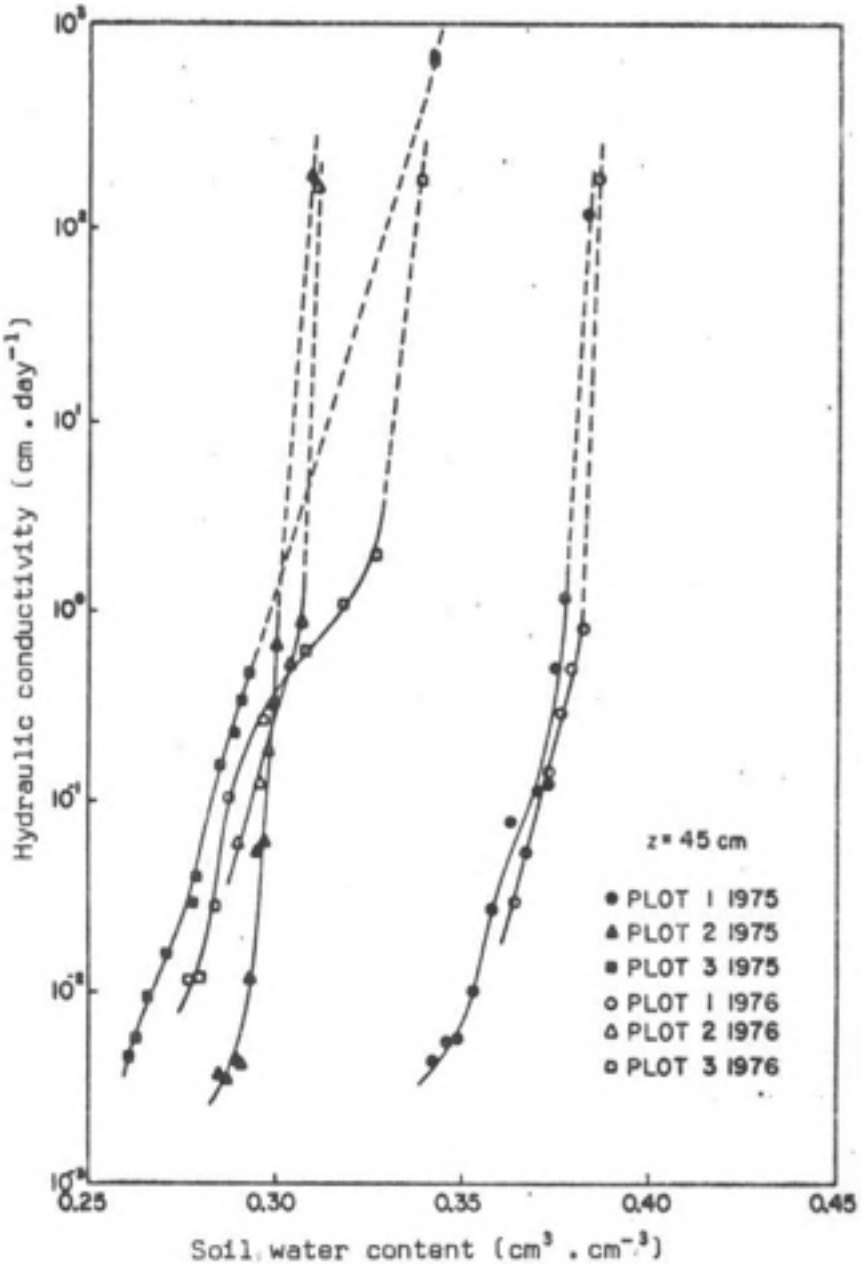

Figure 10 - Hydraulic conductivity as a function of soil water content for z = 45 cm.

