DETERMINATION OF WATER CONDUCTIVITY COEFFICIENT OF SOIL POROUS MEDIA

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1. INTRODUCTION

The most important hydrophysical soil characteristics, the curve of water retention and water conductivity coefficient in saturated and unsaturated zones have a decisive effect on shaping conditions of plants' growth, development and yield. Relation of soil water potential to humidity (the curve of water retention) determines water resources and water availability to plants, while the value of water conductivity coefficient at a given value of soil water potential informs about a possibility of water moving in soil profile and influx of water containing dissolved chemical substances to plants root systems [45,159, 160,162,195,196].

Knowledge about hydrophysical soil characteristics is indispensable when it comes to description, interpretation and forecasting the course of physical, chemical and biological processes that take place in soil – plant – atmosphere system, and modelling these processes requires representative data concerning hydrophysical properties of soil. Utility of simulating and forecasting models describing hydrophysical processes which take place in soil – plant – atmosphere system greatly depends on precision of data concerning water characteristics of soil. Establishing the value of water conductivity coefficient correctly is necessary for obtaining demanded accuracy of models used, owing to a wide variability range of (a few orders of magnitude) water conductivity coefficient values within the whole range of changes of soil water potential values. Therefore methods of measuring and estimating these characteristics are developed intensively [3,4,11,16,20,28,31,35,38,79,83,85,89,93,100,106,107,117,119,132, 14,145,155, 162,166,182,186,188,190,191,211,217,218].

Relation of water conductivity coefficient to soil water potential or humidity is indispensable when it comes to simulation and forecast models of hydrophysical and hydrological processes occurring in soil – plant – atmosphere system. These physical and mathematical models based on constitutive physical equations make up basis for effective forecasting amounts of water resources in plant productivity as well as forecasting and preventing dangers connected with environment degradation and extreme water conditions, e.g. droughts and floods.

Spatial and temporal variability of physical, chemical and biological properties of soil [2,25,26,93,142], including water conductivity coefficient means that a great number of time-consuming measurements involving expensive equipment is necessary for proper characteristic of a specific area. Therefore, there are only fragmentary data measurement collections and water conductivity coefficient is estimated by means of models and algorithms whose results usually feature serious errors.

Verifying models is practically hindered because neither in Poland nor abroad is there any methodically homogenous database concerning physical, chemical and hydrophysical properties of soil.

Therefore the problem of creating databases concerning hydrophysical properties of soil, which has already been deemed crucial, was addressed in 1997 [196, 206, 207, 127] by scientific commissions of the Europen Union within the framework of the project called "Using existing data to derive hydraulic parameters for simulation models in environmental studies and in land use planning." The Institute of Agricultural Physics of PAN (Polish Academy of Science) in Lublin took part in drafting premises of this project.

Within the framework of research project P06B 012 15 a data base data concerning hydrophysical properties of arable lands in Poland was created in the Institute of Agricultural Physics of PAN (Polish Academy of Science) in Lublin, the database includes e.g. information about the curve of water retention and values of water conductivity coefficient at given values of soil water potential [193].

2. WATER CONDUCTIVITY COEFFICIENT OF SOIL POROUS MEDIA

The most important transport property of soil is hydraulic conductivity, K, which is a strong, non-linear function of the volumetric water content.

Relation of water conductivity coefficient to water potential or water content is a basic characteristic conditioning water movements in soil. Its value at a given value of soil water potential plays an important role when it comes to courses of various soil processes and, first of all, when it comes to providing plants with water and nutrients. Knowledge of this characteristic is necessary for description of water movement, forecasting water content layout in soil profile [10,17,76,115,116,146,158,194,197,198], designing irrigation devices and modelling processes of transport and pollution.

Flow of water in soil profile treated as a porous and capillary medium may be described by means of Richard's equation, whose one-dimensional form is as follows:

$$\frac{\partial\theta}{\partial t}=\frac{\partial}{\partial z}\left[k(\psi)\left(\frac{\partial\psi}{\partial z}+1\right)\right] \tag{2.1}$$

where: ψ stands for soil water potential, [cm H_2O], θ stands for medium humidity, [cm^3 cm^{-3}], $k(\psi)$ is water conductivity coefficient, [cm s^{-1}], t stands for time [s], and z is a coordinate, [cm]. In order to solve this equation, either analytically, or numerically, one must know both relation of soil water potential to humidity (water retention curve) $\psi(\theta)$, and relation of water conductivity coefficient to water potential or humidity $K(\psi)$, $K(\theta)$.

Water conductivity coefficient in porous medium is defined by Darcy's equation, which states that water stream density is directly proportional to its potential's gradient:

$$\vec{q}=-k(\psi)\cdot grad\psi \tag{2.2}$$

where: q – water stream density [cm s^{-1}], ψ - soil water potential [cm H_2O], $k(\psi)$ – water conductivity coefficient [cm s^{-1}].

Many methods have been drafted that aim at establishing values of water conductivity coefficient of soil in unsaturated zone, both in the state of set flow and the flow that has not been set [30,38,59,60,61,66,83,85,113,176,200, 201,202,203,205]. Due to fast developments, particularly in recent years, in electronics, mathematical and numerical methods there has been a considerable development in research into methods measuring hydrophysical characteristics of porous media, including soil. It is particularly true about methods of measuring water conductivity coefficient in unsaturated soil. Development of tensiometric methods and TDR method has facilitated modifying Wind's method [179,200,204] as well as Instantaneous Profile Method (IPM) [174,176,199,200,202,203], which are presently most often used in research of this type.

However, measuring water conductivity coefficient both in saturated and unsaturated zones is time-consuming and laborious, moreover it requires specialist equipment. Consequently, there is a general tendency to estimate water conductivity coefficients with acceptable accuracy by means of drafted physical, mathematical and models and algorithms.

3. MEASURE METHODS FOR DETERMINATION WATER CONDUCTIVITY COEFFICIENT

Methods for determination of soil hydraulic conductivity can be divided into: small-scale in-situ methods, large-scale in-situ methods, hydraulic laboratory methods, modelling and correlation methods.

Knowing hydraulic conductivity of K in unsaturated soils (often commonly referred to as capillary conductivity) plays important theoretical and practical roles in water management in soil, particularly whenever water relations are dealt with dynamically. In the light of currently known methods, hydraulic conductivity may be determined both in the field and in a laboratory or calculated on the basis of patterns published in specialist literature. In this literature there are many methods to be found that establish values of K, e.g. Gardner's method [83], Wesseing's infiltration method [202], Czurajew's method [45] and others, such as Wind's method [204] Zawadzki and Olszta's method [217]. Measurements of unsaturated hydraulic conductivity are certainly difficult, particularly in the field where there are frequent changes in soil humidity. It is because of this fact that determining K in field conditions is rarely practiced, as it requires simultaneous measurement of many parameters. Observing and controlling these parameters simultaneously is only possible in laboratories fitted with necessary apparatuses and equipment. There is no doubt that even the most carefully collected sample does not fully mirror conditions of soil in the field. Therefore all measurements obtained from the sample give merely an approximate picture of natural conditions. On the other hand, it is only natural that any projects connected with water in soil should be based on prearranged indicators. Consequently, it is necessary to establish them in laboratories by means of a wide range of methods. Computational methods of establishing water conductivity K are some of the most important of such methods.

This thesis aims at depicting measurement and computational methods of establishing hydraulic conductivity and comparing results acquired by means of these methods.

3.1. Water outflow through shallow pressure according to Gardner.

Determining hydraulic conductivity in a laboratory boils down to forcing water flow through a soil sample by means of creating a difference in pressures φ on the sample's brims. This rule is easy to use whenever it comes to water flow in fully saturated soil ($\Theta = \Theta_S$, where Θ_S – water contents at full saturation). For unsaturated soils the issue becomes more complex, and laborious, as in order to establish function K=f(θ) or K=f(h),

where:

θ – water contents in soil [cm^3 cm^{-3}],

h – value of soil water potential [cm H_2O],

many values K(θ_i) or K(h_i) must be measured that reflect various water saturation of soil Θi or various soil suction pressure h_i (i= 1,2,…….,n). Whenever one has a stated number of matching values K i Θ or K i h, one may draw function K=f(Θ) or f(h).

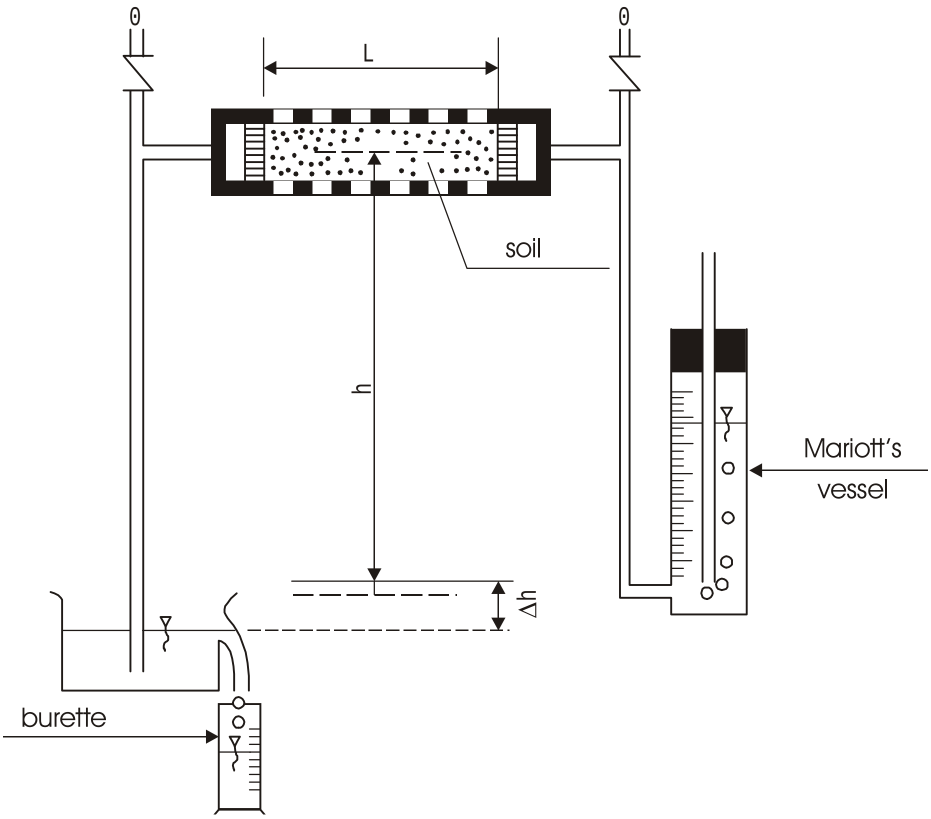

Fig. 3.1.1 Device for measurement unsaturated hydraulic conductivity

Difficulties in measurement are also connected with the time of establishing soil suction pressure (soil running dry prolongs the time of establishing soil suction pressure, Fig.3.1.1).

Gardner's method [83] of measuring K based on pace of water outflow from a vertical soil sample of small height (2-4 cm) assumes quick achievement of balance and h (or θ) being "constant" in the sample. Numerous experiments by Zaradny [214] have proven this method to be both laborious and imprecise.

3.2. Infiltration methods

A method based on constant sprinkling of soil monoliths with water proposed by Wessling [202] was used in Poland by Zawada [215] and others. Here measurement is carried out inside soil columns, which are usually 8 cm in diameter and 0.5 – 1.5 m in length, though there is a possibility of measuring water outflow and inflow as well as a possibility of measuring water suction pressure along the column. Water is supplied to the column from above at constant inflow volume q (the value must be chosen in such a way that water flow in the column takes place at partial water saturation). Having reached the state of balance, i.e. having equalled water outflow from the column with outflow q, one should read layouts of suction pressures using tensiometers installed along the column. Hydraulic conductivity ***K(h)*** is calculated by means of generalised Darcy's law:

$$K(h)=\frac{q}{F\left(1-\frac{dh}{dx}\right)} \tag{3.2.1}$$

where:

$K(h)$ – hydraulic conductivity matching suction pressure h;

q – water flow volume;

F – cross section surface area of the column;

h – pressure in soil;

x – horizontal coordinates (positive values directed upwards).

Ready soil column featuring properly installed tensiometers should be firstly saturated with water ($\Theta = \Theta_S$ for all values of x). Then hydraulic conductivity should be established for full saturation $K(\theta_S)$ (this value is most often called filtration coefficient). To achieve this assumed flow q must be relatively large ($q \geq K(\Theta_S)$). One must remember to choose stated values for calculations in such a way that water inflow equals water outflow. In the next phase of measurements

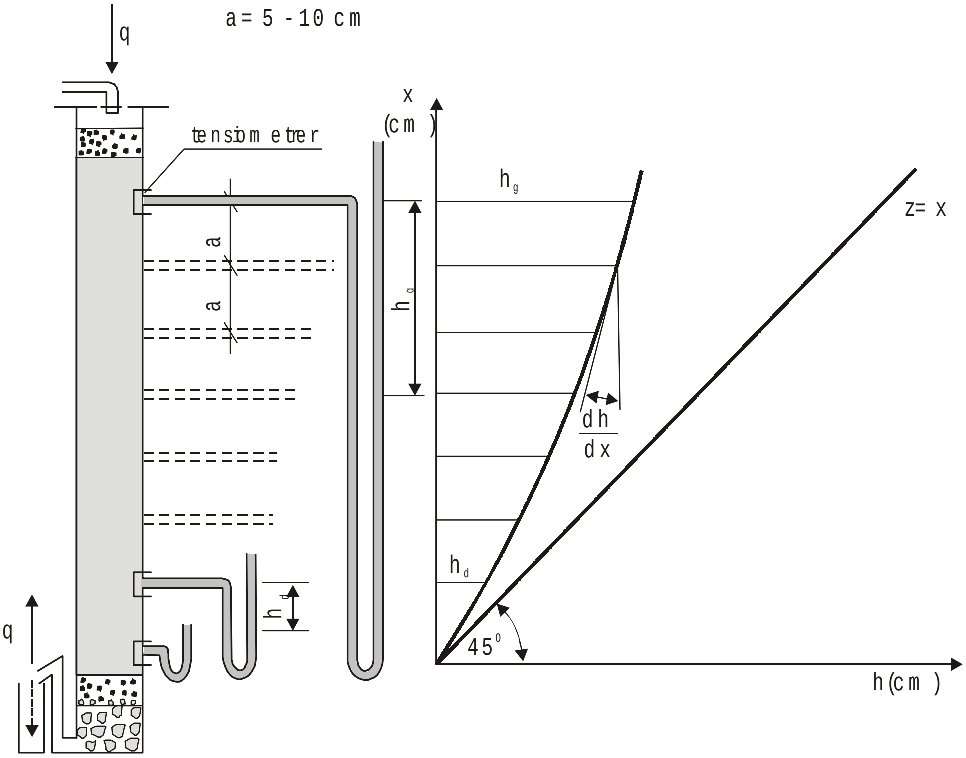

Fig. 3.2.1 Scheme of the device for soil hydraulic conductivity measurement using infiltration method

water inflow to the column is gradually decreased ($q_i<K(\Theta_S)$), establishing (having determined flow conditions) following K(h) values in formula (3.2.1.). Repeating measurement one may acquire many relations of K to h. Measurements should be carried out until curve h=f(x) approaches straight line z=x (straight line slanted at the angle of 45^0 to the horizontal axis),Fig.3.2.1 . Such a constraint is caused by the fact that formula (3.2.1.) cannot drop to zero in denominator. Basic requirement of this method is reducing flows in column q. These requirements as well as the fact that the method facilitates measuring K values for a narrow range of soil suction pressure make up considerable constraints.

3.3. Constant evaporation method

Determining hydraulic conductivity by means of constant evaporation method, which was worked out by Zawadzki and Olszta [216], is done using dust block. Such blocks are quite commonly used for determining retention abilities of water in soil deposits within the range of pF from 0,4 to 2,1. Cylinders measuring 15 cm in diameter and 35 cm in height are used for measurements from intact soil; the cylinders are positioned on ceramic plate covered with a plastic sheet.

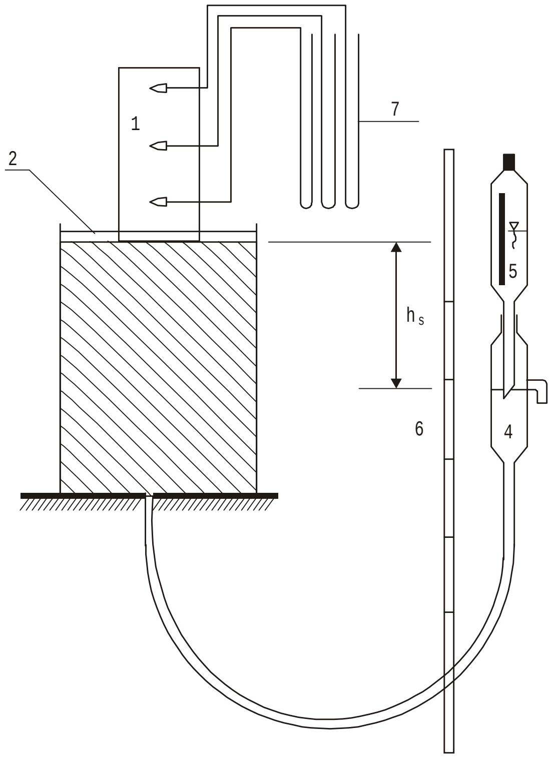

Fig. 3.3.1. Scheme of the device for unsaturated conductivity determination using constant evaporation method, 1-soil sample, 2-plastic foil, 3-ceramic plate, 4, 5-reservoir, 6-measurement scale, 7-manometer

Opening cut out at the top allows the monolith to come into contact with the surface of dust block. From above soil sample comes into contact with atmosphere (Fig 3.3.1).

Soil monoliths inside cylinders, in the state of current humidity, stand on the surface of fully saturated dust blocks. A monolith and a block simulate water relations to be found in soil profile.

Water level in dust block is regulated by means of a so called levelling vessel connected with the block by means of a transparent plastic cable, which allows monitoring to make sure that the cable is not air-locked. Water evaporated from the soil monolith's surface is supplemented from a bottle and its amount is read off a scale placed on the bottle.

Suction force h_S (in cm of water column) at the junction of soil monolith and dust block is determined by positioning of the levelling vessel measured against the block's upper surface. Suction force in soil monolith is measured by means of liquid manometers (connected to tensiometric drains).

Using this method one must account for the fact that the value of humidity potential (suction force equals the absolute value of h) at the junction of the monolith and the block depends on physical properties of the dust block. Values of suction force h_S acquired in the dust block can maximally equal 125 cm, which corresponds to pF = 2,1 value. Despite these constraints, this method facilitates simulating flow in the process of soil dehydration and irrigation, which also facilitates determining hysteresis from relation $h=f(\theta)$. Using this method to acquire evaporation values and suction pressure layout along a monolith facilitates calculating hydraulic conductivity according to formula (3.2.1) resulting from Darcy's law.

3.4. Dried monoliths method

The simplest method of determining hydraulic conductivity has been proposed by Wind [204]. It is much less laborious than the ones described previously, moreover it brings satisfactory results with pressure ranging from 0 to 900 cm ($pF \leq 2,95$).

Measurement is based on determining evaporation values from top surfaces of dried soil monoliths and changes of suction pressure in various layers of studied samples (Fig. 3.4.1). Soil monoliths are taken in the field, in an intact state, inside PVC cylinders measuring 35 cm in height and 15 cm in diameter.

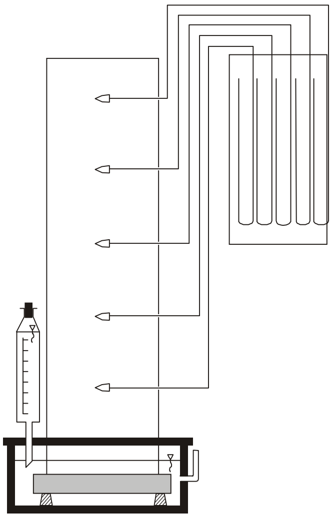

Fig. 3.4.1 Scheme of the device for unsaturated hydraulic conductivity determination using Wind's method

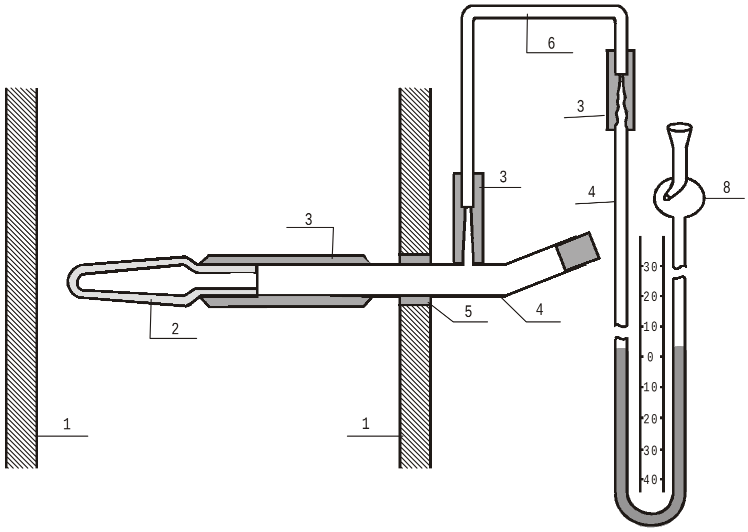

Fig.3.4.2. Scheme of tensiometer with mercury manometer

Holes are drilled in cylinders' side walls at 5-10 cm intervals (1.5 cm in diameter) so as to install ceramic drains in soil samples, each of which is connected to a mercury manometer (Fig.3.4.2).

Cylinders supplied to a laboratory are positioned horizontally inside a water-filled vessel, then ceramic endings of tensiometers are inserted into soil through the holes. Water-filled vessel and the top surface must be protected against water evaporation. At the moment of reaching a balanced humidity layout in the monolith, covers should be taken off cylinders, and at the same time observing pace of water evaporation q should be started as well as observing changes in layout of potentials h.

At the first stage of research, when soil water level is shallow (approx. 35 cm) evaporation volume q is read off a scale placed on a glass tube connected to the water-filled vessel (Fig.3.4.1). Evaporated water is supplemented from a bottle fixed to the vessel's lid. Suction pressure is indicated by manometers. Recording air humidity and temperature in the laboratory is carried out simultaneously by means of self-recording devices. Drying of the samples may be precipitated by changing pace of air flow and by warming the cylinders by means of radiators. However, in such case one must account for influence of temperature gradients on water flow from soil sample to atmosphere.

Whenever soil water occurs shallowly – on account of relatively intensive capillary water permeation in the monolith – range of potential changes of soil water that can be acquired is relatively narrow . This does not match conditions in soil. In order to broaden the range, during the second stage of measurement, having sealed all the holes in the cylinders, one must place them in a dry environment (without any contact with water in the vessel), leaving the top surface open to let water evaporate from the monolith into the atmosphere**.** Water losses which occur are marked by means of weighing the cylinders. In the first phase of drying the cylinders are weighed once a day, while data is read off tensiometers twice a day (e.g. at 8 a.m. and p.m.). As the samples dry the cylinders are weighed every 5 days and in the last stage every 10 days. Water losses through evaporation correspond to flow pace q for the monolith's upper (10 cm) level. For lower levels q is calculated from curves pF, reading suction force after a set time (e.g. every 5 or 10 days).

3.5. Wind method.

Wind method is one of important methods for unsaturated hydraulic conductivity determination.

In this method each soil sample is first wetted to near saturation in the laboratory. Then, the sample is allowed to dry by evaporation from the top surface;

and meanwhile at known times, pressure heads are measured at different depths in the sample using tensiometers, and the mass of the sample is measured. These measurements are continued until air enters any tensiometer; this can take a few days to two weeks depending on the type of soil. At the end of the test, after completing these measurements, the sample is dried and weighed, and its water content is calculated for each of the known times.

The sample is regarded as two or more compartments (sub-samples), one for each tensiometer; and, for each of the known times, the water content of each compartment is calculated from the water content of the whole sample and the tensiometer readings. The soil water retention characteristic and the unsaturated hydraulic conductivity are calculated from these data using an adaptation [96] of Wind's evaporation method [205]. The method treats the soil sample as being homogeneous in its hydraulic properties and assumes one-dimensional flow.

Equipment for sampling undistributed soil samples. Usually metal or plastic sleeves with known dimensions are used together with equipment to push the sleeves into the soil. Usually the sampling sleeves are used to retain the sample throughout the test, and therefore it is necessary to pre-dill holes for the tensiometers. The dimensions of the soil samples are dependent on the soil type and the purpose of the investigation. The height of a sample shall be smaller than or equal to its diameter, to prevent the acquisition data. In most cases a height of 8 cm and a diameter of 10 cm are suitable for stone-free soils.

The height must be large enough to accommodate 2 to 4 tensiometers. But, larger heights delay the drying of the lower compartments unduly, so determination takes too long, and also perhaps, the number of known times has to be in creased. The ratio of the diameter to the height should perhaps be just above unity, 10:8, to provide reasonably uniform conditions across the sample without requiring too broad a sample. Container and polyamide mesh to saturate the soil samples. Balance, capable of weighing to within ± 0,1% of the mass of the soil sample. A balance dedicated to a sample for the duration f the test is preferable to reduce possible disturbances.

Tensiometer system, capable of measuring heads with an accuracy better than ± 1 cm (see ISO 11276). The lengths of the tensiometers shall be smaller than half the diameter of the sample. The diameters of the tensiometers shall be smaller than 0,1 part of the height of the sample. Equipment to install the tensiometers, i.e. an auger or similar device, of suitable dimensions to bore holes into which the tensiometers will fit closely. Materials to effect seals between the sleeve and the tensiometers. Drying oven, capable of maintaining a temperature of 105 ± 2 C.

