# CZECH UNIVERSITY OF LIFE SCIENCES, PRAGUE

FACULTY OF ENVIRONMENTAL SCEINCES

**DEPARTMENT OF WATER RESOURCES AND ENVIRONMENTAL MODELLING**

## MASTER OF SCIENCE (M.Sc.) THESIS

# INFILTRATION MODELLING INTO SAND

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## ČESKÁ ZEMĚDĚLSKÁ UNIVERZITA V PRAZE

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# ZADÁNÍ DIPLOMOVÉ PRÁCE

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Název práce
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**Cíle práce**

Modeling of water flow in sand sample with help of laboratory experiment and the Hydrus model.

## Metodika

1) Build up laboratory apparatus for simulation of water infiltration and water flow in unsaturated zone with help of sand box and tension disk infiltrometer.

2) Realize laboratory experiment and evaluated measured data.

3) Modeling of realized experiment in the Hydrus model

## Doporučený rozsah práce

40 pages

## Klíčová slova

disk infiltrometer, sand box, unsaturated zone

## Doporučené zdroje informací

ŠIMŮNEK, J., RADCLIFFE, D E. Soil physics with HYDRUS : modeling and applications. Boca Raton, FL: CRC Press/Taylor & Francis, 2010.

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# DECLARATION

I hereby declare that I EDMUND ABROKWA solely authored this master thesis as one of the prerequisite requirements for the M.Sc. degree at the Faculty of Environmental Sciences, Czech University of Life Sciences, Prague.

I have carried out different studies connected to my thesis on my own; therefore I declare that I only used those sources that are referenced in the work.

Prague, 22nd April 2015.

EDMUND ABROKWA

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I dedicate this thesis to the Most High God for His gift of life and wisdom and also to the memory of my late mother, Juliana Oduro.

# ABSTRACT

The aim of this research is to assess the accuracy and reliability of the Hydrus-1D model to simulate the flow of water in a sand soil profile (PR13) in the laboratory. The work includes conducting experiments in the laboratory to estimate certain hydraulic properties of the sand sample, such as the bottom flux and hydraulic conductivity. The experimental data obtained were used to test the accuracy and reliability of the Hydrus-1D model. Results showed that the measured data obtained from the laboratory experiments, compared with the modelled data, were almost the same. There were very minute differences due to some errors in the measurements in the pressure and time, in the laboratory experiments. Lateral flow was also observed due to the dimension of the sandbox used for the experiment. The sandbox used was improvised and human errors played a role. The differences between the measured data and modelled data are not very significant. Hydrus-1D model is therefore a more appropriate, flexible and accurate model that can be used to predict accurately the water dynamics in a soil column.

**Key words:** disk infiltrometer, sandbox, unsaturated zone.

# ABSTRAKT

Cílem tohoto výzkumu je posoudit přesnost a spolehlivost modelu Hydrus-1D pro simulaci proudění vody v profilu písčitých půd (PR13) v laboratoři. Práce zahrnuje provádění experimentů v laboratoři pro odhad určité hydraulické vlastnosti vzorku písku, jako je například spodní tok a hydraulická vodivost. Získaná experimentální data byla použita k testování přesnosti a spolehlivosti modelu Hydrus-1D. Výsledky ukázaly, že naměřené údaje získané z laboratorních experimentů, ve srovnání s modelovanými údaji, byly téměř stejné. Velmi malé rozdíly mají původ v některých chybách laboratorního měření tlaku a času. Boční průtok byl také pozorován v důsledku rozměru pískového lože použitého pro experiment. Použité pískové lože bylo ručně vytvořeno dle originálu, proto obsahuje chyby a nepřesnosti způsobené lidským faktorem. Rozdíly mezi naměřenými daty a modelovaných údaji nejsou příliš významné. Hydrus-1D model je tedy vhodnější, flexibilní a přesný model, který může být použit pro přesnou predikci dynamiky vody v půdním profilu.

**Klíčová slova:** disk infiltrometer, sandbox, nenasycené zóny.

# 1. INTRODUCTION

Hydraulic properties of soil that are defined by the hydraulic conductivity and soil water retention functions predict the flow of water in the unsaturated zone. They also show the distinction between infiltration and runoff. These properties have very important applications, some of which are: (i) for modelling available water resources and (ii) for flood forecasting. It is also important in assessing the capacity of soil to retain chemical pollutants and in evaluating the potential of groundwater pollution (Léger *et al*., 2013). Laboratory experiments are usually conducted to determine the parameters involved in the van Genuchten soil water retention function (van Genuchten, 1980).

Numerical models have progressively been used to anticipate and to investigate water flow and solute transport in the unsaturated zone over the last decades. Numerical and theoretical difficulties still prevail even though a number of systematic approaches have been created to solve water flow equations (Bruce *et al*., 1956), especially at the transient scale flow and with multidimensional field applications (Abbasi *et al*., 2004). The movement of water in both vertical and lateral directions have been studied to be greatly related to soil macropores (Lawes *et al*., 1882). Soil hydraulic conductivity is greatly determined by soil porosity (Beven *et al*., 1982).
The objective of this study is to determine the hydraulic properties of the soil profile, to evaluate Hydrus-1D model using pressure heads, and to evaluate the reliability and accuracy of the Hydrus-1D model in simulating the measured dynamics of water flow in the sand soil profile (PR13).

## 1.1 Project Aim/Objectives

### 1.1.1 Aim

The aim of this research is to assess the accuracy and reliability of the Hydrus-1D model to simulate the flow of water in a sand soil profile (PR13) in the laboratory.

### 1.1.2 Objectives

The objectives of this research are to:

1. estimate soil hydraulic properties through laboratory experiments.
2. study the differences between saturated and unsaturated hydraulic conductivities.
3. conduct simulations on the measured dynamics of water flow in a sand soil profile (PR13) in the laboratory with Hydrus-1D.
4. discuss the results obtained from the Hydrus-1D model to determine its reliability accuracy.

## 1.2 Research Questions:

1. Is Hydrus-1D model reliable?
2. How does the experimental flux data differ from the Hydrus-1D model?

These research questions have been answered in this project through laboratory measurements of bottom flux of sand sample PR13 and compared with the modelled bottom flux by Hydrus-1D.

# 2. LITERATURE REVIEW

## 2.1 Soil

Generally, soil refers to the top few feet of the land surface. The soil acts as a natural filter to screen out many substances that mix with the water. Nevertheless water still transports some contaminants into the groundwater. The amount of groundwater recharge, storage, discharge, as well as the extent of groundwater contamination, all depend on the soil properties (Loxnachar *et al*., 1999): texture- grain size, texture classification, soil series classification; porosity; specific yield; permeability; attenuation capacity (soil's filtering ability).

### 2.1.2 Soil composition and texture.

According to Kutilek (1978), soil a polydisperse system composed of a disperse part and disperse environment. The disperse part comprises of the soil particles and organic matter whereas the disperse environment is made up of liquid (soil water) and air (soil air). Soil texture comprises the relative distributions of particles of various sizes such as sand, silt and clay in the soil. Soil is a combination of three soil separates; sands, being the coarsest, silts, and clays, which are the finest.

### 2.1.3 Classification Systems

Organizations such as Food and Agriculture Organization (FAO), United States Department of Agriculture (USDA), Soil Conservation Service (SCS), Massachusetts Institute of Technology (MIT), British Standards Institution (BSI) classify soil particles according to their particle size or grain size. Classification of these separates depends on grain size. Loxnachar *et al*., (1999) summarized the soil separate and its corresponding diameter size as shown in **Table 1**.

**Table 1:** Grain Size Distribution. **(**Loxnachar *et al*., 1999)

| **Name** | **Size Range (mm)** |
|---|---|
| gravel | > 2.0 |
| very coarse sand | 1.0-1.999 |
| coarse sand | 0.500-0.999 |
| medium sand | 0.250-0.499 |
| fine sand | 0.100-0.249 |
| very fine sand | 0.050-0.099 |
| silt | 0.002-0.049 |
| clay | < 0.002 |

**Table 2:** The taxonomic classification system in the Czech Republic (Kutilek, 1978):

**Particles smaller than 2mm**

| Particle | Size range (mm) |
|---|---|
| Clay | Up to 0.002 |
| Silt | 0.002-0.05 |
| Sand | 0.05-2 |

**Particles larger than 2mm**

| Particle | Size range (mm) |
|---|---|
| Coarse sand | 2-4 |
| Gravel | 4-30 |
| Stones | Greater than 30 |
| Boulders | Greater than 300 |

### 2.1.4 Texture Classification

The relative distribution of soil separates in a particular soil determines its soil texture.  The soil texture triangle presents the texture name, which is based on the percentages of sand, silt, and clay contained in the soil sample (Donahue *et al*., 1983). In using the diagram, the points showing the percentages of silt and clay present in the soil under examination are positioned on the silt and clay lines respectively. Then lines are drawn inward, parallel in the first case to the clay side of the triangle and in the second case parallel to the sand side. The name of the chamber in which the two lines converge is the class name of the soil in question. For example a soil composed of 15% clay, 20% silt and 65% sand is sandy loam and a soil composed of equal proportions of sand, silt and clay is clay loam. The percentages of sand, silt and clay in a soil could be estimated in a soil laboratory by two standard methods - pipette method and hydrometer method (Black *et al*., 1965). These two methods depend on the fact that the concentrations of the particles vary with time at any given depth in a settling suspension, as the coarser fractions settle at a faster rate than the finer fractions, as described by the soil texture triangle in **Figure 1** (Donahue *et al*., 1983)

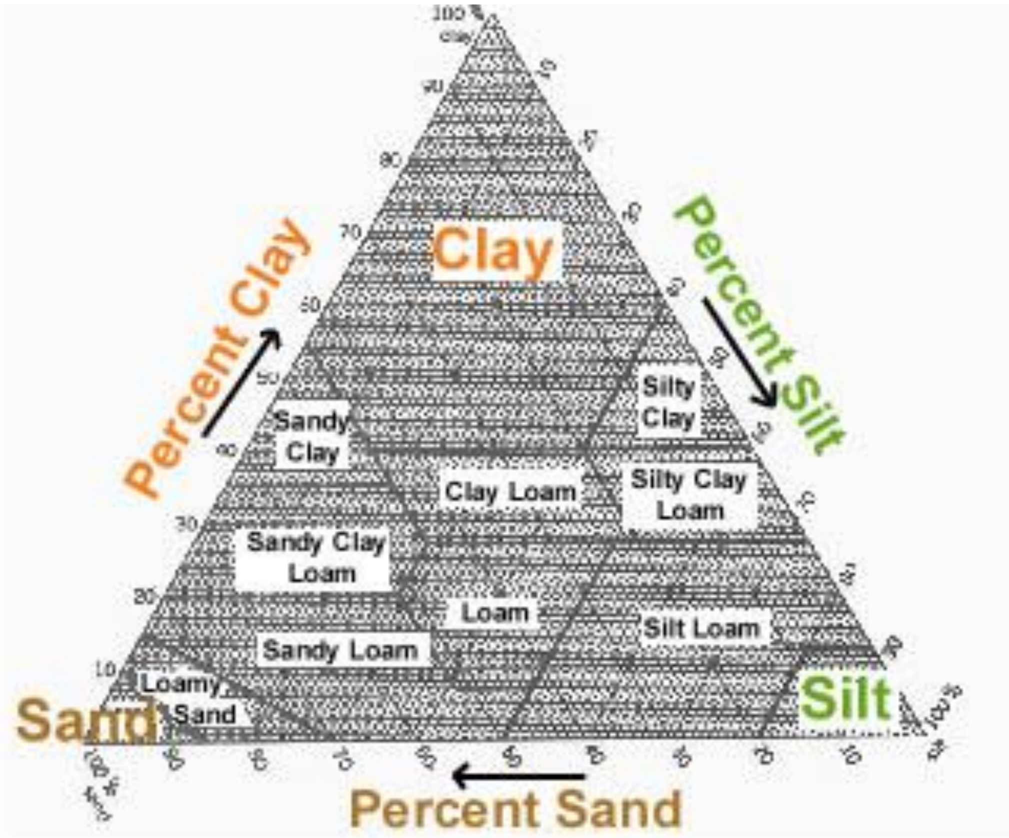

**Figure 1:** Soil Texture Triangle (Donahue *et al.*, 1983)

### 2.1.5 Soil Structure

Soil structure is the term used when the arrangement of the particles of the soil is considered. Structure refers to the aggregation of primary soil particles (sand, silt and clay) into compound particles or cluster of primary particles which are separated by the adjoining aggregates by surfaces of weakness. The availability of nutrients, the effect of texture in regard to moisture and air relationships, action of microorganisms and root growth are adjusted by structure. For instance, a very plastic clay (60% clay) is good for crop product if it has a well-developed granular structure which expedites water movement and aeration (Black *et al.*, 1965). Correspondingly, soil in spite of having a heavy texture, can have a strongly developed structure, hence making it not very suitable for aquaculture. This is because this soil allows high seepage losses.

