INVESTMENTS IN EDUCATION DEVELOPMENT

Jan Evangelista Purkyně University

Faculty of the Environment

Examples of Determining the Hydraulic Conductivity of Soils

Theory and Applications of Selected Basic Methods

University Handbook on Soil Hydraulics

Jakub Stibinger

Ústí nad Labem
2014

INVESTMENTS IN EDUCATION DEVELOPMENT

Title: Examples of Determining the Hydraulic Conductivity of Soils
Theory and Applications of Selected Basic Methods

Author: doc. Ing. Jakub Stibinger, CSc.

Scientific editor: Ing. Martin Neruda, Ph.D.

Reviewers: doc. Ing. Václav Kuráž, CSc.
doc. Ing. Vladimír Švihla, DrSc.

This publication was supported by the OP Education for Competitiveness project: EnviMod – Modernization of education in technical and natural sciences at UJEP with respect to environmental protection.
Reg. No.: CZ.1.07/2.2.00/28.0205

Free copy

ISBN 978-80-7414-837-8 (brož.)
ISBN 978-80-7414-836-1 (online: pdf)

Preface:

I would like to introduce university textbook “Examples of Determining the Hydraulic Conductivity of Soils” which describes and explains some selected methods procedures for determination of hydraulic conductivity (K-value) of soils.

The text of this university handbook comes from the results of research, surveys, investigations, and also from the teaching experiences. Presented material can serve as a very good tool not only for university student, but also for engineers and experts solving the problems connected with wide area of soil hydrology.

It is proved, that the present and future climate dynamics is accompanied with strongly worse hydrology conditions. At the same time it has to be emphasized, that the value of hydraulic conductivity of soil has an influence on all spheres of hydrology areas with direct impact on environmental protection, water regime, agriculture (food security), forestry and other branches.

It means that the suitable selection of the methods and procedures going to the correct determination of the hydraulic conductivity of soils play a key role in soil hydraulics with a large impact on the significant branches of economy.

The value of hydraulic conductivity is one of the most important parameters for design and realization of all type measures for mitigation of negative impact of hydrological extremes as are floods and enduring droughts.

The methods of determining the hydraulic conductivity of soils are divided into two main parts, pedotransfer functions and hydraulic methods, where the main importance is placed on the application of the hydraulic methods in field conditions. The methods are clearly categorized in Figure 1 and are very briefly described at chapter 2. “Methods of Hydraulic Conductivity Determination – Reduced Summary” of this textbook.

The use of the hydraulic methods is based on the mathematic-physical descriptions of water flow in porous media with certain allowed simplification. An application of those assumptions enables the use of analytical solution without undesirable influencing of results, which can suitably approximate real water flow processes.

Analytical procedures are from the pedagogy point of view very demonstrated. The finding parameters of the problem are directly projected in the final formulas and show the impact of individual soil hydrology characteristics on value of hydraulic conductivity (K-value).

With use of Darcy’s Law and Darcy-Buckingham Law in a case of unsteady state flow in unsaturated zone and by equation of continuity for defined boundary and initial conditions we get final formulas for direct estimation of hydraulic conductivity.

Correction and smoothing of measured data were carried out by selected analytical procedures, with application of advanced methods of non-linear regression analysis to describe time series of infiltration and groundwater flow.

Hydraulic conductivity is subject to variation in space and time, what means, that we have to adequately asses a representative value. This is time-consuming and costly, so a balance has to be struck between budget limitations and desired accuracy. Because of variability of soil is better to use large-scale field experiments, e.g. existing field drainage systems methods, what is expensive but the most reliable.

In this context is necessary to say, that not optimum surveying techniques exist. Much depends on the skill of the person conducting the survey. To get representative K-value the surveyor must have knowledge of the theoretical relationship of existing parameters of the problem to be solved.

Which method is the best to select for survey and investigation to determine K-value depends on the practical applicability, and the choice is very often limited. It depends on natural and water regime conditions and on the type of measures which will be designed.

Some advanced apparatuses for laboratory and filed measurement of hydraulic conductivity of soil developed by Research Institute for Soil and Water Conservation (RISWC) Prague-Zbraslav, Czech Republic was not presented in this handbook.

In a frame of current research is going on their next testing and verifications in field conditions of RISWC experimental areas. The students will be introduced with their possibilities of use and with basic principles of apparatus during the lecturing and training lessons of Soil Hydraulics.

Next advanced procedures leading to determining the hydraulic conductivity of soil are showed, besides other possibilities of internet finding, e.g. on the relevant web pages of CULS (Czech University of Life Sciences) Prague.

Very demonstrative is the textbook named "The Multimedial Study Guide of Field Hydropedological Measurements" which is created in Czech and English versions and it is available at http://hydropedologie.agrobiologie.cz/en-index.html. This guide was made at Department of Water Resources, Faculty of Agrobiology, Food and Natural Resources, Czech University of Life Sciences Prague (prof. S. Matula).

Swampy areas, wetlands, peat-lands and also paddy (rice) fields are very important territories with respect to environment and water resources protection, but also for agriculture policy and food security. Therefore was slightly larger attention given to a single ring infiltrometer method, used for measurement with a little water layer on soil surface. In the textbook are also introduced a few practical examples of infiltration trenches and their modifications like swales, which can serve as the measures for mitigation of negative impacts of climate dynamics such as waterlogging and floods.

The procedures of determining of K-value described in the textbook are in the text very often attended by practical examples of application of K-value in water management engineering practice.

Practical application of hydraulic conductivity (K-value) together with derivation of connected process is also presented in a few power-point presentations with theme "Soil Hydraulics in Environmental, Landscape and Water Regime Protection". There are following titles:

- Drainage Retention Capacity (DREC) in Soil Hydrology – Practical Application
- Infiltration Ditches at Slope-land Orchard – Old City Prague
- Hydraulic Conductivity with Clogging Impact of Landfill Leachate Drainage System – Practical Application
- Integrated Water Management System to Mitigate Negative Impacts of Typhoons (case study form Taiwan, Republic of China) Soil Hydraulics in Agriculture Drainage, Food Safety and Environmental Protection

At the end of preface and from the fact mentioned above is possible to state that the hydraulic conductivity (K-value) of soil and its determination has great importance in environment, water management, agriculture, civil engineering and in other branches and the university textbook "Examples of Determining the Hydraulic Conductivity of Soils" can contribute to its reliable determination.

Jakub Stibinger, November 2014, Ústí nad Labem

Contents

PREFACE: **3**

1. DEFINITIONS AND TERMINOLOGY **6**

2. METHODS OF HYDRAULIC CONDUCTIVITY DETERMINATION – REDUCED SUMMARY **10**

2.1. Correlation Methods **11**

2.2. Laboratory Methods **14**
2.2.1. Examples of Hydraulic Laboratory Methods 14
2.2.1.a. Falling-Head Method 14
2.2.1.b. Constant-Head Method 16

2.3. Field Methods **18**
2.3.1. Examples of Small-Scale Field Methods 18
2.3.1.a. Piezometer Methods 18
2.3.1.b. Infiltrometer Methods 19
2.3.1.c. Auger-Hole Method (Hooghoudt's Method) 41
2.3.1.d. Inversed Auger-Hole Method 43
2.3.1.e. Modified Inversed Auger-Hole Method (Porchet's Method) 46
2.3.1.f. Small-Scale Infiltration Trench 51
2.3.1.g. Application of Infiltration Trenches 55
2.3.2. Examples of Large-Scale Field Methods 65
2.3.2.a. Pumping Tests 65
2.3.2.b. Parallel Drains Method 65

2.4 DETERMINATION OF UNSATURATED HYDRAULIC CONDUCTIVITY **70**

REFERENCES **70**

1. Definitions and Terminology

Hydraulic conductivity, marked as K, or K-values, is one of the principal and most important soil hydrology (hydraulic) characteristic (parameter) and it is an important factor in water transport in the soil and is used in all equations for groundwater (subsurface water) flow. Very often is K-value use for simulation of infiltration processes.

Hydraulic conductivity, represented by symbol K ($M.T^{-1}$), is expressed by units of velocity ($M.T^{-1}$), where M is unit of length and T is time unit. The definition of the hydraulic conductivity follows from the Darcy's Law (Darcy, 1856, Kutílek and Nielsen, 1994, Todd and Mayes, 2005). In the saturated flow conditions the velocity of water flow v ($M.T^{-1}$) in the soils or in the other porous media, is directly proportional to the hydraulic gradient I (-).

The coefficient of this direct proportionality is a constant with units of velocity ($M.T^{-1}$), usually is marked by symbol K ($M.T^{-1}$) and is named hydraulic conductivity. The relates, described above, can be generally expressed by equation (1)

$$v = K.I \tag{1}$$

In the saturated flow conditions and according to the Darcy's Law, the flow velocity v ($M.T^{-1}$) can be expressed

$$v = K\frac{\partial h}{\partial x} \tag{2}$$

where x (M) is distance in the direction of groundwater flow, h (M) is hydraulic head.

Hydraulic conductivity K ($M.T^{-1}$) can characterize the hydraulic properties of soils, earths and also the hydraulic properties of the other porous materials and media, from the point of view of the velocity of water flow in their porous fully saturated flow conditions.

Soil hydraulic conductivity is defined as a constant (coefficient) of proportionality in Darcy's Law (see equation (1), (2)). If the hydraulic gradient, which is the difference of h (M) over a small difference of x (M), converges to 1, equations (1) and (2) can be rewrite as

$$v = K \tag{3}$$

What means, that the hydraulic conductivity can be regarded as the groundwater velocity, when the hydraulic gradient equals unity (when the hydraulic gradient is 1).

In water and landscape engineering practice, the value of hydraulic gradient is mostly less than 0.1 (-), hence the velocity of groundwater (subsurface water) flow is generally less than 0.1K.

Very often is the K-value less than 10 $m.day^{-1}$. So the velocity of groundwater (subsurface water) flow is almost always less than 1 $m.day^{-1}$. Of course, the K-value of saturated soils introduces the average hydraulic conductivity, which is dependent mainly on the shape, size and distribution of the pores and also depends on viscosity and density of water and on the soil temperature.

In some structure-less soils (sandy soils) the K-value is the same in all directions, but usually the K-values varies with flow direction. Soil layers vertical permeability is very often different from horizontal permeability because of vertical differences in the structure, texture and porosity. Generalized table with the ranges of K-values for certain soil texture is presented in Table 1 (Ritzema 2006).

Table 1. K-value range by soil texture (Ritzema 2006)

Texture	Hydraulic conductivity K (m.day^{-1})
Gravelly coarse sand	10 – 50
Medium sand	1 – 5
Sandy loam, fine sand	1 – 3
Loam, clay loam, clay (well structured)	0.5 – 2
Very fine sandy loam	0.2 – 0.5
Clay loam, clay (poorly structured)	0.002 – 0.2
Dense clay (no cracks, pores)	< 0.002

Table 1 should be used with care. Smedema and Rycroft (Ritzema 2006) warned: "Soils with identical texture may have quite different K-values, due to differences in structure" and also "Some heavy clay soils have well-developed structures and much higher K-values than those indicated at the Table 1".

Anisotropy plays very important role in soil hydrology. K-values in vertical direction is marked as K_v, hydraulic conductivity in horizontal direction represents symbol K_h and K-value in intermediate direction is K_r. The value of K_r can by approximated by formula

$$K_r = \sqrt{K_r K_h} \tag{4}$$

or

$$\ln K_r = \frac{\ln K_v + \ln K_h}{2} \tag{5}$$

Hydraulic conductivity K-value in the soil profile can be highly variable from place to place and also can be variable with a different depth, what means spatial variability. K-values can be variable not only in connection with different soil layers, but also within a one soil layer.

Soil layer measured K-values have usually a log-normal distribution, but with rather wide variation (Dieleman, Trafford 1976). Representative K-value, marked K*, can be estimated from the geometric mean by equation

$$K^* = \sqrt{K_1.K_2.K_3.....K_i.....K_n} \tag{6}$$

where n is a total number of observations.

Soil is hardly only uniform in vertical directions, and very often varies in the horizontal directions as well. It is necessary to keep in mind that homogeneity in the soil characteristics is exception rather than the rule. Both vertical and horizontal variations are major points of investigation in soil surveys.

A vertical variation in a soil can be partly due to layered composition of the parent material, but more commonly the results of profile development.

Horizontal variations in soil properties are common at any scale, even at less than 1 meter. In some cases the change in colour, salinity, texture, structure at the soil surface is quite sharp, but, more generally is gradual (Braun and Krujin, in: Ritzema 2006).

Detailed description of soil heterogeneity is presented in guide of Ritzema (2006) */Soil Conditions. Braun H. M. H. and Krujine R. / In: H. P. Ritzema (Ed) Drainage Principles and Applications (pp.83-84). ILRI Publ. 16, Wageningen, The Netherlands.*

The term of **hydraulic conductivity K ($M.T^{-1}$)**, or K-value, expresses conductivity of the fully saturated zone (which means fully saturated soil, earth, or any porous environment), unlike **unsaturated hydraulic conductivity k(H)**, which belongs among hydro-physical characteristics of unsaturated zone (see chapter Infiltrometer's method, Theory) and depends on negative soil pressure H (M) and thus on the soil moisture.

The values of the hydraulic conductivities K ($M.T^{-1}$) of some selected soils and rocks are presented in Table 2 (Todd and Mayes, 2005).

Table 2. Hydraulic conductivities K ($m.s^{-1}$) of selected soils and rocks (Todd and Mayes, 2005)

Earth, soil, material	**Hydraulic conductivity K ($m.s^{-1}$)**
Clay, heavy soils	$2{,}0.10^{-9}$
Gravel, medium	3.10^{-3}
Peat	$6{,}6.10^{-5}$
Dune sand	$2{,}3.10^{-4}$
Dolomite	1.10^{-8}

Can you write down the any hydraulic conductivities K ($m.s^{-1}$) of another soils? What about values of hydraulic conductivities K ($m.s^{-1}$) of ordinary loamy soils?

Hydraulic conductivity plays a key role in an environmental and water regime protection. In water engineering practice is necessary to know how to apply corresponding fundamental methods to determine (approximate) K to describe, explain and analyse subsurface and surface water flow not only in landscape, but also in rural and urban areas.

There are many various methods, the laboratory methods and the field's methods, to determine or to approximate the value of hydraulic conductivity. The thematic focusing and the limited range of the script do not allow the detail theoretical and practical description and analysis of the all individual methods of determination hydraulic conductivity.

In the following text are briefly mentioned several methods to determine the hydraulic conductivity, each having its own advantages (applications) and disadvantages (limitations).

Photo 1. Paddy field in Chiang Mai (Thailand). The determination of the value of hydraulic conductivity of surface layers is one of the most important points in a water management of paddy fields. (source: institute Two-highs Worldwide Holydays, Dublin -2, Dawson Street 31, Ireland, published marketing materials).

18.3.2008 time: 10.30 18.3.2008 time: 11.39 18.3.2008 time: 11.40

Photo 2. Time series of water level lowering during terrain experimental measurement of hydraulic conductivity by inverse auger-hole method in agricultural locality Čejkovice (Czech Republic). Photo: J. Štibinger, M. Soukup.

2. Methods of Hydraulic Conductivity Determination – Reduced Summary

Hydraulic conductivity determination of soils can be realized with correlations methods or with hydraulics methods. Hydraulic methods can be divided into laboratory methods or field (in-situ) methods.

Correlation methods come from predetermined relationship between soil property (e.g. grain size distribution, texture, etc.) which can be easily determined and K-value. The advantage of the correlation methods is a fast estimation of the K-value, than the direct measurement.

A deflect is a fact that the application of relationship can be incorrect and can be a reason of random errors.

Hydraulic methods come from supposed certain flow conditions, with boundary and initial conditions and with use of Darcy's Law (respectively Darcy-Buckingham Law in a case of unsteady state flow in unsaturated zone) and with use of equation of continuity. Final formulas for direct calculations (approximation) of K-values were received by analytical solution of initial fundamental equations.

The hydraulic laboratory methods, or **laboratory methods**, still fast and cheap, are used to core soil samples. They have similar deficiency as correlation methods and also the small sample area means the high possibility of a large random error.

The hydraulic fields methods, or **fields methods**, based also on description of the water flow processes, can determine K-values around the hole made in the investigated soils (in a case of small-scale), but the external boundary line of the soil environment is very often not exactly known.

The hydraulic fields methods can be divided into small-scale and large-scale methods, where **small-scale fields methods** serve for fast testing of many locations. They supposed allowable simplifications of groundwater (subsurface water) flow, so that measurements can be realized relatively cheaply and quickly.

The **large-scale fields methods** guarantee the representative K-values, where the problem of variation is eliminated as much as possible. But the large-scale field's methods are rather expensive and time-consuming, than the methods mentioned above.

Ritzema (2006) on Figure 1 shows the scheme of the methods for hydraulic conductivity (K-value) determination.

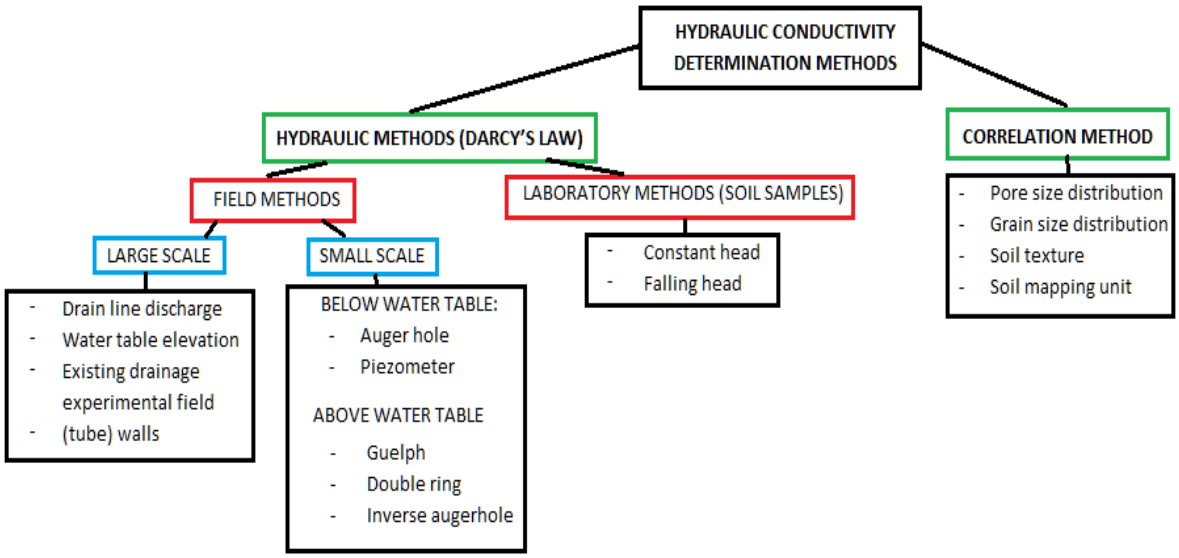

Figure 1. Overview of methods for the hydraulic conductivity determination (Ritzema 2006).