The soil samples should be taken from the field in accordance with ISO 10381-4. It is essential that undistributed soil samples be used, since soil structure has a strong influence on the hydraulic properties. If the soil sample has been compacted or its structure disturbed during sampling or transportation, it shall not be used for this determination.

The bottom of the soil column should be cover by circle of mesh, or similar hydrophilic close woven material. The mesh and sheet will retain the soil sample. After that the sample must be wetting by capillary rise. The structure of some soils is not stable under saturated conditions. Such soils should not be saturated completely. They may be wetted either by placing them in a container in which the water level is maintained at the base of the sample.

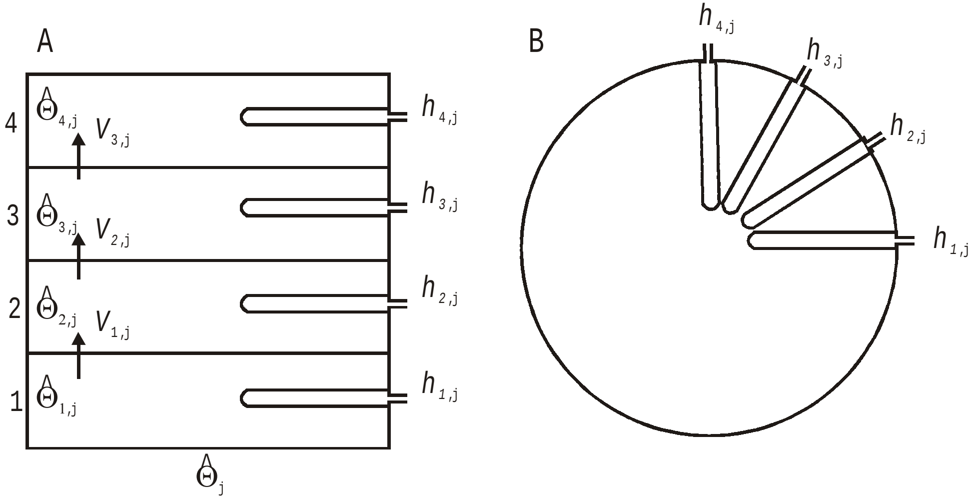

Fig. 3.5.1 Schematic overview of the sample with compartments. A – side view. B – top view.

After watering the top and the bottom of soil sample has to be close and the tensiometers should be installed in two or more compartments (Fig.3.5.1A). Compartments should have equal height. For instance, four tensiometers could be installed at 1,3,5 and 7 cm depth in a sample 8 cm high, thus giving fur compartments which are each 2 cm high (Fig.3.5.1A).

The distance between the centre of a tensiometer and an end of the sample shall be not less than the diameter of the tensiometer.

It is essential to obtain a smooth installation of tensiometers, minimum disturbance and a good contact between the soil and the tensiometer. Tensiometers must be installed horizontally in each compartment and in such a way that their vertical projections do not intersect (Fig.3.5.1B). The measurement is preferably performed in a room with constant temperature and humidity.

During the measurement in evaporation process the sample is located on the balance. The mechanical contact between the equipment on the balance and the other parts of the set-up (e.g. tubing or signal and power cables) influences the measured weight of the sample. Care should be taken that this mechanical contact is as small as possible and invariant during this measurements.

During evaporation process at intervals, the total mass, $m_j(j=1,\ldots..,$number of intervals) the pressure heads, $h_{i,j}(i=1,\ldots..,$number of tensiometers), and time, t_j must be determine. The maximum time between intervals should not exceed 8 hours. The experiment should be stop when air enters any tensiometers. Usually, this happens between h= -800 cm and -900 m. After measurement the total mass of the soil sample at the end of the test, m has to be determined. The measurements require approximately 2 days for clays and 2 weeks for coarse sands in an environment with T=20 C and a relative humidity of 50 %. The measurements require more time for sands due to the larger amount of water that has to evaporate before a pressure head of - 800 cm is reached. After drying the soil at 105 C, the average water content $\overline{\theta}_C$ can be obtained in accordance to ISO 11461.

Using equation (3.5.1) the average water content as a volume fraction of the whole soil sample at interval j can be calculated:

$$\overline{\theta}_j = \frac{m_j - m_e}{\rho_w V} + \overline{\theta}_C \qquad (3.5.1)$$

where:

$\overline{\theta}_j$ - is the average content as a volume fraction of the soil sample at interval j;

$\overline{\theta}_C$ - is the average content as a volume fraction of the soil sample at the end of the test

m_j- is the mass of the soil sample at interval j, kg;

m_e- is the mass of the soil sample at the end of the test, kg;

ρ_w - is the density of water, kg m^{-3} (≈1000 kg m^{-3});

V- is the volume of the soil sample, m^3.

For calculation the mean pressure head in the soil sample the equation 3.5.2 is using:

$$\overline{h}_j = \frac{1}{n}\sum_{i=1}^{n} h_{i,j} \qquad (3.5.2)$$

where:

$\overline{h}_j$ - is the mean pressure head in the soil sample at interval j, cm;

$h_{i,j}$ - is the pressure head of tensiometer i at interval j, cm;

n- is the number of tensiometers, i.e. of compartments.

For every interval j the average water content of the whole sample is calculated.

The mean estimated water content at interval j is calculated from retention curve using the equation:

$$\bar{\hat{\theta}}_j = \frac{1}{a}\sum_{i=1}^{n} a_i \hat{\theta}_{i,j} \qquad (3.5.3)$$

where:

$\bar{\hat{\theta}}_j$ - is the mean estimated water content of the soil sample at interval j;

$\hat{\theta}_{i,j}$ - is the estimated water content of compartment i at interval j;

a – is the total height of the soil sample, cm;

a_i- is the height of compartment i, cm.

For correction of $\hat{\theta}_{i,j}$ for every compartment the following equation is using:

$$\hat{\theta}^*_{i,j} = \frac{\bar{\theta}_j}{\bar{\hat{\theta}}_j}\hat{\theta}_{i,j} \qquad (3.5.4)$$

where:

$\hat{\theta}^*_{i,j}$ - is the corrected estimated water content of compartment i and interval j;

$\bar{\theta}_j$ - is the water content calculated in 3.1.1.

The pairs $h_{i,j}$ and $\hat{\theta}^*_{i,j}$ describe the retention characteristic as a table and as a curve respectively. For interval j, the gradient of the pressure head for adjacent compartments i and i+1 using the equation 3.5.5 is calculated:

$$\frac{\Delta h}{\Delta z} = \frac{-\sqrt{h_{i+1,j+1}h_{i+1,j}} + \sqrt{h_{i,j+1}h_{i,j}}}{z_{i+1} - z_i} \qquad (3.5.5)$$

where:

$\Delta h/\Delta z$ - is the gradient of the pressure head;

z – is the vertical coordinate, cm;

z_i – is the position of the tensiometer in compartment i, cm;

z_{i+1}- is the position of the tensiometer in compartment $i+1$ (the compartment above compartment i), in cm;

$h_{i,j}$- is the pressure head in compartment i and interval j, cm;

$h_{i,j+1}$- is the pressure head in compartment i and interval j+1 (the interval after interval j),cm.

Between all adjacent compartments the gradients can be calculated. A threshold for $\Delta h/\Delta z$ can be estimated when the standard deviation of the measurement noise of h, sd(h), is known. An appropriate method is to calculate an unsaturated hydraulic conductivity only if $\Delta h/\Delta z < -1-3\mathrm{sd}(h)/\Delta z$.

For interval j, the volume flux densities between adjacent compartments i and i+1 using the equation 3.5.6 is calculated:

$$v_{i,j} = \frac{1}{t_{j+1} - t_j} \sum_{k=1}^{i} a_k \left(\hat{\theta}_{k,j} - \hat{\theta}_{k,j+1}\right) \tag{3.5.6}$$

where:

$v_{i,j}$- is the volume flux density from compartment I to i+1 between intervals j and j+1, cm d^{-1};

t_j- is the time interval j, d;

a_k- is the height of compartment k(k=1the bottom compartment), cm;

$\hat{\theta}_{k,j}$ - is the estimated water content of compartment k at interval j.

Using Darcy's equation the unsaturated conductivity is calculated:

$$K_{i,j}\left(\overline{h}_{i,j}\right) = -\frac{v_{i,j}}{\frac{\Delta h_h}{\Delta z}} = -\frac{v_{i,j}}{\frac{\Delta h}{\Delta z} + 1} \tag{3.5.7}$$

where:

$K_{i,j}\left(\overline{h}_{i,j}\right)$- is the conductivity at a pressure head $\overline{h_{i,j}}$, cm d^{-1};

$v_{i,j}$- is the volume flux density, calculated in 3.5.6, cm d^{-1};

$\Delta h_h/\Delta z$ - is the gradient of th hydraulic head;

$\Delta h/\Delta z$ - is the gradient of the pressure head, calculated in 3.5.5.

The matching pressure head , $\overline{h}_{i,j}$ is calculated using the equation:

$$\overline{h}_{i,j} = -\sqrt[4]{h_{i,j} h_{i+1,j} h_{i,j+1} h_{i+1,j+1}} \tag{3.5.8}$$

and the matching water content, $\bar{\hat{\theta}}_{i,j}$, using the following equation:

$$\bar{\hat{\theta}}_{i,j} = \frac{1}{4}\left(\hat{\theta}_{i,j} + \hat{\theta}_{i+1,j} + \hat{\theta}_{i,j+1} + \hat{\theta}_{i+1,j+1}\right) \qquad (3.5.9)$$

The soil water retention characteristic can be expressed either as a table of the pair of values $h_{i,j}$ and $\hat{\theta}^{*}_{i,j}$ or as a curve of the last iteration.

The hydraulic conductivity as a function of the pressure head can be expressed either as a table of the pairs of values $\bar{h}_{i,j}$ and $K_{i,j}$ or analytically with a suitable curve.

The hydraulic conductivity as a function of the water content can be expressed either as a table of the pairs values $\hat{\theta}_{i,j}$ and $K_{i,j}$ or analytically with a suitable curve.

Hydraulic conductivities cannot be measured accurately in the wet part of the conductivity characteristic, due to the smallness of the gradients of the hydraulic head. The results obtained with this standard can be supplemented with independent measurements of the saturated hydraulic conductivity by a van Genuchten curve [132,188].

Measurement noise in the pressure head measurements has a large influence of the calculation conductivity, K, at small gradients of the hydraulic head (or $dh/dz \approx -1$). This can lead to biased results. Therefore, the conductivity can be calculated only when dh/dz differs significantly from -1. the calculation of the retention characteristic is not sensitive to this noise. The type of curve chosen for the water retention characteristic influences the results for both the water retention characteristic and the unsaturated hydraulic conductivity.

Characteristic of soil samples have been calculated using one and two modal van Genuchten curves to describe the retention characteristic[55]. The largest deviation of the measured average water content and the mean of the estimated water contents of the compartments was 0.015. typically, these differences were smaller than 0.005. The largest difference between the conductivities calculated with different curves was a factor of 5.4. Typically, these differences were smaller than a factor of 1.5.

The accuracy of this method has been estimated from computer simulations [180]. At given water content of the retention characteristic, the relative difference between the true and calculated value of the pressure head was smaller than 4% of the true value. After removal of non-significant hydraulic gradients, the inaccuracy of the determination of the conductivity was bounded by its value divide by 3 and 3 times its value.

3.6. Instantaneous Profile Method (IPM)

Work on determining the water conductivity coefficient in an unsaturated zone by means of the Instantaneous Profile Method was started in the 1960's [197], and continued into the 1970's [185]. Owing to the necessity to be able to measure water potential and soil humidity instantaneously, this method used to be considered laborious, requiring expensive specialist measuring devices especially for measuring humidity [127,176]. Due to developments in techniques for measuring capillary and porous media humidity, such as the TDR [126,175], the 'Instantaneous Profile Method' is currently employed by many scientific centres - including the Institute of Agricultural Physics of PAN (Polish Academy of Science) - as the standard method for determining the water conductivity coefficient in an unsaturated zone

The water conductivity coefficient in an unsaturated zone was determined by means of a laboratory TDR (Time Domain Reflectometry) test measuring the humidity, the soil's water potential, the temperature and salinity using the 'Instantaneous Profile Method' (IPM)[16,48,126,127,128,174,176,180,199,200]. In most of the cases described in the literature, soil columns measuring 5.5 cm in diameter around the perimeter and 10 cm in height were used to determine the water conductivity coefficient in an unsaturated zone by means of a TDR measurement test and the 'Instantaneous Profile Method', which facilitated the installation of 5 pairs of sensors measuring humidity and soil water potential at heights of 1, 3, 5, 7 and 9 cm from the bottom of the column. Such measurements have been deemed the standard.

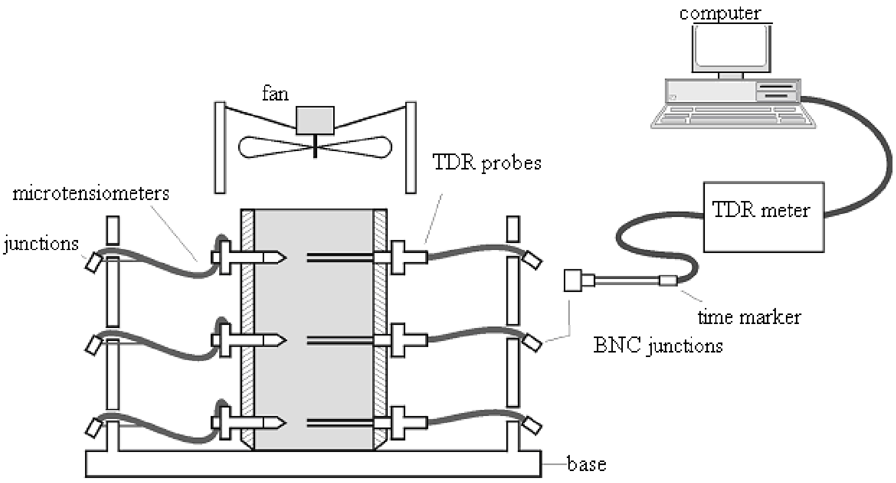

Fig. 3.6.1. TDR test for measuring dynamics of humidity and the soil's water potential [126, 175].

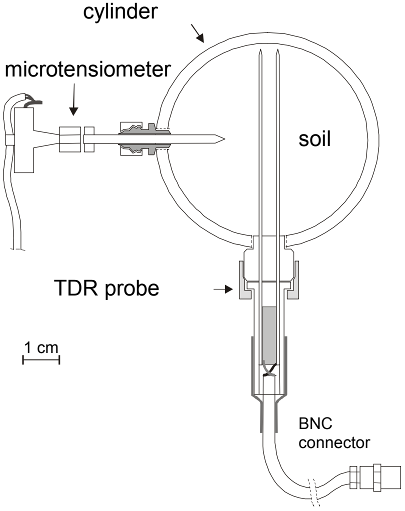

Fig. 3.6.2. Cross-section of soil cylinder with pre-installed sensors measuring humidity and water potential [127,176].

Nevertheless, in most cases it is the field research cylinders measuring 5.5 cm in diameter and 5 cm in height which are used to take samples. These are the cylinders used to take the soil samples stored in the database of the Institute of Agricultural Physics of PAN (Polish Academy of Science) in Lublin. Owing to a column's height it is only possible to install 3 pairs of sensors in it to measure humidity and the soil's water potential at heights of 1, 2.5, and 4 cm counting from the column's bottom. The research which has been carried out and the statistical analysis has shown that using the 'Instantaneous Profile Method' to determine the water conductivity coefficient in an unsaturated zone by means of a column measuring 5 cm and featuring 3 pairs of sensors generates results that are statistically equivalent to the results obtained by means of a column measuring 10 cm in height and featuring 5 pairs of sensors [174].

Therefore, soil columns measuring 125 cm^3 in volume, 5cm in height and 5.5 cm in diameter were used for all measurements. TDR probes measuring humidity and micro-tensiometers measuring the soil's water potential were inserted inside a soil column through pairs of holes at heights of 1, 2.5 and 4 cm counting from the bottom of the column (Fig. 3.6.1, 3.6.2).

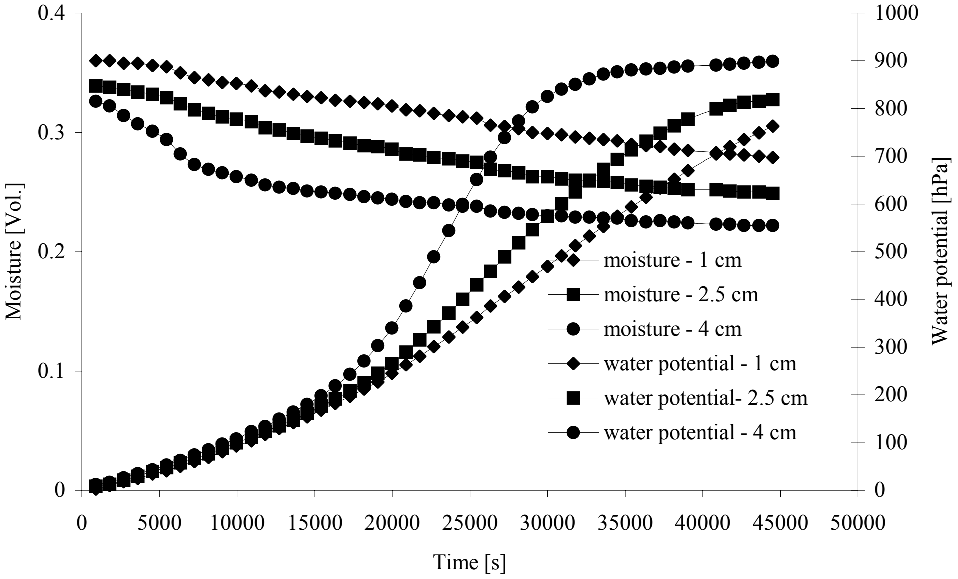

Fig. 3.6.3. Moisture and water potential dynamics in a soil column – an example.

The column was prepared inside a cylinder with three pairs of holes and was fully saturated by capillary action. The top end of the column was then covered and it was carried to measurement stands where humidity and soil water potential sensors were installed; finally it was set for 24 hours so that it reached a thermodynamic equilibrium. After that, the column was uncovered and the values of humidity and soil water potential were measured inside the three layers of the column while the soil sample was dried through evaporation from its surface. The values measured were automatically registered by the computer system controlling the device. The dynamics of the moisture and the soil's water potential are illustrated in Fig. 3.6.3.

As the measurements of water content and soil water potential are separate in time, because of their dispersion, Bezier's function was used to regularise them; the function takes the form of [8,176,199]:

$$\varphi_i^N = \frac{N!}{(N-i)!\cdot i!}\tilde{x}\cdot(1-\tilde{x})^{N-1},\ 0 \le \tilde{x} \le 1,\ \text{i=0, N} \tag{3.6.1}$$

provided that:

$$\sum_{i=1}^{N}\varphi_i^N(\tilde{x}) \equiv 1 \tag{3.6.2}$$

for all values $\tilde{x}$. Using Bezier's function, the humidity and water potential values on the surfaces are obtained interpolated into time and space co-ordinates as described by means of the following equations:

$$\hat{\theta}(\tilde{z},\tilde{t}) = \sum_{i=0}^{N}\sum_{j=0}^{M}\theta(z_i,t_j)\varphi_i^N(\tilde{z})\varphi_j^M(\tilde{t}) \tag{3.6.3}$$

$$\psi(\tilde{z},\tilde{t}) = \sum_{i=0}^{N}\sum_{j=0}^{M}\psi(z_i,t_j)\varphi_i^N(\tilde{z})\varphi_j^M(\tilde{t}) \tag{3.6.4}$$

Assuming that the process of water transport in a soil column takes place in isothermal conditions and that it is one-dimensional, the following form of Darcy's equation may be used in its description:

$$q(z,t) = -k(\psi)\left(\frac{\partial \psi(z,t)}{\partial z} - 1\right) \tag{3.6.5}$$

Alternatively, the water flux may be calculated by means of the following equation:

$$q(z,t) = -\int\limits_{z=z_0}^{z} \frac{\partial \theta(z,t)}{\partial t} dz \tag{3.6.6}$$

Comparing equations (3.6.5) and (3.6.6), the water conductivity coefficient may be calculated in a zone of unsaturated soil by means of the equation:

$$k(\psi) = \frac{\int\limits_{z=z_0}^{z} \frac{\partial \theta(z,t)}{\partial t} dz}{\left(\frac{\partial \psi(z,t)}{\partial z} - 1\right)} \tag{3.6.7}$$

Boundary conditions for the experiment being conducted were as follows: *$q(z_0, t) = 0$* where z_0 – is the closed end of a soil column (in the case of the experiment, the bottom of the column).

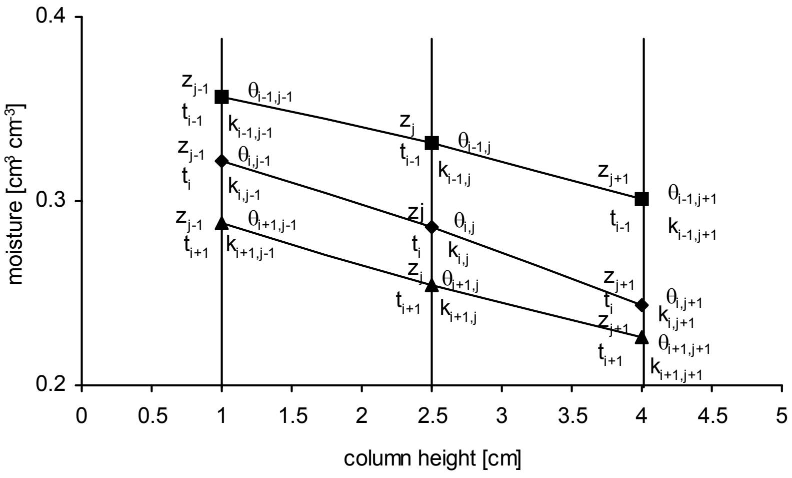

Fig. 3.6.4. Moisture profiles for chosen compartments of a soil column and for chosen times.

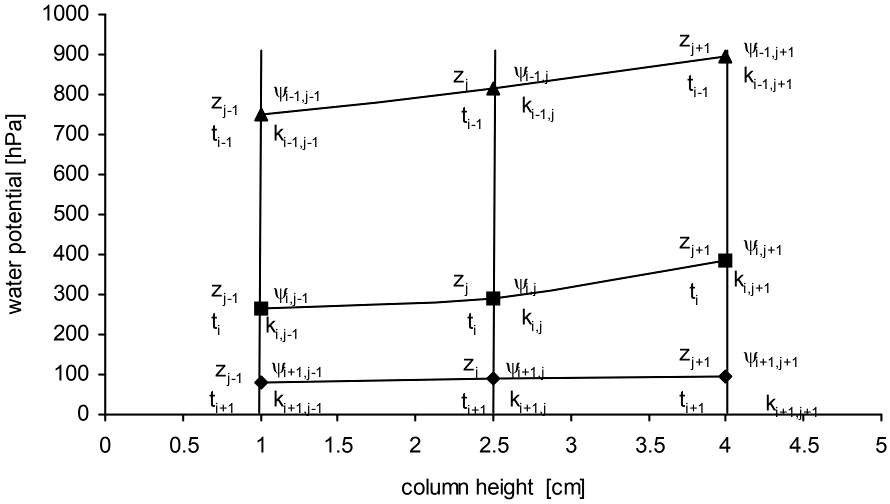

Fig. 3.6.5. The soil's water potential profiles in a soil column for selected compartments and selected times.

The soil's water potential and humidity profiles in individual layers of a soil column for selected times have been illustrated in Fig. 3.6.4 and 3.6.5. Indicator *i* numerates time, while indicator *j* numerates layers. In the knot *ij,* according to equation (3.6.8) the value of the water conductivity coefficient may be calculated by means of the following formula:

$$k_{ij}(\psi_{ij}) = \frac{\sum_{j=1}^{n} \frac{\theta_{i+1,j} - \theta_{i-1,j}}{t_{i+1} - t_{i-1}} \cdot \left(z_{j+1} - z_{j-1}\right)}{\frac{\psi_{i,j+1} - \psi_{i,j-1}}{z_{j+1} - z_{j-1}} - 1} \tag{3.6.8}$$

The soil water potential ψ_{ij} and water content θ_{ij} are ascribed to this value.

Due to the dispersion of the results of the calculations acquired thanks to equation (3.6.8) and the need to interpolate the values of the water conductivity coefficient for set values of the soil's water potential, it is necessary to illustrate the relation of the water conductivity coefficient to the soil's water potential in the form of a function. So, the final step for determining the water conductivity coefficient in a zone of unsaturated soil by means of the 'Instantaneous Profile Method' was smoothing out and interpolating the results of the calculations acquired. Various forms of functions may be used to achieve this aim. Functions in the form of a multinomial, a power function and a logarithmic function as well as van Genuchten's [188] equation have all been analysed. A function in the form of a multinomial has been rejected, as the coefficient of determination increases with the increase of the multinomial degree; nonetheless, the function behaves ambiguously between interpolated points. Statistical analysis pointed out that the highest coefficient of determination R^2 (within the range from 0.95 to 0.99) was acquired for the interpolation carried out by means of van Genuchten's equation [186]:

$$k(\psi) = K_S \frac{\left\{1 - (\alpha\psi)^{n-1}\left[1 + (\alpha\psi)^n\right]^{-m}\right\}^2}{\left[1 + (\alpha\psi)^n\right]^{\frac{m}{2}}} \tag{3.6.9}$$

where: K_S - is water conductivity coefficient in a saturated zone [cm day-1], ψ - is the soil's water potential [hPa], α, n and m – are the equation's parameters.