According to Black *et al.*, (1965), grade of structure is the degree of aggregation and expresses the differential between cohesion within aggregates and adhesion between aggregates. These properties change with the moisture content of the soil and it should be resolved when there is normal moisture content. It should not be determined when the soil is unusually dry or unusually wet. The four major grades of structure graded from 0 to 3 are listed in **Table 3**.

**Table 3:** Grades of soil structure (FAO 1974)

| Grades | Description |
|---|---|
| 0 = Structureless: | No observable aggregation or no definite orderly arrangement of natural lines of weakness. Massive if coherent; single grain if noncoherent. |
| 1 = Weak: | That degree of aggregation characterized by poorly formed indistinct aggregates that are barely observable in place. When disturbed, soil material that has this grade of structure breaks into a mixture of few entire aggregates many broken aggregates and much unaggregated material. |
| 2 = Moderate: | Well-formed distinct aggregates that are moderately durable and evident but not distinct in undisturbed soil. When disturbed, they break down into a mixture of many distinct entire aggregates, some broken aggregates and little unaggregated material. |
| 3 = Strong: | Durable aggregates that are quite evident in undisturbed soil that adhere weakly to one another. When removed from the profile the sokl material consists very largely of entire aggreates and includes few broken ones and little or no nonaggregated material. |

### 2.1.6 Consistence

Consistence is the ability a soil to resist rupture or deformation. It is determined by the adhesive and cohesive properties of the soil volume. The consistence of the soil is a term used to label the revelation of the adhesive and cohesive properties of soil at different moisture contents. A comprehension of the consistence of the soil is necessary for various operations, such as, traffic and pond constructions, and in tillage operations Consistence also gives an explanation of the soil texture (FAO 1974)

FAO (1974) characterized consistence for three moisture levels:

i. Wet soil - non sticky, slightly sticky, sticky, very sticky; non plastic, slightly plastic, plastic and very plastic.
ii. Moist soil - loose, very friable, friable, firm, very firm, extremely firm.
iii. Dry soil - loose, soft, slightly hard, hard, very hard, extremely hard.

### 2.1.7 Soil Colour

Soil colour gives a suggestion of the various operations happening in the soil, and also it gives an indication of the type of minerals in the soil. An example can be given of the red colour in the soil, which is due to the abundance of iron oxide under oxidised conditions in the soil. Oxidised soil conditions mean well-drained soil; yellow colour is as a result of hydrated iron oxides and hydroxide; Abundant pale yellow mottles paired with very low pH are demonstrative of possible acid sulphate soils; dark colour is commonly due to the accumulation of highly decayed organic matter; black nodules are due to manganese oxides; mottling and gleying are connected with poor drainage and or high water table. Colours of soil matrix and mottles present the water and drainage conditions in the soil and consequently suitability of the soil for aquaculture as stated by Munsell Soil Color Charts, (1973).

## 2.2 Particle Density

The particle density or true density of a particulate solid or powder, is the density of the particles that make up the powder, in comparison to the bulk density, which estimates the average density of a large volume of the powder in a specific medium (usually air). The particle density is a somewhat well-defined quantity, because it does not depend on the degree of compaction of the solid. Bulk density on the other hand has different values depending on whether it is measured in the freely settled or compacted state (tap density). Nonetheless, there are a number of definitions of particle density. These definitions differ in terms of whether pores are included in the particle volume, and whether voids are included.

The measurement of particle density can be done in a number of ways such as based on:

### i. Archimedes' principle

The powder is poured into a container whose volume is known, and weighed. The pycnometer is then filled with a fluid. The density of the fluid must be known, and it must not dissolve the powder. The volume of the powder is estimated by the difference between the volume as displayed by the pycnometer, and the volume of liquid added (that is the volume of air removed). Another method, which is also similar, but does not include pore volume, is to suspend a known mass of particles in molten wax of known density, bubbles are allowed to escape, if there are any, the wax is allowed to solidify, and then the volume and mass of the wax or particulate brick is then measured

A slurry of the powder in a liquid whose density is known can also be used with a hydrometer to measure particle density by buoyancy. The weight of the sample in air, and also in a liquid of known density can be measured by another method based on buoyancy.

A column of liquid that has a density gradient can also be prepared: The column should hold a liquid of constantly changing composition, so that the maximum density (at the bottom) is higher than that of the solid, and the minimum density is lower. If a small sample of powder is allowed to settle in this column, at the point where the liquid density is equal to the particle density, it will come to rest.

### ii. Volumetric measurement

The volume of a powder sample can be measured using a gas pycnometer. A sample whose mass is known is poured into a chamber of known volume. The chamber is connected by a closed valve to a gas reservoir, whose volume is also known, at a higher pressure than the chamber. After the valve is opened, the final pressure in the system permits the total gas volume to be estimated by application of Boyle's law.

A mercury porosimeter is an instrument that allows the total volume of a powder to be determined, as well as the volume of pores of different sizes: Powder with a known mass is drenched in mercury. The mercury does not penetrate the interparticle spaces or the pores of the sample at ambient pressure. The mercury penetrates smaller and smaller pores, at increasing pressure, with the relationship between pore diameter and pressure being known. For a complete characterization of the sample's porosity, a continuous trace of pressure versus volume can then be generated.

### 2.2.1 Bulk Density

Bulk density is a property of powders, granules, and other "divided" solids, especially used in reference to mineral components (soil, gravel), chemical substances, (pharmaceutical) ingredients, foodstuff, or any other masses of corpuscular or particulate matter. It is estimated as the mass of many particles of the material divided by the total volume they occupy. The total volume includes particle volume, inter-particle void volume, and internal pore volume (Buckman *et al.*, 1960).

Bulk density is not an intrinsic property of a material; it can alter depending on how the material is handled. For example, a powder poured into a cylinder will have a particular bulk density; if the cylinder is disturbed, the powder particles will move and usually settle closer together, resulting in a higher bulk density. For this reason, the bulk density of powders is usually reported both as "freely settled" (or "poured" density) and "tapped" density (where the tapped density refers to the bulk density of the powder after a specified compaction process, usually involving vibration of the container.)

The bulk density of soil greatly depends on the minerals that make up the soil and the degree of compaction. The density of quartz is around $2.65 g/cm^3$ but the dry bulk density of a mineral soil is usually about half that density, between 1.0 and $1.6 g/cm^3$. Soils that are high in organics and some friable clay could have a bulk densities below $1 g/cm^3$.

To calculate the total bulk density or the wet bulk density of a sample, this sample is weighed, giving the mass $M_t$. For the dry bulk density, the sample is oven dried and weighed, giving the mass of soil solids $M_s$. The relationship between these two masses is $M_t = M_s + M_l$, where $M_l$ is the mass of substances lost on oven drying (usually, mostly water). The dry and wet bulk densities are calculated as:

Dry bulk density = mass of soil/ total volume

$$\rho_b = \frac{M_s}{V_t} \tag{2.1}$$

Wet bulk density = mass of soil plus liquids/ total volume

$$\rho_t = \frac{M_t}{V_t} \tag{2.2}$$

The dry bulk density of a soil is inversely related to the porosity of the same soil, that is, soils that have more pore space have lower bulk densities (USDA 2013).

### 2.3 Water Content

The water content in soils is usually expressed as either a dimensionless ratio of two masses or two volumes, or is given as a ratio of a mass per unit volume. The dimensionless ratios can be expressed either as decimal fractions or percentages if multiplied by 100. To avoid confusion between the two dimensionless water content ratios, their basis (that is, mass or volume) should always be stated (Gardner, 1986). According to Gardner (1986), the water content in soils on a mass basis, $w$, is defined as the ratio of the mass of the liquid phase (water), $M_l$, in the given soil sample to the mass of the solid material, $M_s$, according to the following expression:

$$w = \frac{M_l}{M_s} \tag{2.3}$$

The volumetric water content in the soil (also called the volume wetness or volume fraction of soil water) represents the fraction of the total volume of soil that the water contained in the soil occupies. Assuming that $V_l$ is the volume of the liquid phase (water) in the soil sample and that $V_t$ is the total volume of the sample, the volumetric water content, $\theta$, can then be expressed as follows:

$$\theta = \frac{V_l}{V_t} = \frac{V_l}{V_s+V_p} \tag{2.4}$$

where $V_s$ is the volume of the solid phase and $V_p$ represents the pore space.

From the definition presented in Equations 2.3 and 2.4, the volumetric water content, can be deduced in terms of the mass-basis water content, $w$, according to the following formula:

$$\theta = \frac{V_l}{V_t} = \frac{M_l/\rho_w}{M_s/\rho_b} = \left(\frac{M_l\rho_b}{M_s\rho_w}\right) = w(\frac{\rho_b}{\rho_w}) \tag{2.5}$$

where $\rho_b$ is the bulk density of the soil and $\rho_w$ is the density of the water.

The volumetric water content can also be expressed in terms of the total porosity, $P_t$, and the water saturation (or saturation ratio), $R_s$, according to the following expression:

$$R_s = \frac{\theta}{P_t} \tag{2.6}$$

Where $\theta = P_t R_s$ and $P_t$ is the total porosity and $R_s$, the saturation ratio, is calculated as the ratio of the volume of water, $V_l$, to the volume of the pore space, $V_p$. Therefore, taking the definitions of $P_t$ and $R_s$ into consideration, the expression for the volumetric water content can be rewritten as follows:

$$\theta = P_t R_s = \left(\frac{V_p}{V_t}\right)\left(\frac{V_l}{V_p}\right) = \frac{V_l}{V_t} \tag{2.7}$$

The possible values for the volumetric water content range from near zero for dry soils approaching zero saturation, up to the value of the total porosity for fully saturated soils. The lower limit of zero for the volumetric water content is hardly achievable because it is difficult to completely eliminate the water from the soil. In sandy soils, the upper limit, which is also equal to the total porosity, $P_t$, is also hardly achievable because of the difficulty of eliminating all the air bubbles in the soil in order to saturate it completely. But because clayey soils swell upon wetting, the values for these types of soils can exceed their total porosity.

Soil moisture content is very dependent on soil type. A saturated coarse sandy soil can hold far less water than a saturated heavy silty clay. Sand has large particles which take up a lot of physical space. Also, as sand particles do not bind water, a lot of water will drain out of the sand due to gravity before field capacity is reached. For these two reasons, sand has a much lower maximum and minimum water content than a clay soil does (Gardner, 1986).