2.1. Correlation Methods

Description (principle):

The hydraulic conductivity of granular porous soils depends on the space between soil particles and thus on the grain-size distribution. The relation between hydraulic conductivity and grain-size distribution may be known from lots of laboratory tests with a certain type of soil.

Application (use):

Is possible to estimate the hydraulic conductivity by certain formula, if other methods are difficult to apply, as e.g. for deep soil layers.

Disadvantages (limitations):

This method can be applied only in good permeable soils (e.g. sandy or light loamy soils). This method may be applied only if no other methods of measuring the conductivity can be used. The obtained value of hydraulic conductivity is not an exact value, but gives order-of-magnitude estimates.

Formulas and expressions:

For approximation of hydraulic conductivity (K-value) of soils by correlation methods can be used, besides a lot of other expressions, Hazen's or Terzaghi's empirical formulas in a shape $K(cm.s^{-1}) = 100.(d_{10})^2$ where d_{10} (cm) is grain diameter corresponds in grain size curve at 10% of content. Terzaghi considers the number of porosity e (-) and for K-value ($cm.s^{-1}$) is valid expression $K(cm.s^{-1}) = 200.(d_{10})^2 e^2$.

Example from water management engineering practice:

According to the car park project design from UDIMO /Czech Republic/ with realization in 2009, all surface waters from precipitations should be evacuated by infiltration processes. Hydraulic conductivity (K-value) of soil subsurface layers bellow the permeable pavement was approximated by correlation methods with result K = 105 – 106 $m.s^{-1}$. Direct laboratory or field measurement of K-value based on hydraulic methods was not realized.

Precipitations, which in any case were not extreme intensity or long time duration, periodically flooded car park surface with permeable pavement (see Photo 3), after approximately one year of car park operation.

Photo 3. Flooded surface of car park with continuous water level (2009).

In a frame of following reconstruction of car park was discovered, that the infiltration capacities of the subsurface layers, situated bellow the permeable pave, are very limited. Generally speaking, to estimate K-values of subsurface layers was used correlations methods with formulas in Hazen or Terzaghi type although the investigated subsurface soils evidently could not be characterized as sandy soils or light loamy soils (see Photo 4 and Photo 5).

Photo 4. Beginning of the car park reconstruction. In left corner are evident sandy and clay particles and sticky loamy soils, surface water cannot infiltrate through subsurface layers (2009).

Photo 5. Car park soil profile with coal landfill placed at lower layers. In upper layers are loamy and sandy soils with clay (2009).

From the facts viewed above follows that the determination of K-value by correlation methods requires large experiences and caution and is applicable only for sandy and light loamy soils with good permeability.

At the same time is necessary to realize that the obtained value of hydraulic conductivity (K-value) is not an exact value, but gives order-of-magnitude estimates.

2.2. Laboratory Methods

2.2.1. Examples of Hydraulic Laboratory Methods

2.2.1.a. Falling-Head Method

Description (principle):

The water head at one of the sides of the sample decreases with time. A high initial water head is desirable for low hydraulic conductivities. The calculation of the hydraulic conductivity from the velocity of total flux through the sample can be somewhat complicated, because the head difference is not constant.

Use (application):

This laboratory method is suitable especially for layers with a low hydraulic conductivity, in horizontal or vertical direction.

Disadvantages (limitations):

Small sample area means the high possibility of a large random error.

Problem:

Can you write down final formula for calculation of the hydraulic conductivity? Instruction: Use equation (2) /Darcy's Law/ and expression v=dy/dt.

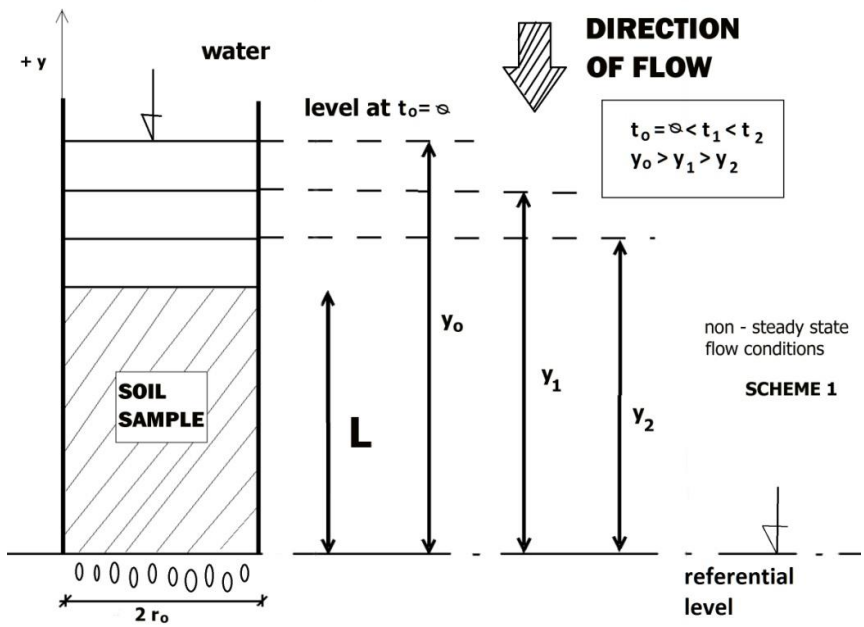

Figure 2. Falling head permeameter, referential level above the drops falling down from the permeameter represents the continuous free water level.

Theory:

After an addition of the certain amount of water into laboratory permeameter (apparatus), the water table goes up to the level y_0 (M) /see Scheme 1/. Corresponding time is t = 0, what is the beginning of measurement.

By gravity the water level in permeameter (apparatus) gradually goes down from the level y_0 to the level y_1, y_2, etc., with the corresponding times t_1, t_2, etc. Nevertheless it is valid $y_0 > y_1 > y_2 > 0$ and $0 < t_1 < t_2$ (see Figure 2).

In saturated flow conditions and according to Darcy's law the flow velocity v ($M.T^{-1}$) in the soil sample of permeameter (apparatus) in the vertical y-direction can be approximated as

$$v = -K \frac{y}{L_s} \tag{7}$$

where L_s (M) represents the length of soil sample in the laboratory permeameter (apparatus). Negative sign in a right part of equation (7) marks the flow, which is running down, it means opposite to the positive direction of y-axis.

The flow velocity v_i ($M.T^{-1}$) in the permeameter (apparatus) can be expressed as:

$$v_i = \frac{\partial y}{\partial t} \tag{8}$$

It is supposed that the flow velocity v ($M.T^{-1}$) in the soil sample, placed in the permeameter (apparatus) in y-direction will be equal as the flow velocity v_i ($M.T^{-1}$) in the permeameter (apparatus) $(v \equiv v_i)$ and we obtain:

$$-K \frac{y}{L_s} = \frac{\partial y}{\partial t} \tag{9}$$

From rearrangement equation (9) we get ordinary differential equation, which can be formed as:

$$\partial t = (\frac{-L_s}{K}).y^{-1} \partial y \tag{10}$$

Let's suppose, that the decreasing of water table level in permeameter (apparatus) will be e.g. from level y_1 (M) to level y_2 (M). The corresponding times are t_1 (T) and t_2 (M) /see Scheme 1/. Integration of equation (10) can be shaped:

$$\int_{t1}^{t2} dt = (\frac{-L_s}{K}) \int_{y1}^{y2} y^{-1} dy \tag{11}$$

After integration from y_1 to y2 in y-axis direction and from t_1 to t_2 in time axis and after other corrections and with use formula $-\int_{y1}^{y2} y^{-1} dy = +\int_{y2}^{y1} y^{-1} dy$ we get:

$$K = \frac{L_s}{t2 - t1} \ln \frac{y1}{y2} \tag{12}$$

The value of hydraulic conductivity K ($M.T^{-1}$) can be calculated directly by formula (12), which was derived with use of Darcy's Law.

By substitution of the corresponding couple of values of y (M) and t (T), we can get the finding value of hydraulic conductivity K ($M.T^{-1}$).

Problem:

In a laboratory falling-head permeameter (apparatus) the water level at the beginning of measurement was y_0 = 35 cm. After 2 hours and 48,5 minutes the water level dropped down to 1,7 cm. Length of the soil sample, placed in the permeameter (apparatus) is L_s = 20 cm.

Can you approximate the value of hydraulic conductivity K ($M.T^{-1}$) of soil sample in permeameter (apparatus)?

Instructions:

Draw simple scheme and determine the corresponding couple of values of y (M) and t (T). Realize, that for y_0 = 35 cm is corresponding time t_0 = 0 minutes.

In the scheme define the rested of couple of values (y_1, t_1), substitute to the formula (8) and calculate the value of hydraulic conductivity.

Result:

(K = $9{,}85.10^{-7}$ $m.s^{-1}$)

2.2.1.b. Constant-Head Method

Description (principle):

A constant difference in head is created over an undisturbed soil sample in a Kopecky steel ring. At certain times, the volume of water that has flowed through the sample is measured. From this discharge, the size of the soil sample, and from the difference of head, can be calculated the value of hydraulic conductivity.

Use (application):

Laboratory method is good tool to measure hydraulic conductivity of a certain layer in horizontal or in vertical direction.

Disadvantages (limitations):

The measured value is valid for the relatively small soil sample area only, so there may be a large error. This laboratory method is not suitable for samples with extremely high or with very low hydraulic conductivity.

Problem:

Can you write down final formula for calculation of the hydraulic conductivity?

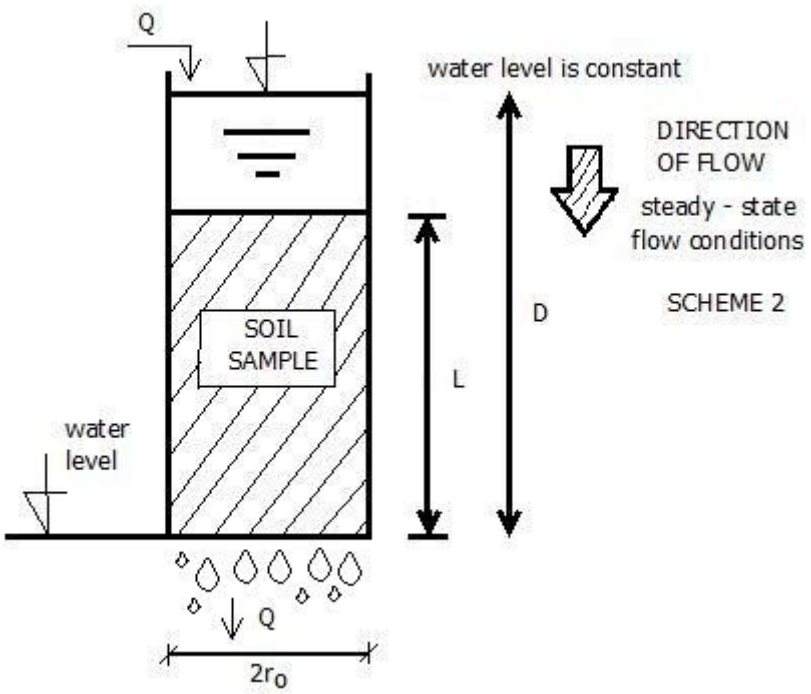

Figure 3. Constant head permeameter, water level is identical with referential level situated just above the drops falling down, is as a matter of fact the continuous free water level.

Theory:

Darcy's Law can be apply as $v = K\dfrac{D}{L_s}$, equation of continuity can be formed as $Q = S.v$, where v ($M.T^{-1}$) is a velocity of water flow in permeameter, S (M^2) is water profile of permeameter in the direction of flow and is valid $S = \pi.r_0^{\ 2}$.

Measured Q ($M^3.T^{-1}$) is the constant flow of water that has flowed through the soil sample. L_s (M) is a length of soil sample (see Figure 3).

By substituting formula of Darcy's Law to the equation of continuity and after correction we get expression for direct calculation of hydraulic conductivity in a shape:

$$K = \frac{L_s}{D}\frac{Q}{S} \tag{13}$$

From the known values of L_s (M), D (M) and Q ($M^3.T^{-1}$) and from the known radius r_0 (M) of permeameter, the value of hydraulic conductivity K can be calculated.

Problem:

From the measurement at the laboratory constant-head permeameter for determination of the value of hydraulic conductivity of soil sample in a steady state saturated flow conditions, following data were obtained:

L_s = 20 cm
D = 30 cm
Q = 2.10^{-4} l/s
r_0 = 5 cm

Try to draw simple scheme of laboratory constant-head permeameter for determination the value of hydraulic conductivity of soil sample in a steady state saturated flow conditions.

Define the referential level with connection of position of soil sample and known values of constant difference of heads D (M), constant water flow (recharge) Q ($M^3.T^{-1}$) to the permeameter, water profile of permeamater S (M^2) and a constant length L (M) of soil sample.

By formula (13) calculate the value of hydraulic conductivity K ($M.T^{-1}$).

Result:

$(K = 5{,}1.10^{-5} m.s^{-1})$

Some other data can be received from laboratory analysis. Depending on the analysis required, disturbed samples can be taken from the auger or by soil pits. If needed, undisturbed soil samples may be taken usually in special sampling cylinders. The disturbed samples are used to analyse texture, organic content, nutrients and soil water mix ratios. Undisturbed soil samples are analysed bulk density, porosity and saturated and unsaturated hydraulic conductivity.

2.3. Field Methods

In a case of a high level of groundwater table, the auger-hole method is used. Infiltration methods, inversed auger-hole method, Guelph permeameter method or another experimental methods based on water infiltration into the soil profile can be used in a case of absence of groundwater.

2.3.1. Examples of Small-Scale Field Methods

2.3.1.a. Piezometer Methods

Description (principle):

Non-perforated pipe is placed into the hole under the water level, leaving only a small cavity at the bottom. This means that water can only flows through this cavity.

The water table in the hole is decreased with a bailer then the rate of rise of the water table is measured. From this water recharge and from the geometry of the cavity, can be calculated the value of hydraulic conductivity.

Use (application):

This field method can be used for measuring the hydraulic conductivity of layers at relatively great depth or of separate soil layers. Piezometer method is not used in practice very often. This method serves for estimation of impact of soils heterogeneity and also for differentiation of horizontal and vertical components.

Disadvantages (limitations):

The value of hydraulic conductivity represents only the direct surrounding of the small cavity.

2.3.1.b. Infiltrometer Methods

Because of large heterogeneity in the soil conditions of Czech Republic, is not very often fulfilled boundary condition of half-infinity homogenous soil profile.

2.3.1.b.1. Double Ring Infiltrometer Method

Hydraulic conductivity can be also estimated from the results of the field infiltration measurements.

Description (principle):

A large steel ring is placed on the soil surface and struck some centimetres into the soil. In the ring a constant water head is maintained above the soil surface. The amount of percolating water is continuously measured until a constant infiltration rate is reached.

Use (application):

The developed infiltration process serves for measurement an infiltration rate (capacity) for irrigation and drainage design or to decide about a surface or subsurface drainage system. Infiltration processes can be also used for determining of hydraulic conductivity of soil layers.

Double ring infiltrometer method

This procedure is very well known and is used very often, when the groundwater table is absent. The principle of double ring infiltrometer method is presented in Photo 6 and Photo 7. In the unsaturated soils two concentric infiltrometer rings are placed at the certain depth, where the infiltration properties (hydraulic conductivity) will be measured (see Photo 6, 7).

Photo 6. Hydraulic conductivity measurement of grassy surface by double ring infiltrometer method (J. Štibinger).

The soil bellow and around the rings is saturated by infiltration. The rings are then filled the water and the rate of fall of the levels in both outer and inner ring is measured. The rings are then refilled by water, the water level in outer ring is kept constant, and the rate of fall in the inner ring being recorded.

Photo 7, 8. Detail view of measuring of hydraulic conductivity by double ring infiltrometer method (J. Štibinger).

It means, that in inner ring is recorded certain infiltrated height of water (e.g. 5,0 mm) and at the same time is recorded continued time. In inner ring is always refilled amount of water, which represented recorder infiltrated height (at this case 5.0 mm). The procedure is repeated. The water level in the outer ring being constant and is kept approximately at the same level as is the water table in inner ring.

The role of the outer ring is to minimize horizontal flow below the inner ring. The results are almost vertical flow path below the inner ring, where the data are measured. This process is known as cumulative infiltration by ponding. The example of the record from the beginning of cumulative infiltration is shown in Table 3.

Table 3. Double ring infiltrometer method – record of infiltration process

Number	**Hour**	**Time t (sec.)**	**Cumulative infiltration i(t) (mm)**
1.	13:10:00	0	0
2.		67	5
3.		245	10
4.		438	15
5.		679	20

After some time the infiltration rate stabilizes and approximates the hydraulic conductivity.

Disadvantages (limitations):

Results depend on actual moisture content of the soil (sorptivity); only values for the top layer (measured layer) can be found. Boundary effects may cause errors.

Theory:

The process of infiltration created by double ring infiltration method is in possession of non-steady unsaturated flow and can be expressed by Richard's equation in a form

$$\frac{\partial\left[k(H).(\partial H/\partial z)\right]}{\partial z}+\frac{\partial k(H)}{\partial z}=\frac{\partial W}{\partial t} \tag{14}$$

where k(H) is unsaturated hydraulic conductivity dependant on negative pressure (suction) H, parameter z respectively t, is vertical axis (positive in down direction) respectively time. Parameter W represents soil moisture.

This expression (14) can be marked as a type of Fokker-Planck Equation, which is significantly nonlinear and in its closed form is very difficult to use analytical solution. Philip's solution of equation (14) by method of perturbation (Philip 1977) comes from diffusion form of equation (14) and after many further modifications is formed as

$$i(t)=St^{1/2}+At \tag{15}$$

and

$$v(t)=(1/2)St^{-1/2}+A \tag{16}$$

where i(t) (M) is cumulative infiltration (see Table T1), v(t) ($M.T^{-1}$) is intensity of infiltration, parameter S ($M.T^{-0.5}$) represents sorptivity and coefficient A ($M.T^{-1}$) characterises long term of infiltration and approximates hydraulic conductivity K ($M.T^{-1}$).