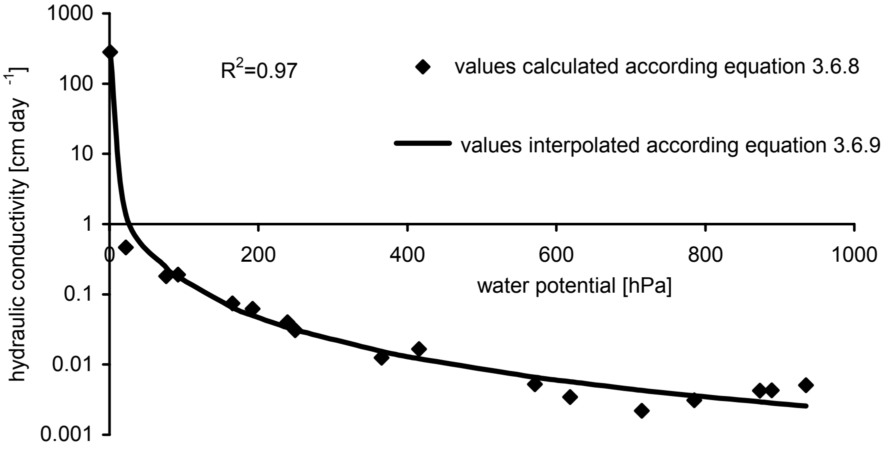

Fig. 3.6.6. An example of a plot of hydraulic conductivity values calculated and interpolated.

The curve described by means of equation (3.6.9) was adjusted to the points calculated by means of equation (3.6.8) using the last square method (Fig. 3.6.6).

Measurement of the water conductivity coefficient in a saturated zone K_{sat}, was carried out by means of a device produced by Eijkelkamp-Agricultural Equipment, which is used as the standard.

All the values of the water conductivity coefficient of the soils studied have been calculated according to the same aforementioned methodology. The ‘Instantaneous Profile Method’ and the measurement of the water conductivity coefficient in a saturated zone, as well as the numerical procedure which has been described above, facilitates the acquisition of the relation of the water conductivity coefficient to the soil’s water potential within the range of the values of the soil’s water potential, changing from 0 to about 900 hPa, which is conditioned by the tensiometers’ measurement range.

4. THE INFLUENCE OF THE SOIL'S PHYSICAL-CHEMICAL PARAMETERS ON THE HYDRAULIC CONDUCTIVITY COEFFICIENT

The soil's water conductivity is one of the basic properties conditioning water flow in soil. This is vital in the shaping of the various soil processes and first and foremost for supplying plants with water. Knowledge of this characteristic is essential to describing water flow, forecasting the humidity layout in a soil's process [10,76,116,146] and designing irrigation devices. The traffic of nutrients, especially nitrogen in the soil profile [144] also largely depends on water conductivity. Determining water conductivity coefficient *k* is lengthy and laborious and, moreover, requires specialist equipment. Consequently, there is a need to find relations which are relatively easy to determine or can be accessed from pedological materials that have already been worked out in order to facilitate the estimation of water conductivity on the basis of the soil's properties. These properties include granulometric composition, which is a relatively constant soil feature and can influence water conductivity both directly and indirectly through affecting aggregation and pore layout in respect of size.

The influence of individual particles on water conductivity varies when it comes to their sizes, shape and physical-chemical properties. Pillsbury [148] states that, in laboratory conditions, the value of coefficient *k* is positively correlated to soil particle size within the range of 110 – 500 µm, and the relation is expressed by means of an exponential function. The considerable influence of coarse-grained sand fractions on increasing water conductivity has been confirmed by the results of other experiments carried out on several chosen soil units [1,94].

Considering the influence of loam particles on water conductivity, it should be taken into account that they are very fine and occur, practically, in the form of small aggregations. Their influence depends on the type of loam mineral in the loam fraction. Fahmy [74] states that the addition of loam aggregations of the silty type (0.5-1 mm) decreased water conductivity only slightly, unlike aggregations of the montmorillonite type. It is so because the former do not swell in humid conditions and do not disintegrate under the influence of water and the way they affect soil structure is similar to sand particles, whereas aggregations of the montmorillonite type absorb considerable amounts of waterand swell. Not only does swelling cause sand particles to rise, but it also presses loam particles into the sand's basic pores, which blocks them. That is the reason why these aggregations contribute considerably to decreasing water conductivity by increasing the contact of soil particles. Loam swelling, causing aggregation stability to decrease, proved to be the cause of decreasing water conductivity in Klages' research [113]. Yet Moldenhauer and Kemper [138] point to the two-pronged character of the influence of loam fractions. On the one hand, it contributes to an increase in water resistance of aggregations by strengthening saturated water flow supplied to the soil in a way that does not result in the mechanical destruction of its structure. On the other hand, irrigation employing the 'falling drops' method which have an energy similar to that of rainfall, causes aggregations to wash away and causes the dispersed loam particles to cover the soil pores, decreasing the pace of water traffic. It is thought that even small amounts of loam cause the saturated water conductivity of sandy soil to decrease considerably [7,74] and its cohesion to increase. Consequently, a considerable increase in loam content in some cases contributes to the formation of an aggregate structure that facilitates water traffic in the soil's profile [23,74,103,166].

Water traffic also largely depends on the loam particle content. This fraction causes an aggregate structure to form and the general porosity to increase although these aggregations feature low water resistance [110,123,178]; the fact that they are washed out results in small soil particles covering the pores taking part in water transport. Krüger and Ehwald [118] proved that the aforementioned fraction in silty soils was negatively correlated to water conductivity saturated to the highest degree. Nielsen and co-authors [144] came to the contrary conclusions, having proven that the silt fraction is not considerably correlated to the water conductivity coefficient and infiltration. The content of silt particles also formed the basis for determining the degree of the water permeability of the soils in Diebold's experiment [58].

The aforementioned lack of coherence in the results may have been caused by the mutual relations of particles of various sizes whose direction depends on a chosen fraction's quota in the granulometric content of individual soils [133]. Consequently, attempts have been made to relate a synthetic index of granularity to saturated conductivity, which would account for total soil graining, [21,72,131]. Therefore, it was established that the saturated conductivity of the soil increases with the increase of the average size of the particles and whenever average sizes are identical – with the increase of the homogeneity of the soil particles [14,131].

Generally speaking, it may be stated that the influence of granulometric composition on water conductivity depends on the size and physical-chemical properties of the individual soil fractions conditioning the degree of so-called soil packing. Water traffic inside any level of a column or in the soil's profile not only depends on the soil's properties in that part, but also on the properties of the neighbouring layers above and below the layer studied [135,136]. Each heterogeneity of the granulometric composition of the profile, compared with the soil in total, diversifies the pores' layout when it comes to size and manifests itself in decreasing water traffic within the range - depending on the soil profile's construction. It was pointed out [137] that a layer which has a sand granulometric content influences water relations of loamy soil in a way that is similar to a low permeability level, yet the presence of a coarse-grain level does not result in water pressure above it. This level quickly transports water from the layers positioned above, at suction pressures close to zero (the state close to the saturation of the soil with water). Consequently increasing suction pressure (decreasing humidity) results in a considerable decrease of the border of heterogeneous phases. Rose and Passioura [164] pointed out that water traffic through porous media, e.g. sand, is mostly conditioned by their low hydrodynamic dispersion. The substantial influence of the presence of sand or gravel levels situated near fine-grained material, on decreasing water traffic through the soil profile was also confirmed by other experiments [6,43,137]. Employing the photographic technique, Taylor [180] and Gardner [80] proved that the gravel and sand layers situated below clay material take part in water transport providing that the clay level reaches a state of saturation. Consequently, relatively insignificant rainfall in regions lying in dry and temperate zones does not penetrate fine-grained material into the underlying sand or gravel levels. In such zones, crusts, consisting of sediments of silica, carbonate of lime and other mineral salts, precipitated through water evaporation, often occur on the border separating various granulometric phases. In such conditions, fine-grained material underlying sand material may also form a barrier, limiting water traffic [179].

The influence of the construction of the soil profile on changes in water traffic and, in consequence, on humidity layout has been used in many experiments in the reconstruction of profiles and the use of artificial levels [63,72,73,102,105,167]. These works prove that the amount of water stored in a profile consisting of heterogeneous layers usually increases, yet this effect decreases with an increase of height above the latter, restricting water filtration. If such a layer is close to the surface, retained water evaporates easily. On the other hand, whenever it is situated deep enough, its influence on the water content is insignificant, although the total water content above such a barrier increases in comparison with similar soil deprived of a level restricting water traffic.

The works discussed suggest that the influence of soil heterogeneity on water traffic and the humidity layout in its profile depends on the mutual positioning of layers of different granulometric composition as well as on the depths at which they occur.

The mutual influence of organic and mineral soil particles leads to the formation of various structural forms that also determine the geometry of the soil pores and condition water traffic in the soil. Most research pertaining to the influence of the size and durability of aggregations on soil water conductivity was carried out in laboratory conditions; the results obtained are not unanimous. However, according to Amemiya [5], aggregations of 0.5-1 mm in diameter including less loam and more sand were characterised by a greater water conductivity coefficient (within a wide humidity range) than aggregations measuring 3-5 mm. This author claims that these differences were caused by the fact that in aggregations measuring 0.5-1 mm, with the same humidity, the soil water potential was greater. Other experiments [44] also show that the water traffic was less intense inside columns filled with aggregations whose diameter was smaller. This phenomenon occurred differently in aggregations of considerable water resistance which exuded from black earth [64]. The value of the water permeability coefficient increased with increases in aggregation size. This relation is directly conditioned by the influence of aggregation size on the number of large pores, measuring over 18.5 x 10^{-6} m. Rengasamy [161] came to similar conclusions and stated that water conductivity decreases with an increase in micro-aggregations forming a larger number of micropores whose participation in the water conductivity of the soil in the unsaturated state is rather inconsiderable.

The value of the water permeability coefficient obtained from field research on black earth was negatively correlated with the content of agriculturally valuable aggregations measuring 1-5 mm in diameter [50], whereas it was positively correlated in loamy and sandy soils [177].

The water conductivity of the soil profile also greatly depends on the degree of the aggregating levels situated in its lower parts [138]. In the case of aggregations measuring 0.5 or 0.5-2 mm in diameter, which underlie soil clumps measuring 8-20 mm, the pace of water traffic was similar, while aggregations measuring 2-4.7 mm situated below caused a considerable decrease in conductivity. The heterogeneity of the structure influenced water traffic conditions by decreasing the hydraulic gradient between the layers of the layout of the various pores.

The stability of the aggregations is an important soil structure element in determining water conductivity. It is generally agreed that the decreased water resistance of aggregations triggers a decrease in the saturated water conductivity of the soil, both in field conditions, [1,77,177] and in laboratory conditions [40,124,175]. The authors of the works quoted above prove that a decrease in water conductivity is mostly caused by the covering of the soil pores with particles of dispersed aggregations and the forming of a crust of low permeability. According to Collis-George and Green [40] aggregations of fine-grained and silty soils disintegrate under the influence of water affecting smaller ones, which can be observed by means of a microscope giving a magnification of 40 times. These micro-aggregations form a completely different structure when compared to levels which have not been washed out and condition water traffic inside a soil's profile [41,42]. It was also pointed out that the phenomenon of the dispersion of loamy levels occurs to a larger extent in soils featuring structures that include rounded aggregations rather than prismatic ones [9,22]. In soils featuring a low water resistance structure, wetting warmth is connected with the disintegration of aggregations. The structure subsequently destroyed limits water traffic to the lower levels [41].

The influence of loam, humus, limestone and iron oxides [39,55] is most often listed among the factors improving aggregation stability, which also result in an improvement of water traffic conditions. Organic substances exert the largest influence on the stability of macro-aggregations [55,98], while iron oxides, influencing mostly by forming films on loam surfaces, are of secondary importance [19]. It has also been pointed out that iron oxides, crystals and hardened zones [173] may form as a result of combining the aforementioned oxides with other soil components. An interest in the use of synthetic chemical substances to improve the stability of the structure has been growing recently due to the possibility of combining them with industrial and agricultural waste [50]. Dobrzański and co-authors [62] have proved that the residue after the flotation of sulphur, flocculated with *rokrysol WF-2,* [water solution of partly hydrolysed (modified) polyacrylamide (5-7% dry substance)] increased water conductivity at low suction pressures thanks to an increase in the share of large pores and the increasing water resistance of the aggregations.

The influence of granulometric aggregation and its water resistance as well as some other soil properties, which have already been discussed, are just indirect influences exerted mostly by forming general porosity, differential porosity and the geometry of the pores which directly determines the possibilities of water traffic in the soil. All these factors are unique to a given type of soil or groups of soils.

There is no homogenous judgement regarding the influence of general soil porosity on its saturated water conductivity. Some research exhibited a positive correlation between coefficient k and general aeration porosity [69,177]; others – negative [141]. These discrepancies probably result from the disparate layout of the pores in respect of their size, primarily from the different content of large pores, when referring to water transport in a saturated zone. Swelling in the aforementioned soil exerts a considerable influence on the layout of the pores in fine-grained soils with an extensive colloidal loam content and high general porosity. During water traffic in the state of water saturation, swelling triggers an increase in general porosity, however, the share of large pores decreases, which results in a drop of coefficient k in a saturated zone. Nevertheless, there is general agreement that saturated water conductivity is proportional to the content of those large pores described as non-capillary [15,33,141,172], and inversely proportional to the content of small pores [141]. Similar conclusions stem from an analysis of Poisseuill's and Darcy's laws, which show that when there are 'same size' soil particles, conductivity increases, the larger the pores among them [74]. This results from assuming that soil particles are round in shape. In fact they have various shapes, which directly influences the form and layout of the pores. Canals produced by earthworms exert an extensive influence on saturated conductivity and are classified as large soil inter-grain spaces. Their influence on water transport inside a profile is larger in fine-grained soils than in coarse-grained soils. The reason is the existence of differences between the diameters of the canals and the diameters of other pores in fine-grained soils [94]. The extensive share of these spaces in saturated conductivity has also been confirmed by other works [51,70,71]. Germann and Beven [87] point out that it is possible to regulate the activity of earthworms and other animals, e.g. by cultivating soils and exerting an extensive influence on soil hydrology.

Water density is an important factor affecting water traffic in the soil. The considerable influence of this factor on the decreasing volume and gas exchange of macro-pores has been confirmed by experiments carried out at the Lublin Research Centre [5,67,68,125,184,185,192]. Each change in soil properties, manifesting itself in decreased pore size, causes saturated water conductivity to decrease [29]. According to Prasad and Perkins [149], the saturated water conductivity of the layer situated directly below the cultivated level, which is usually denser (longitudinal foundation footing), was smaller in comparison to the underlying levels as well as to those situated above. The authors have proved that the below-cultivation level featured smaller general porosity, smaller accessible water content and a smaller share of large pores. Bouma and Hole [23] came to similar conclusions having carried out research on silty soil. A closer correlation of saturated conductivity with density and porosity was perceived in soils with the least dust content [32]. According to Das [47] a soil density increase of 1.25 – 1.56 g/cm^3 resulted in the multiple decrease of water conductivity, yet the change range depended on the physical state of the soil during thickening. Canarache and co-authors [34] state that the value of coefficient k is considerably negatively correlated with density and value r is -0.77.

In the unsaturated zone, unlike the saturated zone, an increase in soil density occasionally triggered an increase of coefficient k. According to Koleva [115], an increase of coefficient k - influenced by density - mostly manifested itself at a density corresponding to the water supply accessible to plants. The drying out of the compact illuvial level, which occurred more quickly than it did in the case of the spongier outer layer during vegetation, was also connected with it. According to Berden and Pavlakis [13] an increase in unsaturated conductivity under the influence of tightening, results from an increase of water traffic through water membranes along surfaces of soil particles. Having researched the influence of soil density on soil water diffusivity Shard [169] came to different conclusions. According to this author, decreasing diffusivity, which accompanies an increase in soil density, results from an increase in the specific surface - and consequently the sinuosity - of the pores.

5. ESTIMATIMATION OF THE WATER CONDUCTIVITY COEFFICIENT

The development of computer technologies has facilitated the drafting of models allowing the values of the water conductivity coefficient to be determined, depending on the values of the soil water potential or humidity.

These models can be fundamentally divided into two groups:

- mathematical-physical, based on principal, physical laws [28,31,35,82,84,88,100,132,151,188];
- statistical-physical, based on statistical correlation relationships [1,4,12,20,79,93,94,106,107,117,142,143,145,162,189,208].

The Basic physical-mathematical model for calculating the water conductivity coefficient of a porous medium is based on the assumption that water flow, water being incompressible liquid, in a porous medium with a given pore layout, is a laminar flow and takes place through cylindrical capillary tubes (Poiseuille's flow).

The theoretical model [99] employed to calculate the water conductivity coefficient has been drafted on the basis of these assumptions. The coefficient is calculated from the following formula:

$$K(S_e)=\frac{k_0}{r_{\max}^{2j}}S_e^l\left(\int_0^{S_e} r^{(2+2j)/i}(x)dx\right)^i \tag{5.1}$$

where: l – is the pores' sinnuousity parameter, r – is a pore's radius [m], r_{max} – is the radius of the pore of maximum size to be found in a sample [m], S_e – is reduced humidity $S_e=(\theta-\theta_r)/(\theta_s-\theta_r)$, θ - humidity [cm^3 cm^{-3}], θ_r – is residual humidity [cm^3 cm^{-3}], θ_s – is saturated humidity [cm^3 cm^{-3}], k_0 [m s^{-1}] is the coefficient of proportionality $k_0=g/(k^*\nu\sigma)$, g – is the acceleration of gravity [m s^{-2}], k^* – is the coefficient of shape, ν - is kinematic viscosity [s^{-1} m^{-2}], σ - is surface tension [N m^{-1}], j and i are the model's parameters.

If data is replaced for $l=a$, $(2+2j)/i=b$, $i=c$ and $k=k_0/r_{\max}^{2j}$ the result will be an equation of the empirical model for calculating the water conductivity coefficient in an unsaturated zone [100]:

$$K(S_e) = kS_e^a \left(\int_0^{S_e} r^b(x)dx \right)^c \tag{5.2}$$

Models to be found in literature, which are usually used for calculating the water conductivity coefficient in an unsaturated zone are specific cases of equation (5.2) and are obtained by replacing parameters *a, b* and *c* with particular values (In table 5.1, the empirically determined parameters have been bracketed [100]).

Table 5.1. Values of a, b and c parameters of equation (5.2) corresponding to indi vidual models [**100**].

Model	a	b	c
Parcell [1949]	0	2	1
Fatt and Dykstra [1951]	0	(3)	1
Burdine [1953]	(2)	2	1
Mualem [1976]	(0.5)	1	2

Determining *r(Se)* is necessary for calculating the water conductivity coefficient. *K(S$_e$).* All models are based on the Young-Laplace equation, which states that pressure under the curved surface of a liquid is inversely proportional to the average radius of that surface's curve.

In the case of a meniscus whose shape is that of a spherical surface: $h = 2\sigma / R_H$, where *h* is capillary pressure [cm H_2O], σ is surface tension [Nm^{-1}], and R_H is the radius of the meniscus' curve [m].

Two models often used in literature to calculate the water conductivity coefficient in an unsaturated [31] zone are Burdine's:

$$K(S_e) = K_s S_e^l \left[\int_0^{S_e} h^{-2}(x)dx / \int_0^1 h^{-2}(x)dx \right] \tag{5.3}$$

and Mualem's [132]

$$K(S_e) = K_s S_e^l \left[\int_0^{S_e} h^{-1}(x)dx / \int_0^1 h^{-1}(x)dx \right]^2 \tag{5.4}$$

There are also specific cases of the general model. A comparison of the compatibility of results acquired by means of Burdine's and Mualem's models with experimental data, allows it to be stated that by means of both models one may assess the water conductivity coefficient with similar accuracy [100]. Lesser known models of a similar type were suggested by Purcell [149], Fatt and Dykstre [75], Wyllie and Gardner [209] and finally by Alexander and Skaggs [4]. There are also models, whose functional relation between the water conductivity coefficient and the soil's water potential or humidity is determined on the basis of experimental data. Brooks-Corey's [27] equations may be listed among models of this type:

$$K = K_{sat} \qquad h>h_b$$

$$K(h) = K_{sat}\left(\frac{h}{h_b}\right)^{-(2+3\lambda)} \qquad h<h_b \tag{5.5}$$

Wind's [203];

$$K(h) = ah^{-n} \tag{5.6}$$

Gardner's [82,83,84];

$$K = K_s e^{\alpha h} \tag{5.7}$$

and

$$K = \frac{a}{h^n + b} \tag{5.8}$$

Gilham's et al [88];

$$K = a\theta^n \tag{5.9}$$

as well as Campbell's [35];

$$K = K_{sat}\left(\frac{\theta}{\theta_s}\right)^{2b+3} \tag{5.10}$$

These are classified as exponential or power models. In these equations, K stands for the water conductivity coefficient in an unsaturated zone expressed in units [cm s^{-1}], h is the soil water potential expressed in units [cm H_2O], h_b is the pressure of the entering air, [cm H_2O], K_{sat} is the water conductivity coefficient in the unsaturated zone [cm s^{-1}], θ is volume humidity [cm^3 cm^{-3}], whereas parameters a, α, b, n, λ are empirical parameters of specific equations. These parameters are determined by means of mathematical methods for matching empirical equations to measurement data, e.g. using the method of the smallest squares.

Only some of the parameters in the models described above (mathematical equations) have a defined physical sense. The conductivity coefficient in saturated zone K_{sat}, the pressure of the entering air h_b and the humidity of saturation θ_s all may be ranked among such parameters. The remaining do not have any physical sense, yet, as a result of empirical and theoretical analyses they are often illustrated as functions of certain physical parameters of the solid soil phase. An advantage of these equations is the fact that a range of their parameter's variability is determined for most types of soil, which considerably facilitates using them, especially in the initial phase of the model physical processes occurring in soil when it is vital to know at least the approximate values of the water conductivity coefficient.

Research pertaining to the possibilities of correlating the humidity value corresponding to the chosen values of the soil water potential (the curve of water retention) with the physical-chemical parameters of the soil's solid phase (soil density, content of specific granulometric fractions, the size of a specific surface, the content of organic matter, etc.) has resulted in the creation of many statistical correlation models [3,12,37,95,99,145,154,155,156,193]. Therefore a basis has been formed for making attempts at correlating the parameters of equations (the models) describing the relation of the water conductivity coefficient to the soil's water potential or humidity with the parameters of the soil's solid phase [11,20,209,217]. Vereecken et al [188] determined the equations of multiple regression between the equations' (the models') parameters illustrated by Wind [203], Gardner [82,83,84], Gilham et al [88] as well as Brooks and Corey [28] (K_{sat}, b, n) and described the relation of the water conductivity coefficient to the soil's water potential and the content of specific granulometric fractions, the content of organic matter and soil density. The determination coefficient of the equations drafted ranged from 20% to 71.5%. Nimo [145] accounted for parameters characterising soil structure and texture in Mualem's [132] model. He suggested that soil humidity and its porosity should be illustrated as the sum of two elements, a so-called "textural" one, connected with a soil's texture, and a structural one. Comparing the values of the water conductivity coefficient measured with the values calculated by means of Mualem's model and a modified model exhibit, there is more agreement with the experiment in the case of those values obtained by means of the modified model. Research into the relations between the water conductivity coefficient and the layout of pore sizes [89] has also been carried out. Brooks and Corey's model has been modified on the basis of relations between fractal size and the layout of the sizes of the pores. Comparing measured values of the water conductivity coefficient with the values calculated by means of both models exhibited, there is more agreement in the case of the values calculated by means of the modified model. Renger et al [162] demonstrated the auto-regression procedure facilitating the calculation of the values of the water conductivity coefficient for 6 selected values of the soil's water potential. The values of the coefficient of the correlation between the measured values of the water conductivity coefficient and those calculated by means of determined auto-regression equations range from 0.73 to 0.94. Using the method of multiple regression, Gnatowski [92] correlated the transformed logarithmic parameters of van Genuchten's equation with selected physical-chemical parameters of the constant phase of peat soils and obtained values of determination coefficient R^2 ranging from 0.4 for parameter n to 0.81 for humidity corresponding to the full saturation of the soil.

The development of IT and numbering techniques has facilitated employment of the method of artificial neuron networks in the description of processes and phenomena in many fields of science and technology. One of the uses of this method has been the determining of the hydro-physical properties of porous materials, including soil. Schaap and Leij [168] suggested using the method of artificial neuron networks to determine the parameters of Gardner's [82,83,84] and van Genuchten's [187] equations on the basis of a knowledge of the physical parameters of the soil.

Research into determining the water coefficient of soil in a saturated zone on the basis of a knowledge of the parameters of its constant phase constitute a separate research group. The basic equation used in this type of consideration is Kozena-Carman's [36,212] equation which takes the form:

$$K_{sat} = C\phi_e^m \tag{5.11}$$

where: K_{sat} – is the water conductivity coefficient in a saturated zone [m h^{-1}], ϕ_e – is effective porosity [cm^3 cm^{-3}], C and m – are empirical constants. Effective porosity is defined as the difference between soil humidity in the state of saturation and humidity at water potential corresponding to 33 kPa. Parameter m, in theoretical considerations, is interpreted as a function of the cumulative distribution function of the pores, their continuity and sinuosity [90]. Parameter C is interpreted in different ways by various authors. Naney et al [142] show it as a function of loam fraction content, whereas Timilin et al [182] as a function of parameter λ, connected with the layout of the pores and the pressure of entering air ψ_a in Brooks and Corey's [28] equation. Gimenez et al [89] suggested using a fractal dimension to interpret this equation's parameters. As a result of theoretical research pertaining to the fitting of Kozeny-Carman's equation to the experimental data, it was proved that, depending on soil type, parameter C ranges from 44 to3,400, while parameter m ranges from 1.59 to 3.98 [2, 78,134]. Lipiec [122] analysed the possibilities of evaluating the water conductivity of the soil in a saturated zone on the basis of a knowledge of their granulometric layout. The correlation coefficient of regression equations ranged from 0.83 to 0.92.

