Soil and Plant Analyses
Vol. I Soil Physics Analyses

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Preface

Although there are several books available on the methods of soil physical analysis, this manual's primary aim is to present a simplified and quick guide to routine methods. In addition, most of the methods presented in this manual have been modified to better suit local tropical conditions such as in Malaysia. They have been modified through experience over a decade after much trial-and-error as well as discussion with other researchers. It is hoped that this manual will be useful to students (both undergraduates and postgraduates) and researchers as a quick reference for specific soil physical laboratory or field method of analysis and measurement. The references at the end of this manual are provided as additional reading for those who require more detailed information on the principles or alternative methods.

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Dr. Christopher Teh Boon Sung and Assoc. Prof. Dr. Jamal Talib
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1. Particle-size Analysis

1.1 Principles

Particle-size analysis is a measurement of the size distribution of individual particles in a soil. The particle-size distribution (*i.e.*, their percentages) in a soil, in turn, is known as the soil texture. Unfortunately, the texture of a soil is classified differently according to the many soil classification schemes that are currently available. One of the most widely known and used soil classification schemes is that developed by the USDA (United States Department of Agriculture) which is represented in Figure 1. In this scheme, soil particles and their size ranges are: clay (<2 µm), silt (2–50 µm) and sand (50-2000 µm).

Example: a soil having 55% sand, 5% silt and 40% clay (their total must be 100%) would be classified as having a *sandy clay* texture in the USDA scheme (as marked in Figure 1). Note that the pre-requisite for a soil texture classification is the percentages of any two particles.

Teh and Rashid (2003) developed and published a software for computing the USDA and other classification schemes of texture classes. This software is available for download at www.agri.upm.edu.my/~chris/tal.

The objective of particle-size analysis is to determine the distribution (by weight) of the different grain size fractions of the mineral part of the soil. Hydrogen peroxide solution is used for eliminating the soil organic matter. The destruction of calcium carbonate and disaggregation of cement (amorphous sesquioxides), which link the colloidal fraction, is accomplished by treating them with hydrochloric acid followed by washing with distilled water. The very fine fraction is separated from sand by sieving under water using a 50 µm sieve. The separation of the various sand fractions is in turn made by dry sieving using the appropriate sieve sizes.

Determination of silt and clay is carried out using a pipette after dispersion of the colloidal suspension in a dispersing reagent. The time and the depth of the sampling are deduced using Stokes' Law (Stokes, 1849). Stokes' law simply describes the balancing of downward force due to gravity and the resisting force due to buoyancy (surface friction and solution movement). With the assumption that smooth rigid spheres are settling in a quiescent (non turbulent) viscous fluid of known density and viscosity (resistance to flow). The mathematical representation for Stokes' Law or the settling equation is

$$V = \frac{D^2(\rho_p - \rho_w)g}{18\eta}$$

where V is the velocity of fall (cm s^{-1}); g is the acceleration due to gravity (cm s^{-2}), usually 980 cm s^{-2}; D is the "equivalent" diameter of particle (cm) ($D^2 = 4r^2$); ρ_p= density of particle (g cm^3) about 2.6 g cm^3; ρ_w= density of solution (g cm^3), about 1.0 g cm^3; η = viscosity of the solution. The density, gravity and viscosity can be expressed by a constant (k); thus,

$$V = kD^2$$

This means that the speed with which particles fall is proportional to the square of their diameters. Any calculation using this law is only approximately correct for soils because the particles are not spherical, are not smooth, and vary in mineral density.

A common method used for particle-size analysis is the pipet method (Gee and Bauder, 1986).

1.2 Materials

1. Weighing balance
2. Mechanical stirrer
3. Mechanical shaker
4. Hot plate
5. Oven
6. Desiccator
7. Aluminium dishes
8. Sieves 1000-, 500-, 250-, 100- and 50-µm
9. 1000 ml beaker and plunger
10. 20 ml pipet
11. Hydrogen peroxide H_2O_2 (30%), HCl (0.2 N), and distilled water
12. Calgon solution (sodium hexametaphosphate 50 g L^{-1}). The Calgon solution is adjusted to pH 8.3 by adding Na_2CO_3 (use H_3PO_4 if pH exceeds pH 8.3)

1.3 Procedure

1. Weigh 20 g of dry soil (< 2 mm) on the weighing balance; bring quantitatively in a tall beaker of 1000 ml.

2. Add 100 ml of hydrogen peroxide H_2O_2 (30%) and leave overnight.

3. Heat the beaker on a hot plate, and add further quantities of H_2O_2 until no more frothing occurs.

4. Add 100 ml 0.2 N HCl and distilled H_2O until the volume equals 400 ml.

5. Heat on the hot plate during 15 minutes and allow to cool.

6. Siphon off the supernatant liquid and wash thrice with 400 ml of distilled water.

7. Transfer the contents of the beaker to the stirrer and add 50 ml Calgon solution.

8. Stir for 15 minutes using a mechanical stirrer.

9. After stirring, transfer the contents into 1000 ml cylinder through a 50 µm sieve. The contents in the cylinder are the clay and silt particles. The particles left on the 50 µm sieve are the sand fraction (see step 17 for the sand separation).

10. Wash the mechanical stirrer blades with distilled water until no traces of the soil are left and introduce the wash water through the sieve in the cylinder.

11. Make up the contents of the cylinder to 1000 ml.

12. The cylinder is introduced in the water bath at constant 23 $^{\circ}$C.
13. The plunger is used to mix the contents thoroughly for 1 minute.
14. Leave the suspension to settle for 6 hours and 39 minutes, then introduce the pipet again at a depth of 10 cm, and pipet an aliquot.
15. Transfer the contents of the pipette into a previously weighed aluminium dish, and oven-dry at 105 ^{0}C for 24 hours. This is the weight of clay particles (A).
16. After oven-drying, the aluminium dishes should be allowed to cool in a desiccator before weighing.
17. From step 9, the sand fraction on the 50 µm sieve is oven-dried (105 $^{\circ}$C), and then transferred onto a nest of sieves arranged from top to bottom with decreasing size in the following order: 1000-, 500-, 250-, 100- and 50-µm.
18. Shake the nest of sieves for 5 minutes using a mechanical shaker.
19. Weigh each sand fraction in the nest of sieves. Total weight of all sand fractions is B.
20. As 50 ml of Calgon solution was added (step 7), there must be a correction for the weight clay (A) determined from step 15. Pipet 50 ml of Calgon into three separate dishes. Place them in oven until dry, cool in desiccator and weigh immediately. Determine the mean weight of the three replicates (C).