The first part of the left side of equation (15) of cumulative infiltration $S.t^{1/2}$ affects the process at the beginning of infiltration. After some time the infiltration rate is stabilized and

the expression $A.t$ begin to predominate. The expression $A.t$ is sometimes called "gravity" part.

Intensity of infiltration is determined by differentiation of equation of cumulative infiltration (15) by time t (M), expression $v(t) = \frac{di(t)}{dt}$ is valid and at the same time it has to be valid expression $i(t) = \int v(t)dt$.

Equations (15) and (16) are very well known as reduced Philip's Infiltration Equations. Both formulas are fully verified and are very often used not only in water engineering practice, but also in scientific sphere (Štibinger, 2006).

Determining the sorptivity S ($M.T^{-0.5}$) and coefficient of long term infiltration A ($M.T^{-1}$):

Determining of sorptivity S ($M.T^{-0.5}$) and coefficient of long term infiltration A($M.T^{-1}$) from the records of cumulative infiltration (see Table P1) is mostly based on the application of the suitable statistical methods. At this case can be used (besides others) the methods of the linear and non-linear regression which are the good tools just for this purpose.

The estimation of S ($M.T^{-0.5}$) and A ($M.T^{-1}$), parameters of formula (15), and also the correctness of the shape of the infiltration equation (15), can be proved by means of the nonlinear regression processing by the following procedure.

Let us have at one's disposal the data from the record of cumulative infiltration i(1), i(2),….i(i),….i(n) at the corresponding time t_1, t_2,….t_i,…..tn.

Formula 13 can be expressed as integral curve by vector Y, vector X, parameter P1 and parameter P2 and as:

$$Y = P1.X^{1/2} + P2.X \tag{17}$$

where:

Y represents the cumulative infiltration i(t) (M)
X represents time t (T)
P1 represents sorptivity S ($M.T^{-0.5}$)
P2 represents coefficient of long term infiltration A ($M.T^{-1}$), which approximates hydraulic conductivity K ($M.T^{-1}$).

The equation (15) represents in reality a time series and can be formed as:

$$Y[i(1), i(2),..i(i),..i(n)] = P1.X[t(1), t(2),..t(i),..t(n)]^{0.5} + P2.X[t(1), t(2),..t(i),..t(n)] \tag{18}$$

where the vector $Y[\ i(1), i(2),....i(i),....i(n)\ ]$ represents dependant variables and vector $X[\ t(1), t(2),....t(i),....t(n)\ ]$ represents independent variables.

By nonlinear regression analysis of the equation 18 can be determined parameter P1, which represents sorptivity S ($M.T^{-0.5}$) and parameter P2, which represents coefficient of long term infiltration A ($M.T^{-1}$).

The vector $Y[\ i(1), i(2),....i(i),....i(n)\ ]$, vector $X[\ t(1), t(2),....t(i),....t(n)\ ]$ and the shape of the equation (18) are known values (known information), unknown values of P1 and P2 will be calculated by nonlinear regression, e.g. with use of the method of Marquardt (Marquardt 1963), known also as the method of Marquardt and Levenberg. The estimation of the values of P1 and P2, that fit the data best, is the goal of the nonlinear regression processing.

By evaluation of the parameters P1 and P2 can be checked not only the time series from the recorded cumulative infiltration of field measurement, but also the correctness of the shape of the equation (15).

Determining sorptivity S ($M.T^{-0.5}$) and coefficient of long term infiltration A ($M.T^{-1}$) – practical example 1:

Infiltration measurement of surface layer - loamy soils with grassy cover. Depth of the placement of inner ring was 10 cm (below the surface). Record of the cumulative infiltration measurement is presented in Table 4, graph of the time series of cumulative infiltration is viewed on Figure 4.

Table 4. Double ring infiltrometer method – record of cumulative infiltration

Number	**Time t (min)**	**Cumulative infiltration i(t) (mm)**
1	0	0
2	6,0	6
3	15,0	12
4	22,0	18
5	32,0	24
6	39,0	30
7	48,0	36
8	56,0	42
9	66,5	48
10	80,5	54
11	95,0	60

For determining of sorptivity S ($M.T^{-0.5}$) and for determining of the coefficient of long term infiltration A ($M.T^{-1}$) according equation (CC) was used non-liner regression analysis with method of Marquardt and Levenberg. For estimation was used commercial software STATISTICA.

Selected final results:

Dependent variable: i(t)
Number of independent variables: 1
Number of estimation: 11
Number estimated parameters: 2
Function: least squares
Final value of function: 22.43278771
Index of determination: R2 = 0.99433
Index of correlation: R = 0.99716

Method of estimation: Levenberg-Marquardt
Number of iteration: 3
Initial values of parameters:
0.1, 0.1

Final values of parameters:
P1 = S = 1.67498379 ($mm/min^{0,5}$) = $2,1.10^{-4}$ ($m/s^{-0.5}$)

P2 = A = 0.48626302 (mm/min) = 8,1.10^{-6} (m.s^{-1})
The estimated value of sorptivity S (M.T$^{-0.5}$) is S = 2.1 .10^{-4} (m.s$^{-0.5}$).

Hydraulic conductivity K (M.T^{-1}) can be approximated as a coefficient of long term infiltration A (M.T^{-1}) and it means that the hydraulic conductivity K (M.T^{-1}) of surface soil layer of depth 10 cm with grass is K = 8,1.10^{-6} (m.s^{-1}).

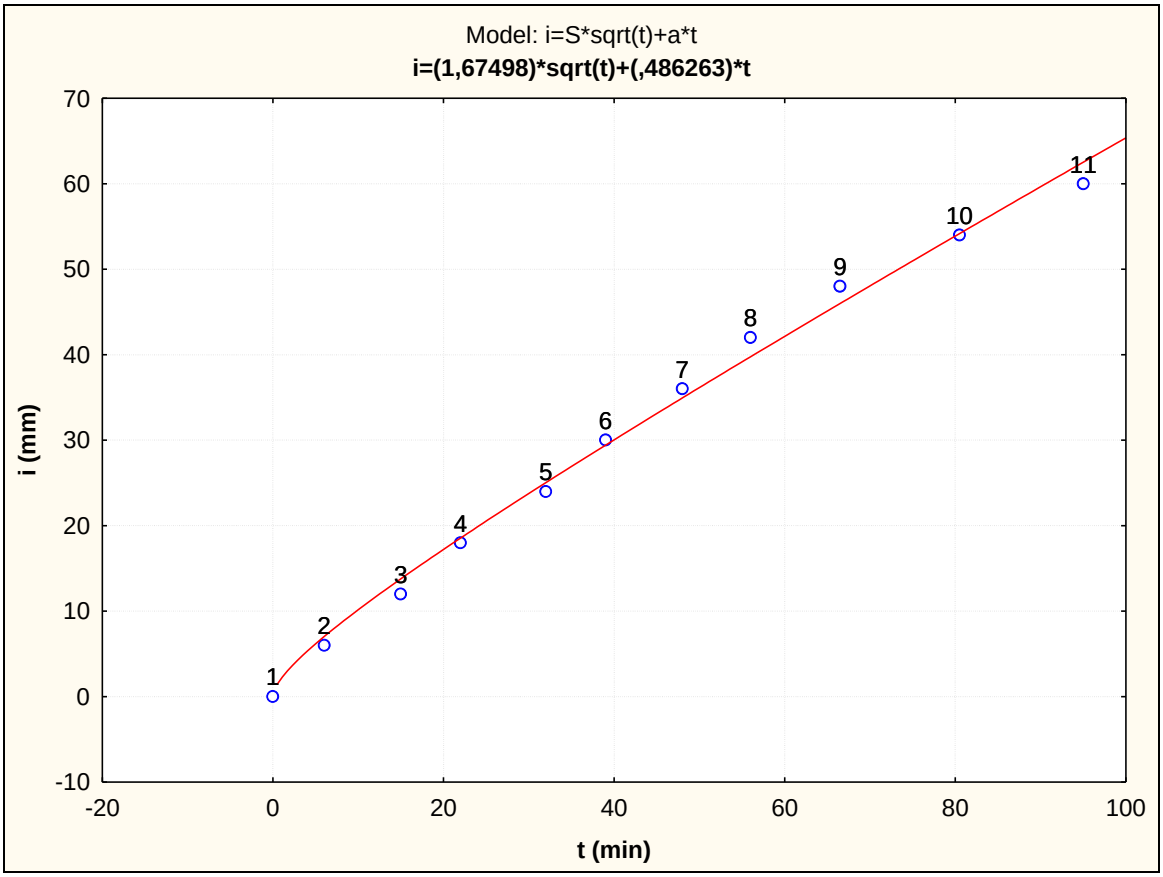

Figure 4. Time series of measured values of cumulative infiltration i(t) (mm).

Approximation sorptivity S (M.T$^{-0.5}$) and coefficient of long term infiltration A (M.T^{-1}) directly from equation of cumulative infiltration – practical example 2:

In this case is sorptivity Sp (M.T$^{-0.5}$) and coefficient of long term infiltration Ap (M.T^{-1}) approximated without any statistical procedures.

The record of cumulative infiltration from the field measurement is available (see Table 4).

The expression (At) approximates the process after some time of infiltration and it means that for the end of the record the formula $i(t) = At$ is (approximately) valid. We get:

i(t) = i(t)n – i(t)n-1 = 60 – 54 = 6 mm

t = tn – tn-1 = 95 – 80.5 = 14.5 min.

i(t) = Ap . t

6 = Ap . 14.5

Ap = 6 / 14.5 = 0.41 mm.min^{-1} = 6.89.10^{-6} m.s^{-1} = 7. 10^{-6} m.s^{-1}.

Ap ($M.T^{-1}$) is coefficient of long term infiltration which is approximated directly from the equation of cumulative infiltration (after some time of infiltration process).

Note:
Infiltration process can be also analysed by a very well-known Green-Ampt method, based on wetting front approximation (Kutílek and Nielsen 1994).

Approximation of sorptivity Sp ($M.T^{-0.5}$) is made by calculation according to equation (8) with use of value of Ap = 0.41 $mm.min^{-1}$ (see above) and with use of the measurement number 3. Then we get $12 = Spx15^{-0.5} + 0.41x15$. This expression is valid for Sp = 1.51 $mm.min^{-0.5}$.

The comparison of these results with the results obtained by non-linear regression analysis with use of the Marquardt-Levenberg method, where sorptivity S = 1.67 ($mm/min^{0,5}$) and long term infiltration coefficient A = 0.48 ($mm.min^{-1}$), shows excellent conformity (almost identity) of the tested values.

It means that method of the directly determining from the equation of cumulative infiltration (7), with respect of specific conditions of infiltration process mentioned and described above, is a good tool for approximation of S ($M.T^{-0.5}$) and A ($M.T^{-1}$) from the terrain infiltration measurements.

Note, that in a case of enormous variances of measured data of cumulative infiltration (it means i(t), versus t) from the equation of cumulative infiltration (15), this way of approximation is not so reliable.

It should be mentioned, that exist a lot of another field infiltrometer methods to determine hydraulic conductivity, K-values, for measurement in soil profile above water table, which were not presented in the previous text. Very important and suitable for use when groundwater level in soil profile is not present is Guelph Permeameter Method.

2.3.1.b.2. Guelph Permeameter Method:

This method is based on steady state flow q ($M^3.T^{-1}$) which is measured with constant head h (M) of water level above the bottom of auger-hole of certain radius r (M).

Soil profile around the hole is not saturated, marriotte flask in permeameter is used to simulate fully saturated zone. It is supposed, that after relatively short time period (20 – 30 minutes) quasi-steady flow is reached in the investigated soil profile.

Hydraulic conductivity K ($M.T^{-}1$) can be approximated by formula

$$K = \frac{q\left[\sinh^{-1}(h/r) - (h^2/r^2 - 1)^{0.5} + r/h\right]}{2\pi h^2} \tag{19}$$

where condition r > h has to be valid.

Problem:

In the auger-hole with diameter DM = 200 mm is by steady recharge q = 0.05 litres per second water level head h = 5 cm above the bottom of hole.

How is the value of hydraulic conductivity K ($m.s^{-1}$)?

How permeable is the soil profile?

2.3.1.b.3. Single Ring Infiltrometer Method to Water Layer on Soil Surface

In some cases can be the soil surface below the relatively thin water layer. Determination of hydraulic conductivity of soil surface with a permanent thin water layer can be carried out a by field experimental measurements and testing with use of the single ring infiltrometer method.

Description (principle):

A single steel ring is placed on the soil surface with the water layer and struck a few centimetres into the soil. Free water level on the surface is situated under the upper rim of the steel ring. By an addition of the certain amount of water into the single ring infiltrometer, the water table level inside the ring goes up, and then by gravity the initial water level in the ring infiltrometer gradually decreases.

Use (application):

The developed process of subsurface water flow in the ring serves for measurement of hydraulic conductivity of surface soil layers under the permanent (continuous) water table.

Single ring infiltrometer method:

Single ring infiltrometer is pushed into the soil with the thin water layer on the surface. The lower rim of the ring is a few centimetres below the soil surface and below the water level. The soil under the surface and around the ring is fully saturated. Into the ring is then added the water and the rate of fall of the level in single ring infiltrometer is measured. The principle of single ring infiltrometer method is presented in Figure 5 and viewed in Photo 9.

Disadvantages (limitations):

Only values for the top layer (measured layer) can be found. Boundary effects may cause errors and relatively small sample area can bring the higher possibility of a large random error.

Photo 9. Single ring infiltrometer in an experimental area of paddy fields for measurement of hydraulic conductivity of top soil layer (J. Štibinger).

Theory:

In Figure 5 is shown single ring infiltrometer in the one-dimensional system with vertical y-direction, positive upward, and horizontal referential level, which is identical with water table level on the soil surface of experimental field.

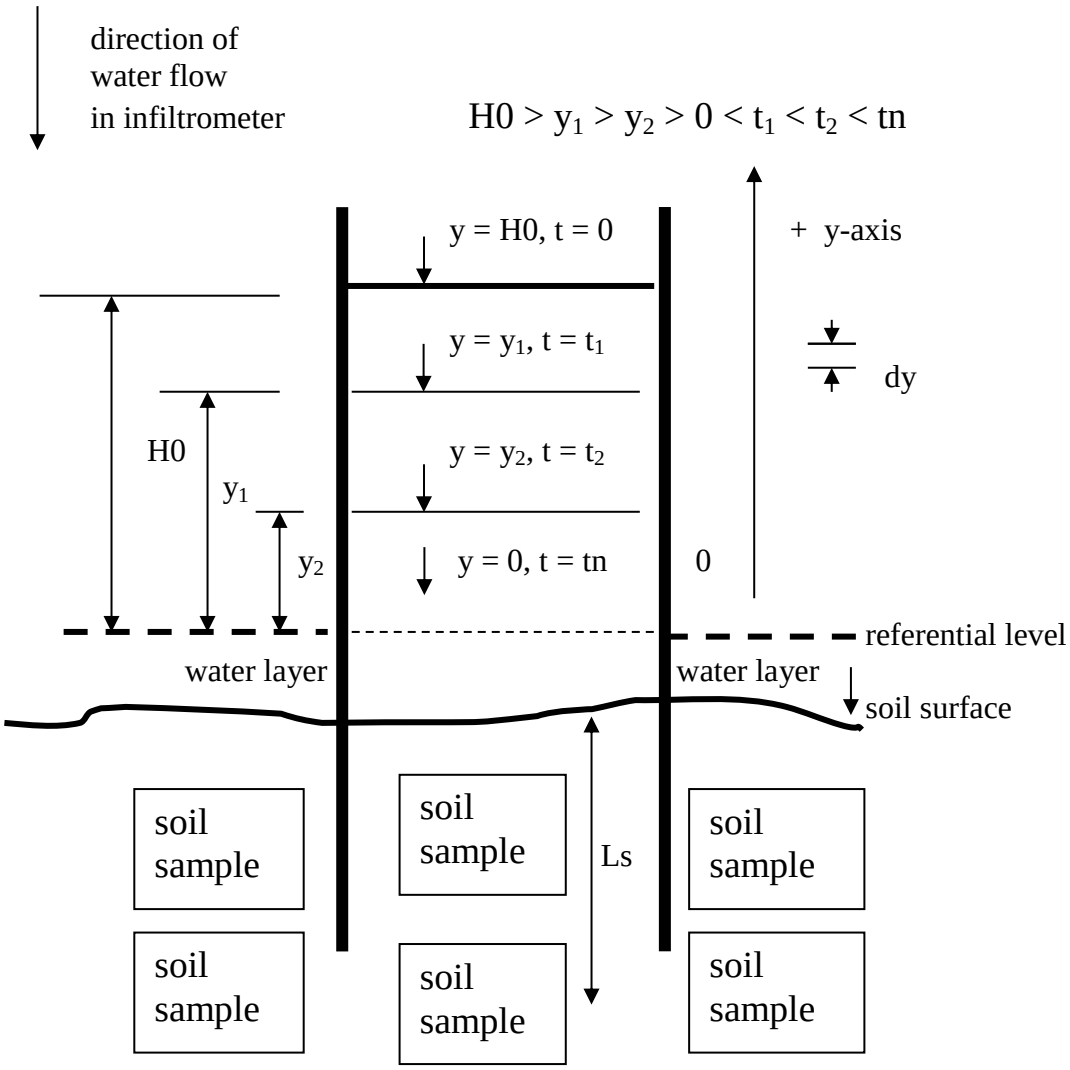

Figure 5. Process of decreasing of water table level in the single ring infiltrometer in an experimental area

By an addition of the certain amount of water into the single ring infiltrometer, the water table level inside the infiltrometer goes up to the level H0 (M), which represents water table level (M) in infiltrometer at the beginning of measurement, at time t = 0. By gravity the water level in infiltrometer gradually goes down from the level H0 to the level y_1, y_2, etc., finally to y = 0, from the time t = 0 to the corresponding times t_1, t_2, etc., finally tn. Nevertheless it is valid H0 > y_1 > y_2 > 0 and 0 < t_1 < t2 < tn (see Figure 5). In saturated flow conditions and according to Darcy's law the flow velocity v_s ($M.T^{-1}$) in the soil sample of experimental infiltrometer in the vertical y-direction can be approximated as:

$$v_s = -K \frac{y}{2L_s} \tag{20}$$

where L_s (M) represents the depth of the infiltrometer under the soil surface. The value of $2L_s$ (M) characterizes the approximation of the supposed trajectory of the water particles in the terrain tested soil. Negative sign in a right part of equation (20) marks the flow, which is running down, it means opposite to the positive direction of y-axis.