Fractal models are often used for quantity descriptions of the soil's structure within the aspect of their relation to the water conductivity coefficient in a saturated zone [8]. Thanks to employing fractal geometry parameters for pore sizes, such as fractal size, the coarseness of the surface of the pore wall as well as the layout of the pores according to size, have all been determined [24,89,108,145,147].

The models illustrated refer to the possibility of correlating the theoretical parameters and empirical equations chosen (the so-called pedotransfer functions), which are widely used in research into the hydro-physical properties of porous materials, with the physical-chemical parameters of their constant phase. The parameters of these equations do not usually have any physical sense and therefore any attempt at connecting them with the physical properties of a constant phase facilitates the better recognition of the influence of specific parameters on the process of water flow in capillary-porous bodies. Another important advantage of these models is the fact that they facilitate the determining of the values of the water conductivity coefficient both in a saturated zone and in an unsaturated zone, on the basis of the easily measured physical parameters of its constant phase.

5.1. Model for estimation saturated hydraulic conductivity

Investigations were performed (Lipiec, 1983) aimed at estimating the structure of the hydraulic conductivity of undisturbed soil. The research included 31 soil profiles studied at three depths of 0-30, 30-60 and 60-90 cm. The determination of water conductivity coefficient K in saturated and unsaturated zones was performed by means of the filtration method, using gypsum-sandy crusts. In this method, steel infiltrometers, measuring 28 cm in diameter and 15 cm in height were applied, into which soil samples, of undisturbed structure, were inserted. On the basis of the statistically significant correlation between soil water conductivity in a saturated zone and the percentage content of sand, silt and clay fractions as well as the grain-size distribution index, regression equations were determined to calculate the water conductivity coefficient in a saturated zone. For particular layers, these equations have the following form, respectively:

$$K_s = 131.32 + 0.089 \cdot sand^2 - 5.75 \cdot sand \quad \text{R=0.71} \quad \text{(0-30 cm)}$$

$$K_s = 118.56 + 0.091 \cdot sand^2 - 5.87 \cdot sand \quad \text{R=0.90} \quad \text{(30-60 cm)}$$

$$K_s = 143.46 + 0.09 \cdot 5sand^2 - 6.32 \cdot sand \quad \text{R=0.87} \quad \text{(60-90 cm)}$$

$$K_s = 29.10 + \frac{1785.62}{silt} \quad \text{R=0.70} \quad \text{(0-30 cm)}$$

$$K_s = 45.51 + \frac{1165.05}{silt} \quad \text{R=0.84} \quad \text{(30-60 cm)}$$

$$K_s = 87.45 + \frac{799.99}{silt} \quad \text{R=0.83} \quad \text{(60-90 cm)}$$

$$K_s = 19.65 + \frac{1753.35}{clay} \quad \text{R=0.76} \quad \text{(0-30 cm)} \tag{5.1.1}$$

$$K_s = 24.66 + \frac{1312.45}{clay} \quad \text{R=0.89} \quad \text{(30-60 cm)}$$

$$K_s = 55.12 + \frac{1042.17}{clay} \quad \text{R=0.82} \quad \text{(60-90 cm)}$$

$$K_s = 50.94 + 318.15 \cdot f \quad \text{R=0.49} \quad \text{(0-30 cm)}$$

$$K_s = 110.78 + 422.02 \cdot f \quad \text{R=0.81} \quad \text{(30-60 cm)}$$

$$K_s = 75.51 + 390.0 \cdot f \quad \text{R=0.80} \quad \text{(60-90 cm)}$$

where: K_s - is the hydraulic conductivity coefficient in a saturated zone [cm day^{-1}], sand – the percentage content of the sand fraction, silt – the percentage content of the silt fraction, clay – the percentage content of the clay fraction and *f* – the grain size distribution index (Giesel, 1972) calculated as:

$$f = \frac{\sum_{i=1}^{k} \left(P_{i+1} - P_i\right) \cdot \frac{\log \frac{P_{i+1}}{Pi}}{\log \frac{S_{i+1}}{S_i}}}{P_k - P_i} \tag{5.1.2}$$

where: P_i – is the constant of *i-th* is grain fraction, which is read off the cumulative grain distribution curve and the smaller grain fractions, S_i – is the maximum diameter of the *i-th* grain fraction, k – is the quantity of grain fractions. Index f is uni-dimensional.

5.2. Model for estimation unsaturated hydraulic conductivity

In the Institute of Agrophysics Polish Academy of Sciences there is a database including the hydro-physical properties of the mineral arable soils in Poland, which also includes information concerning the parameters of soil solid phase [91,196].

The parameters of the soil solid phase and the values of the water conductivity coefficient juxtaposed in the database have been analysed statistically. All statistical analyses carried out, unless stated otherwise, were performed at a confidence level of 0.05.

Values of soil water potential for which statistical analyses have been carried out are as follows and correspond to pF:

0.1 [kJ m^{-3}] – pF 0;
1 [kJ m^{-3}] – pF 1;
3.16 [kJ m^{-3}] – pF 1.5;
9.8 [kJ m^{-3}] – pF 2;
16.0 [kJ m^{-3}] – pF 2.2;
31.0 [kJ m^{-3}] – pF 2.5;
50.0 [kJ m^{-3}] – pF 2.7;
100.0 [kJ m^{-3}] – pF 3.

In subsequent sections of this work, values of the soil's water potential expressed in pF units have been used in order to simplify notation.

The following labels have been adopted, which have been consequently used in the subsequent sections of this work:

K – is the water conductivity coefficient [cm day^{-1}];
F_{sand} – is the percentage content of the sand fraction [%];
F_{dust} – is the percentage content of the dust fraction [%];
F_{loam} – is the percentage content of the loam fraction [%];
D_{cz} – is the statistically average diameter of the particles [mm];
S_g – is the geometric specific surface of the soil particles [cm^2 g^{-1}];
ρ - is soil density [g cm^{-3}];
S_{BET} – is specific surface [cm^2 g^{-1}];
W_G – is the content of gravitational water [cm^3 cm^{-3}];
PPW – is the water content corresponding to the field water capacity (pF 2,2) [cm^3 cm^{-3}];
C_{org} – is the percentage content of organic coal [%];
P – is general porosity [cm^3 cm^{-3}].

Table 5.2.1. Correlation coefficients between values of hydraulic conductivity and analysed parameters of soil structure (bold marks coefficients statistically valid).

Parameter K	F_{il}	F_{pyl}	F_{piasek}	D_{cz}	S_g	P	ρ	W_G	PPW	S_{BET}	C_{org}
pF 0	**-0.11**	-0.07	**0.11**	**0.11**	**-0.14**	0.02	-0.09	**0.19**	**-0.15**	-0.07	**0.12**
pF 1	**-0.24**	-0.10	**0.21**	**0.21**	**-0.21**	0.01	-0.03	**0.26**	**-0.21**	**-0.18**	-0.07
pF 1,5	**-0.21**	-0.07	**0.17**	**0.17**	**-0.17**	-0.01	0.02	**0.22**	**-0.19**	**-0.16**	-0.05
pF 2	**-0.24**	-0.04	**0.17**	**0.18**	**-0.21**	-0.00	0.07	**0.21**	**-0.18**	**-0.23**	**-0.12**
pF 2,2	**-0.26**	0.01	0.16	**0.17**	**-0.26**	-0.00	**0.12**	**0.17**	**-0.18**	**-0.31**	**-0.16**
pF 2,5	**-0.17**	0.06	0.08	0.08	**-0.19**	**-0.11**	**0.17**	0.05	**-0.13**	**-0.29**	**-0.18**
pF 2,7	**-0.12**	**0.11**	0.01	0.02	**-0.15**	**-0.12**	**0.18**	-0.00	-0.08	**-0.27**	**-0.18**
pF 3	-0.04	**0.14**	-0.05	-0.0.5	-0.08	**-0.15**	**0.19**	-0.11	-0.02	**-0.21**	**-0.15**

Table 5.2.2. Correlation coefficients between values of logarithms of hydraulic conductivity and analyzed parameters of soil structure (bold marks coefficients statistically valid).

Parameter Log$_{10}$K	F_{il}	F_{pyl}	F_{piasek}	D_{cz}	S_g	P	ρ	W_G	PPW	S_{BET}	C_{org}
pF 0	**-0.21**	**-0.11**	**0.19**	**0.20**	**-0.22**	0.03	**-0.12**	**0.29**	**-0.23**	**-0.12**	0.07
pF 1	**-0.45**	**-0.21**	**0.40**	**0.41**	**-0.37**	0.05	-0.04	**0.46**	**-0.36**	**-0.31**	**-0.13**
pF 1,5	**-0.46**	**-0.17**	**0.39**	**0.40**	**-0.40**	0.01	0.03	**0.43**	**-0.36**	**-0.38**	**-0.17**
pF 2	**-0.42**	-0.07	**0.31**	**0.31**	**-0.38**	-0.01	0.09	**0.34**	**-0.30**	**-0.41**	**-0.22**
pF 2,2	**-0.39**	-0.01	**0.26**	**0.26**	**-0.38**	-0.05	**0.14**	**0.27**	**-0.27**	**-0.44**	**-0.25**
pF 2,5	**-0.27**	0.08	**0.13**	**0.14**	**-0.29**	-0.09	**0.17**	**0.13**	**-0.18**	**-0.40**	**-0.25**
pF 2,7	**-0.16**	**0.14**	0.02	0.03	**-0.21**	**-0.11**	**0.18**	0.02	-0.10	**-0.35**	**-0.23**
pF 3	-0.00	**0.20**	**-0.12**	**-0.11**	-0.08	**-0.13**	**0.18**	**-0.15**	0.03	**-0.23**	**-0.17**

The particle correlations of the values of the water conductivity coefficient and the decimal logarithms of these values at analysed soil water potential values with selected parameters of the soil solid phase are listed at the beginning of this work.

The results are illustrated in Tables 5.2.1 and 5.2.2 They show that the higher values of the correlation coefficient occur between decimal logarithms of the water conductivity coefficient's values and the parameters of the soil structure analysed. This must stem from the fact that due to the considerable diversity of soils in respect of their physical properties, values for the water conductivity coefficient also feature a wide range of changes at selected values of soil water potential.

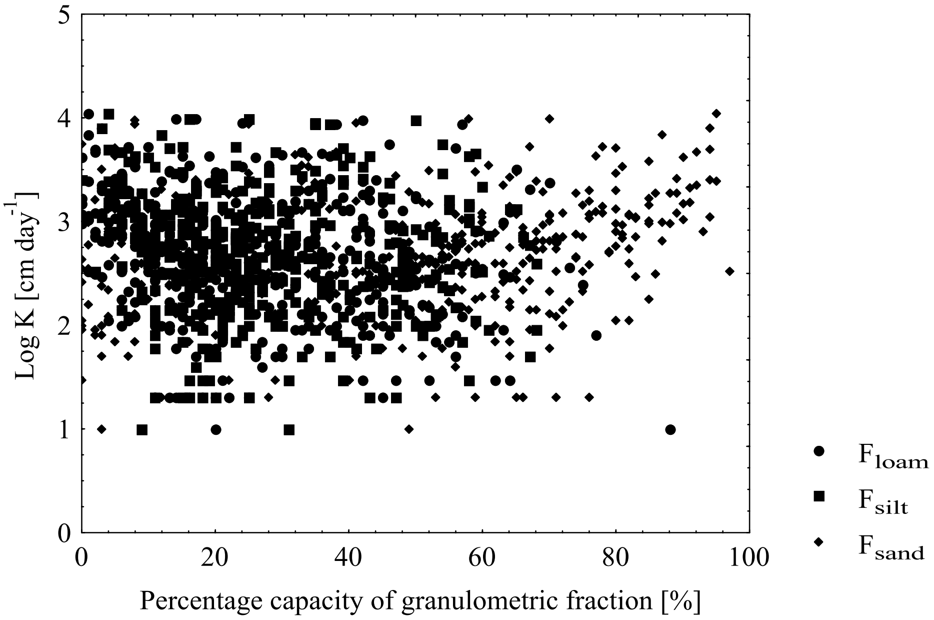

Fig. 5.2.1 Relationship between values of the logarithm of saturated hydraulic conductivity and percentage capacity of granulometric fractions.

Therefore it is the logarithmic values of the water conductivity coefficient that are often analysed. The low values of the coefficients of particle correlations

between the parameters of soil structure and the values of the logarithms of the water conductivity coefficient analysed also confirm the thesis about research material being greatly diversified when it comes to physical properties.

Relations between the water conductivity coefficient logarithm in a saturated zone and the percentage content of specific granulometric fractions have been illustrated in Fig. 5.2.1 as an example of this diversity.

In creating a specific, statistical-physical model, a certain subgroup of parameters was selected from the set of parameters illustrated, the following criteria were used in its choice: a relatively high coefficient of particle correlation between the parameters of the soil's constant phase selected to create the model and the value of the logarithm of the water conductivity coefficient at a given value of soil water potential as well as the lack of functional or correlation relations in the physical parameters of soil which were chosen to create the model. Selected functions of physical parameters of the soil, e.g. logarithms, inverse proportionality, etc. were also used for constructing correlation models.

The model of multiple linear regression was suggested as the first physical-statistical model. The general equation of such a model takes the following form:

$$Y = a_0 + a_1 X_1 + a_2 X_2 + ... + a_n X_n \tag{5.2.1}$$

where: Y – is a dependent (predicted) variable, $X_1, X_2,...X_n$, - are independent variables (predictors), whereas $a_0, a_1, a_2, ..., a_n$, – are the model's parameters. According to accepted criteria for statistical analysis, the following parameters of soil structure were selected: percentage content of loam fraction – F_{loam}, percentage content of sand fraction – F_{sand}, specific surface – S_{BET}, percentage content of organic soil – C_{org}, content of gravitational water – W_G, as well as water content at its potential, corresponding to field water capacity – *PPW.* A multiple regression analysis was carried out on the values of the logarithms of the water conductivity coefficient at defined values of the soil's water potentials, expressed in *pF* units and a selected set of physical parameters of the soil's constant phase. The correlation coefficients acquired for this model (Table 5.2.3) range from 0.33 to 0.52 and prove that the prediction capacity of these models is insignificant.

Multiple regression analysis carried out for other combinations of sets of parameters proved that the highest values of the correlation coefficient for all values of the soil's water potentials analysed are obtained for the sets of parameters selected above.

Table 5.2.3. Correlation coefficients of regression equations between logarithms of hydraulic conductivity and selected parameters of the soil's solid phase.

$Log_{10}K$	Correlation coefficient R
pF 0	0.33
pF 1	0.52
pF 1.5	0.52
pF 2	0.51
pF 2.2	0.52
pF 2.5	0.48
pF 2.7	0.46
pF 3	0.45

Consequently, regression methods were later sought to increase the predictive power of the models. An analysis of factor and multinomial regression was employed.

Models of complete factor regression are defined as systems featuring all possible independent products (predicators) of variables. For instance, the model of complete factor regressions for three independent variables X_1, X_2, X_3 includes not only major effects expressed with predictors X_1, X_2, X_3,,but also their two- and three-factor interaction. In this case, the complete equation of complete factor regression takes the following form:

$$\begin{aligned} Y = a_0 + a_1X_1 + a_2X_2 + a_3X_3 + a_4X_1X_2 \\ + a_5X_2X_3 + a_6X_1X_3 + a_7X_1X_2X_3 \end{aligned} \qquad (5..2.2)$$

Therefore, analysing factor regression facilitates research into the influence exerted by both specific parameters, as well as their products (interactions) on the value of the dependent variable.

The regression coefficients illustrated in Table 5.2.4 were obtained by analysing complete factor regression, carried out for the set of parameters of the soil's structure analysed (independent variables).

Table 5.245. Correlation coefficients of complete factor regression equations between the values of logarithms of water conductivity coefficient and selected parameters of the soil's solid phase.

$Log_{10}K$	Correlation coefficient R
pF 0	0.45
pF 1	0.58
pF 1.5	0.59
pF 2	0.60
pF 2.2	0.60
pF 2.5	0.57
pF 2.7	0.55
pF 3	0.51

The values of correlation coefficients obtained are only slightly better than the values obtained by means of multiple regression analysis. Therefore, it may be stated that the influence of specific products of independent variables on the calculation of the logarithmic values of the water conductivity coefficient is insignificant and the form of regression equations is fundamentally complicated. Consequently, it has been decided that illustrating them in this work is pointless, considering the fact that they comprise from 15 to 23 sections.

In a subsequent research stage, an analysis of multinomial regression was carried out, which is an example of a non-linear regression.. It allows the influence of powers which are higher than the first power of independent variables to be examined against the values of dependent variables In fact, a non-linear model of multinomial regression can be reduced to a linear model by means of the proper transformations of independent variables. Therefore, an analysis of multinomial regression of the second degree was carried out. The correlation coefficients obtained have been illustrated in Table 5.2.5. It must be pointed out that increasing the degree of the multinomial does not significantly influence the increase in the value of the correlation coefficient. Table 5.2.5 shows that the values of the correlation coefficient of multinomial regression equations are comparable with the values of these coefficients obtained for factor regression equations.

Table 5.2.5. Correlation coefficients of multinomial regression equations (2^{nd}. degree of a multinomial) between the values of decimal logarithms of the water conductivity coefficient and selected parameters of the soil's solid phase.

$Log_{10}K$	Correlation coefficient R
PF 0	0.48
PF 1	0.55
PF 1,5	0.59
PF 2	0.57
PF 2.2	0.59
PF 2.5	0.55
PF 2.7	0.52
PF 3	0.49

It may also be supposed that the influence of higher powers of soil structure parameters does not significantly influence the possibility of predicting the logarithmic values of the water conductivity coefficient.

None of the statistical analyses carried out above facilitated the building of a model which would give a satisfactory opportunity to predict the values of the water conductivity coefficient on the basis of knowing the parameters of the soil's solid phase. A natural conclusion would be that the relation of the values of the water conductivity coefficient to the parameters of the soil's solid phase are non-linear in character. Therefore, in the following stage of research, the focus was on searching for non-linear regression models, which would satisfactorily describe relations between the values of the water conductivity coefficient and the parameters of the soil's solid phase.

Non-linear estimation is a general fitting procedure which can be employed for estimating any kind of relations between a dependent (predicted) variable and independent variables. Generally, all regression models take this form:

$$Y = F(x_1, x_2, ..., x_n) \qquad (5.2.3)$$

where: Y – is a dependent variable expressed with any function F of independent variables $x_1, x_2, ..., x_n$.

A non-linear estimate facilitates the defining of any type of regression model. However if any type of relation between independent variables and a dependent variable should be allowed, questions may arise concerning the type of relations that make sense, i.e. how can they be interpreted and how can their relation be calculated precisely, i.e. how can the question of whether the expected relation actually occurs be concluded. Research has been carried out concerning the possibility of building a model to include some functions of the parameters of the soil's solid phase analysed e.g. logarithms, inversions or squares of inversions as independent variables. The correlation coefficients *R,* for regression equations obtained and determined in this way, ranged from 0.51 to 0.73, whereas the very form of these equations is complicated and its interpretation is difficult

Equations obtained from the above statistical analyses of regression equations feature a rather insignificant prediction capacity. Therefore, an analysis was carried out to check the possibility of building regression models for an analysed database introducing a division into classes on account of chosen physical properties of the soil's constant phase. The basic feature differentiating the soils studied in respect of their physical properties was their granulometric layout. Division into granulometric groups made up the basis for establishing the separate types of soils, so the analysed database was divided into four types (Table 5.2.6) so that in the first group there were soils classified as sandy (187 samples), in the second group, classified as loamy soils (115 samples), in the third group, classified as silty soils (86 samples), and in the fourth group, classified as clay soils (27 samples).

Table 5.2.6. Division of soil formations into granulometric groups [65].

Granulometric groups	sand 1-0.1 mm [%]	dust 0.1-0.02 mm [%]	loam <0.02 mm [%]
Sandy soils	40-100	0-40	0-20
Loamy soils	10-79	0-40	21-90
Silty soils	0-59	41-100	0-50
Clay soils	0-9	0-49	51-100

The regression analyses illustrated above have been carried out for each of these classes. The highest values of correlation coefficient R of regression equations that were obtained ranged from 0.37 to 0.83 and were acquired for individual regression methods and analysed values of soil water potentials. As can be seen, this method did not produce satisfactory results concerning the prediction capacity of sought after models either.

A non-linear regression model, so called ‘segmental regression’, was used for further analyses. The general form of this model is as follows:

$$\begin{aligned} Y &= A(b_{01} + b_{11}x_1 + b_{21}x_2 + \ldots + b_{m1}x_m) \\ &+ B(b_{02} + b_{12}x_1 + b_{22}x_2 + \ldots + b_{m2}x_m) \end{aligned} \tag{5.2.4}$$

providing that: *A=1* and *B=0* for $Y \leq PP$ and *A=0* and *B=1* for $Y > PP$, where *PP* is breaking point. Consequently, in this model, two separate linear regression equations are being assessed, one for the values of a dependent variable which are smaller or equal to (PP) breaking point, and the second for values that are higher than this point.

This segmental regression analysis was carried out under special conditions, so that the highest correlation coefficients could be obtained using the smallest possible number of physical parameters of the soils’ constant phase (independent variables) and also so that the structure parameters used for the model could exhibit a relatively high particle correlation coefficient and could be functionally independent. The segmental regression analysis carried out for the logarithms of the water conductivity coefficient proved that for regression equations featuring the following set of parameters: the percentage content of loam fraction – F_{loam}, the percentage content of sand fraction –F_{sand}, the specific surface – S_{BET}, the percentage content of organic coal – C_{org}, the content of gravitational water – W_G and the content of water at a potential corresponding to field water capacity– *PPW*, the values of the correlation coefficient which were obtained were $0.81 \leq R \leq 0.85$. The relatively high correlation coefficients that were obtained caused this model to be employed for individual types of soil. Correlation coefficients $0.86 \leq R \leq 0.96$ were obtained for the following set of parameters: the percentage content of loam fraction – F_{loam}, the percentage content of sand fraction – F_{sand}, the specific surface – S_{BET}, the percentage content of organic coal – - C_{org}, the content of gravitational water– W_G as well as the content of water at a potential corresponding to field water capacity – *PPW*. The general form of this equation’s model is as follows:

$$\begin{aligned} LogK &= A(a_0 + a_1 F_{loam} + a_2 F_{sand} + a_3 S_{BET} \\ &+ a_4 C_{org} + a_5 W_G + a_6 PPW) + \\ &B(b_0 + b_1 F_{loam} + b_2 F_{sand} + b_3 S_{BET} \\ &+ b_4 C_{org} + b_5 W_G + b_6 PPW \end{aligned} \tag{5.2.5}$$

where: *A=1* and *B=0* for $LogK \leq PP$ and *A=0* and *B=1* for $LogK > PP$.

Table 5.2.7. Coefficient of equation 5.2.5 for sandy soils.

Coefficient	a_0	a_1	a_2	a_3	a_4	a_5	a_6	PP
Log_{10} K	b_0	b_1	b_2	b_3	b_4	b_5	b_6	
pF 0	2.8974	-0.0006	-0.0051	0.0059	-0.1010	0.6611	-1.7706	2.7
	1.9913	0.0147	0.0118	0.0038	0.0468	0.6612	-1.3517	
pF 1	1.1425	-0.0027	-0.0039	-0.0035	-0.1576	0.2158	-0.7143	1.15
	3.5215	-0.0612	-0.0124	0.0049	-0.2705	-0.4811	0.5779	
pF 1.5	0.0840	0.0012	-0.0060	-0.0087	-0.1966	1.4222	1.1267	0.4
	2.1618	-0.0363	-0.0166	-0.0085	0.1963	0.1571	0.8982	
pF 2	-0.4448	-0.0054	-0.0041	-0.0054	-0.1609	0.1847	0.0601	-0.7
	0.7887	-0.0113	-0.0085	-0.0045	0.0733	-1.4526	-0.1640	
pF 2.2	-0.6860	-0.0004	-0.0065	-0.0136	0.0180	0.1712	-0.0569	-1.0
	-0.4282	0.0104	-0.0022	-0.0128	0.1496	-0.2580	-0.6285	
pF 2.5	-1.7011	0.0016	-0.0003	0.0002	-0.0332	-0.5508	-0.4244	-1.65
	-1.1552	0.0236	-0.0010	-0.0081	0.1217	-0.9439	-0.9656	
pF 2.7	-2.2776	0.0140	0.0003	-0.0066	0.0979	-0.9885	-0.3409	-2.15
	-1.4087	0.0087	-0.0041	-0.0082	0.0149	-0.5112	-0.2160	
pF 3	-2.1450	-0.0068	-0.0079	-0.0084	0.0836	-1.0038	-0.9920	-2.9
	-2.1131	0.0205	-0.0054	-0.0125	-0.1000	0.3651	-0.4095	

Table 5.2.8. Coefficient of equation 5.2.5 for loamy soils.