### 2.3.1 Porosity

The shape and arrangement of soil particles play a very critical role in determining porosity. Porosity or pore space is the amount of air space or void space between soil particles. Infiltration, groundwater movement, and storage occur in these void spaces. The porosity of soil or geologic materials is the ratio of the volume of pore space in a unit of material to the total volume of material (Fetter, 1994).

A mathematical equation of porosity can be expressed in percentage as:

Porosity or $n = \frac{V_{void}}{V_{total}}\ x\ 100\%$ (2.8)

Where $V_{void}$ is the volume of voids and $V_{total}$ is the total volume of the sample.

The packing or arrangement of the soil particles plays an important role in porosity. In the **Figure 2**, the particles stacked directly on top of each other (cubic packing) have higher porosity than the particles in a pyramid shape sitting on top of two other particles (rhombohedral packing). There are differences in pore space. When smaller particles are mixed with larger particles, as the diagram shows, the smaller particles could fill in the void spaces between the larger particles, which would result in a lower porosity.

Not all particles are round or spheres. Particles come in many shapes and these shapes pack in different number of ways that may increase or decrease porosity. Generally, a mixture of grain sizes and shapes, results in lower porosity. One important point to note is that the diameter size of the grain does not affect porosity. As stated above, porosity is a ratio of void space to total volume. **Figure 2** shows the various forms of porosity arrangement as presented by Fetter (1994).

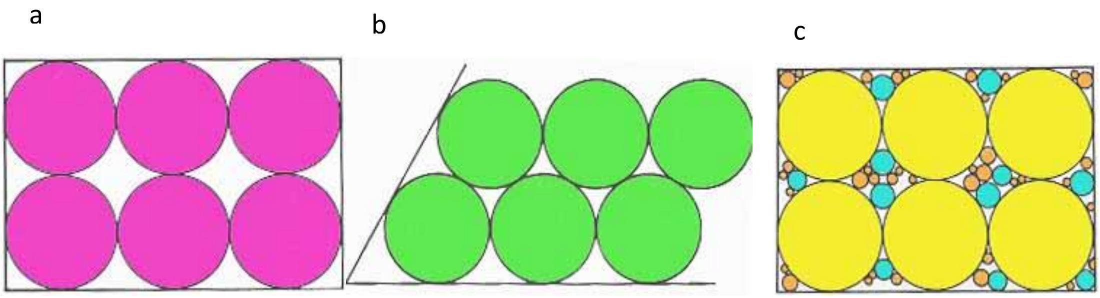

**Figure 2:** sample porosity arrangement; (a) Cubic packing, (b) Rhombohedral Packing, (c) Cubic Packing with Smaller Grains Filling the Void Spaces (Fetter, 1994).

Different sediments have different porosity ranges. These ranges have been presented by Fetter (1994) as shown in **Table 4**.

**Table 4:** Porosity Ranges for Sediments **(**Fetter, 1994)

| **Material** | **Porosity (%)** |
|---|---|
| well-sorted sand or gravel | 25-50 |
| sand and gravel, mixed | 20-35 |
| glacial till | 10-20 |
| silt | 35-50 |
| clay | 33-60 |

Sands have large pore spaces, whereas clays have many small pore spaces. Both sand and clay can have high porosity. According to Holt (1965), the porosity value in the sand-plain department has a value between 32% and 38%.

### 2.3.2 Permeability

The size of pore space and interconnectivity of the spaces help determine permeability, therefore shape and arrangement of grains contribute to permeability determination. Permeability is a measure of a soil's or rock's ability to transmit a fluid, usually water. Water can percolate between granular void or pore spaces, and fractures between rocks. Materials with larger pore space are more permeable. Materials with mixed grain sizes have lower permeability because the smaller grains fill the voids created by the larger grains. "The most expeditious water and air movement is in sands and strongly lumped soils, whose aggregates behave like sand grains and pack to form many large pores" as shown in **Table 5** (Fetter, 1994).

Clays have small grain sizes with large surface areas, due to this fact, clays have low permeability. All these factors result in increased friction in clays. The pore spaces are also not well connected. In the subsurface, confining layers are often created by clay. Permeability determination in rocks with fractures depends on the size of the voids, degree of interconnectedness and the amount of open space. Soils with higher permeability have greater seepage.

**Table 5:** Saturated hydraulic conductivity for Sediments (Fetter, 1994).

| **Material** | **Hydraulic Conductivity (cm/s)** |
|---|---|
| well-sorted gravel | $10^{-2}$ to 1 |
| well-sorted sands, glacial outwash | $10^{-3}$ to $10^{-1}$ |
| silty sands, fine sands | $10^{-5}$ to $10^{-3}$ |
| silt, sandy silts, clayey sands, till | $10^{-6}$ to $10^{-4}$ |
| clay | 10-9 to $10^{-6}$ |

### 2.3.3 Soil Water Potential

Water potential is the potential energy of water per unit volume relative to pure water in reference conditions. Water potential determines the ability of water to move from one area to another due to osmosis, gravity, mechanical pressure, or matrix effects such as capillary action. Capillary action is caused by surface tension and contact angle. Water potential helps to understand water movement within plants, animals, and soil. Water potential is customarily expressed in potential energy per unit volume and very often is represented by the Greek letter $\Psi$. An expression of the ability of soil water to move is expressed by the soil water potential ($\Psi$). The soil water potential ($\Psi$) is defined as the work water can do as it moves from its present state to the reference state. The reference state is the energy of a pool of pure water at an elevation set to be zero. Water potential accommodates a variety of different potential drivers of water movement, which may operate in the same or different directions. It is common for many potential factors to be important within complex biological systems. For example, the addition of solutes to water lowers the water's potential, that is, it makes it more negative. In the same way, the increase in pressure increases its potential (makes it more positive). Water will move from an area of higher water potential to an area that has a lower water potential if the flow is not restricted. A very typical example is water that contains a dissolved salt, like sea water or the solution within living cells. Relative to the pure water reference, these solutions commonly have negative water potentials. Water molecules will advance from the condition of pure water to the more negative water potential of the solution if there is no inhibition on flow. Flow continues until the difference in solute potential is balanced by another force, for example, pressure potential. The total soil water potential ($\Psi_t$) is greatly affected by three important factors. These factors are soil water potentials of $\Psi_g$ Gravitational $\Psi_o$ Osmotic $\Psi_m$ Matric. The general relationship between total soil water potential ($\Psi_t$) and the various factors is expressed as:

$$\Psi_t = \Psi_g + \Psi_o + \Psi_m$$

Gravitational force acts on soil water in the manner it does on all other bodies. The gravitational potential ($\Psi_g$) of water near the soil surface, in a soil profile, is always higher than the gravitational potential ($\Psi g$) in the subsoil, due to heavy precipitation or irrigation, hence, the difference in gravitational potential causes water to flow downward deeper into the soil profile.

The osmotic potential ($\Psi_m$) is attributable to the attraction between a water molecule and various ions (anions and cations) and solutes, example soluble salts, in the soil solution. Large amounts of soluble salts in the water molecule result in osmotic potentials ($\Psi_m$) that reduce soil water potential. Even though water may be present, this makes it difficult for plants to extract soil water. This is referred to as physiological drought and it is the reason why plants wilt and look stunted in saline soil profiles.

Adhesion or attraction of water to the soil matrix presents a matric force (adsorption and capillarity) which reduces energy of water particles near surfaces. It is obvious that the various soil water potentials do not act in the same way; on the grounds that their separate gradients may not be equally effective in causing water flow.

Notwithstanding, the advantage in using the total soil water potential ($\Psi_t$) is that it provides a consolidated measure by which the state of water in soil can be expressed.

Whilst these forces and pressures are important, in distinct field situations the matric potential ($\Psi_m$) is the most important in all unsaturated soil because the interaction between soil and water is universal. The movement and availability of water to move throughout the soil profile is therefore primarily determined by the matric potential ($\Psi_m$.) (UNSW 2007).

### 2.3.4 Water Retention Curve

The relationship between the water content, $\theta$, and the soil water potential, $\psi$, is termed the water retention curve. The water retention curve is peculiar to different types of soil. It is sometimes referred to as the soil moisture characteristic. It is used to anticipate the soil water storage, water supply to the plants (that is field capacity), and soil aggregate stability. Different wetting and drying curves can be distinguished as a result of the hysteretic effect of water filling and draining the pores,

A water retention curve showing the general features is illustrated in **Figure 3** (van Genuchten, 1980), where the volume water content, $\theta$, is plotted against the matric potential, $\Psi_m$. When the potentials get close to zero, the soil gets close to saturation, and water is held in the soil principally by capillary forces. From the diagram, as the water content ($\theta$) decreases, it results in stronger binding of the water, and at small potentials (more negative, approaching wilting point) water is strongly bound in the smallest of pores, at contact points between grains and as films bound by adsorptive forces around particles.

Clayey soils, with osmotic and adhesive binding, will release water at lower potentials or more negative potentials, but sandy soils will comprise primarily capillary binding, and will therefore release most of the water at higher potentials. Peaty soils will usually exhibit much higher moisture contents than clayey soils at any given potential. The water holding capacity of any soil largely relies on the porosity and the nature of the bonding in the soil.

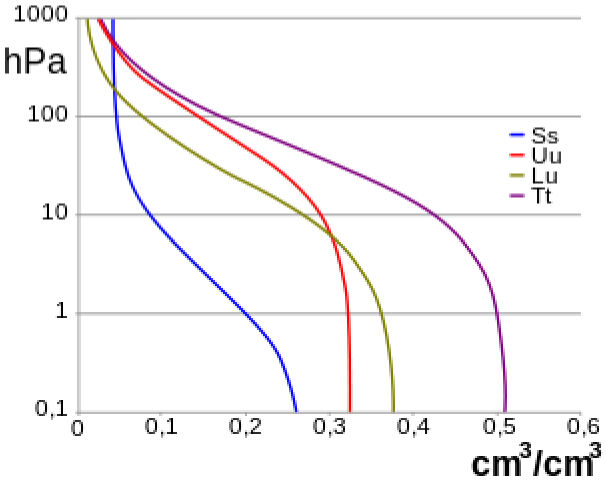

**Figure 3:** water retention curve for a sand, clay loam, clay, and peat. (van Genuchten, 1980).

## 2.4 Darcy’s Law

Darcy's law is a well-researched derived fundamental equation that describes the flow of a fluid through a porous medium. The law was defined by Henry Darcy based on the results of experiments (Darcy, 1856) on the flow of water through beds of sand. It as well forms the scientific grounds of fluid permeability used in the earth sciences, particularly in hydrogeology.

Though Darcy's law, which is an expression of conservation of momentum, was determined experimentally by Darcy, it has since been derived from the Navier-Stokes equations by homogenization (Whitaker, 1986). It is comparable to Fick's law in diffusion theory, Fourier's law in the field of heat conduction and Ohm's law in the field of electrical networks. There are many applications of Darcy's law, one of such is to estimate water flow through an aquifer. The equation of conservation of mass along with Darcy's law is equivalent to the groundwater flow equation. It is one of the basic relationships of hydrogeology. Darcy's law can also be used to define oil, water, and gas flows through petroleum reservoirs.