1.4 Calculations

From step 15, weight of clay = A g

$$\%\ \text{clay} = \left(\frac{A \times 1000}{\text{vol. of pipet}} - C \right) \times \frac{100}{\text{wgt. of soil}}$$

where C is the mean weight of Calgon in the soil solution.

From step 17, total weight of sand particles = B g.

$$\%\ \text{sand} = B \times \frac{100}{\text{wgt. of soil}}$$

Thus,

$$\%\ \text{silt} = 100 - \%\ \text{sand} - \%\ \text{clay}$$

1.5 Texture class determination

The texture of a class can be determined by looking up a texture chart (such as Figure 1) or using the texture software by Teh and Rashid (2003) (Figure 2 and 3).

2. Aggregate Size Distribution and Stability

2.1 Principles

An aggregate is a group of primary particles that cohere to each other more strongly than to other surrounding soil particles (Kemper and Rosenau, 1986). Aggregate stability, on the other hand, is the measure of aggregates resisting disruptive forces. To determine aggregate stability, known amounts of size aggregate size fraction are subjected to a disruptive force. The amount of disruption is measured by determining the portion, by weight, of the aggregates that is broken down into aggregates and primary particles smaller than a selected size.

The aggregate distribution obtained after the dry sieving is an important characteristic because, firstly, resistance to wind and water erosion is related to the size of the aggregates and secondly, it gives and idea of the distribution of the large pores after tillage. However, the relationship between aggregate distribution obtained in the laboratory and in the field is very empirical since the laboratory procedure cannot give a reliable measure of the aggregate distribution in the field. Nevertheless, the aggregate distribution after dry sieving is still useful index for conditioned sandy soils in relation to wind erosion. And because the aggregates > 2 mm are not easily blown away, their percentage is very important.

2.2 Aggregate size distribution

2.2.1 Materials

1. Sieves 8.0-, 4.76-, 2.83-, 2.0-, 1.0-, 0.5- and 0.3-mm
2. Rotary (mechanical) sieve
3. Weighing balance

2.2.2 Procedure

1. Air dry soil sample (± 1500 g) is sieved on a set of nested sieves with mesh widths in the order 8.0-, 4.76-, 2.83-, 2.0-, 1.0-, 0.5- and 0.3-mm. A container is placed below the 0.3 mm sieve to collect the fraction smaller than 0.3 mm.
2. Shake gently for about 3 minutes by hand, or better by a rotary sieve (if available), to divide the dry aggregates into the different fractions without breaking the aggregates.
3. Any aggregates remaining on the topmost sieve (8 mm) should be hand crushed to pass through that sieve.
4. Shake the nest of sieves again for about 5 minutes.
5. Weigh all the aggregates collected on each sieve, including the fraction smaller than 0.3 mm.

2.2.3 Calculations

Aggregate size distribution after dry sieving can be expressed as a single empirical unit called *Mean Weight Diameter* (MWD) which is equal to the sum of products of: 1) the mean diameter $\overline{x}_i$ of each aggregate size fraction, and 2) the proportion of the total sample weight w_i occurring in the corresponding size fraction, where the summation is carried out over all n size fractions (Van Bavel, 1950):

$$MWD = \sum_{i=1}^{n} \bar{x}_i w_i$$

Larger values of *MWD* denote a higher proportion of larger aggregates in the soil sample. The index *MWD* is developed to give greater weightings to larger aggregates than that for smaller aggregates. This reflects the fact that larger aggregates often indicate favourable soil structure for agriculture.

Aggregate size distribution can also be expressed as the percentage of aggregates that are greater than 2 mm.

Sample calculation:

Size (mm)	8.0	4.76	2.83	2.0	1.0	0.5	0.3	<0.3
Weight (g)	0.0	11.0	15.0	20.0	25.0	30.0	10.0	12.0

Total weight of all aggregates = 11.0 + 15.0 + … + 12.0 = 123.0 g

$$MWD = \frac{1}{123}\left[\left(\frac{8.0+4.76}{2}\times 11.0\right)+\left(\frac{4.76+2.83}{2}\times 15.0\right)+\ldots+\left(\frac{0.3+0}{2}\times 12.0\right)\right] = \underline{\underline{1.82}}\text{ mm}$$

$$\text{Percentage of aggregates} > 2\text{ mm} = \frac{11.0+15.0+20.0}{123}\times 100 = \underline{\underline{37.4\%}}$$

2.3 Aggregate stability

2.3.1 Materials

1. Wet-sieving machine (Eijkelkamp, Netherlands) (Figure 4)
2. Sieves 0.25 mm
3. Aluminium dishes
4. Oven
5. Weighing balance
6. Water spray bottle
7. Distilled water

2.3.2 Procedure

Aggregate stability is measured using a wet-sieving apparatus (Eijkelkamp, Netherlands), and the method of wet-sieving is adapted from Kemper and Rosenau (1986).

1. Weigh 4 g of 1- to 2-mm air-dried aggregates into the 0.25 mm sieves.
2. Pre-moisten the aggregates with water vapour using a humidifier or a very fine plant sprayer until the aggregates are saturated.
3. Place the sieves in the sieve holder.
4. Place the sieving cans / bins and add distilled water until three-quarter full.
5. Wet-sieve for 10 minutes.
6. Transfer the remaining aggregates on each sieve into an aluminium dish and oven dry the aggregates at 105 °C for 24 hours.
7. Weigh the oven dry aggregates (*W*).
8. Transfer the aggregates back onto the 0.25 mm sieves. Under gentle running tap water, gently rub the aggregates against the wire mesh of each sieve until all that remains are the sand particles that cannot pass through the sieve.

9. Transfer the sand particles (>0.25 mm) into an aluminium dish and oven dry at 105 °C for 24 hours.
10. Weigh the sand particles (*S*).

2.3.3 Calculations

Aggregate stability is expressed as the percentage of aggregates remaining larger than 0.25 mm. Correction for sand particles larger than 0.25 mm must be done as these particles cannot be broken down further to pass through the sieve.

$$AS\% = \frac{W - S}{4 - S} \times 100$$

Note: larger values indicate greater aggregate stability.