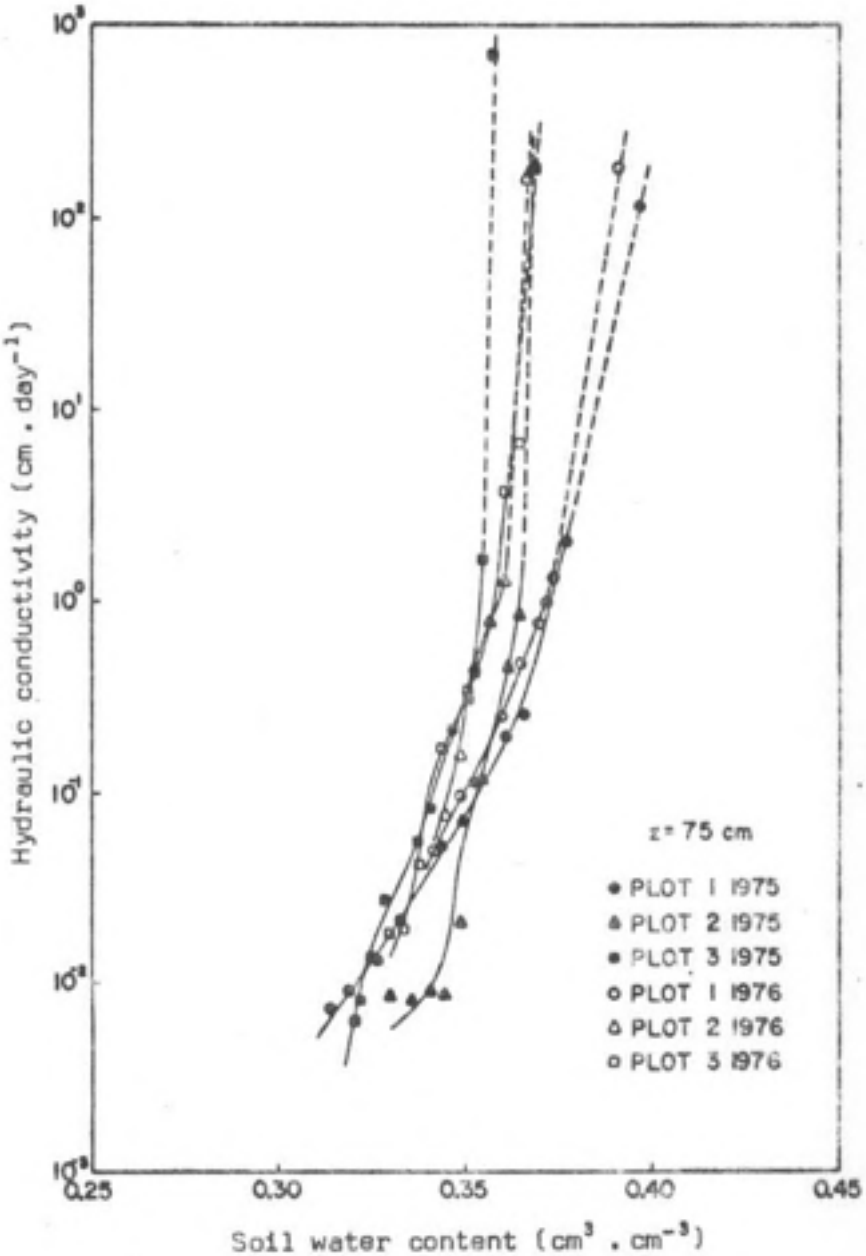

Figure 11 - Hydraulic conductivity as a function of soil water content for z = 75 cm.