Coefficient	a_0	a_1	a_2	a_3	a_4	a_5	a_6	PP
Log_{10} K	b_0	b_1	b_2	b_3	b_4	b_5	b_6	
pF 0	2.0645	-0.0053	0.0013	-0.0009	-0.0572	-0.2332	1.5060	2.5
	4.6765	-0.0275	-0.0306	0.0002	0.0839	2.4754	0.0461	
pF 1	-0.2210	-0.0174	0.0010	0.0059	-0.2517	1.4633	1.5674	0.4
	1.1189	-0.0062	-0.0087	0.0020	0.0696	1.3962	-0.3045	
pF 1.5	-0.4800	-0.0122	-0.0017	0.0028	-0.3116	0.9388	0.7648	-0.4
	0.3746	0.0024	-0.0021	0.0031	-0.1367	0.7259	-1.5509	
pF 2	-1.4171	-0.0078	0.0041	0.0003	-0.2241	0.4248	1.2440	-1.0
	-1.0945	0.0004	0.0069	0.0020	-0.0688	1.0900	-0.1639	
pF 2.2	-1.9970	0.0030	0.0057	-0.0007	-0.0521	0.8186	0.3430	-1.3
	-1.0044	-0.0047	0.0010	0.0026	-0.1256	1.0019	0.0733	
pF 2.5	-2.0237	0.0035	-0.0009	-0.0029	-0.1126	0.4002	0.2134	-1.8
	-1.4776	0.0098	0.0034	-0.0030	-0.1591	-0.4661	-0.6933	
pF 2.7	-2.2900	-0.0037	-0.0039	-0.0014	-0.1415	0.5832	0.6030	-2.3
	-0.6864	-0.0043	-0.0075	0.0001	-0.4245	-0.4469	-1.6003	
pF 3	-2.2094	-0.0061	-0.0078	-0.0042	-0.3315	0.3102	0.6540	-2.7
	-1.1708	0.0072	-0.0082	-0.0047	-0.3810	0.2089	-2.2491	

Table 5.2.9. Coefficient of equation 5.2.5 for silty soils.

Coefficient	a_0	a_1	a_2	a_3	a_4	a_5	a_6	PP
Log_{10} K	b_0	b_1	b_2	b_3	b_4	b_5	b_6	
pF 0	1.6890	0.0068	0.0125	-0.0011	0.1251	-0.2235	0.6567	2.7
	2.2436	0.0114	0.0147	0.0028	0.1204	-0.2114	0.0426	
pF 1	1.2421	-0.0231	-0.0252	0.0035	-0.1982	2.3684	-0.0641	1.05
	1.2624	0.0253	0.0180	-0.0169	0.0388	-0.8633	-0.1711	
pF 1.5	0.8350	-0.0243	-0.0210	-0.009	-0.2978	3.0368	0.1899	0.6
	1.5734	0.0021	0.0003	-0.0066	-0.0368	-0.6883	-1.1336	
pF 2	-1.5179	0.0014	-0.0052	-0.0059	0.0202	2.1899	0.6596	-0.8
	-0.01623	-0.0130	-0.0108	-0.0066	-0.0763	2.1733	1.2488	
pF 2.2	-2.4709	0.0079	0.0075	-0.0143	-0.0696	3.0310	3.3245	-0.8
	-0.6176	-0.0092	-0.0067	-0.0019	-0.2504	1.5941	1.6359	
pF 2.5	-2.8995	0.0072	0.0085	-0.0080	-0.0499	2.1362	2.7317	-1.4
	-1.3194	-0.0041	0.0010	0.0023	-0.2557	1.9305	0.6382	
pF 2.7	-2.6922	0.0069	0.0069	-0.0081	-0.0337	0.9585	1.0575	-1.95
	-1.4429	-0.0004	-0.0004	-0.0015	-0.1310	0.2967	-0.3331	
pF 3	-3.0816	0.0073	0.0050	-0.0085	-0.0618	-0.1323	1.1010	-2.5
	-2.4199	0.0141	0.0061	-0.0055	0.0330	0.2650	-0.7541	

Table 5.2.10. Coefficient of equation 5.2.5 for clay soils

Coefficient	a_0	a_1	a_2	a_3	a_4	a_5	a_6	PP
Log_{10} K	b_0	b_1	b_2	b_3	b_4	b_5	b_6	
pF 0	0.5506	-0.0178	0.0596	-0.0063	-0.3724	2.7844	7.7332	2.6
	5.0276	0.0047	0.0215	-0.0170	0.5791	-3.9030	-3.4348	
pF 1	-2.6130	0.0017	0.0269	0.0108	-0.8212	5.3524	3.9058	0
	3.5972	0.0067	-0.0460	-0.0020	-0.6656	-0.6254	-5.3130	
pF 1.5	-2.0019	-0.0053	0.0063	0.0008	-0.3935	5.4730	2.8790	-0.6
	0.2254	-0.0435	0.0673	-0.0013	-0.8384	3.814	6.9157	
pF 2	-2.1790	-0.0030	-0.0162	-0.0069	-0.0306	5.3572	2.0561	-1.35
	-3.1208	-0.0240	-0.1006	-0.0077	0.6679	4.1315	10.1149	
pF 2.2	-1.6748	-0.0010	-0.0191	-0.0065	-0.0259	2.9821	1.5133	-1.7
	-3.4376	-0.0109	-0.0971	-0.0117	0.6292	0.4903	9.6291	
pF 2.5	-3.5971	0.0002	-0.0067	-0.0065	0.1744	5.3814	2.0592	-2.1
	-2.9011	-0.0210	-0.0891	-0.0121	0.4878	3.1791	7.7954	
pF 2.7	-3.4169	-0.0039	-0.0121	-0.0062	0.1843	3.4015	1.7321	-2.6
	-3.1701	-0.01106	-0.0847	-0.0167	0.5225	1.4274	7.0839	
pF 3	-3.7963	-0.0026	-0.0162	-0.0060	0.1589	2.3341	1.5532	-3.1
	-5.5628	-0.0028	-0.0656	-0.0221	0.0904	2.2514	13.1368	

Tables 5.2.7...5.2.10 show coefficients of the regression equations and the values of the breaking points obtained, whereas charts 5.2.2...5.2..5 illustrate the coefficients measured and calculated from the model of the values of water conductivity coefficient logarithms for individual types of soil and the values of the soil's water potential analysed.

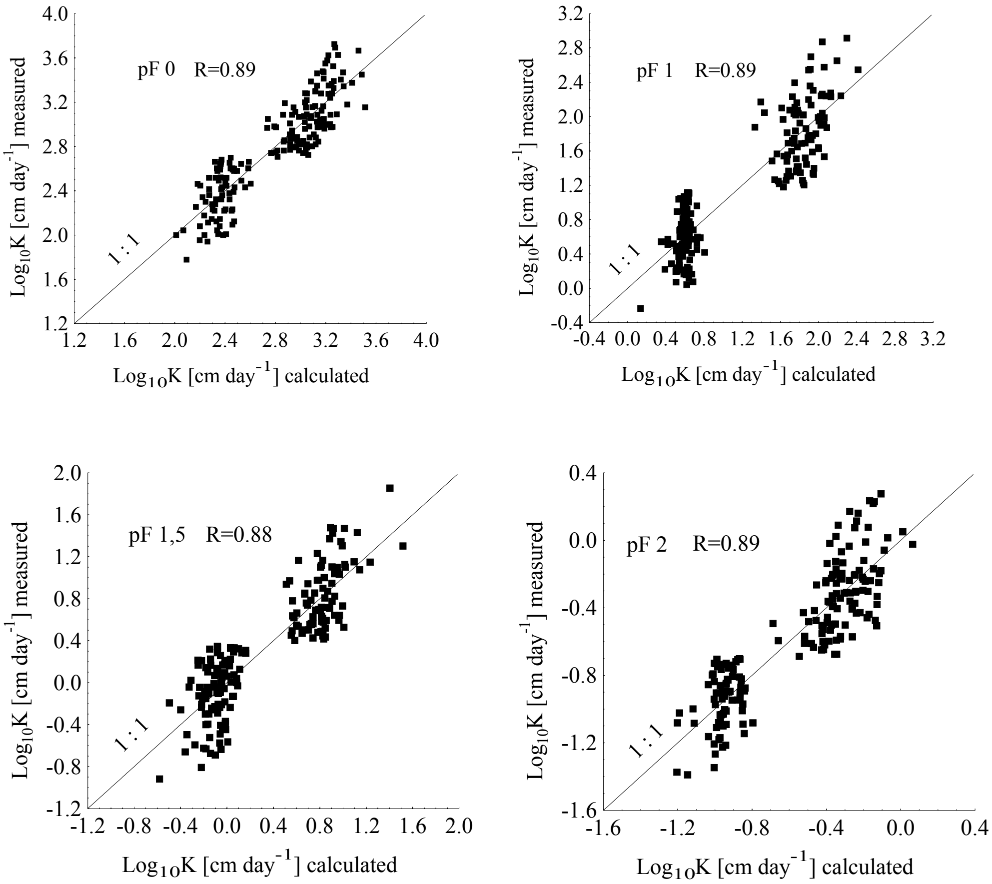

Fig. 5.2.2 Measured and calculated from the model of the logarithmic values of the water conductivity coefficient for sandy soils.

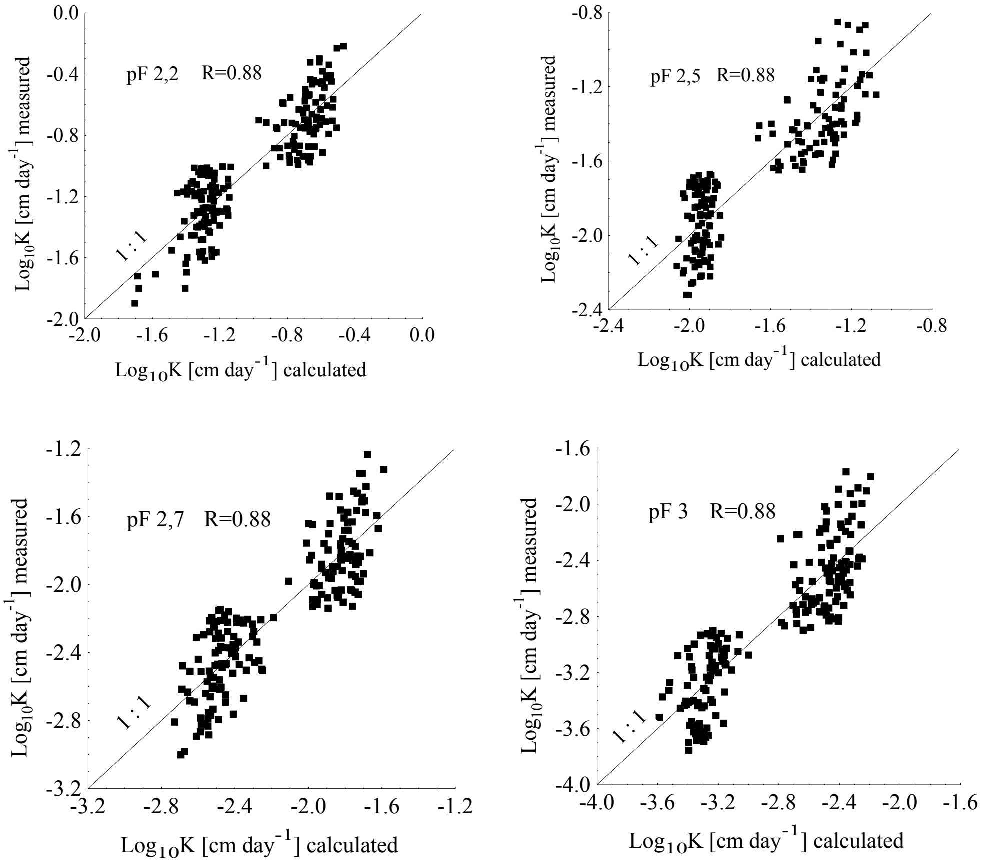

Fig. 5.2.2. Measured and calculated from the model of the logarithmic values of the water conductivity coefficient for sandy soils, continuation.

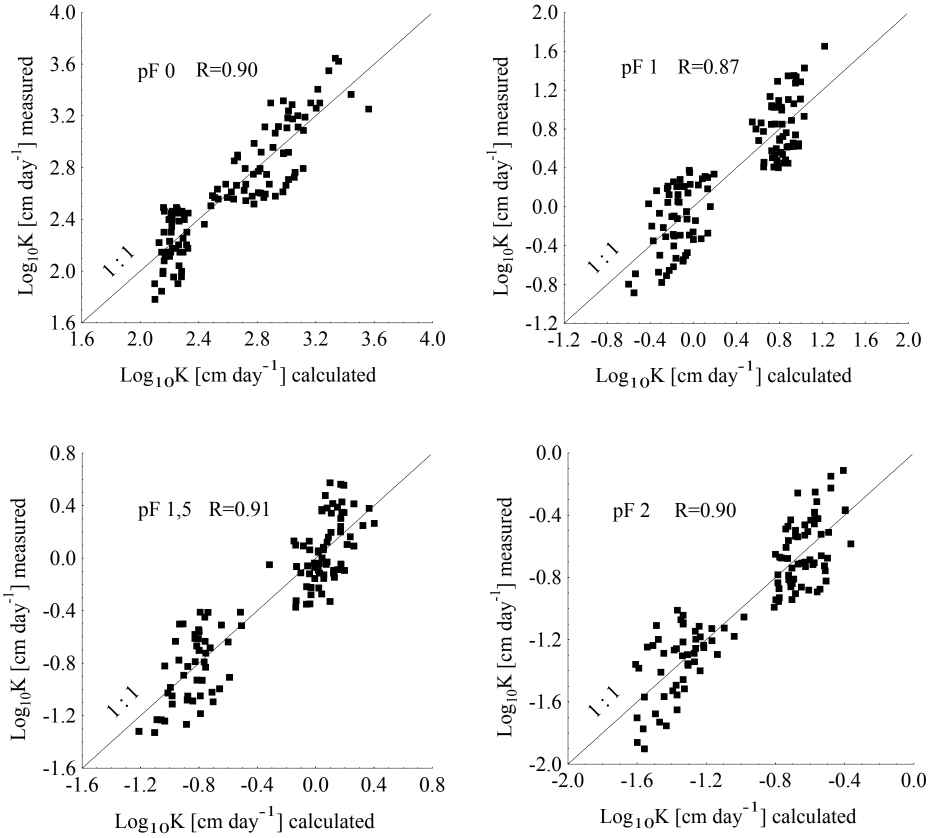

Fig. 5.2..3. Measured and calculated from the model of the logarithmic values of the water conductivity coefficient for loamy soils.

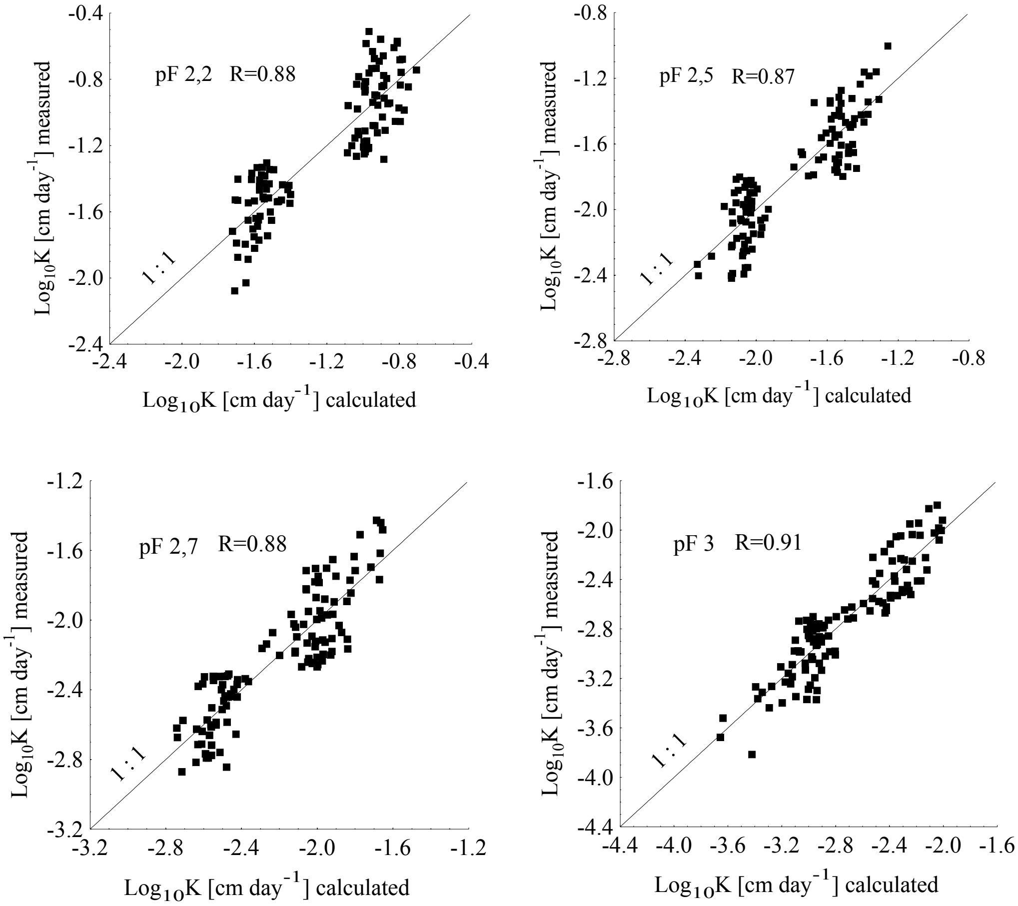

Fig. 5.2.3. Measured and calculated from the model of the logarithmic values of the water conductivity coefficient for loamy soils, continuation.

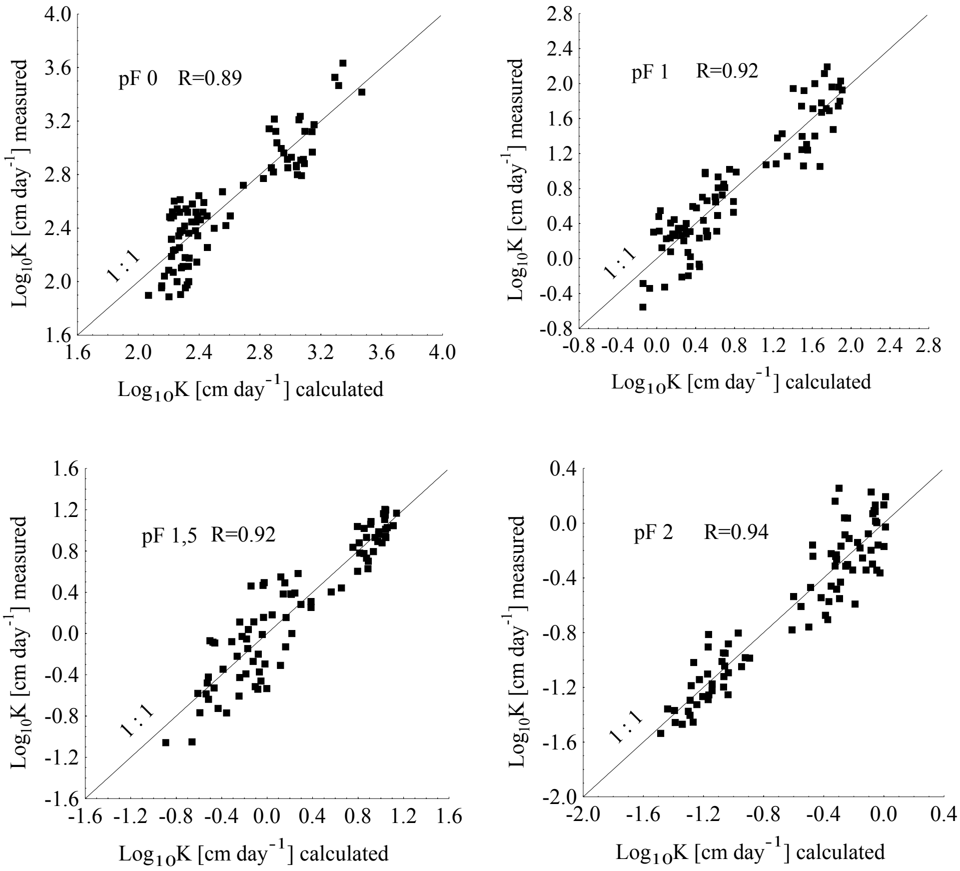

Fig. 5.2.4. Measured and calculated from the model of the logarithmic values of the water conductivity coefficient for silty soils.

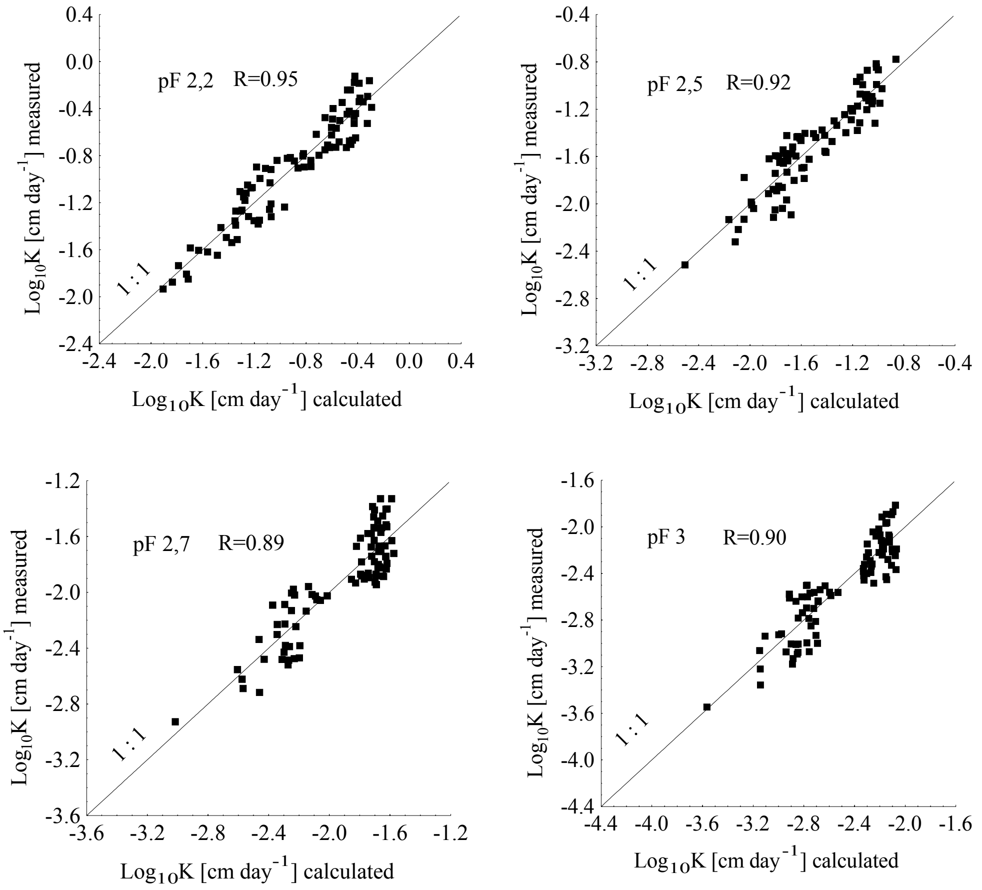

Fig. 5.2.4. Measured and calculated from the model of the logarithmic values of the water conductivity coefficient for silty soil, continuation..

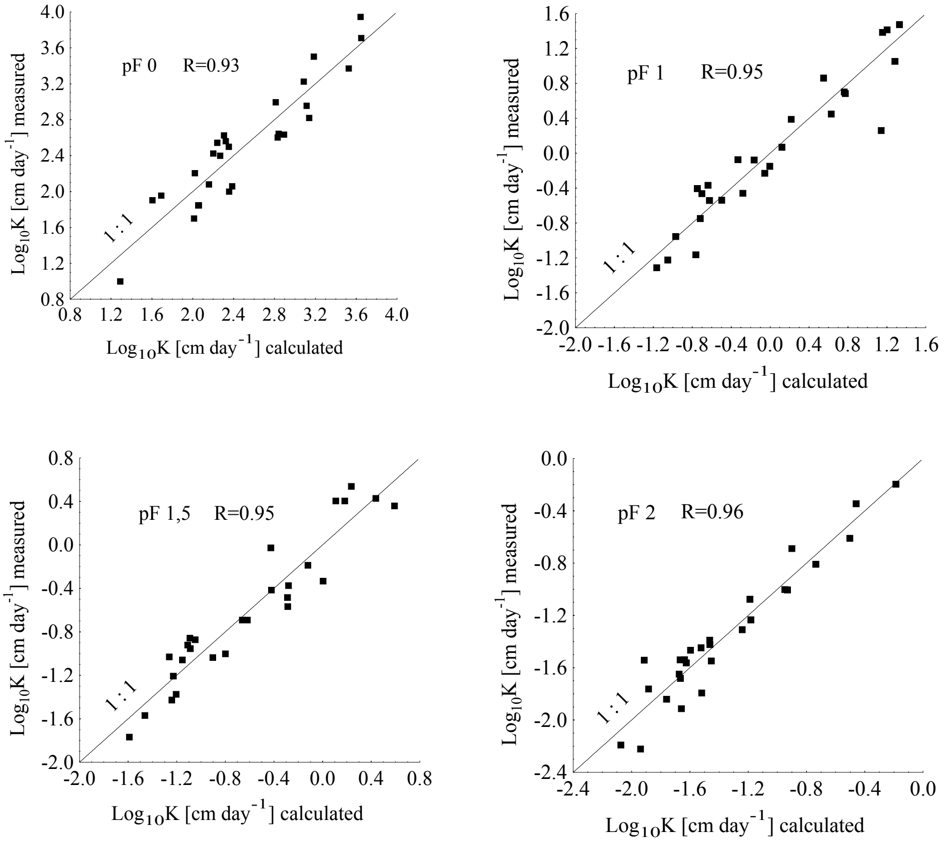

Fig. 5..2.5. Measured and calculated from the model of the logarithmic values of the water conductivity coefficient for clay soils.