Darcy's law at constant elevation is a simple proportional relationship between the instantaneous discharge rate through a porous medium, the viscosity of the fluid and the pressure drop over a given distance.

$$Q = \frac{-KA}{\mu}\frac{(\rho_b - \rho_a)}{L} \tag{2.9}$$

The total discharge, $Q$ (with units of volume per time, example: $m^3/s$) is equal to the product of the intrinsic permeability of the medium, $K(m^2)$, the cross-sectional area to flow, $A$ (with units of area, example: $m^2$), and the total pressure drop $(\rho_b - \rho_a)$, (Pa), all divided by the viscosity, $\mu$ (Pa·s) and the length over which the pressure drop is taking place $(L)$. Fluids flow from high pressures to low pressures, therefore the negative sign is essential. If the inlet and outlet are at different elevations, the elevation head must be taken into consideration. The flow will be in the positive 'x' direction if the change in pressure is negative, that is where $\rho_a$ is greater than $\rho_b$. When both sides of the equation is divided by the area, using a more general notation leads to:

$$q = \frac{-K}{\mu}\nabla p \tag{2.10}$$

where $q$ is the flux, which is the discharge per unit area, (with units of length per time, m/s) and $\nabla p$ is the pressure gradient vector (Pa/m). This value of flux, often referred to as the Darcy flux, is not the velocity which the fluid traveling through the pores is experiencing. The fluid velocity $(v)$ is related to the Darcy flux $(q)$ by the effective porosity $(\phi)$. The flux is divided by the effective porosity to account for the fact that only a fraction of the total formation volume is available for flow. The fluid velocity would be the velocity a conservative tracer would experience if carried by the fluid through the formation.

$$v = \frac{q}{\phi} \tag{2.11}$$

Darcy's law is a simple mathematical statement which accurately encapsulates several familiar properties that groundwater flowing in aquifers exhibits, including:

i. no flow occurs if there is no pressure gradient over a distance (these are hydrostatic conditions),
ii. flow will occur from high pressure towards low pressure (opposite the direction of increasing gradient, thus the negative sign in Darcy's law) if there is a pressure gradient,
iii. the greater the pressure gradient (through the same formation material), the greater the discharge rate, and

iv. the rate of discharge of fluid will usually be different through different formation materials or even through the same material, in a different direction even if both cases have the same pressure gradient.

### 2.4.1 Darcy-Buckingham law

Edgar Buckingham suggested a modification of Darcy’s law in 1907 (Buckingham, 1907), to define the flux of water through unsaturated soil. Buckingham’s modification was essentially based on two assumptions:

i. The driving force for water flow in isothermal, rigid, unsaturated soil containing no solute membranes and zero air pressure potential is the sum of the matric and gravitational potentials.
ii. The hydraulic conductivity of unsaturated soil is a function of the water content or matric potential.

The Darcy-Buckingham law may be expressed in head units, for vertical flow as:

$$J_{w\,=}\; - K(h)\frac{\partial H}{\partial z} = -K(h)\frac{\partial(h+z)}{\partial z} = -K(h)\left(\frac{\partial h}{\partial z}+1\right) \tag{2.12}$$

Where $H = h + z$ (in units of length, $m$) is the hydraulic head in unsaturated soil and $K(h)$ (in units of length/time, $m/s$) is the unsaturated hydraulic conductivity. For saturated flow, the flux $J_w$ (length/time, $m/s$) is the water flow per unit cross-sectional area per unit time.

The Darcy-Buckingham equation, as shown in equation (2.12), is a differential equation that is written across a microscopically thin layer of soil over which $h$ is constant and $K(h)$ is a constant. Unless the water content and matric potential of the layer are uniform, which occurs only under special conditions, the equation may not be written across a finite layer of soil. The derivative in the equation is a partial derivative, because in unsaturated soil $h$ may be a function of both $z$ and $t$. The partial derivative $\frac{\partial H}{\partial z}$ suggests that the derivative with respect to $z$ is taken at constant $t$; it is the instantaneous value of the slope

$$h\,(z,t): \frac{\partial H}{\partial z} = \left(\frac{\partial H}{\partial z}\right)_t = \lim_{\Delta z\to 0}\frac{h(z+\Delta z,t)-h(z,t)}{\Delta z} \tag{2.13}$$

Where $(.)_t$ means the derivative is estimated at constant $t$. partial derivatives are prescribed for the mathematical description of transient (time-dependent) flow. If the system is a steady state, the partial derivative diminishes to an ordinary derivative, since in steady state, $h$ depends only on $z$ (Jury *et al*., 2004).

## 2.5 Hydraulic conductivity

### 2.5.1 Saturated and Unsaturated soil hydraulic conductivity

Saturated hydraulic conductivity (*Ks*) is a quantitative measure of a saturated soil's ability to transmit water when subjected to a hydraulic gradient. It can also be described as the ease with which pores of a saturated soil allow water movement.

Saturated hydraulic conductivity is a constant, or proportionality constant, that describes the linear relationship between the two variables $J$ and $i$ (**Figure 4**). This relation is described by Darcy's law. It is the slope of the line ($J/i$) showing the relationship between flux and hydraulic gradient. When Darcy's equation is solved for *Ks*, it yields $J/i$. Flux is the quantity of water moving in the direction of, and at a rate proportional to, the hydraulic gradient. If two soils have the same hydraulic gradient, the soil from which the greater quantity of water is discharged, that is, the one with the highest flux, is the more conductive (it has the greatest flow rate). In **figure 4**, the sandy soil produces a higher flux (is more conductive) than the clayey soil at the same hydraulic gradient. The soil with the steeper slope (the sandy soil in **figure 4**) has the higher hydraulic conductivity. Hydraulic conductivity (or slope "*K or Ks*" in **figure 4**) describes the proportional relationship between flux and hydraulic gradient, or in this case, of unidirectional flow in saturated soil. Saturated hydraulic conductivity ("*Ks*") is a quantitative expression of the soil's ability to transmit water under a given hydraulic gradient.

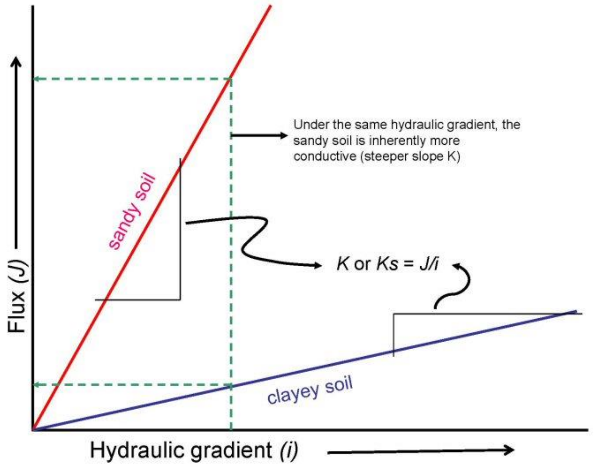

**Figure 4:** Relationship between flux and hydraulic gradient. (USDA 2015).

Hydraulic conductivity (*Ks*) is the slope that defines the relationship, as shown in **figure 4**. At equal hydraulic gradients, soils with higher conductivity have higher flux. This relation is represented by the dotted lines in **figure 4.**

Saturated hydraulic conductivity depends on both soil and fluid properties. It is affected by the soil pore geometry as well as the fluid viscosity and density. When the fluid is more viscous than water, the hydraulic conductivity for a given soil becomes lower. The vertical component of *Ks* can be different from the horizontal component.

Unsaturated soil hydraulic conductivity ($K(\theta)$) is a significant transfer property of soil whose measurement is excessive, difficult, and time-consuming as a result of its large diversities with water content ($\theta$) or matric potential.

## 2.6 Infiltration

Infiltration is the process by which water on the ground surface enters the soil. Infiltration rate in soil science is a measure of the rate at which soil is able to absorb rainfall or irrigation. It is measured in inches per hour or millimeters per hour. As the soil becomes saturated, the rate decreases. Runoff will usually occur unless there is some physical barrier, if the rate of precipitation exceeds the rate of infiltration. Infiltration rate is related to the saturated hydraulic conductivity of the near-surface soil. An infiltrometer can be used to measure the infiltration rate.

Infiltration is driven by two forces: (i) capillary action and (ii) gravity. Although smaller pores provide greater resistance to gravity, very small pores extract water through capillary action in addition to and even against the force of gravity.

The infiltration rate is estimated by some characteristics of the soil including ease of entry, storage capacity, and the rate of transmission through the soil. The soil texture and structure, water content of the soil, vegetation types and cover, soil temperature, and rainfall intensity are all important factors in controlling rate of infiltration and capacity. For example, coarse-grained sandy soils have large voids between each grain and allow water to infiltrate quickly.

The infiltration process can only continue if there is space available for additional water at the soil surface. The available room for additional water in the soil depends on the porosity of the soil (Hogan, 2010) and the rate at which previously infiltrated water can move away from the surface through the soil. The infiltration capacity is the maximum rate that water can enter a soil in a given condition. All of the water will infiltrate if the influx of the water at the soil surface is less than the infiltration capacity. Ponding begins if rainfall intensity at the soil surface occurs at a rate that exceeds the infiltration capacity, and is followed by runoff over the ground surface only when the storage of the depression is filled (Horton, 1942).

## 2.7 The Hydrus software

Hydrus-1D (Šimunek *et al*., 1998) is a windows-based modeling environment used for the analysis of water flow, heat and solute transport in variably saturated porous media (example soil). The Hydrus program solves the Richards' equation for saturated-unsaturated water flow and Fickian-based advection dispersion equations for heat and solute transport numerically. The software has an interactive graphics-based interface for data-preprocessing, discretization of the soil profile, and graphic presentation of the results. As Hydrus-1D does these simulations in one-dimension, Hydrus-2D/3D is an extended version that does the simulations to the second and third dimensions.

# 3. METHODOLOGY

## 3.1 Materials and methods

This experiment was carried out by improvisation of some parts of the equipment. For instance, the sandbox was made from the base of a plastic can. Two holes were drilled by the side of the can, almost close to each other, one a little above the other, and plastic tubes were inserted into the holes. A ring was made from a long plastic tube and placed at the bottom of the plastic can to extract water from the sand in the sandbox. One tube was connected to one hole of a plastic bottle with two holes drilled beneath it, and placed on a level to have the same gradient as that of the sandbox. This tube was connected to the ring in the base of the can acting as the sandbox. The other hole on the side of the can was connected with a tube to a measuring cylinder to measure the water outflow from the sandbox. The other hole beneath the bottle was connected to a jelly can by another tube. The plastic can used for the sandbox had a height of 9cm and a diameter of 50cm.

Sand (PR13) was poured into the plastic container to a height of 6cm and water was poured onto it and left for 24 hours for the sand in the can to be fully saturated. There was a considerable amount of water on the surface of the sand to height of about 2.5cm to form the sandbox, as shown in **Figure 5**.

**Figure 5:** Homemade sandbox filled with water prepared to hold the sand sample to be examined.

The plastic bottle was filled with water to have the same level as the water in the sandbox. When the water in the sandbox rises to a height higher than that in the plastic bottle, due to inflow of water from the sand sample, the gradient between the sandbox and the plastic bottle changes, so water flows from the sandbox to the plastic bottle till equilibrium is reached and both containers have a zero gradient between them. On the other hand, when the water in the plastic bottle rises to a height higher than of the sandbox as a result of water outflow from the sandbox, water flows from the plastic bottle to the sandbox till they both have the same level and a zero gradient.

A hollow cylinder with a diameter of 23.75cm was covered at one end with a clean white cloth and the sand sample (PR 13) was poured into it to a height of 10cm. The sand sample together with the container was placed on the sandbox, as shown in **Figure 6.**

**Figure 6:** Soil sample on sandbox.

The Guelph Permeameter was placed in a basin containing water and a vacuum hand pump was used to draw water into the Guelph Tension Infiltrometer through the porous disk. The vacuum hand pump is connected to the tube at the top of the Guelph reservoir with the base attached to the porous disk. Water is drawn into the reservoir to a desirable height. The process is shown in **Figure 7**.