The flow velocity v_i ($M.T^{-1}$) in the infiltrometer can be expressed as:

$$v_i = \frac{\partial y}{\partial t} \tag{21}$$

It is supposed that the flow velocity v_s ($M.T^{-1}$) in the soil sample of experimental infiltrometer in y-direction will be equal as the flow velocity v_i ($M.T^{-1}$) in the infiltrometer $(v_s \equiv v_i)$ and we obtain:

$$-K\frac{y}{2L_s} = \frac{\partial y}{\partial t} \tag{22}$$

From rearrangement equation (22) we get ordinary differential equation, which can be formed as:

$$\partial t = (\frac{-2L_s}{K}).y^{-1}\partial y \tag{23}$$

Let's suppose, that the decreasing of water table level in infiltrometer will be e.g. from level y_1 (M) to level y_2 (M). The corresponding times are t_1 (T) and t_2 (M). Integration of equation (23) can be expressed as:

$$\int_{t1}^{t2} dt = (\frac{-2L_s}{K})\int_{y1}^{y2} y^{-1}dy \tag{24}$$

After integration from y_1 to y_2 in y-axis direction and from t_1 to t_2 in time axis and after other corrections and with use formula $-\int_{y1}^{y2} y^{-1}dy = +\int_{y2}^{y1} y^{-1}dy$ we get:

$$\frac{y1}{y2} = \exp(\frac{K}{2L_s}[t2 - t1]) \tag{25}$$

Formula (25) can be expressed as integral curve by vector Y, vector X, parameter P1 and constant H0 as:

$$Y(y1, y2....yi.....yn) = \frac{H0}{\exp[P1.X(t1, t2....ti....tn)]} \tag{26}$$

P1 is a function of hydraulic saturated conductivity K ($M.T^{-1}$) and supposed trajectory of the water particles $2L_s$ (M) in the terrain tested soil sample. Parameter P1 can be expressed as $P1 = K/2L_s$ (T^{-1}). H0 represents the water table level (M) in infiltrometer at time t = 0. The vector $Y(y_1,y_2.....y_i.....y_n)$ represents dependant variables and vector $X(t_1,t_2.....t_i,....t_n)$ represents independent variables.

By nonlinear regression analysis of the equation (26) can be determined parameter P1 (T^{-1}). Then from the known approximated value of $2L_s$ (M) is possible to calculate the value of hydraulic saturated conductivity K ($M.T^{-1}$) in the terrain tested soil sample by expression P1 (T^{-1}) = $K/2L_s$, respectively K ($M.\ T^{-1}$) = $P1.2L_s$.

The vector $Y(y_1,y_2.....y_i.....y_n)$, vector $X(t_1,t_2.....t_i....t_n)$, constant H0 and the shape of the equation (9) are known values (information), unknown value of P1 will be calculated by nonlinear regression, with use the method of Marquardt (Marquardt 1963), known also as the method of Marquardt and Levenberg. The estimation of the value of P1, that fit the data best, is the goal of the nonlinear regression processing (GraphPad Software. Inc., 2001).

The value of hydraulic conductivity K ($M.T^{-1}$) can be also approximate directly by formula

$$K = \frac{2L_s}{t2 - t1} \ln \frac{y1}{y2} \tag{27}$$

which was derived from equation (25). By substitution of the corresponding values of (M) and t (T) we can get finding value of hydraulic conductivity K ($M.T^{-1}$).

Area of use:

Swampy areas like wetlands, peat swamps, some part of river landscape situated under the water table, but also paddy fields, can serve as a good tool for mitigation of negative impact of hydrological extremes as are floods and long term droughts. This fact is very important, especially in connection with the present climate dynamics.

In some part in Czech Republic (see Photo 10, 11) is necessary to reclaim and to revitalize the localities mentioned above, to protect and to save their water regime. Good design of water management of swampy areas with focusing on suitable bio-technical structures has to be based, besides others, on the correct description and analysis of groundwater (subsurface water) flow in the saturated soil porous environment.

Photo 10. Bottomland forest with evient water layer on the surface (Břeclav, South Moravia, Czech Republic, J. Štibinger).

Photo 11. Moving barrier on channel to manage subsurface water regime in bottomland forest around Lanžhot city (South Moravia, Czech Republic, J. Štibinger).

Water management of paddy fields (rice fields) belongs to the same category and in the countries like China, Vietnam, Indonesia, Japan, India, Thailand, Philippines, Egypt etc., has also large economical impact (Photo 12, 13).

Photo 12. Water management of paddy fields (rice fields) by system of ditches in surroundings of Red River (Hanoi, Vietnam, J. Štibinger).

Photo 13. Paddy field area divided by concrete wall with controlled opening for water management. Municipality of Taoyuan (Taipei City, Taiwan, Republic of China, (J. Štibinger).

Correct determination of hydraulic conductivity K ($M.T^{-1}$), which comes from direct terrain measurements, is very important and necessary step to reach a good protection of water regime in swampy areas and to set up reasonable water management in paddy fields for environmental and economical agriculture.

Following show in a detail presents the unique experimental terrain field measuring of hydraulic conductivity at the surface of swampy areas, where the soil surface is permanently under the thin water layer.

This method can serve for the determination of hydraulic conductivity especially in a case of a really fully saturated soil porous environment as can be e.g. wetlands, swampy areas, peat swamps, some part of river landscape situated under the water table, paddy fields, with permanent water layer above the surface.

As a typical real example of determination of the hydraulic conductivity was selected the direct terrain experimental measurement by single ring infiltrometer on paddy fields in Taoyuan (Republic of China, Taiwan, Taipei) during the cycle of flooded period (see Photo 14).

Photo 14. Parts of the rice field in Taoyuan (Taipei City, Taiwan, Republic of China) with home-made hydraulic divider during the cycle of flooded period (J. Štibinger).

Hydraulic conductivity is one of the principal soil hydrology characteristics, especially in a case of soils of paddy fields. At the same time, hydraulic conductivity is one of the most important factors in water management design of paddy fields, while it is indispensable for the control of paddy field water regime. In this case, soil layers on paddy fields in Taiwan, measured directly on the fields during the cycle of the flooded period, were used to determine the saturated hydraulic conductivity of the surface. Indirect methods, which use the soil's textural characteristics, do not necessarily provide true values.

The value of hydraulic conductivity K ($M.T^{-1}$) of the surface layers of paddy fields should be relatively small to minimize the seepage losses. But the soils of surface layers of paddy fields should not be definitely impermeable, because it is better if after the flooded period, when the surface waters flayed off from the surface, the water regime of paddy fields can be under the control of the horizontal pipe drainage systems.

The typical picture of the most of the Vietnamese paddy fields from Mekong delta and from basins of Red River is a presence of the very low permeable thin layer with saturated hydraulic conductivity perhaps less than 1.10^{-6} $m.s^{-1}$, placed approximately 60 or 70 cm under the surface of paddy field. Soils surface layers of fluvial deposits above this low permeable layer will be much more permeable (Nguyen, 2007).

Egyptian paddy fields in Sakha (Rycrroft and Mohamed, 1995), situated in the northern part of the Nile delta, dispose of un-drained heavy textured saline-sodic soils of clay, composed mainly montmorillonite, with some kaolinite and illite, with saturated hydraulic conductivity about $1{,}4.10^{-6}$ $m.s^{-1}$.

Another Egyptian paddy fields, located in Mashtul (ILRI, 2008), in the south eastern part of the Nile delta, consists of soils with clay cap, which contains dark brown stiff clay, without cracks the saturated hydraulic conductivity is about $9{,}3.10^{-7}$ $m.s^{-1}$.

Hydraulic conductivities of the soil surface layers of paddy fields in Taoyuan (Republic of China, Taiwan), were measured on selected experimental area during the cycle of flooded period by the direct field experimental measurements, with use the of individual single ring infiltrometers (Kutílek and Nielsen, 1994).

Experimental paddy field area (Taoyuan, suburb of Taipei, Republic of China, Taiwan)

Photo 15. Experimental paddy field area situated at the district of Taoyuan, suburb of Taipei (Taiwan, Republic of China, J. Štibinger).

In Photo 15 is viewed a suburb of Taipei (Republic of China, Taiwan) which is characterized, besides others, by the big amount of greater and smaller paddy fields. They are situated even in the very urbanized parts of the sub-cities. The experimental paddy field area 10 m x 10 m was selected from the paddy fields, placed in the Taipei's suburbs.

The locality of interest approximately from 4 to 6 hectares, with an experimental paddy field area, is situated in Taoyuan (suburb of Taipei), very close to the main street Chungshan Road, around 50 minutes by subway from the second world's highest skyscraper "101" (Štibinger, 2001).

The landscape in the place of experimental paddy field area and in the close neighbourhood is almost flat. The altitude of this place is approximately a few tens of meters above the sea level. The climate is warm and humid, with the annual average temperature about 210 °C. The monthly average temperature is 150 °C in January and 280 °C in July. The total annual precipitation amounts to 2 450 mm, with more rainfall in monsoon season between middle of May and middle of June and in typhoon season between July and September.

The study was performed at the paddy field area on the Anthrosois, which were fine-texture. They have been formed on the alluvial deposit of the local no-named small streams. Permeability of soil layers varied from low to average values. The impervious barrier was approximately from three to four meters below the soil surface. The studied top-layer was fine-texture, homogenous and isotropic.

Results – measured and calculated values

Terrain experiments on the experimental area of paddy field were made during the cycle of flooded period, what means after the rice planting, which was realized after irrigation of paddy field. During terrain experimental measurement of saturated hydraulic conductivities, was the surface of the experimental paddy field permanently under the water table. The thickness of water layer above the surface of experimental paddy field fluctuated approximately from 5 cm to 8 cm. For one's own measurement of saturated hydraulic conductivity was used single ring infiltrometers.

A few centimeters (from 6 cm to 10 cm) under the surface of soil layers of the experimental area of paddy field were vertically pressed down single ring infiltrometers of heights 20 cm, with inside diameters 8,5 cm. Scheme of the single ring infiltrometer stepped at the experimental paddy field views Figure 6.

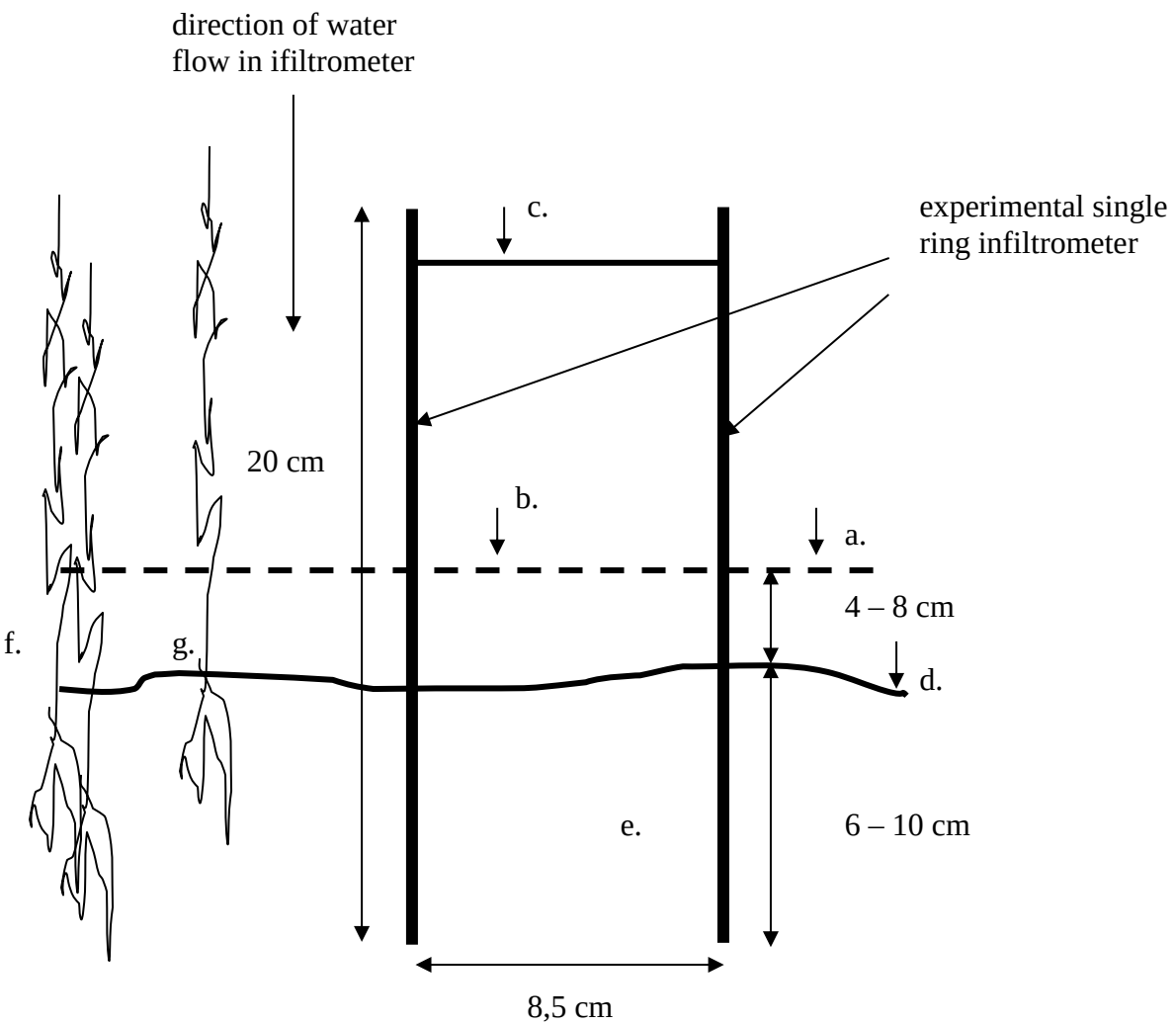

Figure 6. Scheme of single ring infiltrometer stepped at an experimental area of paddy field in Taoyuan (Taipei, Taiwan).

Legend:

a. water table level (w.t.l.) on the experimental paddy field, outside of infiltrometer

b. w.t.l. inside of infiltrometer before of addition of water

c. w.t.l. inside of infiltrometer after of addition of water

d. surface of soil layers of experimental paddy field

e. soil sample of experimental paddy field

f., g. rice plants planted after irrigation and before experimental measurement

By this way, with help of the single ring infiltrometers, were formed individual samples of soil in paddy fields, on which were measured, directly in the terrain, the data wanting for estimation of the saturated hydraulic conductivities. At the beginning of measurement is the height of the water layer above the surface of the area of paddy field (outside of infiltrometers) the same as the height of the water layer inside of the infiltrometers. Usually it is a few centimeters of water (e.g. from 4 cm to 8 cm), which is necessary to keep above the surface of paddy field during the flooded period.

By an addition of the certain amount of water into the infiltrometers, the water table level inside the infiltrometers increases to the level H0 (M). Falling water heads inside of infiltrometers, developed by gravity, in time t = 0 from level H0 (M) gradually to the y_1 (M), y_2 (M) etc. (see Figure 5) are measured by the millimeter's scale and recorded as the time series of decreasing of water table levels inside of individual infiltrometers.

A representative example of the record of the fallings of the water levels inside of the infiltrometers measured directly in the paddy field presents Table 5.

Table 5. Time series of measured and calculated values from the experimental area of paddy field in Taoyuan (Taipei, Taiwan)

20.03. 2001 Number of observation	**Hours**	**Time (minutes)**	**Water table level– measured values (mm)**	**Water table level– calculated values (mm)**
1	7.02	0	90,0	90,0
2	8.22	80	86,0	84,0
3	9.30	148	80,0	79,0
4	10.30	208	76,0	75,0
5	11.30	268	71,0	71,3
6	12.30	328	67,0	67,7
7	13.04	362	65,0	65,7

Single ring infiltrometer of height 20 cm with inside diameter 8,5 cm was pressed vertically to the depth L_s (mm) = 70 under the soil surface of the experimental area of paddy field. The value of $2L_s$ (mm) = 140 characterizes the approximation of the supposed trajectory of the water particles in the terrain tested soil sample of paddy field.

The water table, which is identical with referential level, is approximately 4 cm above the surface of the experimental area of paddy field. A value of the level H0 (M), which represents the water table level in infiltrometer at the beginning of measurement at time t = 0 was H0 (mm) = 90.

A record of the vector of fallings Y (y_1, y_2…..y_i……y_n) is shown in the third column in Table 2 and represents Y-vector as dependent variable (known values). A vector of time X (t_1, t_2…..t_i….t_n) is presented in the second column in Table 5 and represents X-vector as independent variable (also known values).

Relation between vectors Y and X explains equation (26), where only unknown is parameter P1 (T^{-1}).

With use of the nonlinear regression actually with help of the Marquardt search algorithm was determined the value of the parameter P1 (min^{-1}) from the equation (25). The results of the nonlinear regression model fitting show, that P1 (min^{-1}) = 0,000868 and determination coefficient R–squared = 0,987.

It is accepted, that P1 = $K/2L_s$, this means, that according to nonlinear regression analysis will be hydraulic conductivity K ($mm.min^{-1}$) = parameter $P1.2L_s$ = 0,000868.140 = 0,1215 ($m.\ min^{-1}$) = 2.10^{-6} ($m.s^1$).

(The results of model fitting and the analysis of variance for the full regression are presented in Table 6).

Table 6. Results of nonlinear regression analysis from an experimental area of paddy field in Taoyuan (Taipei, Taiwan)

Model Fitting Results				
parameter	estimation		standard error	ratio
P1	0.000868		0.00002467	35.1996
Total iterations = 3			Total function evaluations = 7	
Analysis of Variance for the Full Regression				
source	sum of squares	degrees of freedom	mean square	ratio
Model	41420.245	1	41420.245	36789.271
Error	6.7552703	6	1.1258784	
Total	41427.000	7		
Total (corr.)	537.71429	6		
R – squared = 0.987437		R – correlation = 0.993698		

With help of the known value of P1 (min^{-1}) = 0,000868 and by X-vector (t [min] = t_1, t_2…..t_i….t_n), according equation (26), the new values of the falling water heads in time t, inside of infiltrometers, from the level H0 = 90 mm, were recalculated and placed in the last column in Table TT 1.