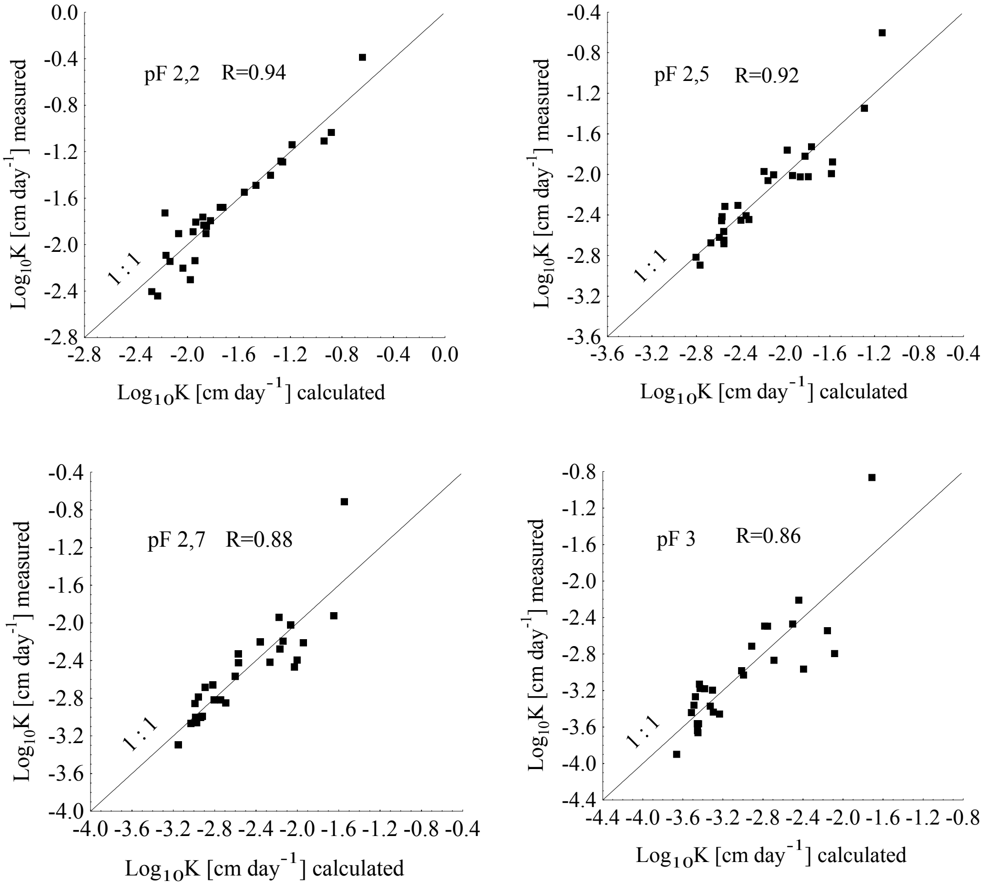

Fig. 5.2.5. Measured and calculated from the model of the logarithmic values of the water conductivity coefficient for clay soils, continuation.

5.3. Model validation

The physical-statistical model drafted was verified by means of the soil material that had not been used in its creation. Verification was carried out for every type of soil. Measurements of the water conductivity coefficient and of the parameters of the soil's solid phase were carried out according to the same methods as those presented in the chapter "Soil Material and Methodology of Research."

145 soil samples, including 45 sandy soils, 32 clay soils, 52 silty soils and 16 loamy soils were selected for verification. Tables 5.3.1...5.3.4 show the physical parameters of the soils employed for verification of the model.

The equation, according to which the value of the water conductivity coefficient should be calculated for a given type of soil and a given value of the soil's water potential, must be specified. This is necessary for verification of the model, i.e. using acquired segmental regression equations. Consequently whether a given value for the water conductivity coefficient lies below the value of the *PP* breaking point or above it, must be determined.

Table 5.3.1. Physical parameters of sandy soils used for verification of the model.

No.	Granulometric group	F_{sand}	F_{loam}	S_{BET}	C_{org}	W_g	PPW
1.	Bwps:pl	89	6	8.6	1.06	0.37	0.09
2.	Bwps.pl	74	7	12.9	0.58	0.44	0.11
3.	Bwps:pl	84	7	47.1	2.28	0.41	0.12
4.	Dzpgl.pl	80	10	10.3	0.36	0.36	0.10
5.	Apgl.gl	66	17	7.1	0.79	0.14	0.20
6.	Apgl.gl	70	19	45.8	0.44	0.12	0.21
7.	Fpgl.ps	72	11	13.9	0.34	0.23	0.15
8.	Fbpglp.ps	52	18	19.4	0.18	0.23	0.16
9.	Bpgm.gl	49	17	22.5	1.18	0.21	0.24
10.	Bpgl.ps	57	19	15.8	0.81	0.20	0.17
11.	Bpgl:gl	65	11	10.6	0.64	0.14	0.26
12.	Bpgl:pl	66	14	17.3	0.88	0.14	0.23
13.	Bpgm:gl	57	17	9.2	0.80	0.22	0.16
14.	Apgm50gl	56	17	6.0	0.67	0.10	0.21
15.	Rbpgm-gs.sk	66	18	25.8	1.08	0.14	0.25
16.	Bps:pl	87	4	11.3	0.88	0.26	0.10
17.	Bps.pl	83	9	12.9	0.81	0.23	0.11
18.	Bps:pl	92	2	35.2	1.00	0.38	0.05

19.	Fpłz:gl	53	17	34.1	0.33	0.21	0.25
20.	Apgl.gl	62	11	6.7	0.80	0.18	0.16
21.	Bwps.pl	88	6	7.5	0.29	0.33	0.09
22.	Bpgm.gl	62	16	19.1	0.71	0.22	0.18
23.	Dpgl.pl	68	9	6.3	0.11	0.29	0.12
24.	Bps.pl	89	3	6.5	0.87	0.31	0.08
25.	Bwpgl.gl	75	11	9.1	0.81	0.22	0.12
26.	Bps80pl	89	4	5.0	0.97	0.28	0.09
27.	Bpgl:gl	91	3	7.6	0.95	0.32	0.10
28.	Apgl50gl	74	11	9.6	0.71	0.20	0.15
29.	Dpgl.pl	87	3	32.8	2.05	0.26	0.25
30.	Bps:pl	88	6	16.0	0.86	0.32	0.13
31.	Bpgl70gl	66	11	7.8	0.67	0.31	0.14
32.	Bpgl60gl	80	11	2.7	0.77	0.23	0.14
33.	Bpgm.gl	59	16	14.8	0.73	0.18	0.22
34.	Bwps.pl	90	1	7.3	0.95	0.31	0.14
35.	Bps:pl	94	3	8.0	0.34	0.26	0.15
36.	Bpgm.gl	85	6	5.0	0.38	0.16	0.22
37.	Bwps:pl	70	6	16.1	0.81	0.28	0.08
38.	Mps.pl	76	4	5.8	0.60	0.28	0.10
39.	Apłz:gl	44	16	9.2	0.59	0.09	0.27
40.	Mps.pl	89	1	15.5	1.48	0.26	0.18
41.	Bpgl60gl	70	11	5.3	0.52	0.20	0.22
42.	Bps:pl	95	1	12.5	0.82	0.32	0.12
43.	Bwpgl.ps	75	1	10.5	0.81	0.31	0.14
44.	Bpgl:gl	66	11	5.3	0.78	0.34	0.18
45.	Bps.pl	72	14	10.8	0.79	0.25	0.17

Table 5.3.2. Physical parameters of loamy soils used for verification of the model.

No.	Granulometric group	F_{sand}	F_{loam}	S_{BET}	C_{org}	W_g	PPW
1.	Rb	57	23	26.6	0.768	0.1716	0.329
2.	Rb	61	24	33.7	0.312	0.24	0.225
3.	Fpłi	38	28	32.7	1.14	0.2415	0.209
4.	Rb	49	29	39	1.62	0.1788	0.268
5.	Rb	53	36	103	0.225	0.1548	0.224
6.	Rc	27	54	115	0.46	0.1708	0.403
7.	Rc s.sk	28	48	80.8	1.78	0.197	0.369
8.	Rc c.sk	30	30	53.2	1.02	0.099	0.426
9.	Bgl	46	33	36.4	0.99	0.0925	0.2595
10.	Rcgs.sk	30	39	80.7	2.61	0.0572	0.4288
11.	Bgs.gc	29	52	63.7	1.1	0.0668	0.3672
12.	Bgc:sk	27	45	45.6	1.39	0.147	0.338
13.	Bgsp:gcp	21	49	45.3	0.94	0.1134	0.3076
14.	Dpłi	25	36	58.5	1.5	0.15	0.285
15.	Fglpłz	13	50	53	1.05	0.0664	0.3486
16.	Bli	36	36	33.3	0.79	0.102	0.272
17.	Bpgm.gl	42	35	38.9	0.8	0.1062	0.3008
18.	Rb pgm-gs.sk	22	62	151	0.41	0.0406	0.1824
19.	Dglp	53	25	7.5	0.09	0.0955	0.2485
20.	Rc gs.sk	28	59	94.1	0.71	0.0762	0.4958
21.	Fglp	33	28	35.2	0.52	0.1332	0.3108
22.	Bgl	51	23	26.7	0.64	0.2325	0.2325
23.	Bgl	59	22	25.5	0.11	0.1274	0.2186
24.	Rbgs.gcp:sk	36	30	18.5	0.16	0.1068	0.2852
25.	Rbpgm-gs.sk	66	23	21.3	0.56	0.2736	0.2014
26.	Rc	28	49	93.3	0.426	0.192	0.348
27.	Rbpgm-gs.sk	36	48	125	0.66	0.165	0.561
28.	Rbpgm-gs.sk	46	43	71.9	0.38	0.1001	0.4289
29.	Rbpgm-gs.sk	68	21	45.4	1.41	0.364	0.208
30.	Rbpgmp	31	56	84.1	0.12	0.221	0.338
31.	Rcs	19	63	60.6	0.61	0.2318	0.378
32.	Bgs.gc	38	37	41.6	0.82	0.0728	0.291

Table 5.3.3. Physical parameters of silty soils used for verification of the model.

No.	Granulometric group	F_{sand}	F_{loam}	S_{BET}	C_{org}	W_g	PPW
1.	Apłz.gl	38	15	18	1.12	0.147	0.338
2.	Apłz.gl	30	17	13.5	1.09	0.2736	0.228
3.	Apłz.gl	20	21	8.5	0.67	0.14	0.297
4.	Bli	8	27	21.2	0.71	0.1924	0.311
5.	Cli	10	29	37.6	2.1	0.2622	0.331
6.	Bwli	1	31	20.4	0.97	0.1029	0.367
7.	Dpłi	8	32	2.0	0.76	0.16	0.32
8.	Bli	5	35	17.3	0.81	0.204	0.286
9.	Bwli	1	36	19.3	0.88	0.145	0.348
10.	Bli	2	37	31.3	0.64	0.1904	0.354
11.	Bwli	4	38	15.9	0.73	0.1216	0.35
12.	Bwli	1	40	25.0	0.76	0.2025	0.337
13.	Bwli	4	40	39.0	0.72	0.1859	0.315
14.	Cli	2	43	34.0	0.62	0.2596	0.33
15.	Cli	5	43	31.7	1.44	0.1905	0.317
16.	Bwli	1	46	23.5	0.73	0.0775	0.372
17.	Dpłi	3	47	5.6	1.97	0.1755	0.351
18.	Bwli	3	47	36	0.71	0.21	0.294
19.	Bwli	1	48	41.2	0.89	0.1742	0.348
20.	Cli	0	32	35.9	1.06	0.2604	0.3226
21.	Fgsp	8	45	54.8	0.59	0.112	0.32
22.	Bgc:sk	6	49	40.2	0.87	0.0624	0.3586
23.	Fgsp	6	38	46.3	1.28	0.0157	0.3143
24.	Fpłi	13	45	53.1	0.72	0.1127	0.3223
25.	Bli	3	49	37.3	0.87	0.128	0.288
26.	Adłx.gl	30	23	16.4	0.89	0.086	0.31
27.	Agsp.gcp (ip)	25	30	23.3	0.9	0.19	0.24
28.	Dpgl.pl	41	11	27.0	1.45	0.1704	0.3266
29.	Bpgm:pgl	43	12	61.8	0.98	0.2067	0.2063
30.	Fbpglp.ps	34	13	25.1	0.77	0.0725	0.2035
31.	Apłz.gl	20	21	8.5	0.7	0.14	0.298
32.	Bwpgl:gl	41	10	6.5	0.8	0.198	0.162
33.	Bli	0	45	46.0	0.8	0.1386	0.3694
34.	Bpłz	9	23	15.3	0.79	0.112	0.336
35.	Bgsp	16	36	32.2	1.24	0.077	0.323

36.	Bpłi	1	32	26	1.15	0.132	0.356
37.	Agsp.gl	6	48	30.4	0.85	0.054	0.36
38.	Cli	9	25	36.5	2.07	0.208	0.312
39.	Bpłz	6	35	52.1	0.9	0.09	0.3
40.	Bli	4	46	31.7	0.92	0.1668	0.3202
41.	Bpłz	12	30	33.1	0.67	0.048	0.288
42.	Bli	0	42	29.6	0.88	0.08	0.34
43.	Bgc:sk	4	32	69.5	2.31	0.0768	0.3842
44.	Fpłi	5	40	52.8	1.67	0.2	0.313
45.	Bip:sk	4	38	66.7	1.85	0.0333	0.3667
46.	Bgs.gc	21	37	67.3	2.64	0.13	0.32
47.	Bgs.sk	34	24	34.1	1.53	0.0966	0.3174
48.	Bbgsp	38	19	40.8	1.85	0.139	0.375
49.	Cli	0	27	40.1	1.13	0.172	0.372
50.	Bwli	2	36	19.7	0.80	0.185	0.355
51.	Cli	3	45	45.6	0.32	0.115	0.393
52.	Cli	1	41	51.4	0.79	0.219	0.302

Table 5.3.4. Physical parameters of clay soils used for verification of the model.

No.	Granulometric group	F_{sand}	F_{loam}	S_{BET}	C_{org}	W_g	PPW
1.	Bwli	1	52	54.2	0.65	0.1099	0.377
2.	Rcgs.sk	2	71	78.7	0.45	0.144	0.419
3.	Fip	2	79	78.6	0.82	0.083	0.473
4.	Fgsp	6	52	86.7	0.98	0.165	0.345
5.	Bglp:gsp	3	88	75.2	0.9	0.1272	0.3338
6.	Bip	2	77	83.9	1.95	0.0417	0.4453
7.	Bpłi	4	57	69.7	0.7	0.068	0.374
8.	Bpłi	1	51	46.0	0.76	0.1113	0.3337
9.	Rcgs.sk	9	75	75.1	1.05	0.2033	0.3637
10.	Fc/Fcgzp	8	56	106	2.04	0.0592	0.5178
11.	Rcgs.sk	0	52	88	1.63	0.1428	0.3692
12.	ZFc/Fpi	6	52	104	2.11	0.0314	0.4706
13.	Fpłi	1	63	107.5	1.86	0.1161	0.3869
14.	Fpłi	4	50	54.8	0.62	0.1332	0.3408
15.	Bgs.sk	4	70	54.5	1	0.081	0.365
16.	Fpłi	3	65	74.3	0.84	0.0608	0.3492

In order to achieve this aim, multiple regression equations have been drafted for each soil type and for each value of the soil water potential analysed. Their general form is as follows:

$$\begin{aligned} LogK &= c_0 + c_1 F_{loam} + c_2 F_{sand} + c_3 S_{BET} \\ &+ c_4 C_{org} + c_5 W_G + c_6 PPW \end{aligned} \qquad (5.3.1)$$

The coefficients of these equations as well as the coefficients of the correlation are illustrated in Tables 5..3.5...5.3.8 Using these equations, one may generally assess the value of the logarithm of the water conductivity coefficient for a given type of soil and a given value of the soil's water potential. A value obtained in this way allows one to determine which segmental regression equation should be used depending on whether this value is smaller or higher than the breaking point. The calculation method may be shown in the form of the algorithm illustrated in fig. 5.3.1.

Table 5.3.5. Coefficient of equation 5.3.1 for sandy soils.

Coefficient Log$_{10}$ K	c_0	c_1	c_2	c_3	c_4	c_5	c_6	R
pF 0	2.9068	-0.0055	0.0018	0.0045	0.0491	1.0280	-3.2526	0.56
pF 1	0.7614	-0.0261	-0.0015	-0.0082	0.0429	2.3803	1.5793	0.30
pF 1.5	-0.1843	-0.0150	-0.0031	-0.0110	0.0404	2.3796	2.6835	0.28
pF 2	-0.4552	-0.0043	-0.0053	-0.0186	0.0411	0.6492	2.3627	0.37
pF 2.2	-0.6111	-0.0002	-0.0085	-0.0168	0.0128	0.5508	1.8692	0.42
pF 2.5	-0.9958	0.0161	-0.0072	-0.0155	-0.0629	-0.5878	0.4825	0.53
pF 2.7	-1.2875	0.0223	-0.0080	-0.0167	-0.0490	-0.9093	-0.1036	0.58
pF 3	-1.3561	0.0240	-0.0122	-0.0150	-0.0815	-1.7238	-1.0625	0.61

Table 5.3.6. Coefficient of equation 5.3.1 for loamy soils.

Coefficient Log$_{10}$ K	c_0	c_1	c_2	c_3	c_4	c_5	c_6	R
pF 0	2.8396	-0.0207	-0.0093	-0.0020	0.0173	-0.1413	3.0672	0.35
pF 1	-1.3668	-0.0044	0.0172	0.0074	-0.0373	2.8944	1.7073	0.45
pF 1.5	-1.1247	-0.0121	0.0119	0.0049	-0.2278	2.1676	1.7717	0.50
pF 2	-1.4857	-0.0111	0.0071	0.0028	-0.1979	2.8384	1.0539	0.55
pF 2.2	-1.6252	-0.0098	0.0070	0.0020	-0.2938	2.2468	1.0117	0.57
pF 2.5	-1.4542	-0.0037	-0.0028	-0.0032	-0.02999	0.8493	0.6675	0.48
pF 2.7	-1.8031	-0.0084	-0.0009	-0.0023	-0.3454	0.7141	0.7789	0.50
pF 3	-1.4674	-0.0130	-0.0095	-0.0038	-04382	0.1963	0.1884	0.52

Table 5.3.7. Coefficient of equation 5.3.1 for silty soils.

Coefficient Log_{10} K	c_0	c_1	c_2	c_3	c_4	c_5	c_6	R
pF 0	1.5850	0.0041	0.0101	-0.0014	0.3674	0.5151	1.0492	0.53
pF 1	0.8618	-0.0289	-0.0147	-0.0021	-0.2281	6.2694	1.4039	0.59
pF 1.5	-0.2772	-0.0176	-0.0127	-0.0124	-0.1172	5.9737	2.7528	0.70
pF 2	-1.1931	-0.0046	-0.0054	-0.0144	-0.0862	4.8574	2.1954	0.73
pF 2.2	-1.3489	-0.0033	-0.0062	-0.0159	- 0.1011	3.6787	2.1671	0.72
pF 2.5	-1.7242	0.0033	-0.0012	-0.0106	-0.1715	2.0635	1.0708	0.69
pF 2.7	-2.0963	0.0089	0.0062	-0.0105	-0.2279	1.1735	0.7440	0.69
pF 3	-2.7046	0.0094	0.0041	-0.0127	-0.2525	0.3621	1.5518	0.62

Table 5.3.8. Coefficient of equation 5.3.1 for clay soils.

Coefficient $\log_{10} K$	c_0	c_1	c_2	c_3	c_4	c_5	c_6	R
pF 0	3.0984	-0.0146	0.0182	-0.0037	0.2577	-0.1350	0.6014	0.34
pF 1	0.1459	-0.0235	-0.0650	-0.0120	-0.5788	10.84	5.3248	0.67
pF 1.5	-0.2900	-0.0208	-0.0418	-0.0092	-0.4956	7.8611	4.0037	0.71
pF 2	-0.4674	-0.0153	-0.0388	-0.0114	-0.3094	4.8818	2.2841	0.71
pF 2.2	-0.5651	-0.0135	-0.0351	-0.0100	-0.2690	3.0038	1.4156	0.66
pF 2.5	-0.6531	-0.0119	-0.0320	-0.0151	-0.1446	1.6644	1.1092	0.66
pF 2.7	-0.9180	-0.0073	-0.0277	-0.0161	-0.0746	0.3975	0.4485	0.60
pF 3	-1.3418	-0.0001	-0.0219	-0.0175	0.0262	-1.4564	0.5138	0.53

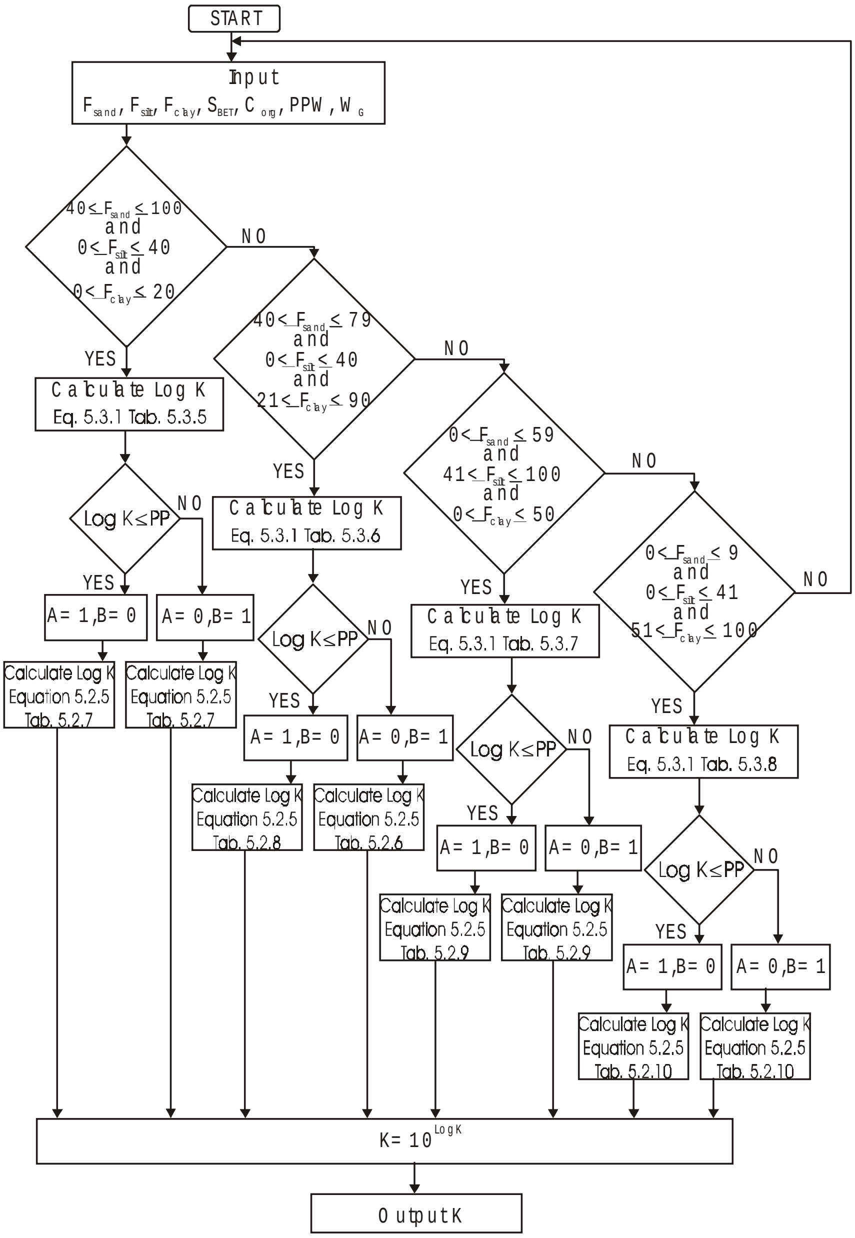

Fig. 5.3.1. Algorithm for calculating the water conductivity coefficient.

A comparison of the values of water conductivity coefficient logarithms for individual soil types which were measured and calculated by means of the suggested model has been illustrated in figures 5.3.2...5.3.5.

Figure 5.3.6 shows a comparison of the values of the water conductivity coefficient logarithms which were measured and calculated for the whole of the soil material studied.

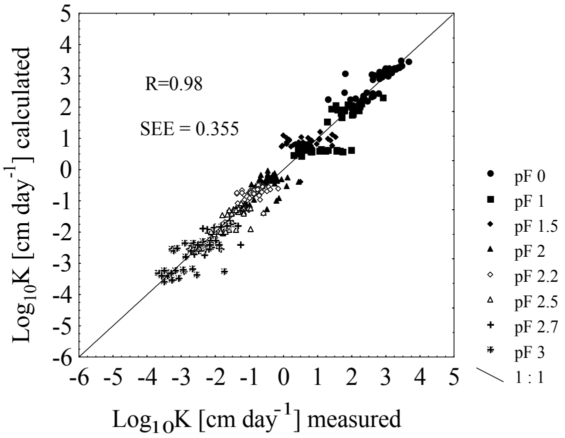

Fig. 5.3.2. Verification of the model: values of logarithms of the water conductivity coefficient measured and calculated from the model for sandy soils.

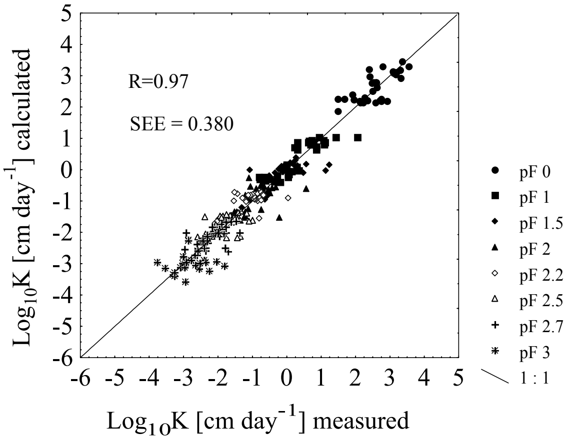

Fig. 5.3.3. Verification of the model: values of logarithms of the water conductivity coefficient measured and calculated from the model for loamy soils.