**Figure 7:** Filling the Guelph reservoir with water.

After filling the Guelph reservoir with water, using the vacuum hand pump, the tube on the top of the reservoir is closed with the “O” ring and the rubber seal, the pressure condition is set with the Marriot Bubbler and the tube on the top of the Marriot Bubble is also closed with the rubber seal. The Guelph Tension Infiltrometer is then lifted and placed on the soil surface, making a very good contact with the soil surface by the porous disk.

The various pressures used for this experiment were: +2cm, +1cm, 0cm, -1cm, -2cm, -3cm, -4cm, -6.1cm and -8cm. At each of these pressures, before the water inflow into the soil could start, the “O” ring at the Bubbler top cap together with the rubber seal plugged in the cap needed to be removed to enable the pressure condition to be set in the soil to ensure the water infiltration into the soil.

**Figure 8** shows how the Guelph Permeameter must be placed on the top of the soil sample to ensure enough hydraulic contact between the porous disk and the soil sample.

**Figure 8:** Guelph Permeameter on the top of the soil to initiate the process of infiltration.

Time readings were taken for the water inflow and outflow (discharge) at each pressure, using a stopwatch.

For this research, two sets of experiments were carried out; two pressure conditions were set in the soil. The first set of experiments were carried out for variably saturated soil sample. A constant pressure condition of +2.5cm was set at the bottom of the soil sample whereas the top of the soil sample was set to varying pressures (+2cm, +1cm, 0cm, -1cm, -2cm, -3cm, -4cm, -6.1cm and -8cm). The second set of experiments was carried out for fully-unsaturated soil sample. A pressure condition of 0cm was set at the bottom of the soil sample and the pressure at the top of the soil was varied from 0cm to -3cm to -6cm.

Using Darcy's law, $Q = KA(\frac{\Delta H}{L})$ (3.1)

Where: $Q$ is the discharge in $m^3/s$, more general form $(L^3T^{-1})$

$K$ is hydraulic conductivity in $m/s$, more general form $(LT^{-1})$

$A$ is the cross-sectional area of the soil sample in $m^2$, more general form $(L^2)$

$\Delta H$ is the vertical drop or change in hydraulic pressure head in $m$, more general form (L),

$L$ is length of soil column in $m$, more general form (L),

all parameters can be calculated for using the known values or parameters.

**Figure 9** is a schematic representation of the soil sample profile.

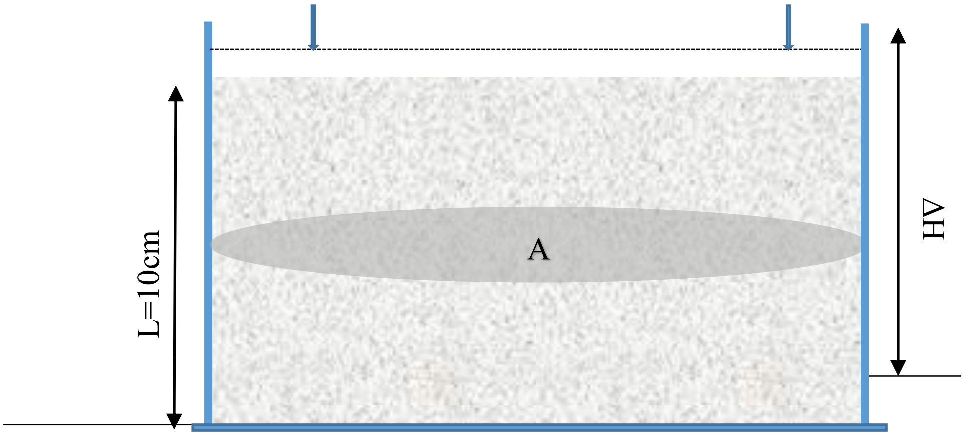

**Figure 9:** A schematic representation of the sand sample profile.

**Figure 10** shows the various apparatus used for the experimental measurements in the laboratory.

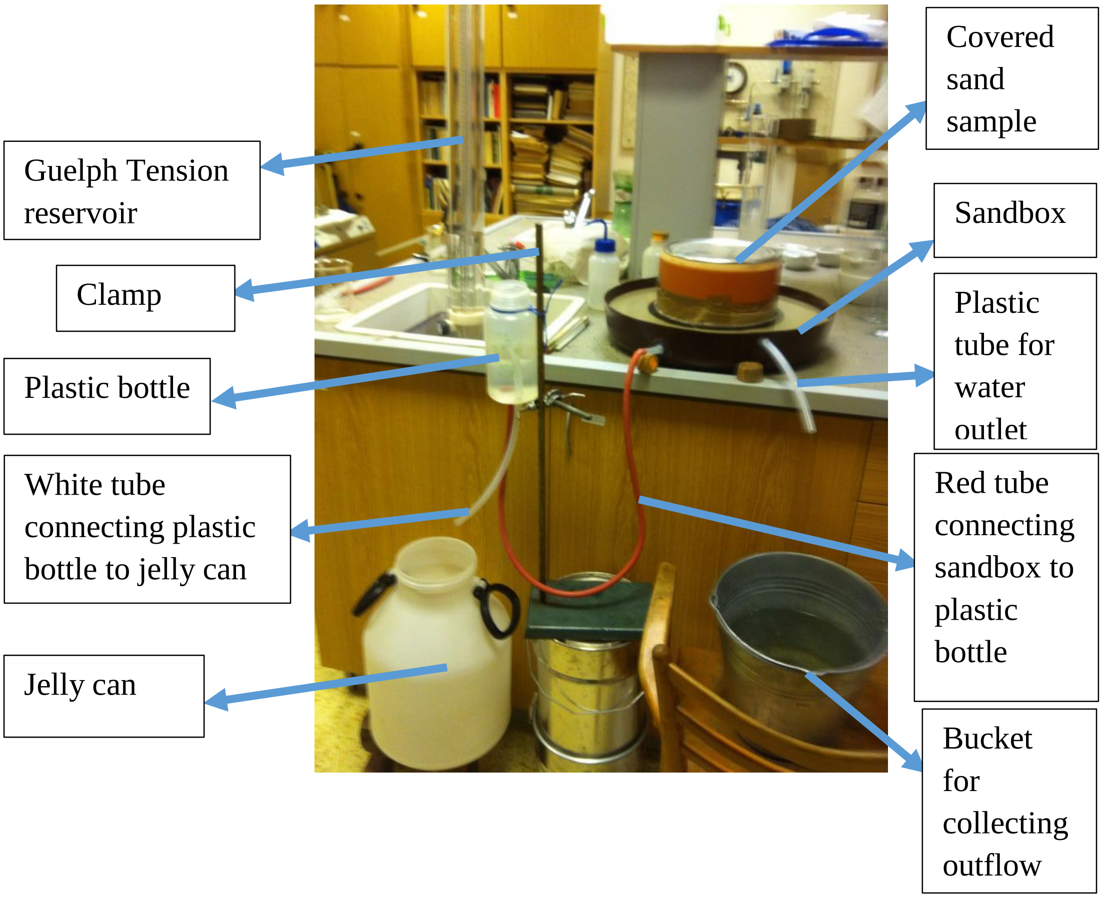

**Figure 10:** The various apparatus used for the experiment in the laboratory.

The dark brown plastic can on the table with the sand is the sandbox, the orange covered cylinder in the sandbox is the container with the soil sample used for the experiment. The soil sample is covered when not in use to prevent evaporation of water from the soil. The plastic bottle on the clamp is set to maintain uniform gradient with sandbox. The red tube from the sandbox to the plastic bottle serves to draw water from the sandbox to the plastic bottle when the gradient of water of the sandbox is higher than that of the plastic bottle. The white tube inserted beneath the plastic bottle reaches to about half the length of the plastic bottle and serves to transport excess water from the half point to the jelly can on the floor.

On the other hand, when the gradient of the water in the plastic bottle is higher than that of the sandbox, water flows from the plastic bottle to the sandbox till a uniform gradient is obtained. The white tube leading from the sandbox to the bucket serves to conduct the water outflow from the soil sample or the discharge from the soil. A calibrated cylinder or flask or measuring cylinder was put in the bucket to measure the discharge from the soil. **Figure 11** shows the overall experimental setup in the laboratory used for this research.

**Figure 11:** The overall experimental setup.

The water inflow can be observed from the drop in level of water in the Guelph reservoir whereas the water outflow can be observed from the white tube leading from the sandbox into the bucket.

### 3.2 Saturated hydraulic conductivity measurement

The volume of water flowing from the sand sample used for the experiment (PR13) was kept constant at $250cm^3$. For each of the upper boundary pressures used for the experiment, (+2cm, +1cm, 0cm, -1cm, -2cm, -3cm ,-4cm, -6.1cm, -8cm), the time taken to fill the $250cm^3$ of the measuring cylinder varied. At each of these pressures, the volume of water was divided by the time taken to get the discharge ($Q$). The discharge ($Q$) was divided by the cross-sectional area of the sand sample ($A$) to get the bottom flux ($V$). The bottom flux obtained was multiplied by the length of the soil column ($L$) and the resulting value was divided by the change in the hydraulic pressure head ($\Delta H$) to obtain the saturated hydraulic conductivity ($Ks$).

## 3.3 Method of comparison of the measured and modelled data

Hydrus-1D was used for the purpose of this experiment (Šimunek *et al*., 1998). The Hydrus-1D model (Šimunek *et al*., 1998) was used to compare the measured bottom flux of the sand sample. In this model, the pressure conditions, time intervals and the measured hydraulic conductivities were fed into Hydrus-1D and the corresponding bottom fluxes were determined.

# 4. RESULTS

## 4.1 Bottom flux and hydraulic conductivity measurements.

The bottom flux of the sand sample was measured for different pressures at the upper boundary, keeping the lower boundary pressure constant. The cross-sectional area of the sand was estimated to be $443.07cm^2$. The time taken to collect a volume $250cm^3$ of water flowing out of the sand was recorded and used to estimate the bottom flux. Similarly, the corresponding hydraulic conductivities were calculated for different pressures. The upper boundary pressure at -6.1cm gave an irregular flow. The flow was not consistent, it stopped and started again at irregular times. There was no flow at an upper boundary pressure of -8cm.

**Table 6** shows the results of the bottom flux and the corresponding hydraulic conductivities.

**Table 6:** Measured hydraulic parameters.

| Pressure (cm) | Cumulative time (s) | Average Bottom Flux (cm/s) | Hydraulic Conductivity (cm/s) |
|---|---|---|---|
| **+2** | 331 | -0.0091728 | -0.00958 |
| **+1** | 331 | -0.008546813 | -0.010005049 |
| **0** | 445 | -0.006944 | -0.00926 |
| **-1** | 513 | -0.0053661 | -0.00826 |
| **-2** | 591 | -0.0045057 | -0.00819 |
| **-3** | 656 | -0.00389 | -0.00858664 |
| **-4** | 825 | -0.00299 | -0.00854 |

**Figures 12 and 13** show the results of the measured bottom flux and the modelled flux, respectively at an upper boundary pressure of +1cm.