Measured data and their smoothing by nonlinear regression (equation /26/) are presented in Figure 7.

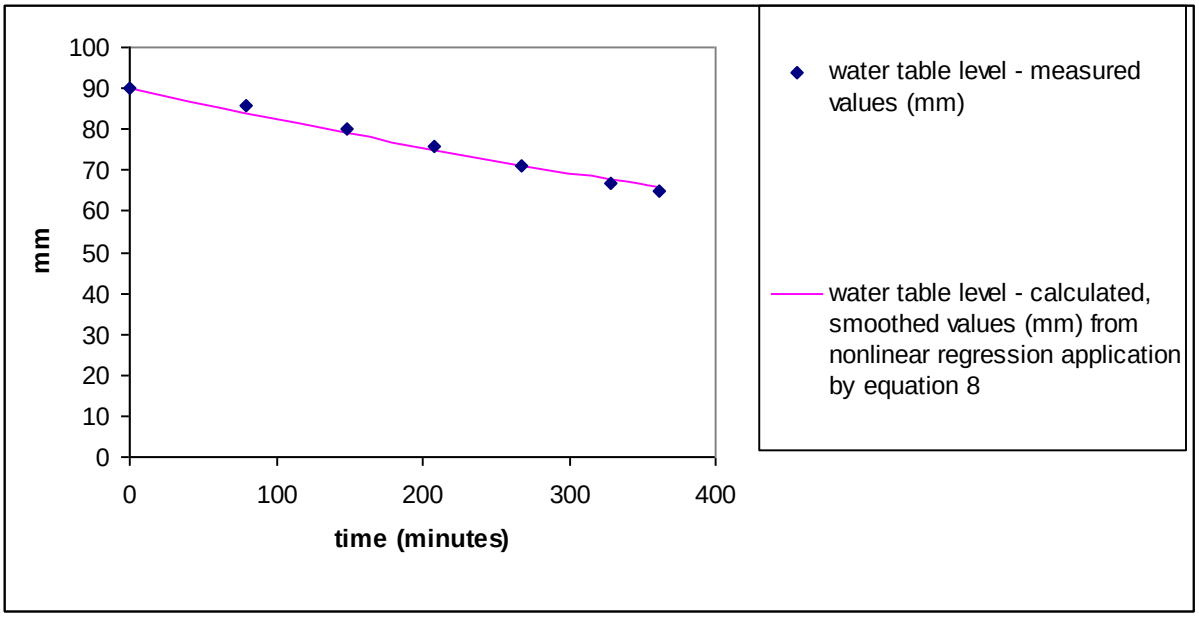

Figure 7. Measured values and calculated, smoothed values from nonlinear regression application by equation (26), from the experimental area of paddy field in Taoyuan (Taipei, Taiwan).

High value of the determination coefficient R–squared = 0,987 reflects particularly good predicating ability of tested model, represented by equation (26).

The terrain measurements on the experimental paddy field proceeded from March 15 to March 25, 2001, when were realised, all in all, the tens of experimental measuring. During the measurements of the saturated hydraulic conductivities, was not recorded any sudden recharge (e.g. typhoons, rain storms or massive floods) to the water table level on the experimental paddy field.

The values of hydraulic conductivities K ($m.s^{-1}$), which were approximated by the direct experimental measurement on the paddy field, varied from 5.10^{-6} to 10^{-7} ($m.s^{-1}$) and were successfully verified by the laboratory measurements on the "undisturbed" core samples in the non-steady state conditions by the falling head permeameters.

It is possible to suppose that the values of hydraulic conductivity are true enough, because they were approximated from the data obtained by the direct experimental measurements on paddy fields.

The data obtained to determine the hydraulic conductivities by direct field measurements during the cycle of the flooded period on paddy fields, that is after agricultural activities such as rice planting and paddy field maintenance were carried out, already reflect the impact of some surface layers compacting, which significantly influences the values of hydraulic conductivities.

Therefore the results of saturated hydraulic conductivities obtained by the way described above are very valuable and objective, and are useful for the control of water management of paddy fields. The values of hydraulic conductivities obtained from the laboratory by the verification of the experimental terrain measurements on paddy fields were fairly similar with the values of hydraulic conductivities obtained by the direct paddy field measurements.

All processes of non-linear regression that were applied to determine the hydraulic conductivities of surface layers of paddy fields yielded extremely high values of the determination coefficients R–squared, which fluctuated about 0,98, some even converged to 1. It not only shows the suitability of the method applied for the solution of the problem, but also a particular forecasting ability of the tested model, in this case represented by equation (26). It ensues from the above that this method is relevant to the given problem, and that the K ($M.T^{-1}$) value is the best and the most plausible basis upon which to design and control the rice fields water management. This is a very significant finding in respect to the rice growing and water saving.

The values of hydraulic conductivities, K-values, of surface layers of paddy fields determined in this way are credible enough. Such data are necessary for the design of basic parameters of paddy fields irrigation and drainage systems, while they can also be used for the verification of the correctness and hydraulic efficiency of the designed systems.

Last but not least, the data of hydraulic conductivities represent a suitable and simple tool for operative hydraulic control and optimization of paddy field water management, which will in result be friendly to water resources and will mitigate their vulnerability.

And just a method of determination of hydraulic conductivities of the surface layers in river landscape, situated under the water table (wetlands, peat swamps), can serve as a good tool to control and to manage the water regime in some parts of river landscape, as are e.g. wetlands or peat swamps, in the hydrological conditions of the Asia (Taiwan, Vietnam), Lithuania or Czech Republic.

Problem:

Verify, if the value of hydraulic conductivity $K = 2.10^{-6}$ ($m.s^{1}$) obtained by equation (9), with using of nonlinear regression (by help of the Marquardt), will be corresponding with hydraulic conductivity received by equation (10) for direct calculation.

Solution:

From Table TT 1 will be chosen corresponding couple of values of t (minutes) and y (mm) and substitute into equation (27).

First testing:

t_1 = 148 minutes, t_2 = 328 minutes and y_1 = 80 mm, y_2 = 67 mm. $2L_s$ = 140 mm.

$$K_1 = \frac{2L_s}{t2 - t1} \ln \frac{y1}{y2} = \frac{140}{328 - 148} \ln \frac{80}{67} = 0{,}138 mm/\min = 2{,}3.10^{-6} m/s$$

Second testing:

t_1 = 0,0 minutes, t_2 = 362 minutes and y_1 = 90 mm, y_2 = 65 mm. $2L_s$ = 140 mm.

$$K_2 = \frac{2L_s}{t2 - t1} \ln \frac{y1}{y2} = \frac{140}{362} \ln \frac{90}{65} = 0{,}125 mm/\min = 2{,}1.10^{-6} m/s$$

The differences among the hydraulic conductivity obtained by equation (26) /K = $2{,}0.10^{-6}$ ($m.s^{1}$)/ and the hydraulic conductivities received by direct calculation with equation (27) /$K_1 = 2{,}3.10^{-6}$ ($m.s^{1}$), $K_2 = 2{,}1.10^{-6}$ ($m.s^{1}$)/ are negligible.

At this case, the direct calculation, represented by equation (10), can be used for approximation of the value of hydraulic conductivity.

Question (note):

Can be used also another corresponding couple of values of t (minutes) and y (mm) from Table 2 with substituting into equation (10) to approximate the value of hydraulic conductivity?

How will be the results, respectively differences? Make simple analysis for this case.

Problem:

To protect water regime of bottomland forest situated at the margin of swampy area of Slokas Lakes (see Photo 16) in the locality of Kemeri National Park (Jurmala city, Lithuania), during summer the period and in connection with a present climate dynamics, it is necessary to determine the value of hydraulic conductivity K ($M.T^{-1}$) to approximate the water losses by infiltration.

Photo 16. Slokas Lakes Location – creation and reclamation some selected parts this area was financed by counsil of Jurmala City (Lithuania, Europe, J. Štibinger).

The water level in the locality of interest is approximately 5 cm above the soil surface. The single ring infiltrometer with length 35 cm (radius r_0 = 10 cm) was 20 cm pressed into the soil surface.

The water level dropped down from the level of full up water infiltrometer to the 1,0 cm above the original water level during 2 hours and 35 minutes.

From these measurement data, from the infiltrometer parameters and from original water level at the tested locality, determine value of hydraulic conductivity K ($M.T^{-1}$).

Instructions:

Make a simple scheme of the measurement. By Figure 1 determine the referential level and corresponding measured values of y (M) and t (T), then with help of equation (10) calculate the value of hydraulic conductivity K ($M.T^{-1}$).

Problem:

Try to design simple laboratory falling-head permeameter for determination the value of hydraulic conductivity of soil sample in a non-steady state saturated flow conditions.

Define the referential level with connection of position of soil sample, define variable head h (M) and a constant length L (M) of soil sample.

Make a simple scheme.

Derive final formula for calculation of value of hydraulic conductivity, use (apply) the same method based on equations (20), (21), (22), (23), (24), (25), (26) and (27).

Explain, where and how will be the differences and why.

Explain step by step the method of measuring and the use of the measurement data for estimation of the value of hydraulic conductivity.

Problem:

Existing wetland at the Kemeri National Park's zone (see Photo 17) will be increased over 0,75 hectares, wetland area is situated in a homogenous isotropic soil porous environment with 5 cm of water layer on the surface. There is no any extra (special) recharge, nor discharge of surface respectively subsurface water to the new designed wetland area because of climate dynamics do not suppose any precipitations in the summer time.

New wetland area is in the same soil and hydrological conditions as the present wetland. In the summer time the forecasted value of daily evapotranspiration Et ($mm.day^{-1}$) of the wetland is Et = 12,5 $mm.day^{-1}$.

To save the water regime in a new wetland, especially at the summer time, is necessary to arrange a sufficient recharge (inflow) to the new wetland.

The value of K of a surface of new wetland area, which is essential for approximation recharge, will be the same as the K in the existing wetland.

In a new conditions also the water table above the surface, so the value of K was measured in the existing wetland area with water layer 5 cm above the surface.

The single ring infiltrometer with length 45 cm (radius r_0 = 10 cm) was 25 cm pressed into the soil surface. The water level dropped down from the level of full up water infiltrometer to the 1,5 cm above the original water level during 3 hours and 55 minutes.

Make simple scheme of the field experiment and use suitable formula to calculate the value of hydraulic conductivity of the wetland surface.

Set up balance equation to determine recharge to the new wetland.

Photo 17. Kemeri National Park - Lake is a lagoon formed after digression of sea. Lake's average depth is 60 cm, in deepest places 1.5 m (J. Štibinger).

2.3.1.c. Auger-Hole Method (Hooghoudt's Method)

Description (principle):

A hole is made to a certain depth below the groundwater table. The water table in the hole is lowered with a bailer and then the rate of rise of the water table is measured. From this and from the geometry of the auger hole, can be calculated the value of hydraulic conductivity.

Use (application):

Common procedure may be used in the field surveys for the design of subsurface pipe drainage system.

Disadvantages (limitations):

Mainly horizontal hydraulic conductivity is measured. This method is not suitable for strongly layered soils or for soils with irregular pore space distribution.

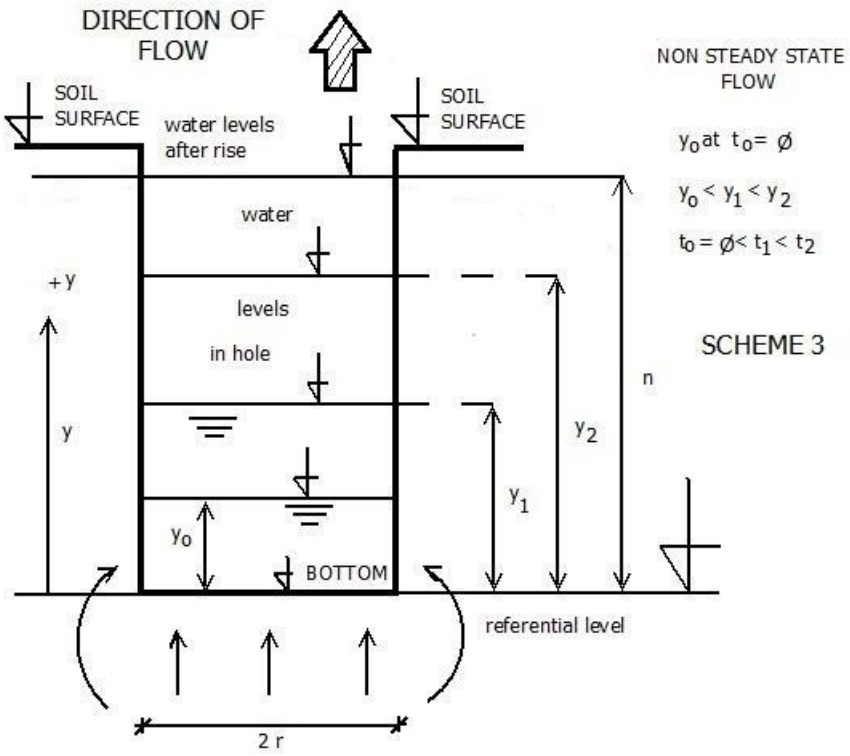

Figure 8. Auger-hole profile with water table of variable (falling) head

Theory:

The change of water flow dQ ($M^{-3}.T^{-1}$) during the rise of water level in a change of time dt (T) can be expressed by formula

$$dQ = \pi r^2 \frac{dy}{dt} \tag{28}$$

where y (M) is vertical y-axis (positive in a up direction), referential level is a bottom of the hole and r (M) is a radius of hole (see Figure 8).

The seepage area S (M^2), through the sides of the hole and through bottom of the hole can be formed as $S = 2\pi.r.n + \pi.r^2$, where n (M) is a initial water level in a hole, when the rising is finished (see Scheme 3).

With use of Darcy's Law and equation of continuity the change of water flow dQ ($M^{-3}.T^{-1}$) during the rise of water level in a hole can be also formed as

$$dQ = S.K.I \tag{29}$$

where symbol I (-) represents hydraulic slope, I = y / L_d and L_d (M) is supposed length of flow. At this case is for determination of L_d (M) used Hooghoudt's empirical approximation (Luthin 1957), which was verified in the sandy tanks. By S. B. Hooghoudt can be used approximation L_d (m) = (r.n)/0,19. The value of 0,19 (m) is a constant and is expressed in meters.

It means, that the radius r (m) of a hole and parameter n (m) have to be expressed also in meters.

According Hooghoudt is valid I (-) = y/L_d = y.0,19 / r.n. With use of equation (28) and (29) and with use of Hooghoudt's approximation can be defined equation

$$\pi r^2 \frac{dy}{dt} = (2\pi.r.n + \pi r^2)K\frac{y.0{,}19}{r.d} \tag{30}$$

after corrections we get

$$\int_{y1}^{y2} y^{-1}dy = K(\frac{2d}{r}+1)\frac{0{,}19}{r.d}\int_{t1}^{t2} dt \tag{31}$$

By integration and with certain corrections we get final equation for direct calculation of hydraulic conductivity K ($M.T^{-1}$)

$$K = (\frac{r.n}{0{,}19})(\frac{2n}{r}+1)^{-1}.(\frac{1}{t2-t1})\ln(\frac{y2}{y1}) \tag{32}$$

where y_1 (M), y_2 (M) are measured water levels in a hole at the corresponding times t_1 (T), t_2 (T) (see Figure 8).

2.3.1.d. Inversed Auger-Hole Method

Description (principle):

A hole is augered to a certain depth well above the groundwater table. Water is flowed into the dry hole, and then the rate of lowering of the water table is measured. From this rate of decreasing and from the geometry of the borehole, is calculated the value of hydraulic conductivity.

Use (application):

Common procedure in field surveys for surface or subsurface drainage design, if groundwater table is not present.

Disadvantages (limitations):

As auger-hole method.

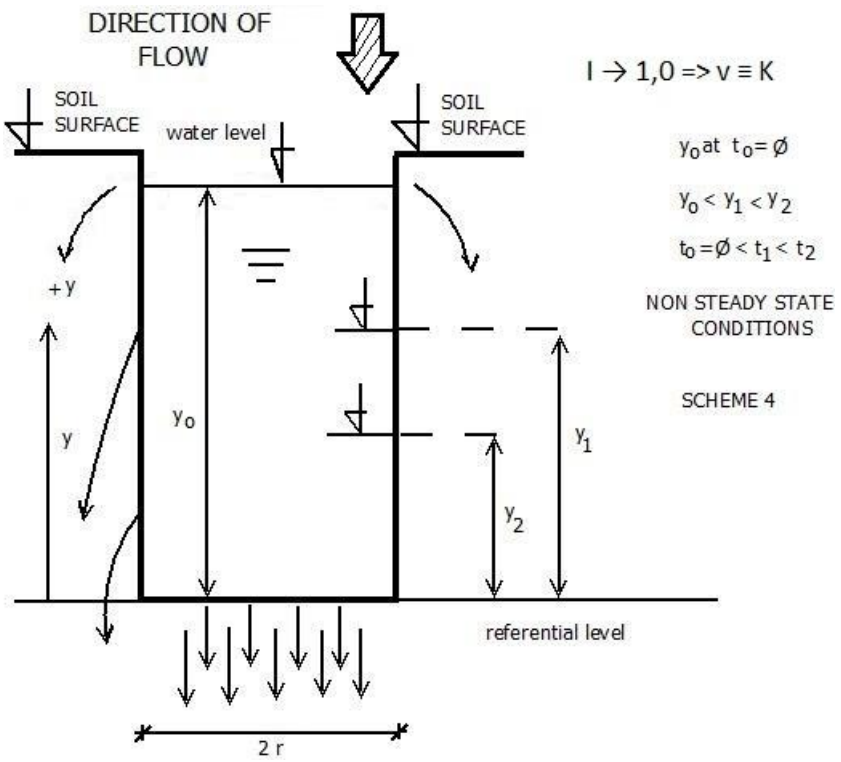

Figure 9. Inverse auger-hole profile with falling head, groundwater table is not present

Theory:

Water is flowed into the dry hole, and the rate of lowering of the water table level is measured. In a fully saturated soil porous environment it is supposed, that the hydraulic slope I (-) is converging to 1,0 (Ritzema 2006), so by Darcy's Law the hydraulic conductivity K ($M.T^{-1}$) is approximately v ($M.T^{-1}$).