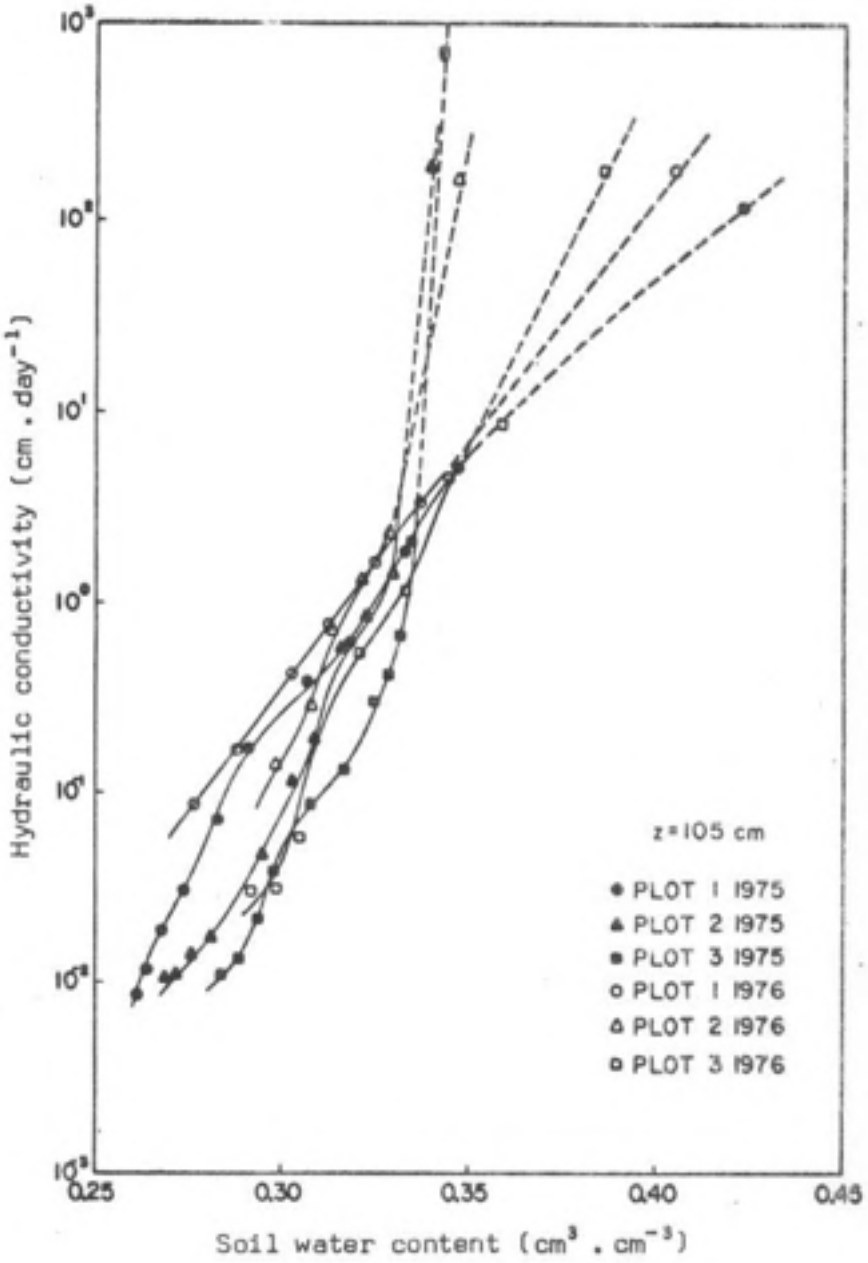

Figure 12 - Hydraulic conductivity as a function of soil water content for z = 105 cm.

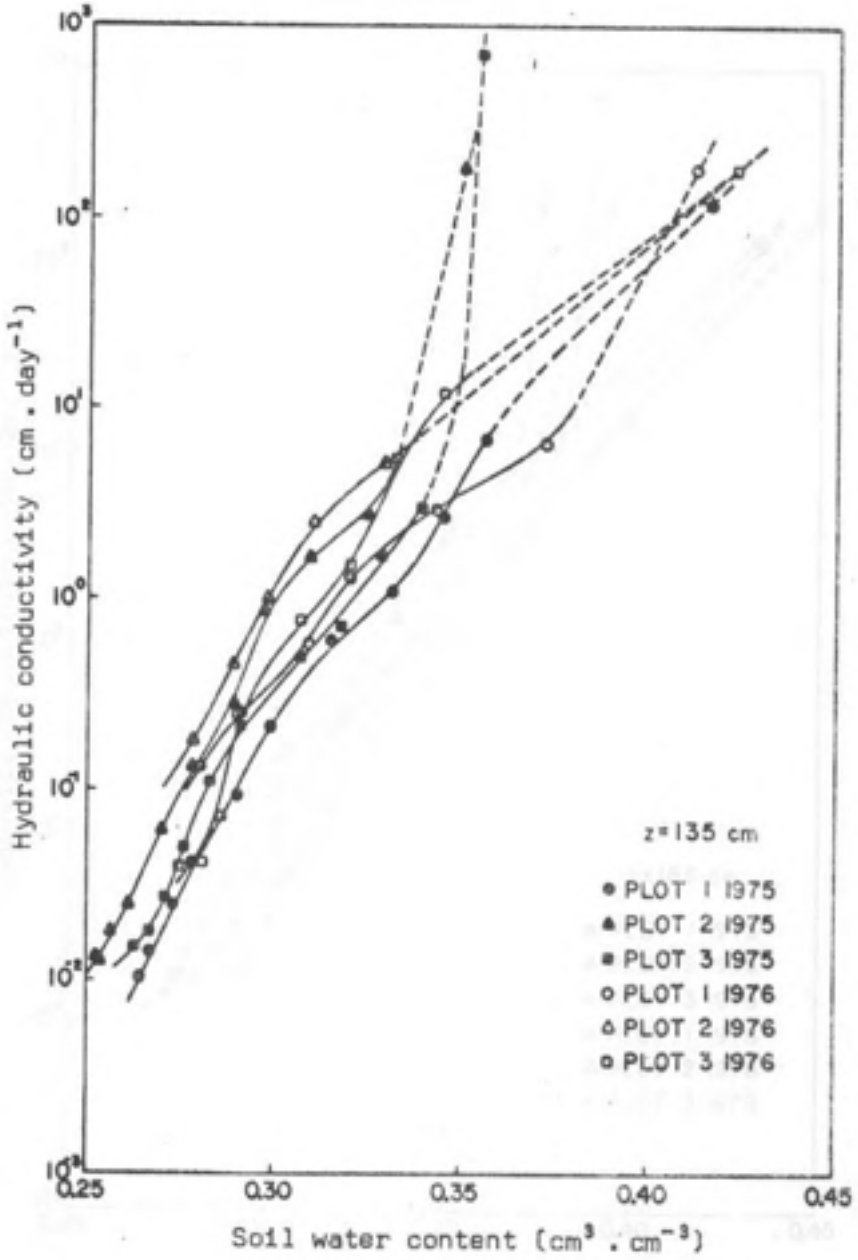

Figure 13 - Hydraulic conductivity as a function of soil water content for z = 135 cm.