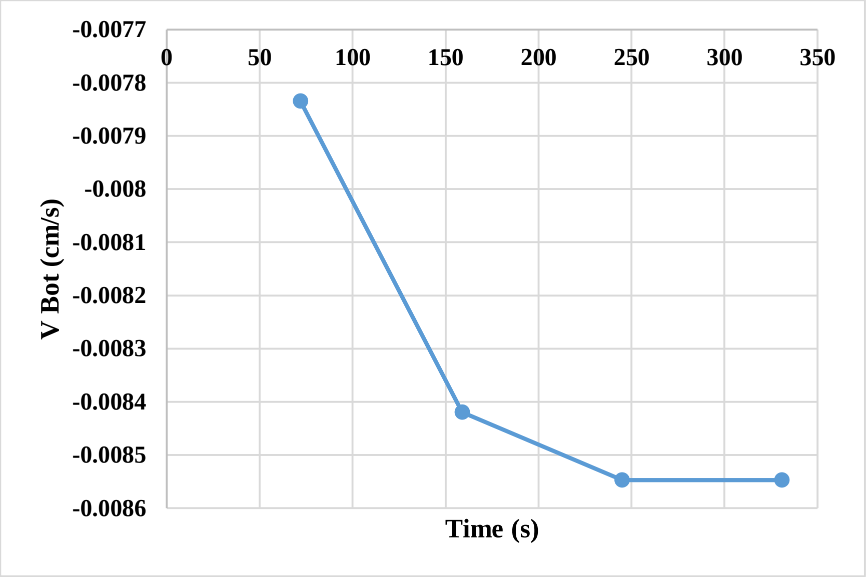

**Figure 12:** Measured Bottom flux against Time at pressure of +1cm.

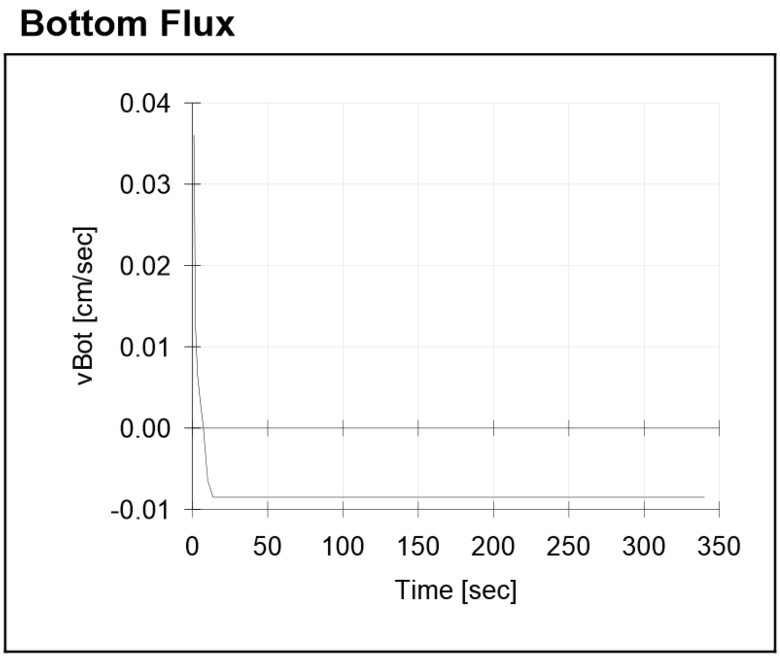

**Figure 13:** Bottom flux against Time by Hydrus-1D model at pressure of +1cm

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Similarly, at a pressure of 0cm, the measured flux and the modelled flux by the Hydrus-1D are shown in **Figures 14 and 15**, respectively.

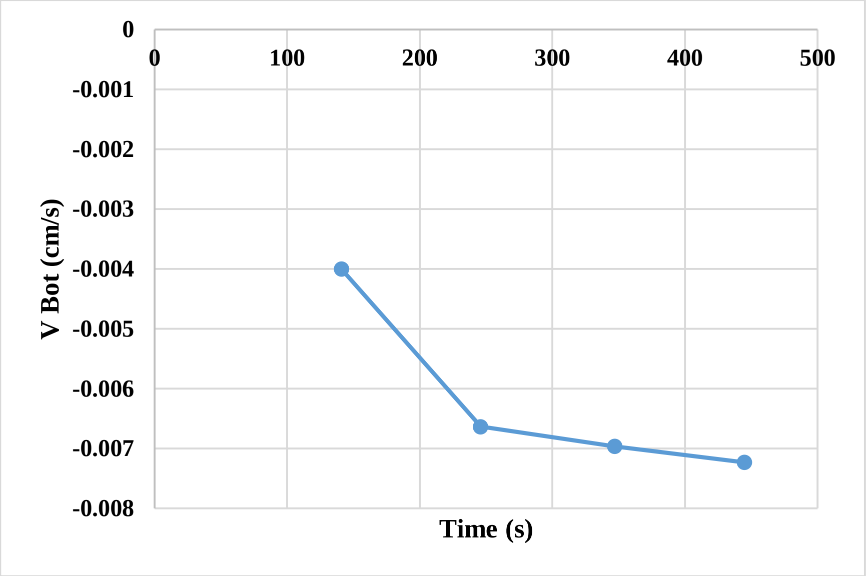

**Figure 14:** Measured Bottom flux against Time at pressure of 0cm.

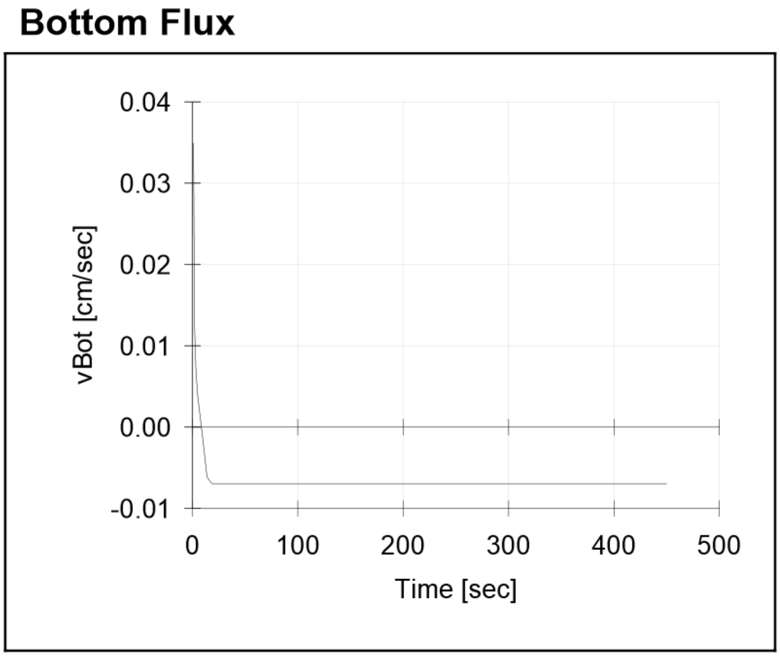

**Figure 15:** Bottom flux against Time by Hydrus-1D model at pressure of 0cm

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Also, at a pressure of -1cm, the measured flux and the modelled flux by the Hydrus-1D are shown in **Figures 16 and 17**, respectively.

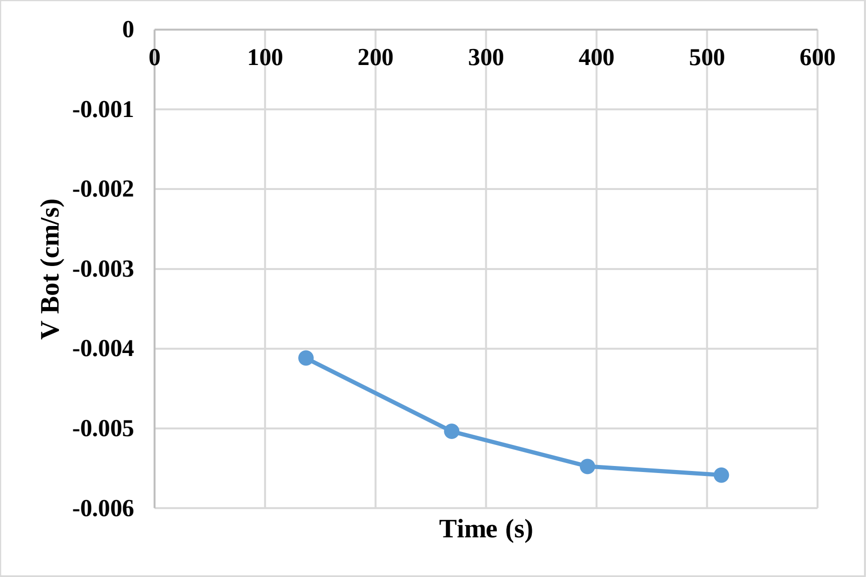

**Figure 16:** Measured Bottom flux against Time at pressure of -1cm

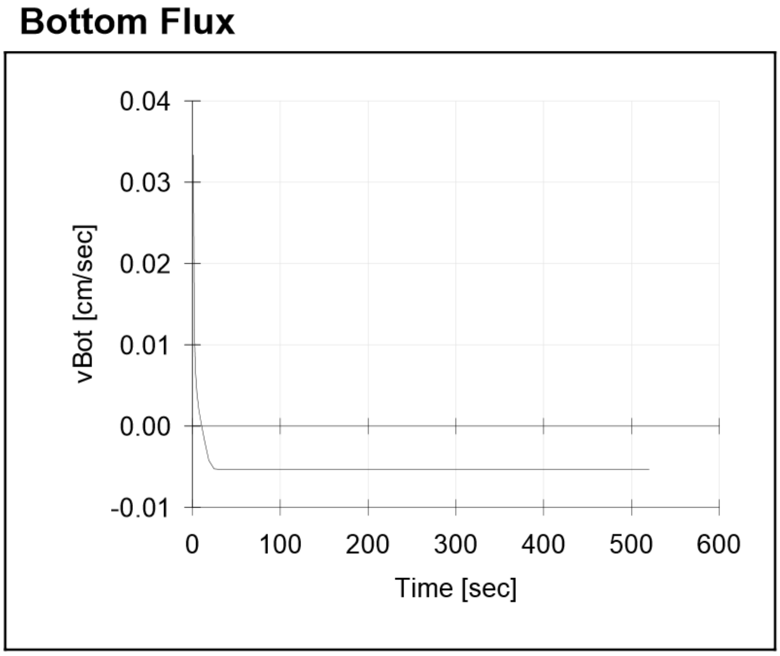

**Figure 17:** Bottom flux against Time by Hydrus-1D model at pressure of -1cm

The main objective was to achieve a steady state condition of the bottom fluxes, and from the results presented, that was achieved. Results for measured and simulated fluxes at upper boundary pressures of +2cm, -2cm, -3cm and -4cm are shown in appendix 8, **Figures 18, 19, 20, 21, 22, 23, 24** and **25**, in their respective order.

**Table 7** illustrates the measured bottom fluxes at the various pressures and their respective modelled bottom fluxes with their corresponding differences.

| **UPPER B.C** | **MEASURED (cm/s)** | **MODEL (cm/s)** | **DIFFERENCE (cm/s)** | **PERCENTAGE DIFFERENCE** |
|---|---|---|---|---|
| **+2cm** | -0.0091728 | -0.009101 | 0.0000718 | 0.78% |
| **+1cm** | -0.008546813 | -0.0085043 | 0.0000425 | 0.50% |
| **0cm** | -0.006944 | -0.006945 | 0.000001 | 0.01% |
| **-1cm** | -0.0053661 | -0.0053182 | 0.0000479 | 0.89% |
| **-2cm** | -0.0045057 | -0.0042607 | 0.000245 | 5.44% |
| **-3cm** | -0.00389 | -0.0033108 | 0.000579 | 14.88% |
| **-4cm** | -0.00299 | -0.0021924 | 0.000798 | 26.69% |

**Table 7:** Comparison of measured bottom fluxes and modelled bottom fluxes**.**

# 5. DISCUSSION

## 5.1 Measured and modelled bottom fluxes.

The measured bottom flux was highest in an upper boundary pressure condition of +2cm and lowest in -4cm. From **Table 7**, it can be noted that the measured bottom flux was decreasing with decreasing upper boundary pressure. This is due to the fact that water takes a longer time to flow through a soil profile that is unsaturated, or less saturated. The more saturated a soil sample is, the less time it takes for water to flow through, or the faster the movement of water through it. The more unsaturated a soil sample is, the slower the movement of water through it, as the water molecules need to fill the pores in the unsaturated zone first before excess (if any) flows through the soil.