The change of water flow dQ ($M^{-3}.T^{-1}$) in a hole during the lowering of water level in a change of time dt (T) can be expressed by formula

$$dQ = \pi r^2 \frac{dy}{dt} \tag{33}$$

where y (M) is vertical y-axis, referential level is a bottom of the hole and r (M) is a radius of hole (see Figure 9). The seepage area S (M^2), through the sides of the hole and through bottom of the hole can be formed as $S = 2\pi.r.y + \pi.r^2$ (see Scheme 4).

With use of Darcy's Law and equation of continuity and with approximation I = 1 (respectively K = v) the change of water flow dQ ($M^{-3}.T^{-1}$) during the decreasing of water level in a hole can be also formed as

$$dQ = -S.K \tag{34}$$

Positive in a up direction, so that is why the right part of equation (34) has negative mark. The right parts of equations (33) and (34) must be identical, so we get

$$-\pi r^2 \frac{dy}{dt} = (2\pi.r.y + \pi r^2)K \tag{35}$$

Now we used substitution Y = y + (r/2) and dY = dy and equation (35) can be rewritten as $(-\frac{r}{2}\frac{dY}{dt} = Y.K)$ and after corrections we get

$$-\frac{r}{2}\int_{y1}^{y2} Y^{-1}dY = K\int_{t1}^{t2} dt \tag{36}$$

With use of the formula $-\int_{y1}^{y2} Y^{-1}dY = +\int_{y2}^{y1} Y^{-1}dY$, by integration, after back-substitution Y = y + (r/2) and with certain corrections we get final equation for direct calculation of hydraulic conductivity K ($M.T^{-1}$)

$$K = (\frac{r}{2}).(t2 - t1)^{-1}\left(\ln\frac{y1 + \frac{r}{2}}{y2 + \frac{r}{2}}\right) \tag{37}$$

where y_1 (M), y_2 (M) are measured water levels in a hole at the corresponding times t_1 (T), t_2 (T) (see Scheme 4).

Problem:

At the beginning of the terrain experimental field measurement by inverse auger-hole method, is water level 25 cm above the bottom (see Photo 18).

Exactly after 2 hours and 15,5 minutes was hole (with radius r = 6 cm, depth 30 cm) empty.

Can you approximate the value of hydraulic conductivity K ($M.T^{-1}$)?

Photo 18. Terrain experimental field measurement of hydraulic conductivity of surface soil layers by inverse auger-hole method (Photo J. Stibinger, spring 2007. Lednice, district Břeclav, South Moravia, Czech Republic).

Instructions:

Draw simple scheme and determine the corresponding couple of values of y (M) and t (T). Realize, that at the beginning of measurement will be for y_1 = 25 cm corresponding time t_1 = 0 minutes.

In the scheme define the rested of couple of values (y_2, t_2), substitute to the formula (19) and with use of radius of hole r (m) calculate the value of hydraulic conductivity.

Result: ($K = 8{,}24.10^{-6} m.s^{-1}$)

Demonstrate, that in a case of infiltration trench (ditch) of cuboid shape (see Photo 4a) with length a (m) and width b (m) equation (38) can be used as

$$K = (\frac{r}{2}).(t2 - t1)^{-1} \left(\ln \frac{y1 + \frac{r}{2}}{y2 + \frac{r}{2}} \right) \tag{38}$$

2.3.1.e. Modified Inversed Auger-Hole Method (Porchet's Method)

Theory:

Basic input assumptions (approximations):

We can suppose a point just above the wetting front at a distance L_w (M) below the bottom of the ditch in the area where the water infiltrates. The matric head of the soil (ditch) at this point has a very small value hm. The head at the bottom of the ditch equals L_w + h, where h is height of water level in the ditch (see Figure 10).

The head difference between the point at depth L_w (M) and a point at the bottom of ditch equals L_w + h + /hm/ so and the average hydraulic gradient (hydraulic slope) between the two points can be written as I (-) = (L_w + h + /hm/) / L_w.

If L_w (M) is large enough (compare h) hydraulic gradient I approximates unity (I = 1). From Darcy's Law (v = I.K) figures that the mean flow velocity v in the wetted soil below ditch approaches the hydraulic conductivity K (v = K), assuming the wetted soil is practically saturated.

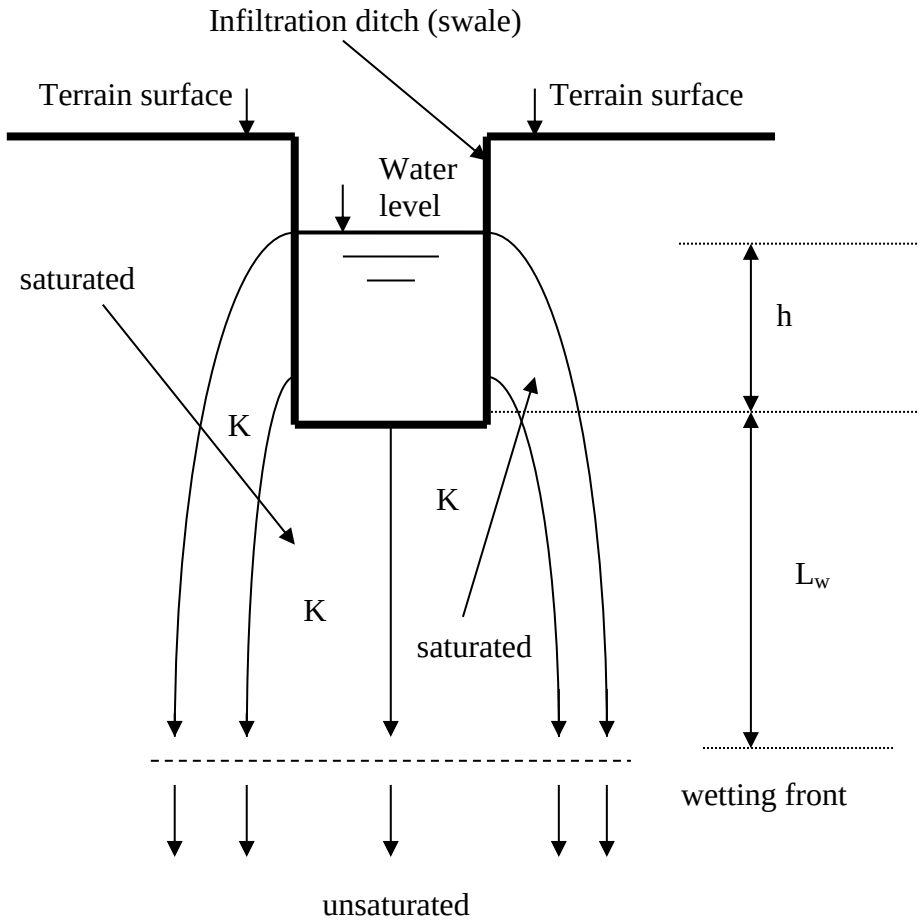

Figure 10. Scheme of the infiltration from a water-filled ditch (swale) to the soil.

Practical example from terrain experimental testing (see Figure 11):

y_0 = 20 cm	t_0 = 0 min.
y_1 = 15 cm	t_1 = 15,5 min
y_2 = 6,5 cm	t_2 = 27,0 min.
.	
y_j	t_j
.	.
.	.
y_m	t_m
.	.
.	.
y_n = 1,5 cm	t_n = 43,5 min

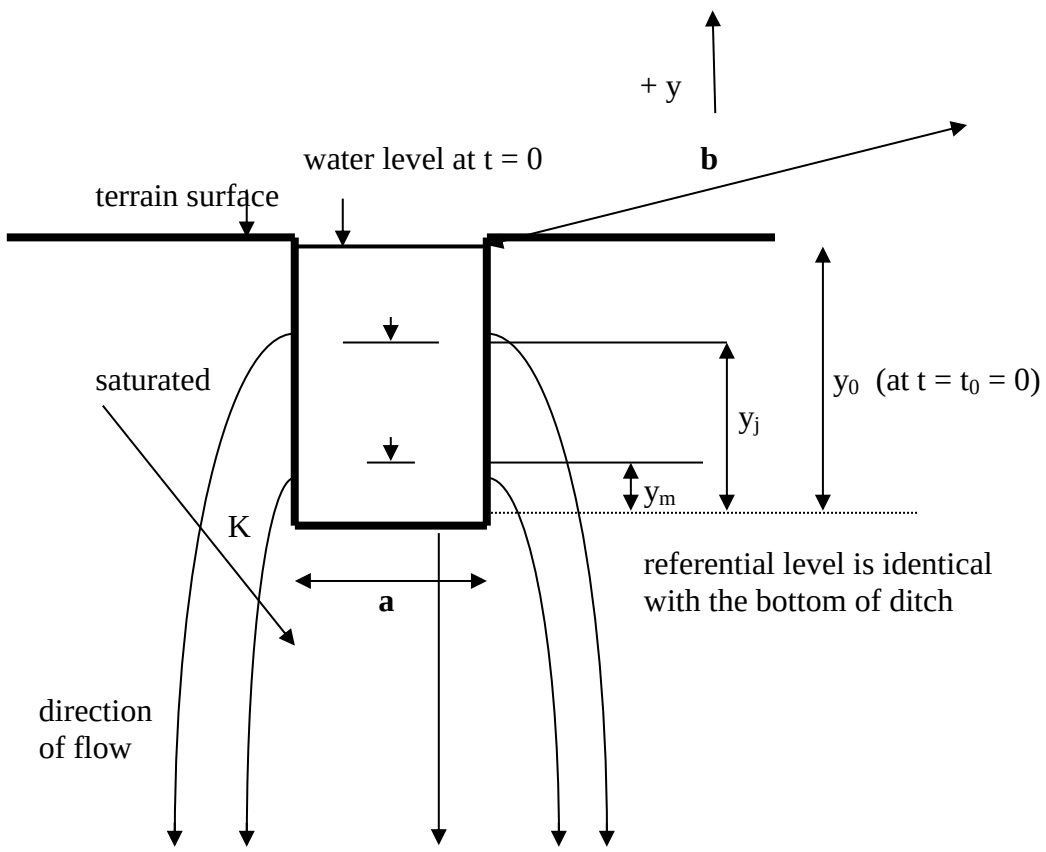

Figure 11. Scheme of the water level decreasing (lowering) during the infiltration process in the ditch.

We have an infiltration ditch of rectangular shape with width a (M) and length b (M) (see Photo 19). Q ($M^3.T^{-1}$) from total ditch infiltration = area x velocity = S x v_l. If the velocity vl ($M.T^{-1}$) of water level in the ditch will be $vl = \frac{dy}{dt}$ then Q (total ditch infiltration) = $(a.b)\frac{dy}{dt}$.

In the same time is valid, that

Total ditch infiltration = bottom infiltration + side walls infiltration.

For velocity v ($M.T^{-1}$) of infiltrating water from the ditch to the soil with Darcy's Law will be v = K, because (see assumption placed above) we get

Bottom infiltration + side walls infiltration = $-(a.b + 2(a+b)y)K$

Because of Q (total ditch infiltration) = bottom infiltration + side walls infiltration, /(with negative mark – because the y axis is positive in the upward direction, but water flow direction is down (negative)/ we get initial differential equation:

$$(a.b)\frac{dy}{dt} = -(a.b + 2(a+b)y)K \qquad (39)$$

Equation (39) can be formed in a shape

$$-\frac{ab}{2(a+b)}dy = \left[\frac{ab}{2(a+b)} + y\right]K.dt \tag{40}$$

After substitution $\frac{ab}{2(a+b)} + y = Y$ and $dy = dY$ we get

$$-\frac{ab}{2(a+b)}Y^{-1}dY = K.dt \tag{41}$$

Then can be written

$$-\frac{ab}{2(a+b)}\int_{yj}^{ym} Y^{-1}dY = K\int_{tj}^{tm} dt \tag{42}$$

With use of $-\int_a^b dY = +\int_b^a dY$. After integrating and back substitution $Y = y + \frac{ab}{2(a+b)}$ we get final equation

$$\frac{ab}{2(a+b)}\ln\left[\frac{y_j + \frac{ab}{2(a+b)}}{y_m + \frac{ab}{2(a+b)}}\right] = K(t_m - t_j) \tag{43}$$

From the terrain measurement are known corresponding data y_j (M), t_j (T) and y_m (M), t_m (T). Also the parameters of infiltration ditch a (M), b (M) are known, so by formula (15) can calculated the value of hydraulic conductivity K ($M.T^{-1}$).

The simple scheme of the water level lowering during the infiltration process in the ditch with individual parameters for calculation is viewed in Figure 11.

The type of equation (43) can be also used for approximation of emptying of infiltration ditch. We have full up water infiltration ditch, where the depth of the ditch is identical with y_0 (M), what is water level from the bottom of ditch to the terrain surface at a corresponding time $t_0 = 0$ (see Figure 11).

To calculate time t_E (T), time, when the infiltration ditch will be empty, can be used expression

$$\frac{ab}{2(a+b)}\ln\left[\frac{y_0 + \frac{ab}{2(a+b)}}{\frac{ab}{2(a+b)}}\right] = K.t_E \tag{44}$$

For this infiltration process it is supposing unit gradient and dominant vertical flow direction. The horizontal flow, which is not prevailing, expresses the right part of equation (39).

Question:

How will be the value of the corresponding head y_E (M) to the emptying time t_E (T)?

Equation (43) can be rewritten by substitution B = a.b / 2(a+b) to the general form in a shape

$$B.\ln\left[\frac{y_j + B}{y_m + B}\right] = K.(t_m - t_j) \tag{45}$$

where a.b (m2), /respective 2(a+b) (m)/ is area of the bottom of ditch (swale) respective perimeter of the ditch (swale) bottom.

It is easy to demonstrate, that if we describe all infiltration process from full up water infiltration ditch (swale), it is from y_0 (M) (t_0 = 0,0) to the empty ditch (swale), y_E = 0,00 ($t_E = T_E$) we can use the vector y* (y_0, y_1, y_2, … yj, …. ym, ….. y_E) and vector t* (t_0, t_1, t_2, … t_j, …. t_m, ….. t_E). Then equation (17) can be formed as

$$B.\ln\left[\frac{y^* + B}{B}\right] = K.t^* \tag{46}$$

After correction

$$y^* = \frac{y_0 + B}{\exp\dfrac{K.t^*}{B}} - B \tag{47}$$

Where:

y_0	it is head of the full up water ditch (swale),
B = [a.b /2(a+b)]	parameter B represents a geometry of the ditch (swale)
y*	it is a vector of water decreasing of the ditch (swale)
t*	it is a vector of time T, corresponding with y*
K	it is hydraulic conductivity of soil (ditch, swale)

Equation (47) represents in reality a time series, where the vector y* is dependant variable and vector t* is independent variable.

Known values in formula (47) is vector y* (M) and vector t* (T) and parameters y_0 (M) and B (M).

Unknown is founded parameter K ($M.T^{-1}$), represented hydraulic conductivity.

By non-linear regression analysis, with use of the method of Marquardt (Marquardt 1963), known also as the method of Marquardt and Levenberg, can be determined unknown parameter K ($M.T^{-1}$), which shows the value of hydraulic conductivity of soil around the ditch (swale).

This way of determination of hydraulic conductivity is correct and more precise then the approximation of hydraulic conductivity by use of equation (43).

Photo 19. Infiltration trench (ditch) experiment with falling water table (J. Štibinger).

2.3.1.f. Small-Scale Infiltration Trench

Approximation of hydraulic conductivity by measured data:

Practical example from terrain experimental testing to use equation (47):

y_0 = 20 cm	t_0 = 0 min.
y_1 = 15 cm	t_1 = 15,5 min
y_2 = 6,5 cm	t_2 = 27,0 min.
.	.
.	.
y_j	t_j
.	.
.	.
y_m	t_m
.	.
.	.
y_n = 1,5 cm	t_n = 43,5 min

Following photos are from field experimental testing of modified trench (ditch, swale) infiltration in locality Pardálov, Všešímy village (Prague east), Czech Republic (see Photo 20, 21, 22, 23 and 24). Date: 23.02.2008, + 12C, loamy soil with sand, grassy surface.

Photo 20. Overview of the experimental terrain measurement of infiltration process in a trench (swale, ditch), locality Pardálov, Všešímy village (Prague east), Czech Republic (J. Štibinger).

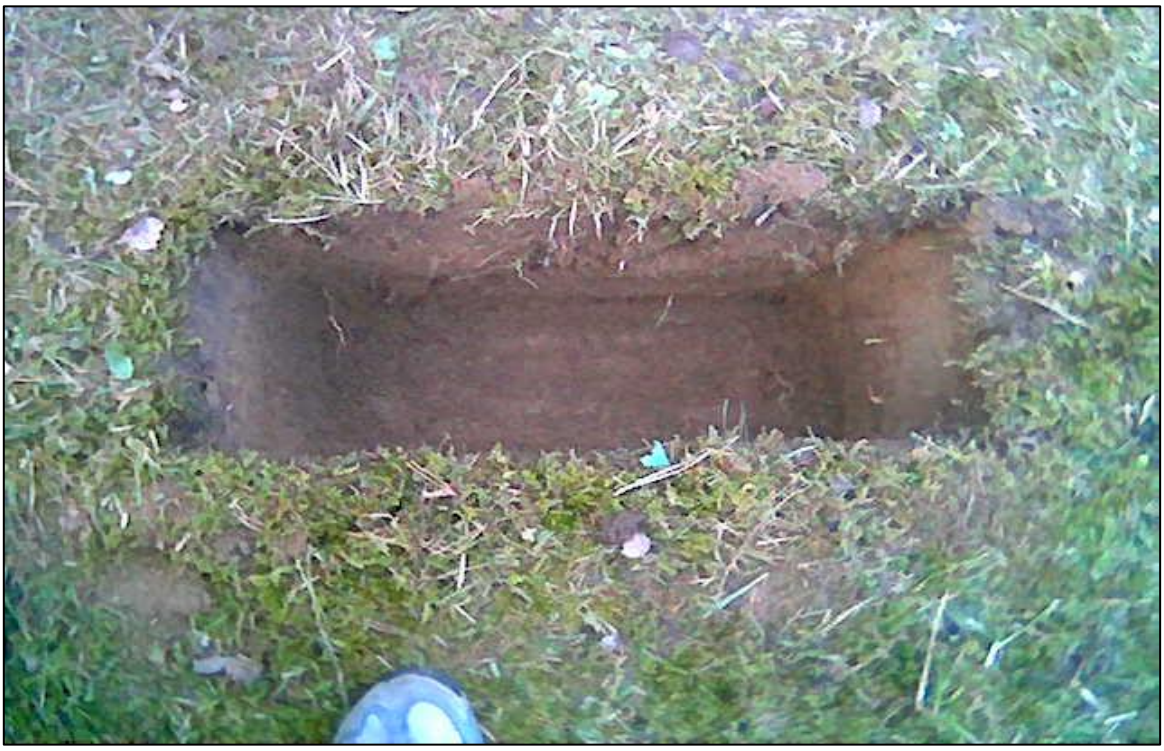

Photo 21 Infiltration trench (ditch, swale), locality Pardálov, Všešímy village (Prague east), Czech Republic, (J. Štibinger).