3. Bulk Density, Water Content and Porosity

3.1 Principles

Bulk density is the ratio of the mass of dry solids to the bulk volume of the soil. This bulk volume includes the volume of the solids and pore space. A common method to measure bulk density is the core method, where undisturbed soil samples are taken in a container of a constant and known volume (Blake and Hartge, 1986). With this method, a core ring (*i.e.*, a cylindrical metal ring) is pressed or hammered into the soil to the desired depth and then carefully removed to preserve a known volume of sample as it existed in the field. The sample is then dried to 105 °C and weighed to determine the field moisture content. The bulk density (BD) can be calculated with the formula:

$$BD = \frac{\text{weight of soil, dried at 105 °C}}{\text{volume of soil}}$$

BD is often expressed as g cm^{-3} or Mg m^{-3} which are both numerically equivalent.

Bulk density of a soil is not constant but varies with soil structure, and it is for this reason that bulk density is often a measure of soil structure. The greater the value of bulk density, the greater the soil compaction.

Usually the measurement of BD allows two other properties of the soil to be measured as well. There are the soil water content at the moment when the soil was sampled and the total porosity.

3.2 Materials

1. Core ring (*i.e.*, a cylindrical metal ring) with plastic caps
2. Hammer
3. Oven
4. Weighing balance

3.3 Procedure

1. Push or hammer a core ring (measuring about 7.6- and 4.0-cm in diameter and depth, respectively, and having a known weight) into the soil to the desired soil depth (Figure 5). Note that the core ring must be pushed into the soil at least as deep as the core ring height or depth.
2. Carefully remove the core ring from the soil and trim excess soil to the level of both ends of the core ring. Clean the outside of the ring of any remaining soil.
3. Close both ends of the core ring with the plastic caps and bring back to the laboratory for analysis.
4. Weigh the soil with the core ring (without the plastic caps), then oven-dry at 105 °C for at least 24 hours. Weigh the soil with the core ring again.

3.4 Calculations

1. Bulk density ρ_b (g cm^{-3}):

$$\rho_b = \frac{W_b - W_r}{\pi h d^2 / 4}$$

where W_b is the weight of the soil and core ring after oven-drying (g); W_r is the weight of the core ring (g); h is the core ring height or depth (cm); and d is the core ring diameter (cm).

2. Soil water content θ_w, expressed on a dry mass basis:

$$\theta_w = \frac{W_a - W_b}{W_b - W_r}$$

where W_a is the weight of the soil and core ring before oven-drying (g).

3. Soil water content θ_v, expressed on a volume basis:

$$\theta_v = \theta_w \rho_b$$

4. Soil porosity e:

$$e = 1 - \frac{\rho_b}{\rho_p}$$

where ρ_p is the particle density (g cm^{-3}). For most mineral soils, particle density is 2.65 g cm^{-3}.

4. Water Retention

4.1 Principles

Soil water retention is the relation between the soil water content and the soil water suction, and it is related to the hydraulic properties of the soil. Soil water retention is often also called as the soil moisture characteristic or the capillary pressure-saturation curve.

Establishing a soil water retention curve involves establishing a series of equilibria between water in the soil sample and a body of water at known potential. At each equilibrium point, the volumetric water content, θ, of the soil is determined and paired with a value of the matric pressure head, h_m, determined from the pressure in the body of water and the gas phase pressure in the soil. Thus, the data pair (θ, h_m) is one point on the retention curve. As the suction is further increased, more water is drawn out of the soil, and more of the relatively large pores, which cannot retain water against the suction applied, will empty out first. In short, the greater the applied suction (higher h_m), the drier the soil (lower θ). A series of plotted (θ, h_m) points then gives the soil water retention curve.

The relation between matric head and soil wetness is not generally a unique and single-valued one. This can be obtained in two ways: 1) in desorption, by taking an initially-saturated sample and applying increasing suction to gradually dry the soil while taking successive measurements of water content versus suction, and 2) in sorption, by gradually wetting up an initially dry soil sample while reducing the suction. Each of these two methods yields a continuous curve, but the two curves will in general not be identical. The equilibrium soil wetness at a given suction is greater in desorption (drying) than in sorption (wetting) – and this effect is called hysterisis.

In this manual, the pressure plate and pressure membrane method by Richards (1947) will be presented.

4.2 Materials

1. Pressure chambers (100, 500 and 1500 kPa) (Figure 6), pressure control manifold, air compressor, and ceramic plates (100, 500 and 1500 kPa plates)
2. Soil sample retaining rings
3. Core ring (i.e., a cylindrical metal ring) with plastic caps
4. Hammer
5. Oven
6. Weighing balance

4.3 Procedure

The following description is the procedure to determine the soil water content at five suction levels: 0, 1, 10, 33 and 1500 kPa.

1. Push or hammer a core ring (measuring about 7.6- and 4.0-cm in diameter and depth, respectively, and having a known weight) into the soil to the desired soil depth (Figure 5). Note that the core ring must be pushed into the soil at least as deep as the core ring height or depth.
2. Carefully remove the core ring from the soil and trim excess soil to the level of both ends of the core ring. Clean the outside of the ring of any remaining soil.

3. Close both ends of the core ring with the plastic caps and bring back to the laboratory for analysis.
4. Place the sample retaining rings on each of the 4 porous plates for 1, 10, 33 and 1500 kPa pressures.
5. Break up the core (undisturbed) samples into 5 pieces of roughly equal sizes. Place one piece of sample in the retaining ring on each of the porous plates (see step 4). For the 0 bar pressure (saturation) the sample is placed in a retaining ring on a coarse wire mesh.
6. Saturate all ceramic plates for 24 hours by keeping the water level just below the edge of the ring (Figure 7).
7. Place the plate with samples inside the corresponding pressure chamber. Connect the plate to the outflow tube. Close the chamber and apply pressure (note: for 1, 10, 33 kPa pressure use the 100 kPa ceramic plate, for the 1500 kPa pressure use the 1500 kPa plate).
8. Equilibrium is attained when no more outflow occurs. A period of 4 to 7 days is usually sufficient to achieve this for most soils. Release the air pressure in the chamber.
9. Open the chamber, remove the samples, and weigh each of them (W_a).
10. Oven-dry the samples at 105 °C for 24 hours, and weigh each of the soil samples again (W_b).
11. Plot a log-scaled chart with the volumetric water content on the y-axis and the soil suction on the x-axis.