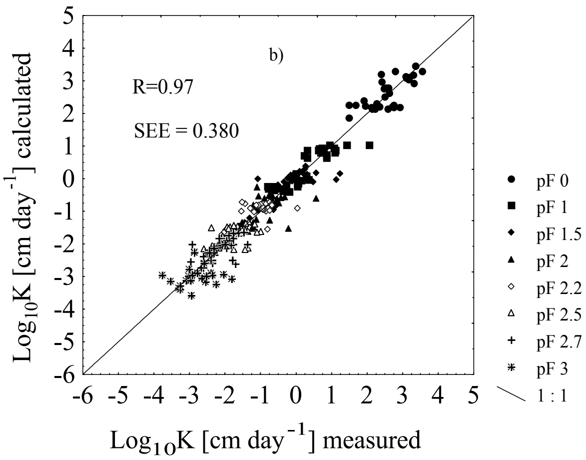

Fig. 5.3.4 Verification of the model: values of logarithms of the water conductivity coefficient measured and calculated from the model for silty soils.

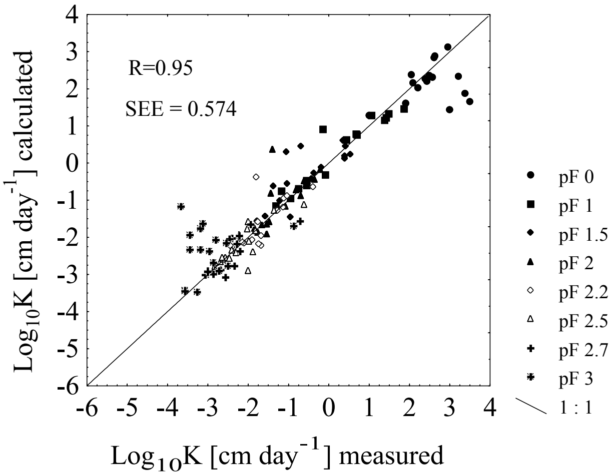

Fig. 5.3.5. Verification of the model: values of logarithms of the water conductivity coefficient measured and calculated from the model for clay soils.

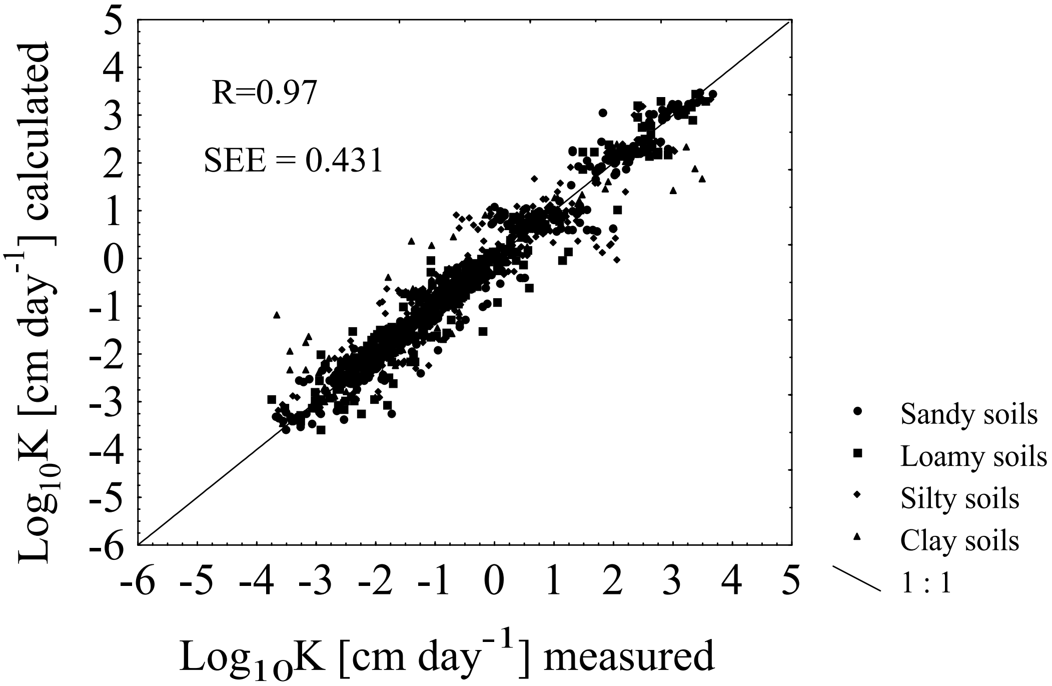

Fig. 5.3.6. Verification of the model: values of logarithms of the water conductivity coefficient measured and calculated from the model.

6. DETERMINATION OF HYDRAULIC CONDUCTIVITY FOR CHOOSEN SOIL MATERIALS

6.1. Hydraulic conductivity of peat and sand mixtures

The effect of mineral soil addition on hydrophysical properties of the peat soil was studied in laboratory experiments[195]. As the model systems of soil samples which correspond to the different stages of peat layer enrichment (intermixing) with mineral materials, the mixtures on the base of peat and sand have been used. The following materials have been used :

- the shallow dried peat soil from a typical landscape of Polesye (Rogóźno, the Lublin Region), formed on

the sedge peat, medium degree of decomposition (35-40%), ash content 42.6%, pH_{KCl} =4,6;

- medium quartz sand , organic matter content 0,1% , pH_{KCl} =4,0.

The determination of the organic matter content in the sand was carried out by Tiurin method. Ash content in peat soil was determined by ignition of the dried peat soil samples in a muffle furnace at about 550^{O} C until their weight was constant .The ash content was expressed in terms of the percentage ignition residue from the quantity of dry material. The soil samples were prepared by hand mixing of the fixed quantities of peat and sand material and as a result, the mass parts (which numerically equal the ratio of the dry peat mass to the whole sample mass, in %) in the peat –sand mixtures were 5%, 20%, 40%, 60% and 80%. The physical properties of the investigated samples are given in table 1.

The total porosity is considered to be equal to the water content at saturation. The content of different size pores has been determined from water retention curves received using the pressure chambers, made by Soil Moisture Equipment Corp., Santa Barbara, USA. The border values between macropores and mezopores were taken as 30 μm (pF 2) and between mezopores and micropores as 0.2 μm (pF 4.2) .

Unsaturated hydraulic conductivity was determinated using a TDR (Time Domain Reflectometry) device of water content, temperature and salinity measurements and instantaneous profile method IPM .

Hydraulic conductivity is affected by the soil structure as well as by its texture. It depends not only on the total porosity but also on the sizes of conducting pores. The hydraulic conductivity of the peat soils depends on a nature of peat, decomposition degree, ash content, bulk density and porosity. In peat materials, the particle size, the structure, and the resulting porosity are determined from the state of decomposition. Hydraulic conductivity varies greatly for different peat types and organic soils. The increase

Table 6.1.1. Hydrophysical properties of peat, sand and their mixtures

Sample number	Composition	Organic matter	Mineral matter	Bulk density [g cm^{-3}]	Total porosity [%]	Content of different sizes pores, [%, m^3 m^{-3}]			K_{sat} [cm day^{-1}]
		Content in % dry matter				Macropores > 30 μm	Mezopores 30-0.2 μm	Micropores < 0.2 μm	
1.	Peat	57.4	42.6	0.33	90	37	9	44	$3.40\ 10^3$
2.	80% of peat+ sand	45.9	54.0	0.41	88	37	8	43	$5.53\ 10^3$
3.	60% of peat+ sand	34.5	65.5	0.51	87	38	7	42	$1.83\ 10^3$
4.	40% of peat+ sand	23.0	77.0	0.68	84	38	7	39	$3.09\ 10^3$
5.	20% of peat+ sand	11.6	88.4	1.05	75	40	5	30	$8.10\ 10^2$
6.	5% of peat+ sand	3.0	97.0	1.57	55	39	3	13	$1.60\ 10^3$
7.	Sand	0.1	99.9	1.86	38	31	2	5	$1.60\ 10^3$

of mineral part in peat-sand mixtures leads to the increase of bulk density and decrease of total porosity as well as to the change of the content of different size pores (Table6.1.1).

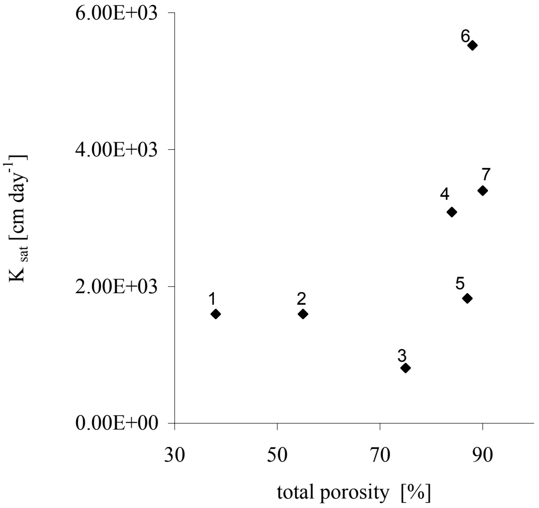

Fig. 6.1.1 Relation between saturated hydraulic conductivity and total porosity of soil material investigated.

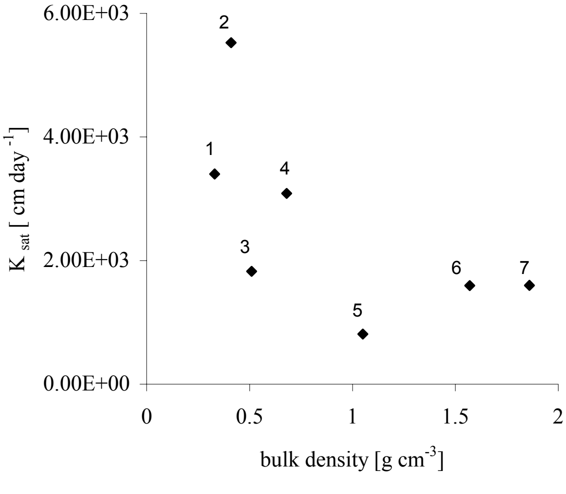

Fig. 6.1.2. Relation between saturated hydraulic conductivity and bulk density of soil material investigated.

The relationships between the soil hydraulic conductivity and total porosity and bulk density for the investigated samples are shows in Fig. 6.1.1 and 6.1.2.

The analysis of the received data has not revealed an essential dependence between the sand content in peat-sand mixtures and the hydraulic conductivity coefficient at saturation. The saturated hydraulic conductivity of investigated systems changes from 1.8 102 cm day-1 for sample 6 to 5.5 103 cm day-1 for sample 2. It should be noticed that the total porosity decrease and the bulk density increase lead to the decrease of saturated hydraulic conductivity (Fig.6.1.1 and 6.1.2). All the peat–sand mixtures, except sample 2, have lower values of the saturated hydraulic conductivity as compared with peat. That makes us suppose, that addition of sand to peat sedge in quantity higher than 20%, leads to deterioration of hydraulic conductivity of the sedge peat at saturated zone.

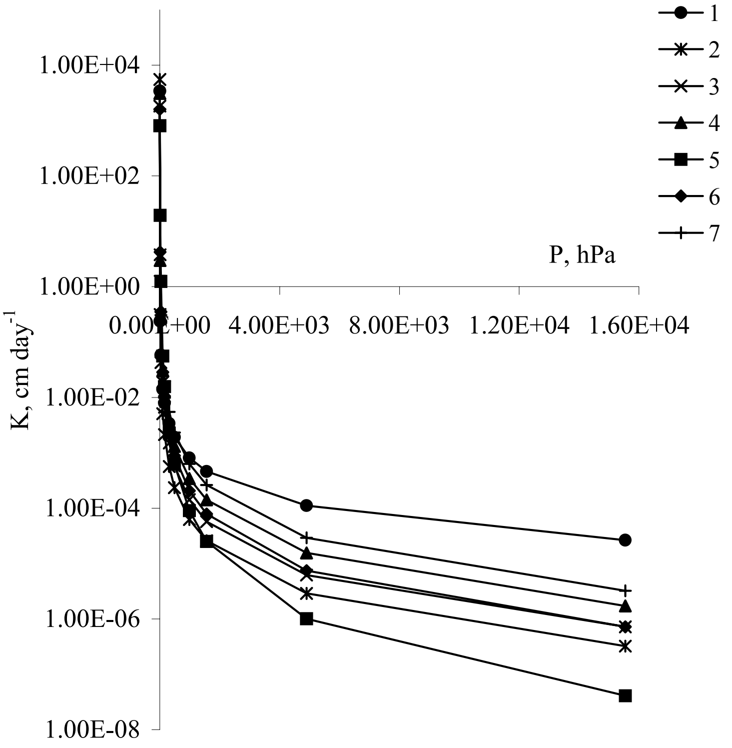

Fig. 6.1.3. Relation between hydraulic conductivity and water potential for soil material investigated.

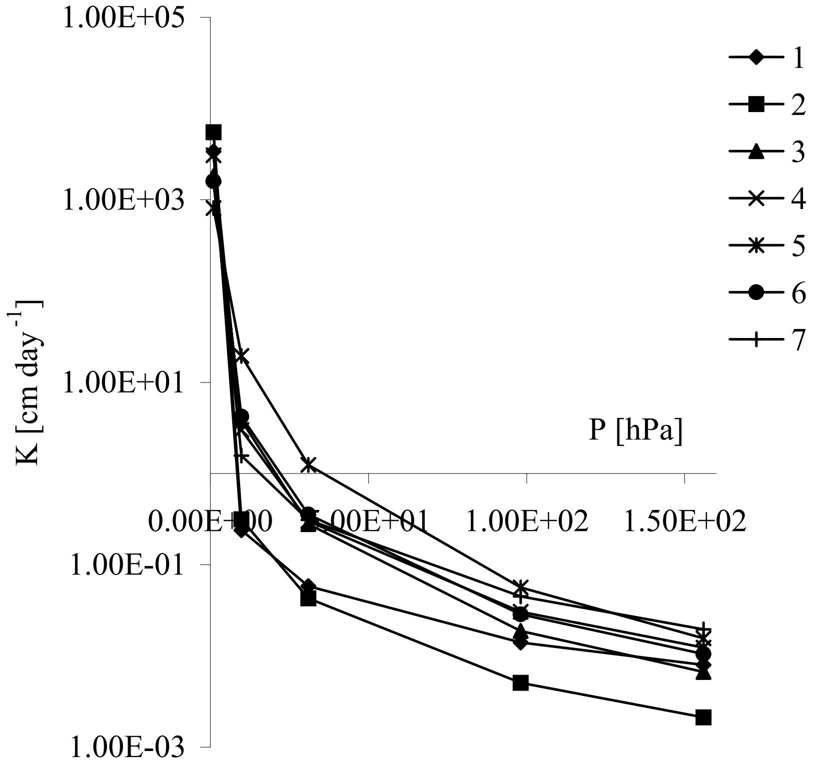

Fig. 6.1.4. Relation between hydraulic conductivity and water potential for soil material investigated.

The previous research has revealed that the instantaneous profile method advantageous for the investigation of hydraulic conductivity of mineral soils. In this paper we used this procedure to investigate hydraulic conductivity of organo–mineral formations. The results of the investigations of unsaturated hydraulic conductivity obtained by the instantaneous profile method IPM as a functions of water potential are given in Fig. 6.1.3 and 6.1.4.

The hydraulic conductivities of the investigated materials were found to cover a wide range under the change of water potential from 0.981 hPa till 15548 hPa. At water potential values from 9,81 to 31 hPa the hydraulic conductivity values of all the peat-sand mixtures are higher than a corresponding values for peat (Fig 6.1.4.). Thus, at water potential 98.1 hPa, the hydraulic conductivity for peat is 0.239 cm day-1, for sand 1.56 cm day-1, for sample 2-0.314 cm day-1, for sample 3-3.79 cm day-1, for sample 4-2.98 cm day-1, for sample 5-1.95 cm day-1 and for sample 6-4.21 cm day-1.

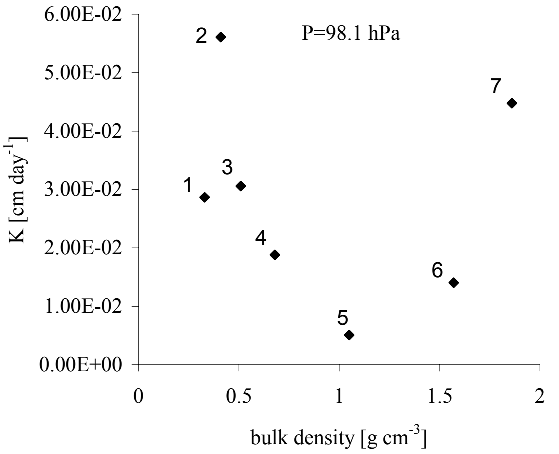

Fig.6.1.5. Relation between unsaturated hydraulic conductivity and total porosity of soil material investigated at P= 98.1 hPa.

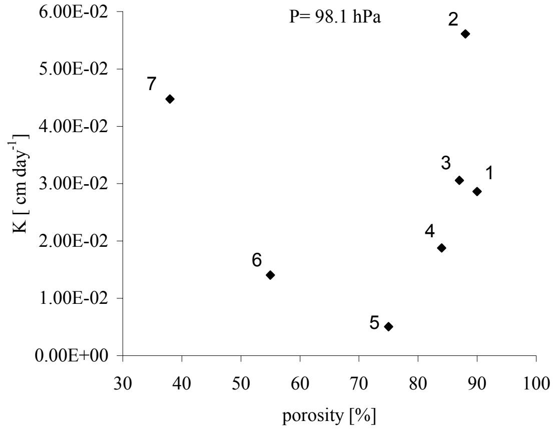

Fig.6.1.6. Relation between unsaturated hydraulic conductivity and bulk density of soil material investigated at P= 98.1 hPa.

It should be remarked, that the dependence of unsaturated hydraulic conductivity of investigated systems shows a minimum at certain value of total porosity and bulk density (Fig.6.1. 5 and 6.1.6).

As water potential decreases (increase on absolute value), the ratio between hydraulic conductivity values of peat and peat–sand mixtures changes. In the region of water potential higher than 310 hPa, the peat has higher conductivity value as compared with other organo-mineral systems. At 310 hPa, hydraulic conductivity value for peat is 3.38 10-3 cm day-1, for sand-2.32 10-3 cm day-1, for sample 2-5,69 10-4 cm day-1, for sample 3-1.51 10-3 cm day-1, for sample 4-3.21 10-3 cm day-1, for sample 5-2.27 10-3 cm day-1 and for sample 6-2.37 10-3 cm day-1.

The performed investigations of the influence of the organic mater content in peat-sand mixtures on their hydraulic conductivity lead to the following conclusions.

1. The hydraulic conductivity of peat-sand mixtures depends to a large extend on the relation between their organic and mineral parts and is affected by the structure as well as by the texture of a soil system.
2. The saturated hydraulic conductivity of peat–sand mixtures decreases if the sand part content is higher than 20%. The addition of sand to a peat sedge in quantity higher than 20% leads to deterioration of hydraulic conductivity of the sedge peat at saturated zone.
3. The obtained hydraulic conductivity values of peat, sand and their mixtures have been generalized using Mualem-Van Genuchten equation and the model parameters have been defined.

6.2. Influence of human activity on peat hydrophysical properties

The soil material was taken from the central part of Polish Polesye [165]. This territory was drained in the second part of the XXth century and now it is situated in the Wieprz-Krzna Canal area where the drainage-drying program is realized.

The studied peat soils represent the material at the different stages of organic matter transformation as well as mineral residual peat soil and mineral soil (Table 6.2.1).

Table 6.2.1. Basic properties of the investigated soils.

SAMPLE	Layer [cm]	Ash content [%]	Humus [%]	Specific surface area $[m^2g^{-1}]$	pH in		Fe $[g\ kg^{-1}]$ of dry matter
					H_2O	KCl	
1.	5-10	43.4	-	197	7.0	6.7	13.7
2.	30-35	39.6	-	195	7.3	6.8	20.6
3.	5-10	42.6	-	207	5.2	4.6	10.0
4.	40-45	34.1	-	254	5.9	5.3	4.8
5.	5-10	46.5	-	177	6.8	6.2	5.0
6.	30-35	40.5	-	165	6.5	6.1	4.5
7.	5-10	92.7	6.4	27	6.8	6.4	6.7
8.	30-35	-	0.10	2	4.8	4.5	1.7

The choice of sample plots was caused also by the way of utilization. The samples were taken from different depths (5-10 and 30-45 cm) to have organic soils with the same genesis but with different stages of the organic matter transformation. All investigated peat formations have a high value of ash content (more than 30%).

Site I (samples 1 and 2) is located in a floodplain. It is a deep organic soil, strongly drained, transformed, which is used as an opencast peat mine. Sample 1 (depth – 5-10 cm) is the proper moorsh derived from the reed-sedge peat with inclusions of iron . Sample 2 (depth 30-35 cm) is the proper moorsh formed from the reed-sedge peat with inclusions of calcium and iron.

Sites II and III (samples 3, 4 and 5, 6) are shallow reclaimed peat soils. Both of them are used as the permanent grass. They have the same botanic composition, but differ in the degree of draining. Site II is heavy humified and site III is moderate humified meadow. Sample 3 (5-10 cm) is the peaty moorsh from a sedge peat and sample 4 (30-35 cm) is the medium decomposed sedge peat. Sample 5 (5-10 cm) is the humic moorsh and sample 6 (40-45 cm) is the sedge peat of high degree of decomposition.

Site IV (samples 7 and 8) is an example of the end stage of anthropogenic evolution of the meliorated peat soil with a sand bedrock. In this site the full destruction of peat soil took place as a result of the drainage and agricultural cultivation with the field crop rotation. The mixture of arable peat layer (sample 7) with underlying sand bedrock (sample 8) led to the formation of the mineral soil which belongs to the residual peaty gleysolic soil according to Bambalov classification or the humic mineral soil according to Okruszko classification . It is used as an arable field.

The ash contents of the studied peat soils were determined by igniting the dried peat soil samples in a muffle furnace at about 550 0C until their weight was constant . The ash content was expressed in terms of the percentage ignition residue from the quantity of dry material (34.1-92.7%).The determination of the humus content in sand and mineral residual peat soil were carried out by Tiurin method. The specific surface area of soil samples (from 2 m2g-1 for sample 8 to 254 m2g-1 for sample 4) was evaluated from adsorption-desorption isotherms of water vapour which were measured by gravimetric method using a vacuum chamber. The bulk densities of the investigated soils varied from 0.24 g cm-3 (sample 4) to 1.86 g cm-3 (sample 8) and the total porosity (determined at saturation) – from 38% to 94%, respectively.

The measurements of static hydro-physical characteristics of the studied soils, i.e. relation between soil water potential and water content, were made within the range from 1 kJ m-3 (pF 1) to 100 kJ m-3 (pF 3) in the drying process according Richards procedure. The standard pressure chambers, manufactured by SOILMOISTURE Equipment, Santa Barbara, California USA, were used (Catalog Nos. 1500 and 1600). It was assumed that effective useful retention (EUR) is defined as an amount of water which is bound in the soil with the potential ranging from 16 kJ m-3 (pF 2) to 50 kJ m-3 (pF 2.7) in the pores with diameter between 6 and 30 mm. The amount of water, which is bound in pores with diameters smaller than 6 mm, is the water not easily accessible and inaccessible for plants and this one, in the pores with diameter bigger than 30 mm, is gravitational water .

Determination of water conductivity coefficients was carried out by a method of instantaneous profiles based on the measurements of water content and water potential in the chosen layers of the soil sample by means of a TDR measuring set in the process of soil drying . The measurements were carried out in the cylinders filled with soil in which holes were drilled at heights of 1, 2.5 and 4 cm from the bottom and TDR water content measuring probes together with micro-tensiometers measuring soil water potential were installed. The soil samples were filled with water till saturation was reached and then left under cover for 24 hours in order to reach a state of thermodynamic equilibrium. Then the samples were uncovered, their water content and soil water potential were monitored during evaporation. The TDR gauge was linked to a PC which enabled automatic measurements, and the values of water content and water potential taken, were recorded on the computer carrier. The measurements of the dynamics of soil water content and soil water potential profiles obtained, rendered possible the coefficient of unsaturated water conductivity, among other things.

During the anthropogenic evolution caused melioration processes and different utilization, the apparent changes of the structure and morphology of peat soils take place. The most significant changes can be observed in the upper layer of peat soils which undergo more strong transformation depending on the mineralization and humification processes. It was noticed that the ash content of peat soils is higher (from 4% to 8%) at an advanced stage of decomposition than in less decomposed peats (Table 6.1.1).

Table 6.2.2. Bulk density, porosity, distribution of pores and saturated water conductivity of the investigated soils.

Sample	Bulk density [g cm^{-3}]	Porosity [%]	Amount of pores [% m^3 m^{-3}]			k_{sat} [cm day^{-1}]
			$\phi > 30\ \mu m$	$30\ \mu m < \phi < 6\ \mu m$	$\phi < 6\ \mu m$	
1.	0.42	80	26	10	44	1720
2.	0.41	83	15	9	57	360
3.	0.38	85	18	10	57	5
4.	0.24	94	23	16	55	393
5.	0.52	82	12	9	61	31
6.	0.41	84	14	9	61	265
7.	1.16	55	21	11	23	370
8.	1.86	38	31	2	5	1600

The transformation of a peat layer leads to changes of its density and porosity. The increase of upper layer peat bulk density was observed in the investigated Sites I, II and III. In Site IV the bulk density of sand (sample 8) is higher than upper mineral residual peat layer (sample 7). The increase of the surface layer peat bulk density simultaneously induced the decrease of total porosity. The changes of bulk density between upper and lower layers of peat soils, from 0.01 to 0.14 g cm-3 for (Sites I, II and III), were the cause of the changes in their total porosity from 2 to 9% (Table 6.2.2).