Photo 22. Infiltration trench with a scale and with the spray ring to eliminate the silting (chogging), locality Pardálov, Všešímy village (Prague east), Czech Republic.

Photo 23. Field experiment in infiltration trench (ditch, swale) with falling water level, locality Pardálov, Všešímy village (Prague east), Czech Republic, (J. Štibinger).

Photo 24. Inside the trench is placed cylinder for control measurement of the hydraulic conductivity after infiltration. Locality Pardálov, Všešímy village (Prague east), Czech Republic, (J. Štibinger).

Record of the water level (y) decreasing above the bottom of the trench (swale, ditch), locality Pardálov, Všešímy village (Prague east), Czech Republic (see Figure 12).

Table record:

Time (hours)	**y (m)**
16:21:34	0,20
:22:06	0,19
:22:35	0,18
:23:06	0,17
:23:48	0,16
:24:25	0,15
25:03	0,14
25:53	0,13
26:43	0,12
27:42	0,11
28:43	0,10
29:54	0,09
31:17	0,08
32:40	0,07
34:18	0,06
36:02	0,05
37:42	0,04
39:20	0,03

16:21:34 is start of the infiltration process,
t = 0,00

t=0 (s)	**y_0=0,2(m)**
32	**0,19**
66	**0,18**
92	**0,17**
134	**0,16**
171	**0,15**
209	**0,14**
259	**0,13**
308	**0,12**
368	**0,11**
429	**0,10**
500	**0,09**
583	**0,08**
666	**0,07**
764	**0,06**
868	**0,05**
968	**0,04**
1066	**0,03**

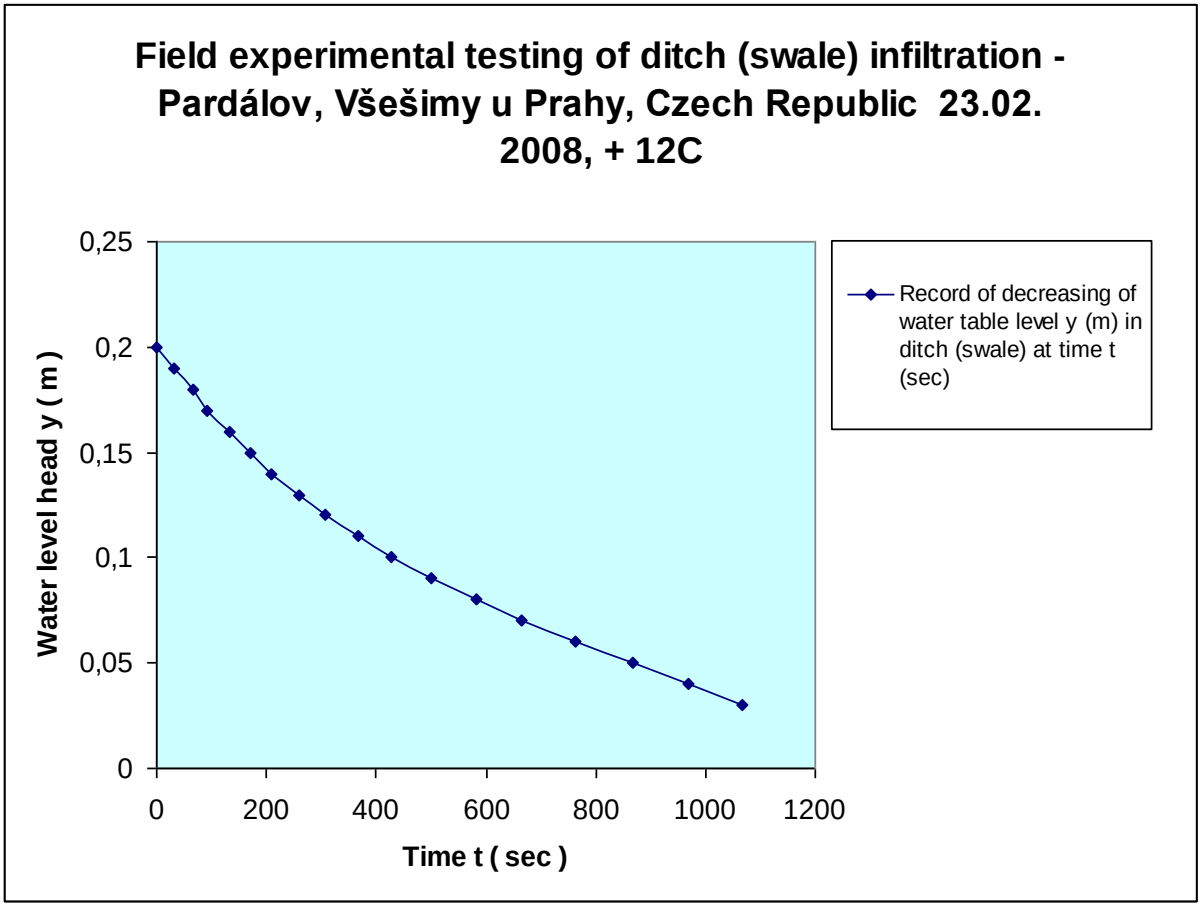

Figure 12. Graph of the water level (y) decreasing above the bottom of the trench (ditch, swale).

Problem:

Ks ($M.T^{-1}$) is only one unknown value, saturated conductivity of soil (ditch, swale, with a = 0,25 m , b = 0,5 m)

Solution:

By non-linear regressions Ks = 7.10^{-5} $m.s^{-1}$ (with Rd = 0,97 and Rk = 0,98) then t is approximated as t = 23,7 minutes.

2.3.1.g. Application of Infiltration Trenches

System of infiltration trenches, designed and realised at the slope garden area next to world famous church Prague Holy Baby (see Photo 25) serves as device for runoff and waterlogging protection of elementary school building, situated below the slope (see Photo 26)

Photo 25. Prague Holy Baby Church in Old Town Prague City. In the foreground a part of infiltration trench system is viewed.

Photo 26. Elementary school below the slope, in the right part is placed infiltration trench system, at the background is a part of Prague Holy Baby Church.

Infiltration trench systems which are inseparable part of infiltration devices serve as a protection structures against negative impact of climate dynamics as can be runoff waterlogging.

The Peter's (Petřín) Slope (the name the interest locality) is very aqueous and landslide vulnerable, therefore are infiltration trenches refilled by supporting walls against landslide.

Detail of infiltration trench with supporting walls is viewed on Photo 27, 28 and 29. Overview of infiltration trench system is displayed on Photo 30.

Photo 27. Detail of infiltration trench with supporting wall, at the upper part on the left is placed slide sprinkler.

Photo 28. Detail of infiltration trench with supporting wall and apple, going along the pavement.

Photo 29. Detail of infiltration trench with supporting wall near the steps, the smudge on the wall shows the very bad drainage capacity at the base and in surroundings.

Photo 30. Overview of infiltration trench system, situated near by Prague Holy Baby Church, in the slope garden area.

Infiltration trench system in slope garden area near by Prague Holy Baby Church is placed just at the middle historical part of Prague (see Photo 31, 32).

Photo 31. Infiltration trench system with Prague Castle at the background.

Photo 32. Infiltration trench system with Saint Nicolas Church at the background, (J. Štibinger).

Infiltration devices (see Photo 33, 34), besides waterlogging protection, can increase by infiltration the groundwater store and by that way save water regime and water resources. Paradoxically in this case, where the Peter's (Petřín) Slope is a very aqueous area, the water from the infiltration trenches must be evacuating by another way.

To apply infiltration devices Czech Standards requires for soil (earth) environment around the infiltration structure $K > 10^{-5}$ $m.s^{-1}$ to be valid.

Photo 33. View on the surroundings of infiltration trench system from the steps (J. Štibinger).

Photo 34. Surface runoff is generated at the strong slope garden area with grassy surface (J. Štibinger).

Simplified process of runoff water flow with infiltration to the trench (which is expressed by coefficient of hydraulic conductivity) is describes by balance equation – in intensity form and in capacity form (see Figure 13).

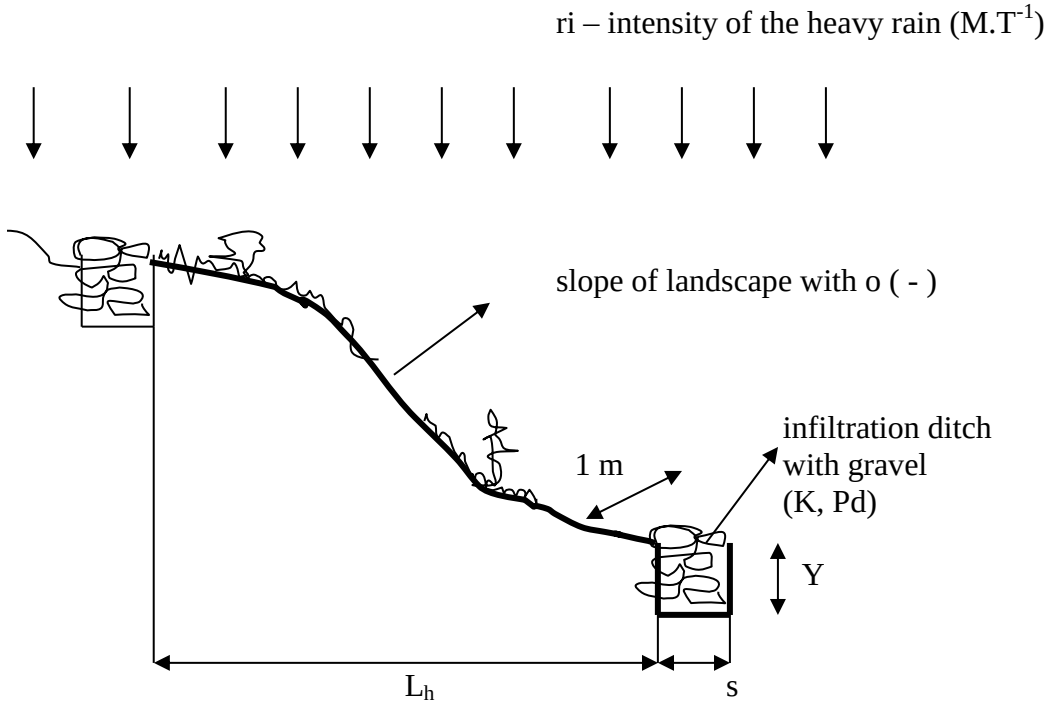

Figure 13. Simulated process of runoff water flow with infiltration to the trench – infiltration process is expressed by hydraulic conductivity of gravel.

Basic balance equation – intensity form

$$L_h.ri.o.1 + s.ri.1 = K.s.1 \tag{48}$$

$$L_h.ri.o = s.(K - ri) \tag{49}$$

Basic balance equation – capacity form

$$L_h.H_r.o.1 + s.H_r.1 = P.s.Y.1 \tag{50}$$

$$L_h.H_r.o = s.(P.Y - H_R) \tag{51}$$

The right part of equation (50) (P . s . Y . 1) (M^3) represents the capacity of the infiltration ditch to the length unit from the point of view of the quantity of the gravity water in the ditch (to the length unit of the ditch).

Where:

L_h horizontal projection of the slope (M)
ri intensity of the heavy rain ($M.T^{-1}$) o runoff coefficient (-)
1 length unit (M)
s width of the infiltration ditch (M)
K hydraulic saturated conductivity of the gravel (gravel-sand) in the infiltration ditch ($M.T^{-1}$)
H_r head (height) of the heavy rain (M), H_r = r . tr , where tr (T) is the time of duration of the rain

P drainable pore space (effective drainage porosity) of the of the gravel (gravel-sand) in the infiltration ditch (-)

Y depth of the gravel (gravel-sand) infiltration ditch (M), can be identical with the real (total) head of the infiltration ditch

runoff coefficient o (-) = quantity of flowed of water / quantity of rainfall

Please define the conditions of the validity for equation (A2) and for equation (A4). Formulas (A2) and (A4) allows to receive information (approximations) which are useful for projects of the infiltration trench systems and for solutions of similar problems in the water management practice.

Problem:

For the volume of very huge (enormous) rain of 100 mm per day, design (approximate) the width s (m) of infiltration trench filled by gravel with depth Y = 85 cm, the average value of trench spacing L = 8.0 m. For this type of green garden slope the runoff coefficient o = 0.25 (-).

Can be infiltration process on the trench gravel surface generated by rain and runoff a problem? Comment the results.

Instructions:

Use balanced formula in a capacity form, substitute corresponding parameters (drainable pore space for gravel can be around 40 % of volume) and calculate width s (m). For infiltration analysis calculate rain intensity, runoff intensity and come from hydraulic conductivity of gravel which can be approximately about 6.10^{-3} $m.s^{-1}$ (e.g.).

Use balance equation in an intensity form. The solution will be presented in the training lectures.

Problem:

To enhance the natural ground water capacity to store drain water was designed infiltration trench system with gravel-sandy material. For L = 9 m, Y = 1.0 m, s = 0.45 m, o = 0,15 (-) approximate the head H_r (mm) of the one day precipitation. Comment the results. Will be infiltration process of this “one day rain” a problem in connection with trench system?

Instructions:

Approximate K ($M.T^{-1}$) and P (-) of the gravel-sand material, use corresponding formula $L.H_r.o.1 + s.H_r.1 = P.s.Y.1$ and solve the problem.

Solution:

The P (-) of the gravel-sand material in the trench is approximately 35 % of volume, from the equation (A3) is valid H_r = 0.013 m = 13 mm.

It means that the volume of infiltration trench system is save (reliable) for one day rain with height of 13 mm or less.

To increase the available volume of infiltration trench system can be recommended following points:

- Decrease trench spacing L (m)
- Increase the parameters of infiltration trench system
- Increase of drainable pore space of gravel-sand material or make suitable substitution of filled porous material

The hydraulic conductivity K ($m.s^{-1}$) of the gravel-sand material in the trench can be for example approximately $K = 5.10^{-3}$ $m.s^{-1}$, from the equation (49) is valid $ri = 1.25.10^{-3}$ $m.s^{-1}$.

Rain intensity, which is save for infiltration trench system, $ri = 1.25.10^{-3}$ $m.s^{-1}$ is big enough. Then infiltration trench system is reliable with connection to infiltration capacity.

The infiltration processes in the ditch should simulated infiltration in the swales. Swales - are grassed depressions which lead surface water overland from the drained surface to a storage or discharge system, typically using the green space of a roadside margin (see following figures, Figure 14, Photo 35, Figure 15 and Photo 36).

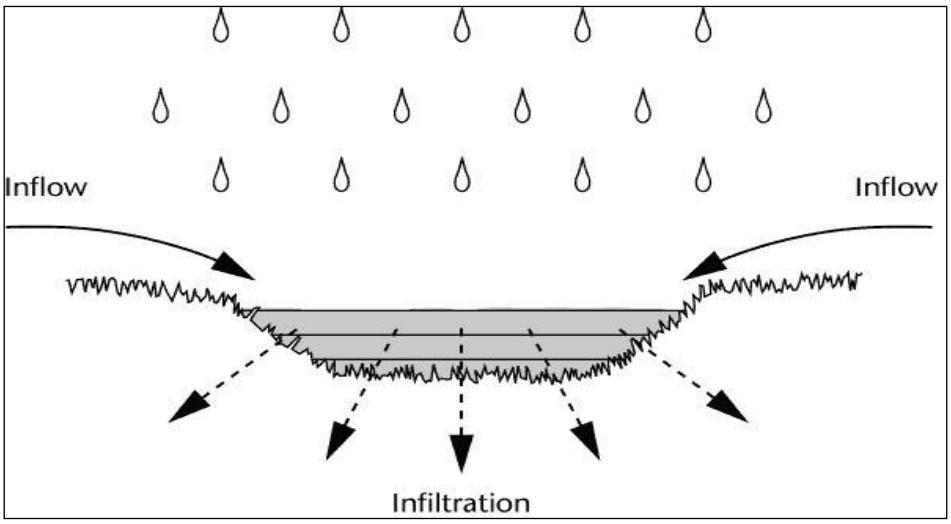

Figure 14. Profile of swale with grassy surface during infiltration (The University of Abertay Dundee).

Photo 35. Infiltration swale (strip, trench, ditch) filled up by gravel around the roadway (Dundee, Scotland, UK, The University of Abertay Dundee)

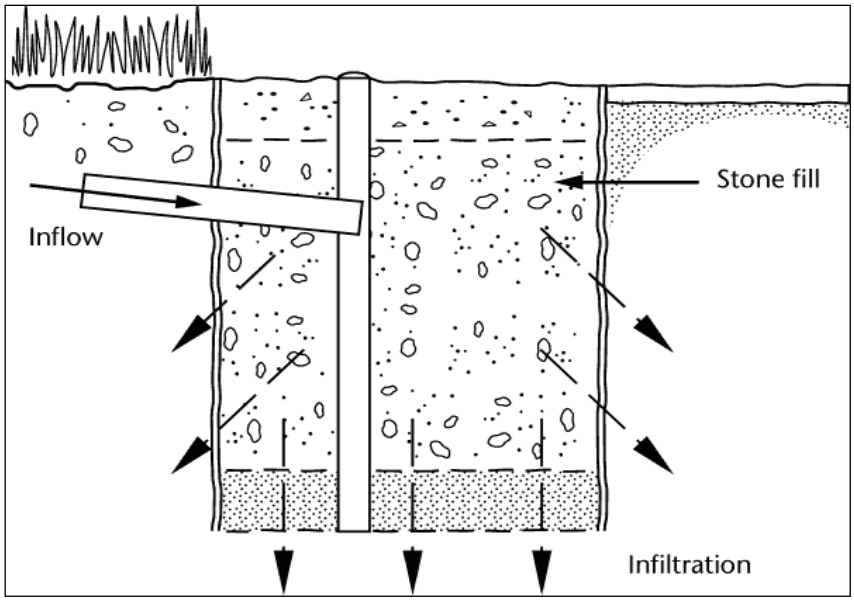

Figure 15. Profile of infiltration measure filled up by gravel sand (The University of Abertay Dundee)

Photo 36. Infiltration trench (swale) in a grassy surface in front of family house (Dundee, Scotland, UK, The University of Abertay Dundee).