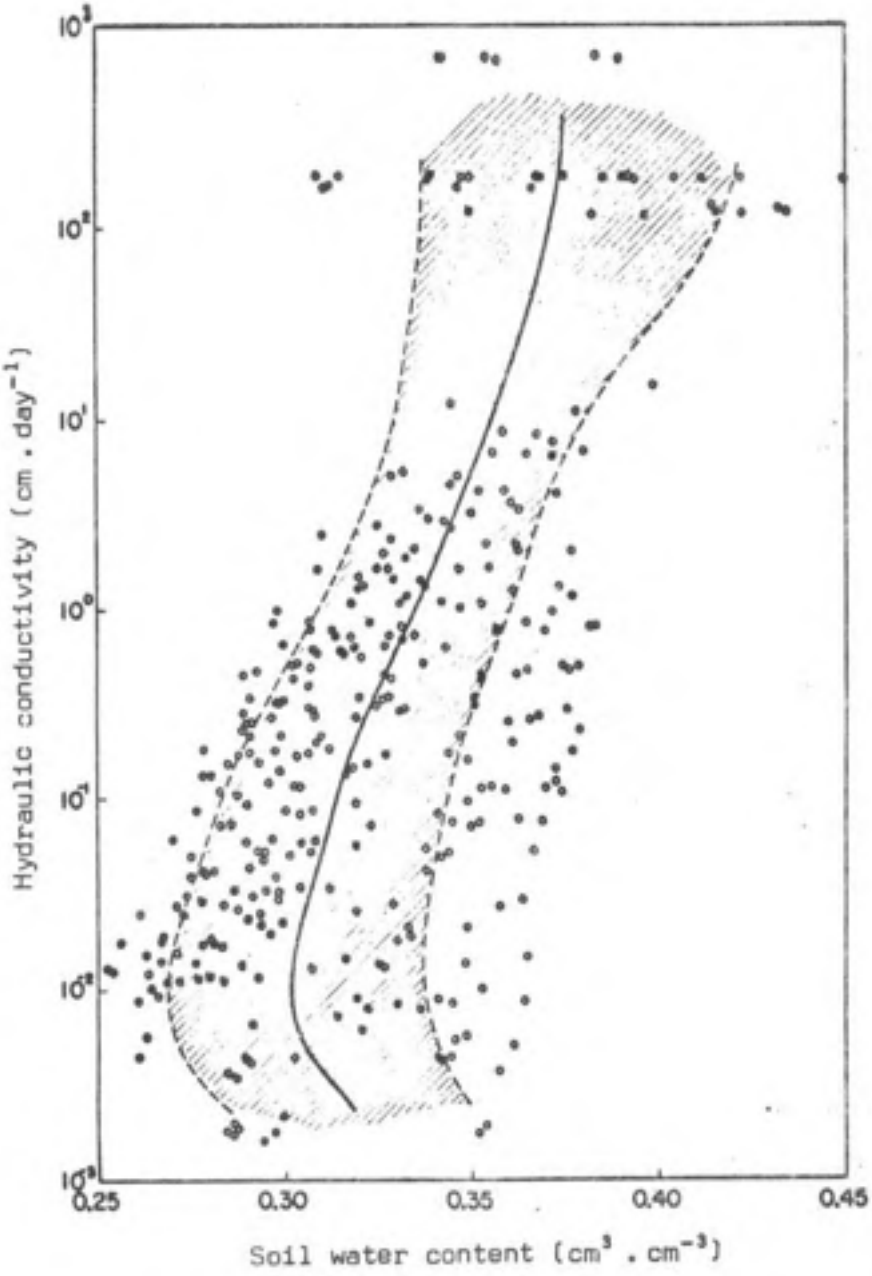

Figure 15 - Hydraulic conductivity data as a function of soil water content for the entire field at all depths. The solid line represents an average curve obtained averaging soil water contents corresponding to fixed values of hydraulic conductivity. The shaded area represents the interval $m \pm \sigma$.