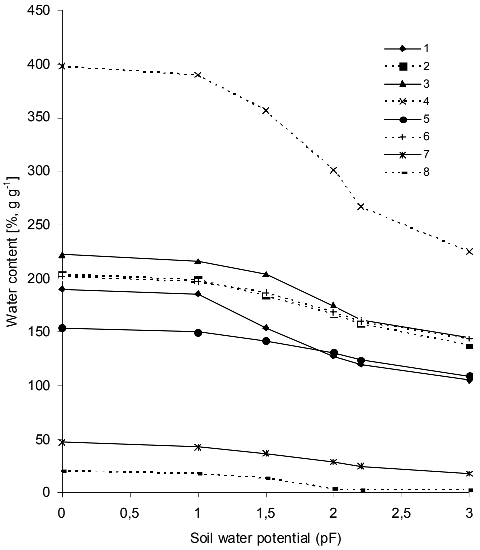

Fig. 6.2.1. Water retention curves for investigated soil samples (Nos 1-8) (water content in mass units [g g^{-1}])

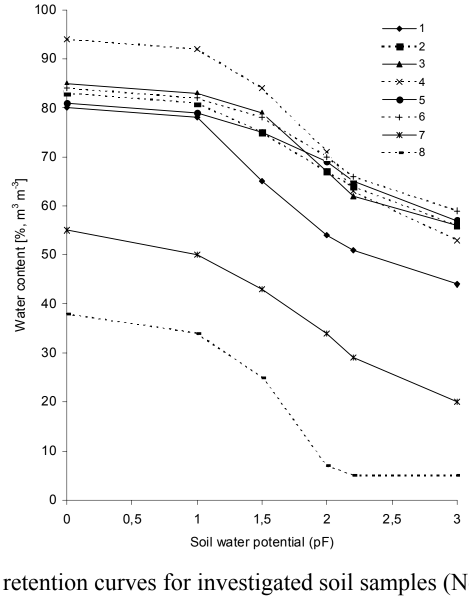

Fig. 6.2.2 Water retention curves for investigated soil samples (Nos 1-8) (water content in volume units [m^3 m-^{3}])

The soil water potential (pF)-water content characteristics for the investigated peat soils are presented in Fig.6.2.1 in mass units and in Fig.6.2.2 in volume units. It can be seen from Figs.6.2.1 and 6.2.2 that the shapes of these characteristics are similar and the water content values in the upper layers are lower than in the bottom layers for samples 1-6. In case of samples 7-8 (mineral residual peat soil) the relation is opposite.

The values of water content in the both cases (mass and volume units) show that less transformated peat soils (samples 2,4,6) have higher moisture than those more transformated (samples 1,3,5). For less transformated peat soils the moisture at saturation varies from 202 to 398% and at pF 3 - from 138 to 224 % in mass units (Fig. 6.1.1) and from 83-94% at saturation to 44-59% at pF 3 in volume units (Fig.6.2.2). The mineral residual peat soil (samples 7-8) retains the significantly lower amount of water, i.e. at saturation - 20 to 47% and at pF 3 - 3 to 18% in mass units, and in volume units at saturation – 38 to 55% and 5 to 20% at pF 3. Water content in the investigated soils between saturation and pF 1 is nearly the same that is an indicator of the absence of very large pores (diameters bigger than 300 mm). The sharp decrease of water content between pF 1 and pF 2-2.2 shows the significant amount of large pores (diameters between 300 and 30-18,5 mm). The differences in water content between pF 2 and pF 3 are significantly lower than between pF 1 and pF 2 what is expressed by the smaller inclination of water retention curve to pF axis. The amount of small pores (diameters smaller than 6 mm), which bind the water not easily accessible or inaccessible, is significantly higher in less transformated peat soils and changes from 44 to 61% m3 m-3 whereas in peat soil with higher degree of transformation – 5-23% m3 m-3.

The effective useful retention (EUR), defined as an amount of water hold in the soil between pF 2 and 2.7, is one of the most important properties of soils for efficient use of water, because it directly influences the growth of plants and their yield . It was noticed that values of EUR for sites I and III were nearly the same (9-10%), for site II the transformation caused the EUR decrease (6%), but for site IV – its increase (9%). It means that upper layer of mineral residual peat soil characterizes with the similar EUR value as organic soils and creates more favorable conditions for agrophytocenozis.

Under natural conditions the water retention of peat soils is very high. Peat soils are characterized by a high amount of small pores (Table 6.2.2) and a very heterogeneous pore structure formed from plant residues in different stages of decomposition . In the peat, different pore categories are identified: large multiple and simply connected open pores, dead-end pores, completely isolated pores and pores in cell structures .

The degree of transformation of organic matter in peat soils under drainage changes with depth. This feature was found to be the most evident for the sedge peat soil (site II, samples 3,4). Two major hydrologic zones exist within this peat soil. An upper aerated zone of fluctuating water conditions and an underlying anaerobic zone of a constant waterlogged state. Thus, peat layer at the depth 30-35 cm (sample 4) is the sedge peat, medium decomposed in anaerobic condition. Figs. 1 and 2 show that of all the peat studied, the sample 4 contains the greatest quantity of water at saturation - about 400 % g g-1 or 94 % m3 m-3 and has the big macropores content –23 %. But at the same time this peat soil gives up its water more readily with increasing soil water potential (pF) that is shown by a sharp change of shape of water retention curve. Water retention curves for both layers of the site III (samples 5, 6) have the same character, but they are more smooth than the water retention curves for site II. The 30-35 cm peat layer retains more water at all pF values as compared with 5-10 cm surface layer. The strongly transformed reed sedge peat soil profile (site I) revealed the same regularity in water retention characteristics as the above organic soil profiles. The top soil layer (sample 1) holds more water at any pF as compared with the deeper soil layer (sample 2), but the differences in water content for these layers from saturation to pF 1 did not exceed a few percent and they increase with the increase of pF values. In case of peats which have reached a more advanced stage of decomposition (samples 1,3,5) the water content at saturation was lower and the decrease of water content with the increasing pF were also smaller. Results of a similar kind have been obtained previously for peat soils . The mineral residual peat soil (site IV, sample 7) has been characterized by a significantly smaller water retention under all the soil water potentials as compared with above mentioned organic soils (samples 1-6). The difference in water content between them reached 100% g g-1 and more than 20% m3 m-3. The high water retention of a upper layer of residual peat soil is stipulated probably by organic content (7%). The water retention of underlying sandy layer (sample 8) is the lowest among all the investigated soils. This sandy layer characterizes the highest content of macropores (31 %) and the lowest content of mezo- and micropores (7%).

The saturated hydraulic conductivity characterises the movement of water in the soil when all the soil pores are filled with water. The values of saturated hydraulic conductivity for investigated soils are presented in Table 6.2.2. It can be seen that for the Site I the value of saturated hydraulic conductivity is considerably higher for the sample 1 – 1720 cm day-1 than for the sample 2 – 360 cm day-1, whereas for the Sites II, III and IV the values are lower for upper layers (samples 3, 5 and 7 – 5, 32 and 370 cm day-1 respectively) than for deeper layers (the samples 4, 6 and 8 – 393, 265 and 1600 cm day-1 respectively). It can be due to the fact, that Site I is strongly decomposed peat (proper moorsh) and it characterises with the big content of large macropores (26%) in the top layer (sample 1) and the Fe content in sample 2 is very high – 20.6 g kg-1. The saturated hydraulic conductivity of organic materials changes with their state of decomposition . The drainage leads to the increase of peat density and speeds up the decomposition. It means that saturated hydraulic conductivity is higher for deeper peat layers which have a small degree of decomposition (samples 4, 6 and 8) than for upper layers where the decomposition reaches the higher level (samples 3, 5 and 7).

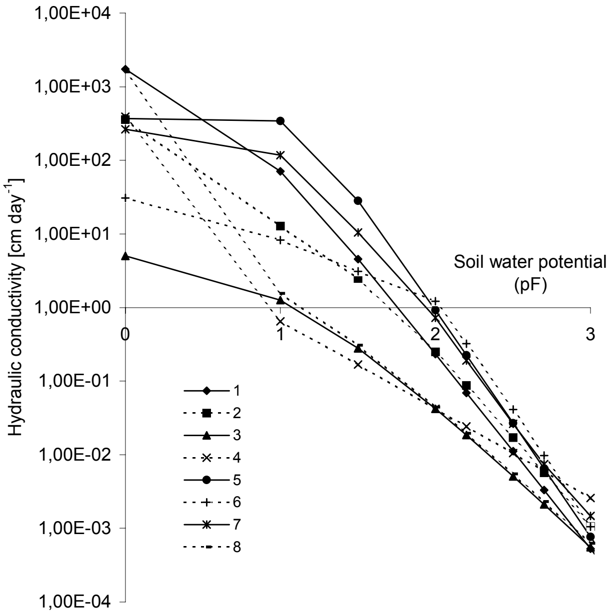

Fig. 6.2.3. Hydraulic conductivity at chosen soil water potentials for investigated soil samples (Nos 1-8)

The results of investigations of unsaturated hydraulic conductivity carried out with the use of the instantaneous profile method and smoothed using Mualem-van Genuchten's model as a function of water potential are given in Fig.6.2.3. It is known that the unsaturated water conductivity of peat soils depends on the nature of peat, decomposition degree, ash content, bulk density (consolidation) and flux direction, so it is clear that the drainage and agricultural utilization, i.e. anthropogenic influence, lead to the changes of physical properties of peat soils and at the same time they change the unsaturated hydraulic conductivity.

By the analysis of the courses of curves presented in Fig. 3 it can be stated that at low soil water potentials in the Sites I and III and at all the investigated potentials in the Site IV (except pF<0.2) the values of unsaturated hydraulic conductivity (kunsat) are lower in the deeper layers of peat soils whereas in the Site II the values of kunsat reversed. For example at pF 1 the kunsat for sample 1 equals 71 cm day-1, sample 2 –12.8 cm day-1, sample 6 – 8.3 cm day-1, sample 7 – 1.2 .102 cm day-1 and for sample 8 –1.6 cm day^{-1}.

In the case of samples 3 and 4 the values of kunsat at pF 1 are nearly the same – 1.2 and 0.7 cm day-1. For the samples 1 and 2 at pF 1.7 and for the samples 3 and 4 at pF 2 the values kunsat are the same. Further on, up to pF 3 they decrease then reaching 6.3.10-4 and 1.10-3 cm day–1 respectively. It can be stated that the anthropogenic evolution of peat soils leads to the considerable increase of kunsat values at low soil water potentials (Sites I and III) or in all range of soil water potentials (Site IV). Only in the case of the Site II (heavy humified shallow reclaimed peat soil) this relationship reversed.

The performed investigations of the influence of human activity on in peat hydrophysical properties lead to the following conclusions.

1. Transformation of organic formations as result of drainage and agricultural utilization leads to changes of their physical properties, i.e. it causes the increase of bulk density and ash content and the decrease of total porosity as well as the quantity of macro- and micropores.
2. Water retention of the drained peat soils which have reached a more advanced stage of decomposition is lower and the loss of water with the increase of the water potential is smaller.
3. Anthropogenic evolution does not cause significant changes in effective useful retention (EUR) in the investigated organic soils.
4. Saturated hydraulic conductivity is higher for deeper peat layers which have a small degree of decomposition than for upper layers where the decomposition reaches the higher level.
5. Anthropogenic evolution of peat soils leads to the considerable increase of unsaturated hydraulic conductivity at low soil water potentials or in the whole range of soil water potentials, only in the 25% of cases of the investigated organic soils this relationship reversed.

6.3 Saturated hydraulic conductivity of soil aggregates

One of the most important functions of the soil in the water circulation of the biosphere is its ability to soak up water and transfer it to neighbouring layers [208]. This property, together with the gradient of the terrain, decides the amount of water stored in the soil profile, provides the underground water supply and, at the same time, prevents evaporation from the surface runoff and protects the soil from erosion . The physical properties of the arable layer, which change during tillage, include, first of all, bulk density and soil structure. Tillage causes the occurrence of the dynamic quantitative changes between micro- and macro-aggregates and their characteristic distribution in the soil profile . The improvement of the soil's structure or the soil's aggregation quality has been a subject of interest for several years. It is determined not only by the necessity for the intensification of agricultural production, but also by the importance of environmental protection, water balance, the re-cultivation of former industrial areas, the utilisation of waste land and erosion control .

Water movement in the soil can take place in conditions of full or incomplete saturation. Soil, fully saturated with water, usually occurs directly after rainfall or irrigation (wetting). The water distribution and the time of its occurrence depend on the soil's properties and, of course, on its structure. Soils which have a stable and well organised structure have the characteristic of good water permeability enabling the water supply in the soil profile and limiting evaporation from their surface. On the other hand, soils without an aggregate structure are characterised with water stagnation on their surface and excessive water runoff .

The aim of this study is to determine the dependence between soil aggregate size and saturated hydraulic conductivity.

The object of investigations were aggregates of Eutric Cambisol, Orthic Luvisol, Calcaric Cambisol Haplic Phaeozem and Stagnogleyic Phaeozem formed of loess, samples of which were taken from the arable layer (Table6.3.1).

Tabela 6.3.1. Properties of investigated soils in natural state.

Soil type	Granulometric composition [%] [mm]			$C_{org.}$ [%]	$CaCO_3$ [%]	pH in KCl	Surface specific area (H_2O vapour) [$m^2\ g^{-1}$]	Bulk density [$g\ cm^{-3}$]
	1-0.1	0.1-0.02	<0.02					
Eutric Cambisol	3	58	39	0.9	4.1	7.4	39	1.33
Orthic Luvisol	14	47	39	1.8	0.5	7.3	51	1.26
Calcaric Cambisol	11	47	42	1.7	1.0	7.3	46	1.32
Haplic Phaeozem	4	50	46	3.0	0.12	6.8	79	1.15
Stagnogleyic Phaeozem	3	57	40	4.0	0.19	5.8	81	1.09

Following the drying of soil samples in the laboratory to an air-dry state, the aggregation of the soils was determined by the standard sieve method. Then, soil cylinders were filled with aggregates of the particular fractions: < 0.25; 0.25-0.5; 0.5-1; 1-3; 3-5 and 5-10 mm. Next, the aggregates were subjected to successive wetting-drying cycles. This procedure allowed soil materials with stable physical characteristics to be obtained . The aggregation of soil samples after the cycles was determined and the mean weight diameter of aggregates in each sample was calculated as an index of the changes in the size of soil aggregates. The mean weight diameter of aggregates for each soil samples for initially mono-aggregate probes increased and varied in the range from 0.12 to 1.98 mm for Eutric Cambisol; 0.13-3.54 mm for Orthic Luvisol; 0.13-2.48 mm for Calcaric Cambisol; 0.13-3.32 mm for Haplic Phaeozem and from 0.13 to 2.64 mm for Stagnoglyic Phaeozem. An amount of large pores was found from the water retention curves. It is an accepted fact in the soil sciences to consider pores of diameters higher than 18.5·10-6 m as large pores. These are pores in which water is bound with potential values lower than the potentials representing the soil water field capacity (15,6 kJm-3, pF 2.2). The saturated hydraulic conductivity was measured by standard apparatus made by Eijkelkamp, Agrisearch Equipment, the Netheralnds.

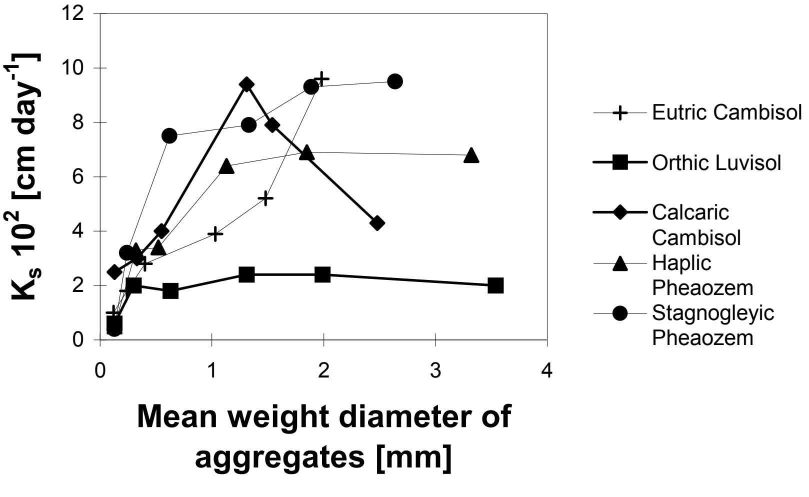

Fig. 6.3.1 Relationship between the mean weight diameter of aggregates and the saturated hydraulic conductivity.

The saturated hydraulic conductivity (Ks) for soil samples composed of different size aggregates are presented in Fig. 6.3.1. Generally, it can be seen from the figure, that the minimum values of the saturated hydraulic conductivity coefficient were noticed for aggregates smaller than 0.25 mm and that it increased with the increase of the mean weight aggregate diameters for all the soils investigated, except Calcaric Cambisol. The lowest values of the saturated hydraulic conductivity coefficient were noticed for all aggregate fractions of Orthic Luvisol; these changed from 0.6·102 to 2.4·102 cm·day-1. The highest values of the saturated hydraulic conductivity coefficient reached for the soil samples initially composed of the largest aggregates of Eutric Cambisol – 9.6 ·102 cm·day-1, the aggregates of 1-3 mm fraction of Calcaric Cambisol – 9.4·102 cm·day-1 and for the samples created from aggregates larger than 3 mm of Stagnogleyic Phaeozem - 9.3·102 cm·day-1 and 9.5·102 cm·day-1 respectively. When analysing the courses of the curves for particular soils it can be observed that the saturated hydraulic conductivity increases with the increase in aggregate size for Eutric Cambisol, Haplic and Stagnogleyic Phaeozem, reaching respectively, from 1.0·102, 0.5·102 and 0.4·102 cm·day-1 for fractions smaller than 0,25 mm to 9.6·102; 6.8·102 and 9.5·102 cm·day-1 for the fraction 5-10 mm. For Calcaric Cambisol, the increase of the saturated hydraulic conductivity was noticed to be increased with an increase in aggregate size to 1-3 mm, from 2.5·102 cm·day-1 to 9.4·102 cm·day-1. Later, it decreased considerably and reached 4.3 cm·day-1 for aggregates of 5-10 mm. For Orthic Luvisol, the coefficient's values increased strongly with the increase of aggregate sizes from the smallest diameter to up to a value of 0.25-0.5 mm from 0.6·102 to 2·102 cm·day-1, further on they varied between 1.8 – 2.4 cm·day-1.

The water flow in the soil saturated zone in natural conditions takes place mainly due to gravity. Thus, it can occur in pores with large diameters (PL), where the inter-particle and capillary forces are less important. The number of large pores in the soil is connected with its aggregate size distribution. Therefore Fig. 2 presents the dependence between them and the mean weight diameter of aggregates of the soils studied. It results from the figures presented, that generally, the amount of large pores in the soil increases with the increase of the mean weight diameter of aggregates. An especially rapid increase in their amount is observed for the smallest aggregates with sizes of about 1 mm, i.e. from several per cent to 0.35-0.40 [m3 m-3], except from the aggregates of 0,5-1 mm fraction of Haplic Phaeozem. The soil samples initially composed of aggregates with diameters higher than 1 mm, i.e. the fractions 1-3, 3-5 and 5-10 mm, contain from 0.31 to 0.43 [m3 m-3] of the large pores. This means especially strong impact of the soil structure on the amount of large pores in the soil. A similar type of the dependence between the saturated hydraulic conductivity and the mean weight diameter of aggregates in the soil samples is shown in Fig. 6.3.1.

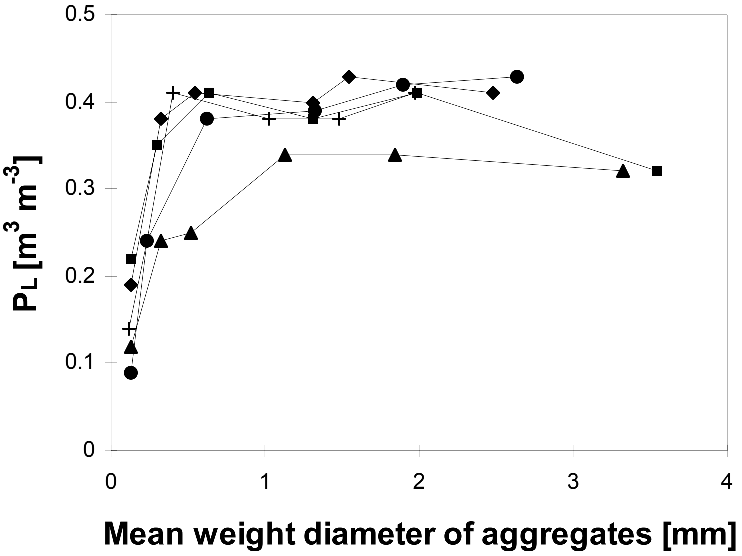

Fig. 6.3.2. Relationship between the mean weight diameter of aggregates and the amount of large pores.

The dependence between the amount of large pores and the mean weight diameter of aggregates in the soil samples is given in Fig.6.3.2. The relationship between the saturated hydraulic conductivity and the amount of large pores is presented in Fig.6.3.3.

The points in this figure indicates that the increase of the amount of large pores in the soils studied results in the increase of the saturated hydraulic conductivity. An increase in the saturated hydraulic conductivity from 0.4·102 to 9.5·102 cm·day-1 with an increase in the amount of large pores from 0.09 to 0.35-0.43 [m3 m-3] was noticed for the Haplic and Stagnogleyic Phaeozem. For Eutric Cambisol, Orthic Luvisol and Calcaric Cambisol the saturated hydraulic conductivity increases slightly with the increase of the amount of large pores from 0.14 to 0.30-0.40 [m3 m-3]. After exceeding 0.30-0.40 [m3 m-3] of the large pores the saturated hydraulic conductivity rapidly increase; for Eutric Cambisol from

1.0·102 to 9.6·102 cm·day-1 and for Orthic Luvisol and Calcaric Cambisol from 0.6·102 to 1.8·102 cm·day-1and from 2.5·102 to 9.4·102 cm·day-1 respectively. For these experimental data the following relationship was found:

$$K_s = 0.356 \cdot e^{6.852 \cdot P_L} \qquad (6.3.1)$$

where:
Ks – saturated hydraulic conductivity [cm day-1],
PL – amount of large pores (f> 18.5 m-6) [m3 m-3].

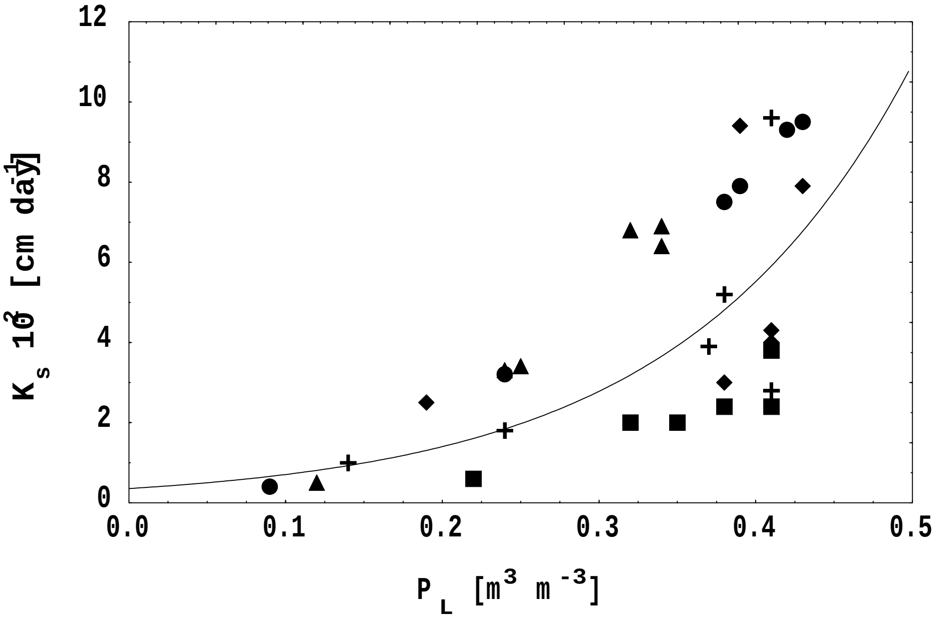

Fig. 6.3.3. Relationship between the amount of large pores and the saturated hydraulic conductivity.

The above described relationship between the saturated hydraulic conductivity and the amount of large pores contained in the soil samples investigated, indicates that narrowing between large pores and probably their tortuosity also have an influence on the water flow velocity in a saturated soil zone. This is because in soil samples where the amount of large pores is considerable, the narrowing between the pores decreases and the large continuous pores of small tortuosity are created causing a rapid increase of the saturated hydraulic conductivity and at the same time unrestricted water flow.

Summarising the results obtained, it can be stated that the increase of soil aggregate sizes causes an increase of the saturated hydraulic conductivity. It is especially clear for soil samples whose mean weight diameter of aggregates is smaller than about 1.5 mm. This influence is directly conditioned by the dependence between soil aggregate size and the amount of large pores in the soil, which is a consequence of the water-resistance of the aggregates. The relation between the aggregate size and the saturated hydraulic conductivity is modified by the soil type.

The results obtained by performed investigations, lead to the conclusion that the aggregates size and the soil type determine the saturated hydraulic conductivity as follows:

- the saturated hydraulic conductivity increases with the increase of the mean weight diameter of aggregates in soil samples, in particular in the range below 1.5 mm;
- the increase of the mean weight diameter of aggregates causes an increase in the amount of large pores, especially for aggregates whose diameter varied from 0.12 to 0.7 mm;
- the relation between the amount of large pores in aggregated soil samples and the saturated hydraulic conductivity may be expressed by the equation .

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