4.4 Calculations

	Suction (kPa)				
Weight (g)	0.1	1	10	33	1500
W_a	15.2	16.2	14.2	15.9	12.9
W_b	12.1	13.1	11.9	14.0	11.9
Water (W_a-W_b)	3.1	3.1	2.3	1.9	1.0
Volumetric water content (m^3 m^{-3}), $\theta_v = \frac{W_a - W_b}{W_b} \rho_b$	0.38	0.35	0.29	0.20	0.13

Note that soil suction at saturation is taken as 0.1 kPa (instead of 0 kPa) so that a log-scaled chart can later be plotted (Figure 8). ρ_b is the soil bulk density. In this example, bulk density is taken as 1.5 g cm^{-3}. Unit conversion: 1 kPa = 0.1 m H_2O = 0.01 bar = 0.01 atm.

Figure 8 shows that with increasing soil suction, the soil becomes progressively drier in a non-linear manner. Soil moisture retention curves typically show a “S-mirror image” shape. In this example, the soil moisture content at saturation, field capacity and permenant wilting point are 0.38, 0.20 and 0.13 m^3 m^{-3}, respectively. The available water content for plant use is 0.20 – 0.13 = 0.07 m^3 m^{-3}.

5. Water Infiltration

5.1 Principles

Water infiltration is the downward movement of water into the soil when rain or irrigation water is on the soil surface. Infiltration is rapid for soils having large, continuous pores, but is slow for compact soils or soils that have their pore sizes reduced due to aggregate breakdown or pore clogging by lodged particles. The factors that control the rate of water movement into the soil include:

1. Percent of sand, silt, and clay. Coarse sands permit rapid infiltration.
2. Soil structure. Fine-textured soil with large water-stable aggregates (granular structure) have higher infiltration rates than massive (structureless) soils.
3. Amount of organic matter in the soil. The greater the amount of organic matter and the coarser it is, the more water enters the soil.
4. Depth of the soil to hardpan, bedrock, or other impervious layers. Shallow soils do not permit as much water to enter as do deep soils.
5. Amount of water in the soil. Wet soils do not have as high an infiltration rate as do dry soils.
6. Soil temperature. Warm soils take in water faster than cold soils.
7. Compaction. This usually reduces pore space which slows infiltration.

Water infiltration rate into the soil is often measured in the field using a double ring (or cylinder) infiltrometer, where the rings are metal cylinders that are pushed or driven into the soil (Bouwer, 1986). The area inside the cylinder is flooded with water and the rate at which water moves into the ground is measured. The flooding and the measurements normally are continued until the infiltration rate has become constant over time.

The purpose of using two rings instead of one is that the water which flows from the outer ring acts to prevent lateral flow of the water from the inner ring. Consequently, water infiltration being measured from the inner ring is that for the vertical (downward) flow only.

In this manual, the double ring infiltrometer (Bouwer, 1986) will be presented.

5.2 Materials

1. Two metal rings. The diameter of the inner (smaller) ring is 30 cm, that of outer ring is 60 cm and both are 24 cm high. The rings are made of galvanized iron sheet (Figure 9)
2. Ten-pound hammer
3. Plastic sheet
4. Water level indicator
5. Stopwatch
6. Water

5.3 Procedure

1. Both rings are driven into the soil 8-10 cm deep with a ten-pound hammer. The inner ring is driven first followed by the outer ring; both rings should be in concentric position. Care is taken to cause minimum disturbance to the soil.

2. The rings must be allowed to remain in place for an extended period (1-2 weeks) prior to measurement so as to allow for resettling of the soil, especially around the ring edges.

3. Before each run, place a piece of plastic sheet within the inner ring as an anti puddling device. The buffer area (*i.e.*, area between the two concentric rings) is first filled with water to a depth of about 10 cm and maintained approximately so throughout the observation period.

4. Immediately after adding water to the buffer area, introduce water into the inner ring to a certain desired depth, say, 2 cm from the top edge of the ring. Remove the plastic cover quickly, start stopwatch and measure the water surface elevation by a water level indicator or gauge.

5. Take the water level measurement at regular intervals: every minute for the first 15 minutes, then every 5 minutes thereafter. Tabulate the results as shown in Table 1.

Table 1. Measurement of water infiltration using the double ring infiltrometer method

Time, T (min)	Water level in inner ring, H (cm)	Time change, ΔT (min)	Height change, ΔH (cm)	Total infiltration, Q (cm)	Infiltration rate=$\frac{\Delta H}{\Delta T}$, I (cm/min)
T_1	H_1				
		T_2-T_1	H_2-H_1		
T_2	H_2				
		T_3-T_2	H_3-H_2		
T_3	H_3				
.	.	.	.		
.	.	.	.		
.	.	.	.		
T_{n-1}	H_{n-1}				
		T_n-T_{n-1}	H_{n-1}-H_n		
T_n	H_n				

6. When the water level in the inner ring has dropped to about 7 cm, add sufficient water to return the water approximately to its original level. Record the depth of water added for the necessary corrections to H and Q. Record also the time just before and just after

filling. The intervals between these two readings are kept as short as possible to avoid errors caused by water intake during the refilling period. Maintain the water level in the outer ring as close to that in the inner ring by adding water to it from time to time.

7. Continue the run until a steady state infiltration is established or up to two and half hours if the steady state has not yet been reached.

5.4 Calculations

Table 2 shows sample data collected using the double ring infiltrometer. It is often useful to plot charts showing the changes in total infiltration and infiltration rate with time as well (Figure 10).

Table 2. Tabulated sample data of water infiltration

Time, T (min)	Water level in inner ring, H (cm)	Amount of water added (cm)	Time change, ΔT (min)	Height change, ΔH (cm)	Total infiltration, Q (cm)	Infiltration rate= $\frac{\Delta H}{\Delta T}$, I (cm/min)
1	20.0					
2	17.1		1	2.9	2.9	2.9
3	14.3		1	2.8	5.7	2.8
5	8.9	20.0	2	5.4	11.1	2.7
10	16.6		5	12.3	23.4	2.5
15	5.5	20.0	5	11.1	34.5	2.2
20	15.4		5	10.1	44.6	2.0
25	6.3	20.0	5	9.1	53.7	1.8
30	18.1		5	8.2	61.9	1.6
35	10.6	20.0	5	7.4	69.4	1.5
40	23.9		5	6.7	76.1	1.3
45	17.8		5	6.1	82.2	1.2
50	12.3		5	5.5	87.7	1.1
55	7.3		5	5.0	92.7	1.0
60	2.8		5	4.5	97.2	0.9

6. Saturated Hydraulic Conductivity

6.1 Principles

Soil hydraulic conductivity is a measure of a soil's ability to transmit water. Water flows through a porous material (such as soil) as a result of different forces, principally the pressure gradients, gravitation, adsorptive and osmotic forces. The flow can be described by Darcy's law (Darcy, 1856) in the form of

$$V = -K.i \qquad \text{or} \qquad Q = -K.i.A$$

where V is the velocity of flow or flux; K is the hydraulic conductivity; i is the hydraulic or potential gradient; Q is the quantity of flow per unit time or the discharge; and A is the cross-sectional area of the soil, perpendicular to the direction of flow. The negative sign indicates that the direction of flow is against the potential gradient, *i.e.*, high to low potential. For a vertical flow, the driving force i is expressed as the *negative* gradient of the hydraulic head H composed of gravitational head z and the pressure head h so that:

$$H = h + z$$

The hydraulic conductivity depends on the properties of fluid in soil and the size and distribution of pores and is thus a measure of soil structure. The two fluid properties are viscosity and density, and the soil properties are texture and structure. In soils with much clay, for example, the composition of soil solution can significantly affect hydraulic conductivity because of the interactions between the soil solution and soil matrix.