The positive upper boundary pressures set in the sand sample created a fully saturated state of the sand sample during the infiltration process. In the case of a pressure of +2cm, this means that a 2cm layer of water was created on the top of the soil sample. The soil sample at this upper boundary pressure is fully saturated. The measured bottom flux was -0.0091728cm/s, as shown in **Table 7**. The negative sign shows the direction of flow, meaning the flow was a vertical downward flow.

Comparably, the bottom flux observed with the Hydrus-1D model gave a value of -0.009101cm/s, as shown in **Table 7**. The difference in bottom flux between the measured and simulated values was 0.0000718cm/s. The difference in bottom flux between the measured and modelled at a pressure of +1cm was 0.0000425cm/s.

The negative pressures created an unsaturated state in the top few centimeters of the soil column depending on the magnitude of the negative pressure condition set at the upper boundary. One of such comparisons is the comparison between the measured and simulated bottom flux at a pressure of -2cm. At an upper boundary condition of -2cm, it means the top 2cm of the soil profile was unsaturated, and the lower 8cm was saturated. So a considerable amount of time will be needed to fill the pores in the top 2cm of the soil profile. The measured bottom flux at -2cm was -0.0045057cm/s and the bottom flux observed with the Hydrus-1D model at pressure -2cm was -0.0042607cm/s, as shown in **Table 7**. The difference between the measured value and the simulated value was 0.000245cm/s.

Some minute differences occurred between the measured bottom flux and the modelled bottom flux when the upper boundary conditions of the soil sample were set at the following pressures, +2cm, +1cm, 0cm, -1cm, -2cm, -3cm and -4cm. The lower boundary condition was kept constant, at a pressure of +2.5cm. **Table 7** presents the various measured bottom fluxes and modelled bottom fluxes and the differences. **Table 7** again shows that, the higher the unsaturated state of the soil, the greater the difference between the measured and modelled bottom flux.

The infinitesimal differences between the measured and simulated bottom fluxes could be due to the following factors:

i. Minor errors in calibrations and measurements: The calibrations in the Marriott Bubbler used in the pressure measurements had error margins of +1mm and -1mm. These error margins could play a role in the final measurements of the bottom flux. Also due to human error, the time taken for each measurement could not be very accurate.

ii. The flow was not totally vertical: Due to the dimensions of the sandbox, the flow of water from the sand sample to the sandbox was not totally vertical. Lateral flows were observed as well. Therefore the bottom flux measured could have a little bit of inaccuracy.

## 5.2 Inverse Modelling

A simulation using the Hydrus-1D was done and the results were compared with the measured data. The main parameter looked for in the Inverse Modelling was the hydraulic conductivity. The average saturated hydraulic conductivity of the Hydrus-1D model was 0.0097894cm/s. The measured average saturated hydraulic conductivity was 0.0097925cm/s. The difference between these two values is 0.0000031cm/s.

Similarly, the average unsaturated hydraulic conductivity was determined with the Hydrus-1D model and that value was compared with the measured average unsaturated hydraulic conductivity. The modelled value of the average unsaturated hydraulic conductivity was 0.008393cm/s and the measured value was 0.00839416cm/s. The difference in the two values was 0.00000116cm/s.

# 6. CONCLUSION

Results showed that the measured data obtained from the laboratory experiments, compared with the modelled data, were almost the same. The very minute differences were due to the errors that might have occurred in the measurements of the pressure and time, in the laboratory experiments; and the lateral flow observed due to the dimension of the sandbox used for the experiment. These differences between the measured data and modelled data are not very significant. The simulation of Hydrus-1D using the van Genuchten equation was adequate to estimate the bottom fluxes and hydraulic conductivities of the sand sample PR13 used for this study. Hydrus-1D model is a more preferred model, because it has flexible water flow boundary conditions, a minimum calculation interval of 1s, it is very easy to use and very understandable. Hydrus-1D model is therefore a more appropriate, flexible and accurate model that can be used to predict accurately the water dynamics in a soil column because the results obtained were not far from reality.

# 7. REFERENCES

Abbasi, F., Feyen, J., and Van Genuchten, M. T. 2004. Two-dimensional simulation of water flow and solute transport below furrows: model calibration and validation. Journal of Hydrology, vol. 290, no. 1-2, pp. 63–79.

Beven, K. and Germann, P. 1982. Macropores and water flow in soils. Water Resources Research, vol. 18, no. 5, pp. 1311–1325.

Black, C. A., Evans, D. D., White, J. L., Ensminger, L. E. and Clark, F. E. (Editors) 1965. Methods of Soil Analysis. Part 1, Agronomy No. 9. American Society of Agronomy. Inc. Madison. 770 pp.

Bruce, R. R. and Klute, A. 1956. The measurement of soil moisture diffusivity, Proceedings-Soil Science Society of America, vol. 20, pp. 458–462.

Buckingham, E. 1907. Studies on the Movement of Soil Moisture, U.S. Department of Agriculture, Bull. No. 38, Bureau of Soils, Washington, DC, USA.

Buckman, Harry, O., Brady and Nyle C. 1960. The Nature and Property of Soils - A College Text of Edaphology (6th ed.), New York: MacMillan Publishers, New York, NY.

Darcy, H. 1856. Les Fontaines Publiques de la Ville de Dijon, Dalmont, Paris.

Donahue, Roy, Miller, Raymond and Shickluna, John. 1983. Soils: An Introduction to Soils and Plant Growth. Englewood Cliffs, NJ: Prentice-Hall, Inc.

FAO. 1974. Guidelines for Soil Profile Description. Soil Survey and Fertility Branch, Land and Water Development Division, Food and Agriculture Organisation of United Nations, Rome, 53 pp.

Fetter, C. W. 1994. Applied Hydrogeology, 3rd ed. Upper Saddle River, NJ: Prentice Hall, Inc.

Gardner, W.H. 1986. Water content. Chapter 21 in Klute, A., ed. Methods of soil analysis. Part 1. Physical and mineralogical methods. 2nd Ed. Soil Science Society of America, Inc. Madison, Wisconsin.

Hogan, M.C. 2010. Abiotic factor. Encyclopedia of Earth. eds Emily Monosson and C. Cleveland. National Council for Science and the Environment. Washington DC.

Holt, C. L. R., Jr. 1965. Geology and Water Resources of Portage County, Wisconsin. (Geologic Survey Water-Supply Paper 1796 and Plate 1). Washington, DC: US Government Printing Office.

Horton, R. E. 1942. Remarks on hydrologic terminology, Eos, Transactions, American Geophysical Union, vol. 23, pp. 479–482.

Jury, W.A. and Horton, R. 2004. Soil Physics, 6th ed, John Wiley & Sons.

Kutilek, M. 1978. Vodohospodárská pedologie. druhé vydání, SNTL Bratislava. 296 s. SNTL 04-721-78[5].

Lawes, J. B., Gilbert, J. H. and Warington, R. 1882. On the Amount and Composition of the Rain and Drainage Water Collected at Rothamsted, Williams, Clowes and Sons Ltd, London, UK.

L éger, E., Saintenoy, A. and Coquet, Y. 2013. Saturated hydraulic conductivity determined by on ground mono-offset Ground-Penetrating Radar inside a single ring infiltrometer.

Loxnachar, Thomas, E., Brown, Kirk, W., Cooper, Terence, H., Milford, Murray, H. 1999. Sustaining Our Soils and Society. American Geological Institute, Soil Science Society of America, USDA Natural Resource Conservation Service publication.

Munsell Soil Color Charts, 1973. Macbeth Colour and Photometry Division, Kollmorgen Corporation. Maryland.

Šimunek, J., Huang, K., Sejna, M. and van Genuchten, M. T. 1998. The HYDRUS-1D software package for simulating the one-dimensional movement of water, heat and multiple solutes in variable-saturated media. Version 1.0, Tech. Rep. IGWMC-TPS-70, International Ground Water Center, Colorado School of Mines, Golden, Colorado, 162p.

UNSW. 2007. Online: http://www.terragis.bees.unsw.edu.au/terraGIS_soil/sp_water-soil_and_water.html, cit. 3.4.2015.

USDA. 2015. National Leader for Soil Classification and Standards, National Soil Survey Center, Lincoln, NE. Online: http://www.nrcs.usda.gov/wps/portal/nrcs/detail/soils/ref/?cid=nrcs142p2_053573, cit. 2.4.2015.

USDA. 2013. Natural Resources Conservation Service Soil quality indicators - Bulk density. Online: http://www.nrcs.usda.gov/Internet/FSE_DOCUMENTS/nrcs142p2_053256.pdf, cit. 3.4.2015.

van Genuchten, M. T. 1980. A closed-form equation for predicting the hydraulic conductivity of unsaturated soils, Soil Science Society of America Journal, vol. 44, no. 5, pp. 892–898.

Whitaker, S. 1986. Flow in porous media I: A theoretical derivation of Darcy's law. Transport in Porous Media 1: 3–25.