Around the ditch it is to be supposed homogenous isotropic soil environment with the one value of the hydraulic conductivity K ($m.s^{-1}$).

2.3.2. Examples of Large-Scale Field Methods

2.3.2.a. Pumping Tests

Description (principle):

A hole is drilled into the soil below the groundwater table. Then water is pumped from the hole with a constant rate. The drawdown of the water table is measured at the certain several distances from the borehole from which the value of the hydraulic conductivity is calculated.

Use (application):

This field method can be applied for deep, thick soil layers with a high value of hydraulic conductivity (aquifers); for calculation of seepage flow; for assessment of the effects of groundwater extraction.

Disadvantages (limitations):

This laborious method is applicable only in the thick, homogeneous, permeable soils.

2.3.2.b. Parallel Drains Method

Description (principle):

Calculation of the value of hydraulic conductivity comes from measurements of drain discharge and groundwater table level midway between the drains. The value of hydraulic conductivity estimated by this way is representative of the all area of groundwater flow to the drains.

Use (application (use):

In pilot areas to get representative values of the hydraulic conductivity, so as to guide for the determination of drain spacing in similar soils in larger project areas.

Disadvantages (limitations):

This large field method is only applicable after the installation of a pipe or ditch drainage system.

Theory:

We have a system of open drain ditches, placed on approximately horizontal, very low permeable layers, under the steady state drainage flow conditions, where recharge R ($M.T^{-1}$) is identical with drainage discharge q ($M.T^{-1}$) and no any change of the shape of groundwater table.

Drain ditches spacing is L (m), free water level in the ditches is D (m) and the maximum of water table at the middle between the ditches (from the impervious layer) is H (m) /see Figure 16/.

Soil porous environment is homogenous isotropic with one value of hydraulic conductivity K ($M.T^{-1}$).

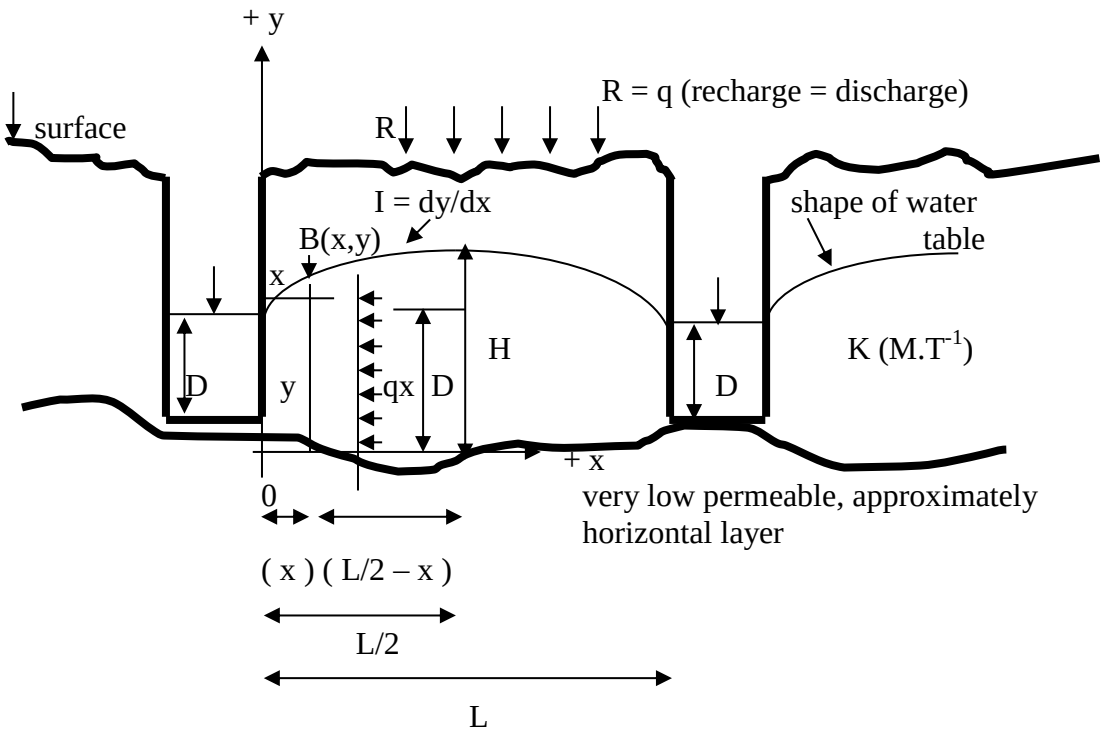

Figure 16. Scheme of the steady state drainage flow to the open ditches, placed on the low permeable, approximately horizontal layer with drain spacing L (M).

Horizontal x-axis is positive to the right direction and is identical with very low permeable, approximately horizontal layer. Vertical y-axis is positive to the upward direction and is identical with the side of the ditch.

All the process of steady state drainage flow is symmetrical with axis of symmetry placed at the midway between the ditches, perpendicularly to the low permeable, approximately horizontal layer.

The problem was solved for variable x (M) from x = 0 to x = L/2 (and for variable y (M) from y = D to y = H) /see Figure 16/.

We supposed above the approximately horizontal low permeable layer, horizontal one-dimensional steady drainage flow in direction to the ditches. The shape of the water table is changeless (constant).

Steady state drainage flow is symmetrical with axis of symmetry placed at the midway between the ditches, perpendicularly to the low permeable, approximately horizontal layer.

Recharge (infiltration) to the groundwater table is going to the down direction, that means in the negative direction of y-axis, is constant and is represented by symbol R ($M.T^{-1}$) /see Figure 1/.

Let's define an arbitrary point B (x,y) on the water table level in a horizontal distance x (M) from the side of ditch, with height y (M) above the low permeable layer (see Figure 1).

The unit steady vertical flow to the point B (x,y) can be expressed by formula $-R(\frac{L}{2}-x)$ with negative value, which shows that the flow goes against positive direction of the y-axis (see Figure 16).

Unit horizontal one-dimensional steady state flow q_x ($M^2.T^{-1}$) from B (x,y) to the open ditch (see Figure 1) can be expressed with Darcy's Law by equation:

$$q_x = -K.y\frac{dy}{\partial x} \tag{52}$$

Negative value signalizes, that the flow goes against positive course of the x-axis. By the principle of continuity is valid:

$$-K.y\frac{dy}{\partial x} = -R(\frac{L}{2} - x) \tag{53}$$

Formula (53) represents an ordinary differential equation, which can be formed as:

$$K.ydy = R(\frac{L}{2} - x)dx \tag{54}$$

With boundary conditions x = 0, y = D and x = L/2, y = H. After integration and further corrections we get:

$$L^2 = \frac{4K(H^2 - D^2)}{R} \tag{55}$$

In steady state drainage flow conditions recharge R ($M.T^{-1}$) is identical with discharge q

($M.T^{-1}$), R = q /see Figure 1/), so equation (55) can be formed:

$$L^2 = \frac{4K(H^2 - D^2)}{q} \tag{56}$$

Equation (55) /respectively (56)/ which was derived in 1936 by S.B. Hooghoudt is very well known as Donnan's Equation (Donnan 1946).

Problem:

In a flat area (see Photo 37) of 2,5 ha, with open ditches (L = 35 m) placed on an horizontal impervious layer at the depth of bottom 2,0 m under the surface, is water level in the ditches 1,0 m.

Water table level at the midway between the ditches is 0,5 m below the surface. To manage and to save water regime at this area, is necessary to determine the value of hydraulic conductivity K ($M.T^{-1}$).

Soil porous environment is homogenous isotropic, the total steady drainage discharge from all area is measured 0,25 litres per second.

Can you approximate the value of hydraulic conductivity K ($M.T^{-1}$)?

Photo 37. System of open ditches to control water regime in Friesian pasture in The Netherlands (source: International Course on Land Drainage, Alterra Wageningen UR, H. Ritzema, 2008, The Netherlands).

Instructions:

Draw simple scheme by Figure 16, determine input parameters D (M) and H (M), transform q in a meter per second and with L = 35 m and by equation (56) calculate the hydraulic conductivity K ($M.T^{-1}$).

Result:

$(K = 2{,}45.10^{-6} m.s^{-1})$

Equation (55) /respectively (56)/ was corrected by S.B. Hooghoudt for "ideal" drain pipe in steady-state drainage flow conditions with dominant horizontal subsurface flow to drains (Ritzema 2006) in a shape

$$L^2 = \frac{8K_l.h.d + 4K_u.h^2}{q} \tag{57}$$

where K_l ($M.T^{-1}$) (respectively K_u /$M.T^{-1}$/) is lower hydraulic conductivity (conductivity under the drains), respectively upper hydraulic conductivity (conductivity above the drains).

Parameter d (M) is equivalent depth of impermeable soil layer below the level of the drains (Hooghoudt's equivalent depth), parameter h (M) is ground water table above drain level between middle the drains. r_0 (M) is radius of lateral drains. All symbols are viewed in Figure 17.

Note:

This problem can be also solved by method of L. F. Ernst and by D. Kirkham's equation (Ritzema 2006).

The value of d (M) can be determined by a lot of formulas, e.g. Dieleman (1976) presents:

$$d = \frac{D}{(8/\pi).(D/L).\ln(\frac{D}{\pi.r_0})+1} \tag{58}$$

If D > L/2), then D = L/2.

Problem:

In a drained agricultural area area (approximately of 25 ha) with subsurface pipe drainage system (L = 20 m, r_0 = 0,05m, drain depth h_d = 1,2 m) keeps water level at the midway between the drains 0,7 m below the surface.

From the field measurements and from the results of hydrology, soil hydrology and geology investigation is valid R (q) = 1,5 mm.day-1, K_u = 0,25 m.day^{-1} and D = 7 m.

Can you approximate the value of lower hydraulic conductivity K_l (M.T^{-1})?

Can you approximate the value of transmisivity of soil (earth) environment under the drains?

Instructions:

With use of equation (58) approximate the value of equivalent depth d (m). By substitution d (m) to equation (57) and with known input data, is possible according to equation (57) get K-value of lower hydraulic conductivity K_l (M.T^{-1}).

Draw simple scheme of the solved problem.

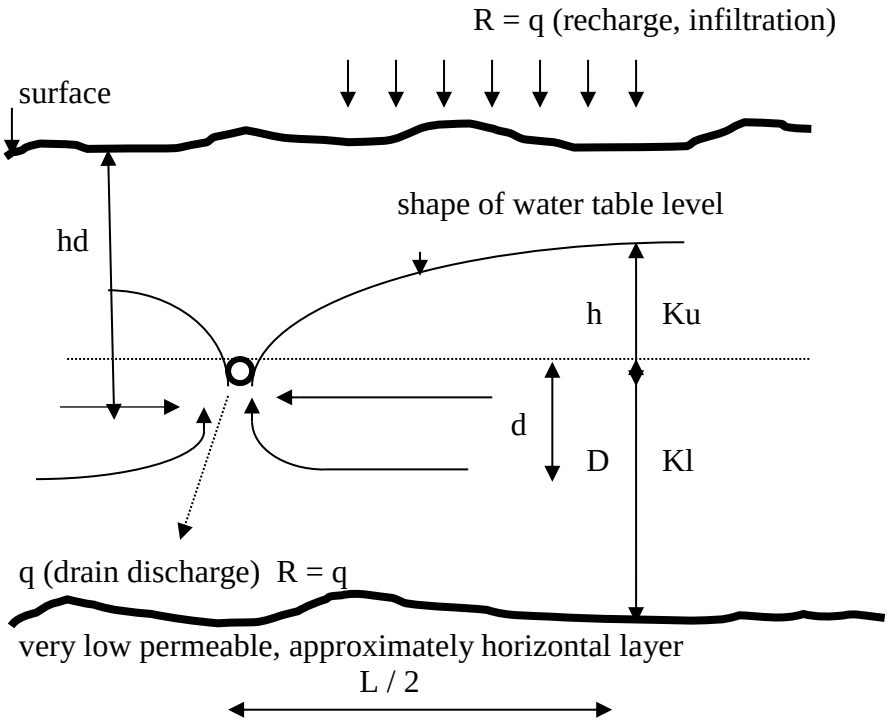

Figure 17. Scheme of the subsurface pipe drainage system in steady-state drainage flow conditions with equivalent layer d (M).

2.4 Determination of the unsaturated hydraulic conductivity

Unsaturated hydraulic conductivity k(H) can be measured by Mini Disk Infiltrometer (see Photo 8), which measures the infiltration rates in unsaturated zone (Decagon Devices, Inc. 2005).

Photo 8. Mini Disk Infiltrometer for estimation of unsaturated hydraulic conductivity k(H) (Kbelské louky, Břeclav, South Moravia, Czech Republic, M. Sůva).

Determination of the unsaturated hydraulic conductivity k(H) may be realized with use of combination of Philip's procedure (Philip 1957, Kutílek 1975, Kutílek a Nielsen 1994) and Van Genuchten's soil parameters (Ippisch et al. 2006, Sůva, Štibinger 2013).

References

Alterra-ILRI / Wageningen UR. (2008). Pre Drainage Investigation (Mashtul, Egypt). Study materials of the International Course on Land Drainage (ICLD), module 3.: Design, Implementation and Operation of Drainage Systems (IDSD), Wageningen, The Netherlands.

Buckingham 1907. Studies on the movement of soil moisture. Bull. 38. US Dept. of Agriculture Bureau of Soils, Washington, D.C., USA.

Darcy H. 1856. Les fontaines publique da la ville de Dijon. Dalmont, Paříž, Francie

Decagon Devices, Inc. 2005. Minidisk infiltrometer, User's manual.

Dieleman P. J. and Trafford B. D. 1976. Drainage testing. Irrigation and Drainage Paper 28. FAO. Rome, Italy.

Donnan W. W. 1946. Model test of a tile-spacing formula. Soil Science Society of America Proceedings 11, pp. 131-136.

GraphPad Software. Inc. (1995-2001). S755 Oberlin Drive, # 110 San Diego, California 921 21. USA.

Ippisch T., Vogel T. Bastian T. 2006. Validity limits for Van Genuchten-Mualem model and implications for parameter estimation and numerical simulation. Advances in Water Resources 29, pp. 1780-1789.

Kovář P., Štibinger J. a kol. 2008. Metodika návrhu a výstavby optimální varianty protipovodňových a protierozních opatření (PPPO) pro zmírnění extrémních hydrologických jevů – povodní a sucha v krajině. Číslo grantu: NPV-Mze 2005. VRK1/TP3-DP6 (1G 577040). Výroční zpráva za r. 2007. ISBN 978-80-213-1743-7. Vydavatel: ČZU Praha, FŽP, KBÚK, ČR.

Kulhavy Z. et al. (2008). Optimalizace krajinné struktury z hlediska hydrologických režimů. Periodická zpráva za r. 2007 z národního programu výzkumu II., projektu č. 2B06022 pro MŠMT ČR, koordinátor VÚMOP Praha-Zbraslav, spoluřešitel ČZU Praha, FŽP, KBÚK, (in Czech, abstract in English).

Kutílek 1975. Aplikovaná hydropedologie. Skriptum ČVUT Praha, FS, Praha, ČR

Kutílek M. and Nielsen D. R. 1994. Soil hydrology. Geo-ecology textbook, Catena Verlag, 38162 Cremlingen Destedt, Germany. ISBN 3-923381-26-3, pp. 98-102.

Luthin J. N. 1957. Drainage of Agricultural Lands. The American Society of Agronomy, Madison, Wisconsin, US.Luthin J. N. 1973. Drainage Engineering. New York: Robert E. Krieger Publishing Company, Huntington, US.

Marquardt D. W. 1963. An algorithm for least square estimation of nonlinear parameters. Journal of Society of Industrial Applied Mathematics 11: 431-441.

Nguyen Duy Binh. (2007). Irrigation of Paddy Fields in Mekong Delta. Materials, reports and documents of Department of Water Resources, Hanoi Agriculture University, Hanoi, Vietnam.

Philip, J. E.: The theory of infiltration: 1. The infiltration equation and its solution. Soil Science 83: 1957. 345-357.

Rektorys K. a kol. 1995. Přehled užité matematiky. Prometheus s.r.o., 117 01 Praha-1, Žitná 25, 6. vydání, ISBN 80-85849-72-0, Praha,ČR.

Richards L. A. 1931. Capilary conduction of liquid through the porous media. Physics 1:318 – 333

Ritzema H. P. 2006. Determining the Saturated Hydraulic Conductivity (Oosterbaan R.J and Nijland H. J.) In: H. P. Ritzema (Ed) Drainage Principles and Applications (pp. 283-294). ILRI Publ. 16, Wageningen, The Netherlands.

Rycroft D. W., Mohamed H. A. (1995). Drainage of heavy soils. FAO Irrigation and Drainage Papers, M-56, ISBN 92-5-103624-1, Rome, Italy.

Štibinger J. (2001). Paddy Fields in Taoyuan (Taipei) during the Cycle of Flooded Period. Papers from ICLPST, Taoyuan, Taiwan, The 84-th Regular Session on Land Tenure and Rural Development, Dept. of Land Use and Improvement, Faculty of Environmental Sciences, Czech University of Life Sciences Prague, Czech Republic.

Štibinger J. 2006. Použití tříparametrické infiltrační rovnice při odhadu retenčních schopností povrchových vrstev krajiny. Sborník referátů „Meliorace v lesním hospodářství a v krajinném inženýrství. Vydalo ČZU Praha, Fakulta lesnická a environmentální a VÚMOP Praha Zbraslav, ISBN 80 – 213 – 1446 – X, str.51-60, Praha, ČR.

Sůva, Štibinger 2013. Určování nenasycené a nasycené hydraulické vodivosti půd. ČZU Praha, software, http://fzp.czu.cz/vyzkum/software.html

Todd D. K. and Mays L. W. 2005. Groundwater Hydrology. John Wiley and Sons, Inc. pp. 125-142, ISBN 0-471-45254-8 (WIE), ISBN 0-471-05937-4 (cloth), US.