In calculations of saturated flow, K is generally assumed to be constant although in reality it decreases with time as due to the gradual deterioration in soil structure. As water content decreases from saturation, the large pores, which are the most effective to conduct water, are first to drain. This in turn increases path flow tortuosity of water and the average properties of the soil solution, such as density and viscosity, may change. All these ultimately reduce hydraulic conductivity with decreasing water content.

When the hydraulic gradient is expressed in cm water height per cm height of the soil, then the hydraulic conductivity has the dimension of velocity (cm s^{-1}). O'Neal (1949) proposed a practical system of classifying the hydraulic conductivity of soils into seven classes (Table 3).

Table 3. Soil hydraulic conductivity classes (O'Neal, 1949)

Class	Hydraulic Conductivity K (cm h^{-1})
Very slow	< 0.125
Slow	0.125 – 0.500
Moderately slow	0.500 – 2.00
Moderate	2.00 – 6.25
Moderate rapid	6.25 - 12.50
Rapid	12.50 – 25.00
Very rapid	> 25.00

Laboratory methods for measuring hydraulic conductivity is often done using undisturbed core samples that are taken from the field. A fixed head of water is supplied and the flux after saturation is measured. The hydraulic conductivity is the difference in head at the inlet and outlet faces of the core divided by the length of the core.

In this manual, saturated hydraulic conductivity is measured using the constant head method (Klute and Dirksen, 1986).

6.2 Materials

1. Glass column (20- and 4-cm in height and diameter, respectively)
2. Nylon cloth
3. Filter paper
4. Mariotte bottle
5. Measuring cylinder
6. Funnel
7. Rubber tubes
8. Weighing balance
9. Water
10. Stop watch

6.3 Procedure

1. Set up the apparatus as shown in Figure 11 and 12.
2. Measure the internal diameter of the glass tube and calculate the cross section area of the tube (Figure 13).
3. Tie a piece of nylon cloth at the bottom end of the tube.
4. Weigh the tube and fill it with air-dry soil (0.3 – 1.00 mm fraction) leaving at least the top 10 cm empty. To do this add the soil slowly and gradually via a funnel at the same time knocking and turning the tube around in order to obtain a uniform packing. Measure the height of the soil column and reweigh the tube and packed soil.
5. Saturate the soil in the tube through capillary rise by immersing the bottom in the water up to a depth of several mm. Leave the soil column for 24 hours.
6. Fix the tube on a clamp as shown in Figure 11 and 12. Cut the filter paper to fit the glass tube and place the paper on the surface of the soil column.
7. Collect the discharge consisting of a funnel and graduated cylinder and then allow the water from a Mariotte bottle to flow to the top of the soil column. Maintain a constant pressure head of 2 cm above the column. Start measuring the quantity of flow per unit time (Q) as soon the first drop falls into the cylinder and continue the measurement until the flow rate remains essentially constant for at least 10 minutes. To check this, record the volume of water collected every 5 minutes.
8. Now empty the measuring cylinder and repeat the measurements with successively increasing pressure heads of 4.0-, 6.0-, 8.0- and 10.0-cm above the soil column.

6.4 Calculations

Radius of glass tube = r cm

Cross section area of soil column = πr^2

Height of soil column = z cm

Weight of soil = W g

Bulk density of soil column = $W/\pi z r^2$ g cm^{-3}

Table 4. Calculation of saturated hydraulic conductivity

Pressure head, h (cm)	Mean volume of water collected every t minutes (ml)	Q (ml/min)	H/z	Saturated hydraulic conductivity, K (cm/min)
h_1	v	$\frac{v}{t}$	$\frac{z+h_1}{z}$	$\frac{Qz}{AH}$

6.5 Note

The above method requires "repacking" the soil whereby the soil sample is packed into the glass tube. If an undisturbed soil is required, the soil could be sampled using a core ring. An empty core ring is then placed above the sampled core ring and taped (Figure 14). These two core rings could then replace the glass tube, where the height of soil column is the height of the bottom core ring.

7. Unsaturated Hydraulic Conductivity

7.1 Principles

The water content of well-drained field soil approaches saturation only infrequently. Most of the processes involving soil-water flow in the field, and in the rooting zone of most plant habitats, occur while the soil is in an unsaturated condition.

Experiments have shown that under such conditions the velocity of flow is still proportional to the potential gradient but that the proportionality “constant” is a rapid changing function of the matric potential of the unsaturated soil.

$$V = -K(\theta)i$$

where V is the velocity of water flow; $K(\theta)$ is the unsaturated hydraulic conductivity; and i is the hydraulic potential gradient. The above Darcy equation is extended for use in unsaturated soils by assuming that hydraulic conductivity is a function of the degree of saturation of the soil. Characterisation of the water-conducting properties of partially saturated media requires measurement of the conductivity-water content relationship $K(\theta)$, or the conductivity-capillary pressure head function.

Unsaturated flow processes are in general complicated and difficult to describe quantitatively, since they often entail changes in the state and content of soil water during flow.

Consider an unsaturated soil in which water is flowing under a potential gradient (Figure 15). The potential difference between the inflow and outflow ends is maintained by different imposed suctions. In general, as the suction varies along the sample, so will the wetness and the conductivity. If the suction heads at both ends of the sample are maintained constant, the flow process will be steady.

Since the gradient along the column is not constant, as it is in uniform saturated systems, it is often not possible to divide the flux by the overall ratio of the head drop to the distance to obtain the conductivity $K(\theta)$. Rather it is necessary to divide the flux by the exact gradient at each point to evaluate the exact conductivity, and its variation with suction.