# 8. APPENDICES

## 8.1 Comparison between measured and modelled bottom fluxes

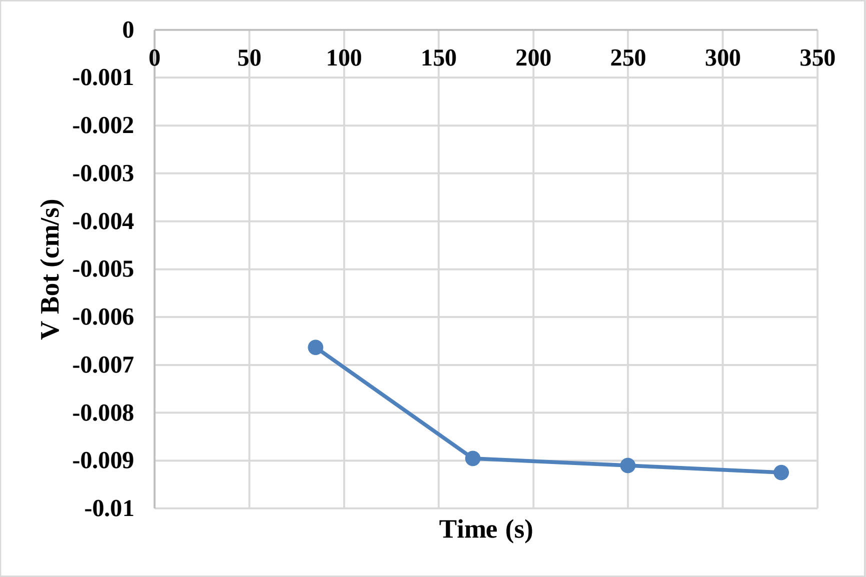

**Figure 18:** Measured Bottom flux against Time at pressure of +2cm

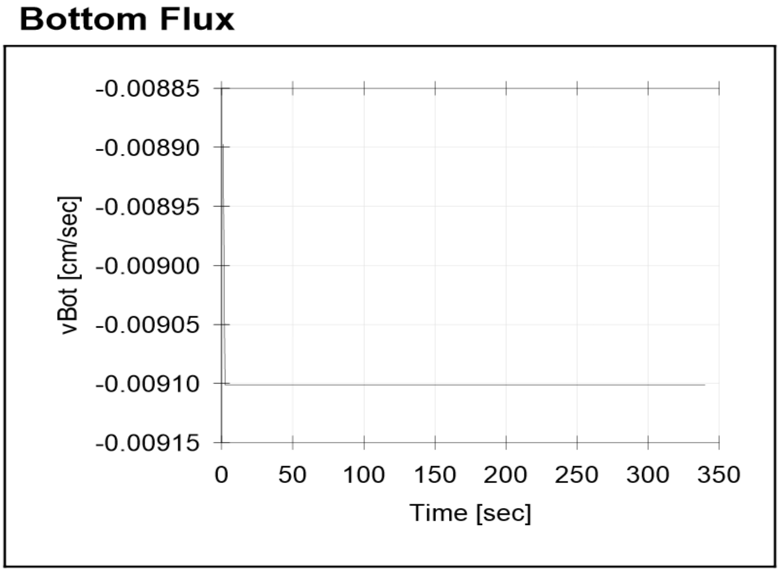

**Figure 19:** Bottom flux against Time by HYDRUS 1D model at pressure of +2cm

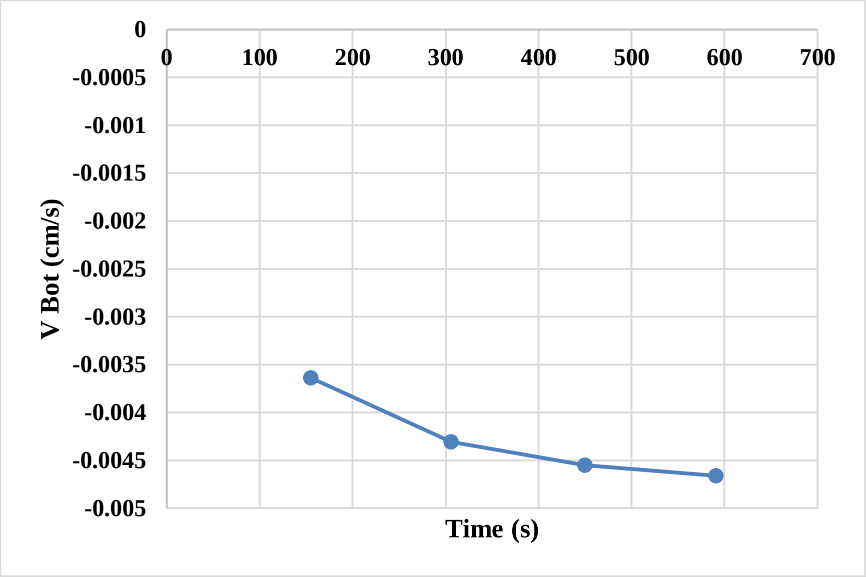

**Figure 20:** Measured Bottom flux against Time at pressure of -2cm

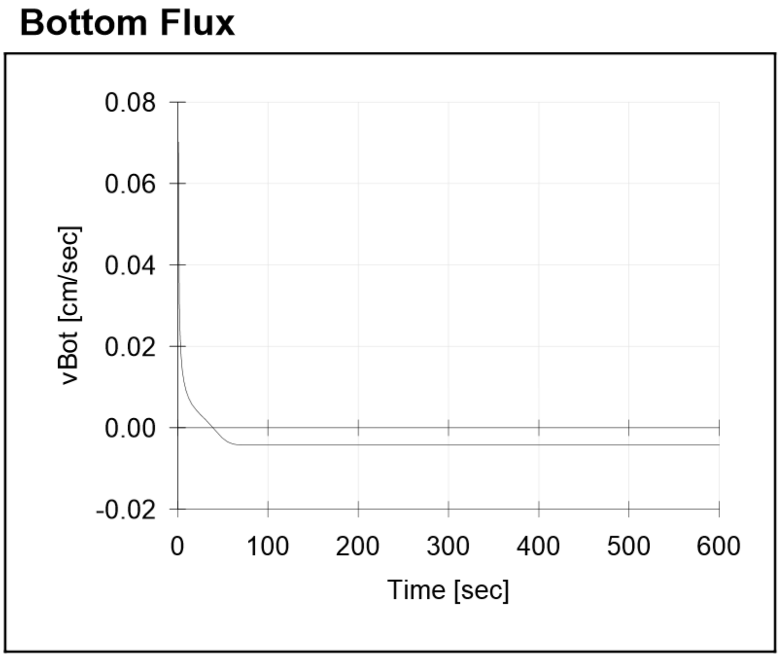

**Figure 21:** Bottom flux against Time by HYDRUS 1D model at pressure of -2cm

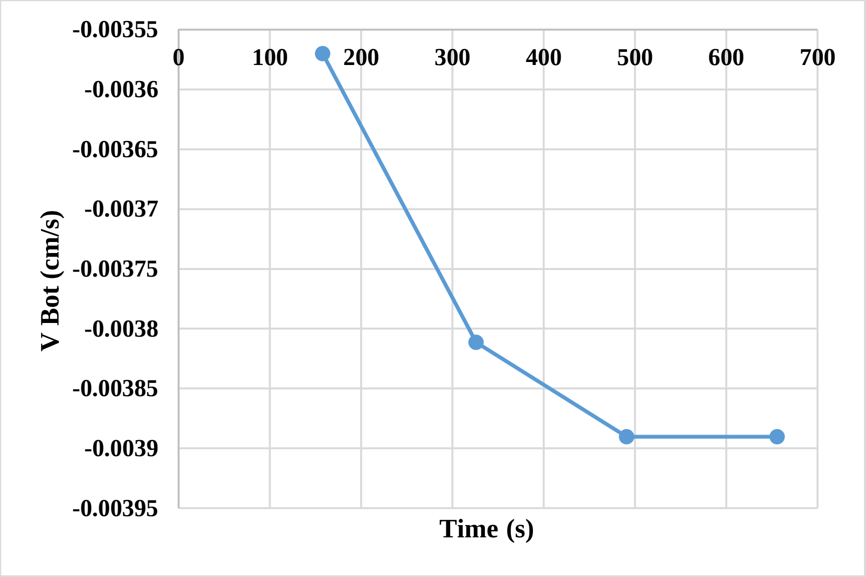

**Figure 22:** Measured Bottom flux against Time at pressure of -3cm

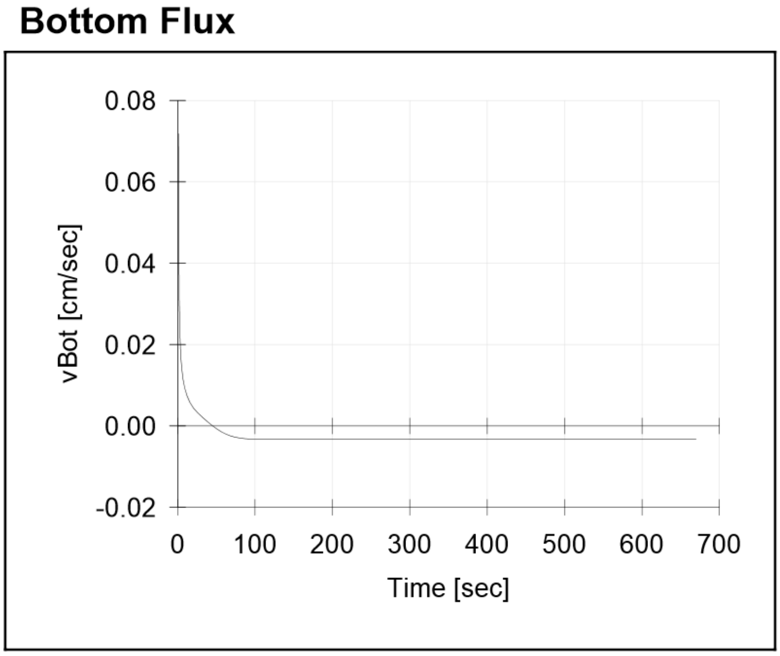

**Figure 23:** Bottom flux against Time by HYDRUS 1D model at pressure of -3cm

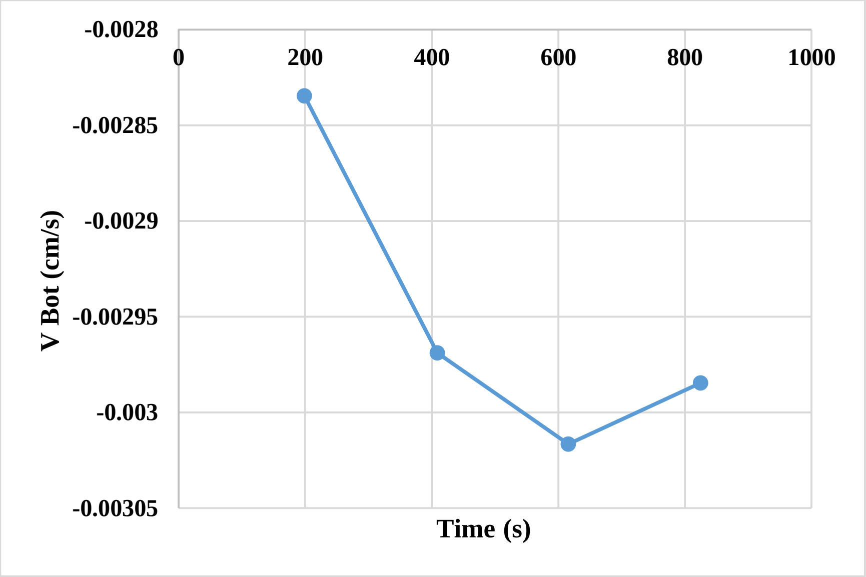

**Figure 24:** Measured Bottom flux against Time at pressure of -4cm

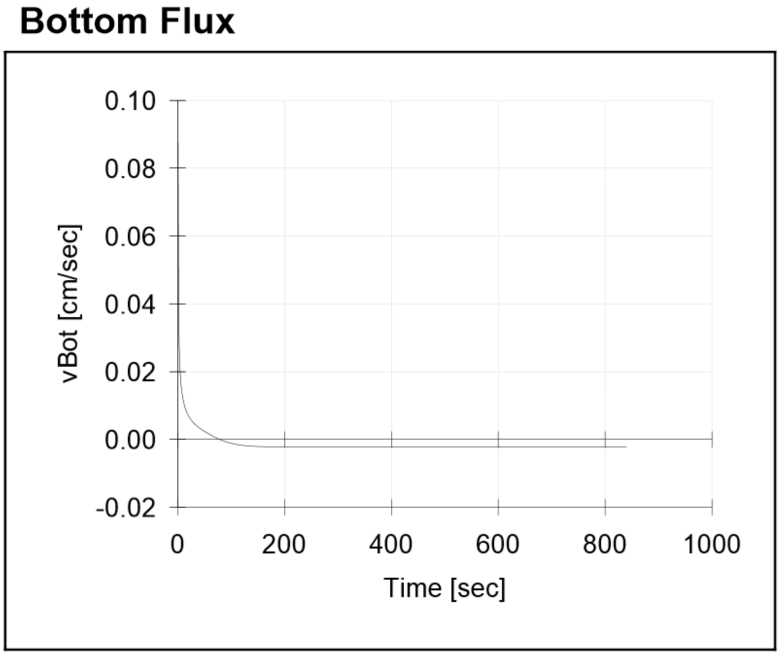

**Figure 25:** Bottom flux against Time by HYDRUS 1D model at pressure of -4cm