In the following case, however, it is assumed that the soil column is sufficiently short to allow one to evaluate at least an average conductivity for the sample as a whole. The $K(\theta)$ can be calculated from Darcy’s law

$$K(\theta) = \frac{Q}{At}\frac{\Delta x}{\Delta H}$$

where Q is the volume of water flowing through the column in time t; A is the cross sectional area of the soil column; Δx is the length of the soil column; and ΔH is the hydraulic head drop. The average negative head or suction, acting in the soil column is

$$H = \frac{H_1 + H_2}{2}$$

In this manual, unsaturated hydraulic conductivity is measured using the steady state method adapted from Klute and Dirksen (1986).

7.2 Materials

1. Glass column (20- and 4-cm in height and diameter, respectively)

2. Porous plates
3. Mariotte bottle
4. Measuring cylinder
5. Rubber tubes and adhesive tape
6. Weighing balance
7. Water
8. Stop watch

7.3 Procedure

1. Set up the apparatus as shown in Figure 15 and 16.
2. The end chamber A with porous plate is fixed to the cylindrical column C by means of an adhesive tape (Figure 17).
3. Measure the diameter of the tube.
4. Fill the tube with soil (1.0-0.3 mm size fraction). The tube is gently knocked to obtain a homogeneous bulk density.
5. Connect the soil column to the end chamber B with the porous plate by means of an adhesive tape.
6. Saturate the soil sample with distilled water and remove the trapped air by means of the air trap valve. The enclosed air can escape through the small openings in the horizontal tube.
7. After the saturation, introduce water tensions of 10 cm (H_1) and 20 cm (H_2) at the two ends respectively.
8. Measure the volume of water flowing through the column at fixed times, until the volume is constant.
9. Repeat this procedure for H_1 = 45-, 95-, 145-cm and the corresponding H_2 = 55-, 105- and 155-cm of water tension. In that way $K(\theta)$ can be calculated for several average water tensions acting in the soil column. It is then possible to determine the ratio between the flux (Q/A.t) and the potential gradient $\Delta H / \Delta x$. The slope of the flux versus gradient line, being the hydraulic conductivity, varies with the average suction (H)

7.4 Calculations

Radius of glass tube = r cm

Cross section area of soil column = πr^2

Length of soil column = x cm

Weight of soil = W g

Bulk density of soil column = $W/\pi x r^2$ g cm^{-3}

Table 5. Calculation of unsaturated hydraulic conductivity

H_1 (cm)	H_2 (cm)	$\overline{H}$ (cm)	Mean volume of	Q (ml/min)	H/x	Hydraulic conductivity

			water collected every t minutes (ml)			$K(\theta)$ (cm/min)
h_1	h_2	$\frac{h_1 + h_2}{2}$	v	$\frac{v}{t}$	$\frac{\overline{H}}{x}$	$\frac{Qx}{A\overline{H}}$

7.5 Note

The above method requires "repacking" the soil whereby the soil sample is packed into the glass tube. If an undisturbed soil is required, two core rings could be used in place of the glass tube, as shown in Figure 14.

8. Soil Loss

8.1 Principles

In agriculture, top soil is a primary resource because this layer of soil contains the highest amount of nutrients and organic matter. Top soil, however, is vulnerable to loss by erosion either due to wind or water. Factors that affect soil loss by water are soil texture, organic matter level, soil compaction, length and slope of the field, and management practices. Generally, soil erosion is enhanced when a soil is compacted, has low organic matter, and a disproportion ratio of clay to sand (*i.e.*, too much of clay or sand). Water erosion is also enhanced in hilly areas, in particular when the length and gradient of the hill slope are high. Besides soil characteristics, the climatic factors in the area also affect soil loss. Strong winds and intense rainfall increase the risk of high soil loss.

Simulation of natural storms, however, is difficult because the complete rainfall characteristics of a given region is often known insufficiently. Even if it was known in great detail, it is difficult to find or develop equipment to reproduce all these characteristics. Consequently, researchers often have to decide the characteristics of the rainfall most important, given the nature and importance of a research project.

Most rainfall simulators have established several criteria of rainfall to be important with respect to estimating soil loss:

1. Drop size distribution similar to that of natural rainfall
2. Drop velocity at impact near terminal velocity
3. Rainfall intensity corresponding to natural conditions
4. Uniform rainfall and random drop-size distribution
5. Total energy approaching that of natural rainfall
6. Reproductive storm patterns.

Other criteria deemed important are: a) research area must be of sufficient size to satisfactorily represent treatments and conditions to be evaluated, b) raindrop application nearly continuous, and c) angle of impact must not be significantly different from vertical crops.

In this manual, the method by Kamphorst (1987) is adapted. The rainfall simulator (Figure 18) essentially consists of three parts namely:

1. a sprinkler with a built in pressure regulator, based on Mariotte bottle principle, for the production of standard rain shower;
2. a support for sprinkler, which also functions as a wind shield in the field; and
3. a stainless steel frame, which is hammered into the soil and is meant to prevent lateral flow of water from the test plot to the surrounding soil.

8.2 Materials

1. Rainfall simulator (Eijkelkamp, Netherlands) (Figure 19)
2. Bucket to collect soil runoff
3. Hammer
4. Stop watch

5. Water

8.3 Procedure

1. Setup the system as shown in Figure 18 and 19.
2. Adjust the sprinkler system so that it produces an 18-mm rain shower in three minutes through 49 capillaries, giving a rain intensity equivalent to 36 cm hr^{-1}.
3. Ensure that the raindrops fall from an average height of 0.40 m onto 0.0625 m^2 surface area of the test plot having 20% slope. The amount of soil loss is dependent on kinetic energy of the falling raindrops. Since the average height of fall is 0.40 m (too small for raindrops to attain terminal velocity) and the plot size is small, a high intensity shower is used (*i.e.*, 36 cm hr^{-1}). With this intensity and height, the rainfall simulator will produce the total kinetic energy of 35.4 J mm^{-1} that is adequate to produce measurable erosion loss within a short duration.
4. Turn on the sprinkler system.
5. Collect runoff and soil sediment in sample bucket or bottle placed in a small trench dug at the lower end of test plot. After three minutes, the shower is stopped and sediment left behind in the gutter is added to the content of the sample bottle with the aid of a wiper.
6. The samples are then taken to the laboratory where the volume of runoff is measured using a measuring cylinder. Soil sediment is determined by evaporating the collected runoff to dryness using a steam bath.
7. Cool the dried samples in a desiccator, and weighed.

8.4 Calculations

Runoff volume	= A cm^{-3}
Dried sediment weight	= B (g / 100 cm^3)
Soil loss	$= C = \frac{A \times B}{100}$ g m^{-2}

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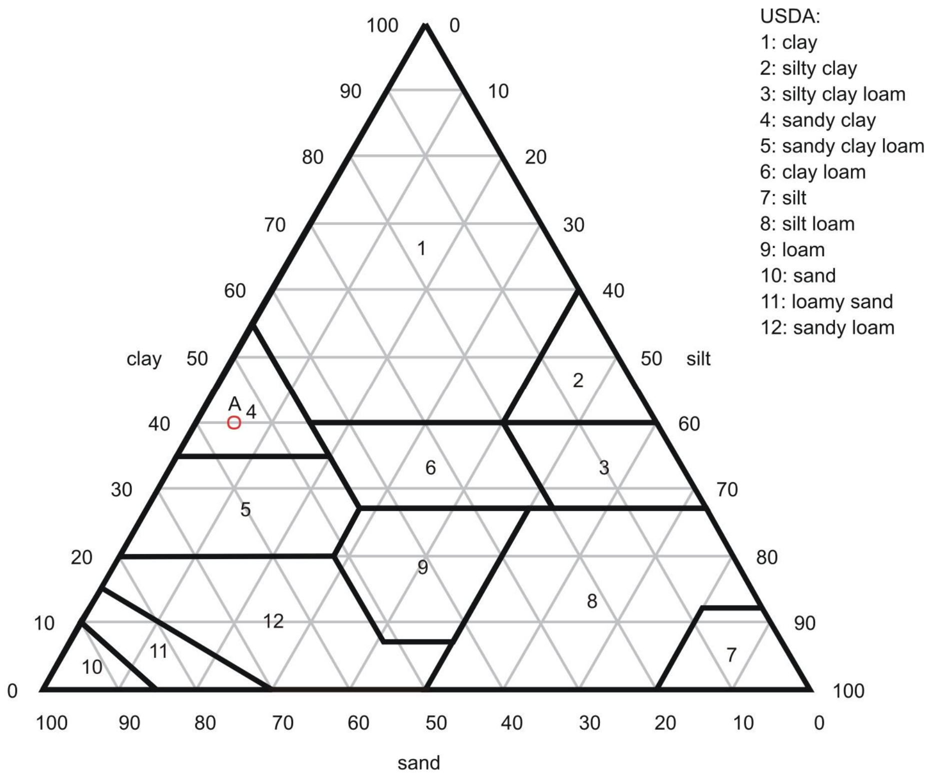
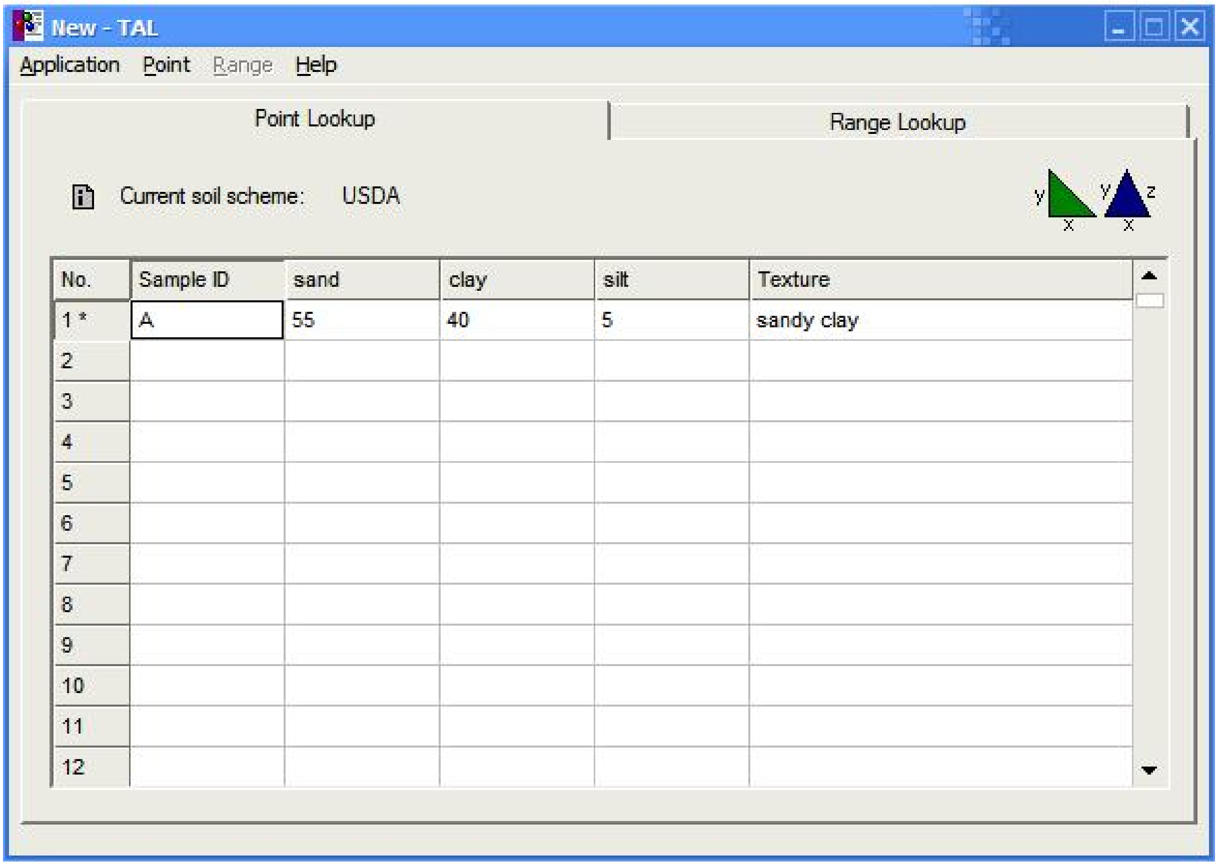


Figure 1. The USDA soil classification scheme, where the plotted point (A) indicates a soil with 55, 5 and 40% sand, silt and clay, respectively (sandy clay texture class)



Current soil scheme: USDA

No.	Sample ID	sand	clay	silt	Texture
1 *	A	55	40	5	sandy clay
2					
3					
4					
5					
6					
7					
8					
9					
10					
11					
12					

Figure 2. Data entry of the soil texture software by Teh and Rashid (2003)

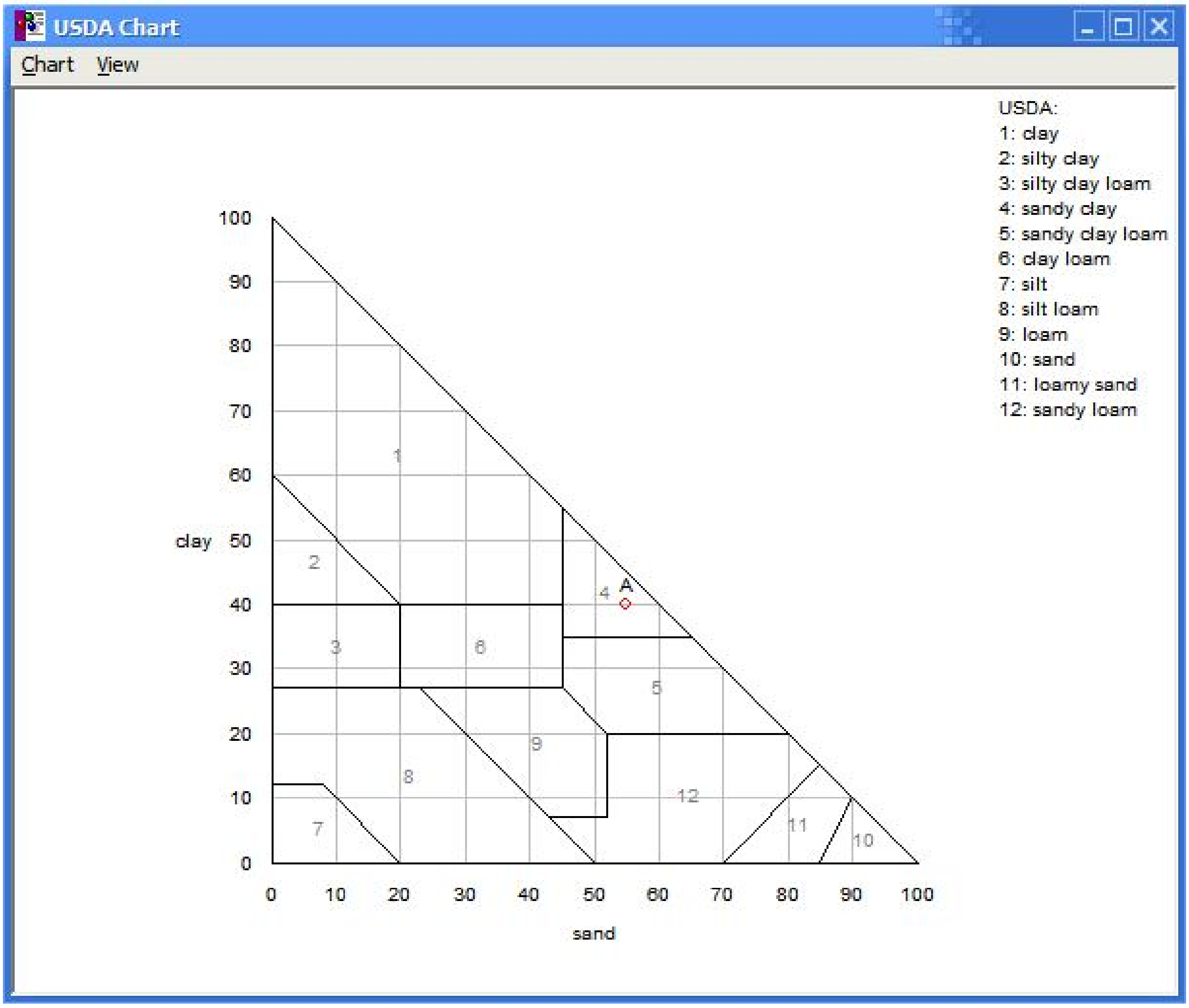


Figure 3. Data plotted on a biaxial texture chart by the soil texture software by Teh and Rashid (2003)

Figure 4. Wet sieving apparatus to determine aggregate stability

Figure 5. The core ring is hammered evenly into the soil

Figure 6. A pressure chamber to dry out a soil at a fixed constant suction

Figure 7. Soil samples and the ceramic plate are saturated

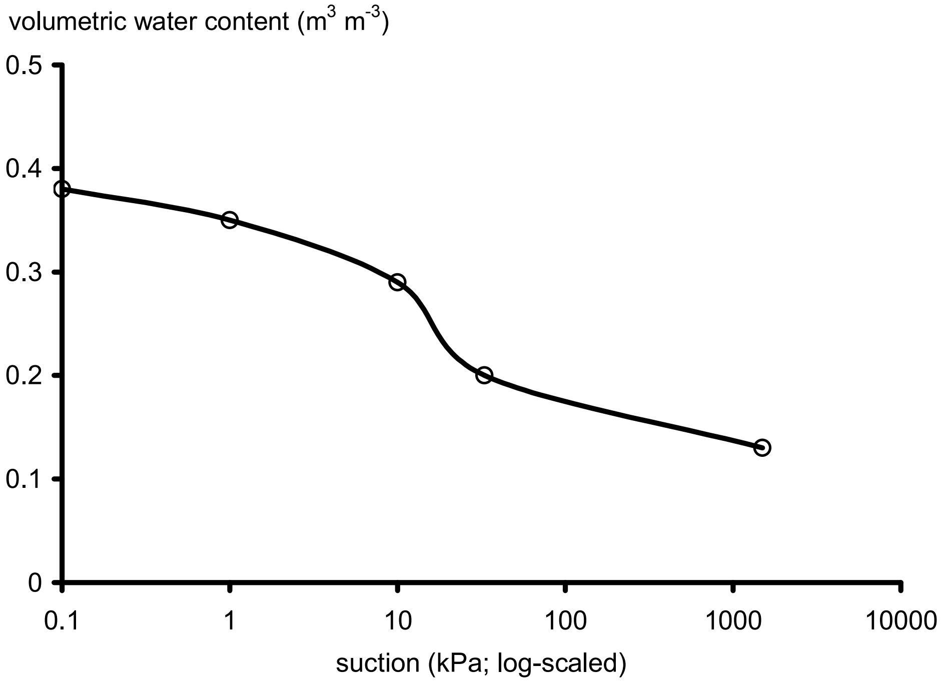


Figure 8. Soil water retention curve

Figure 9. Double ring infiltrometer

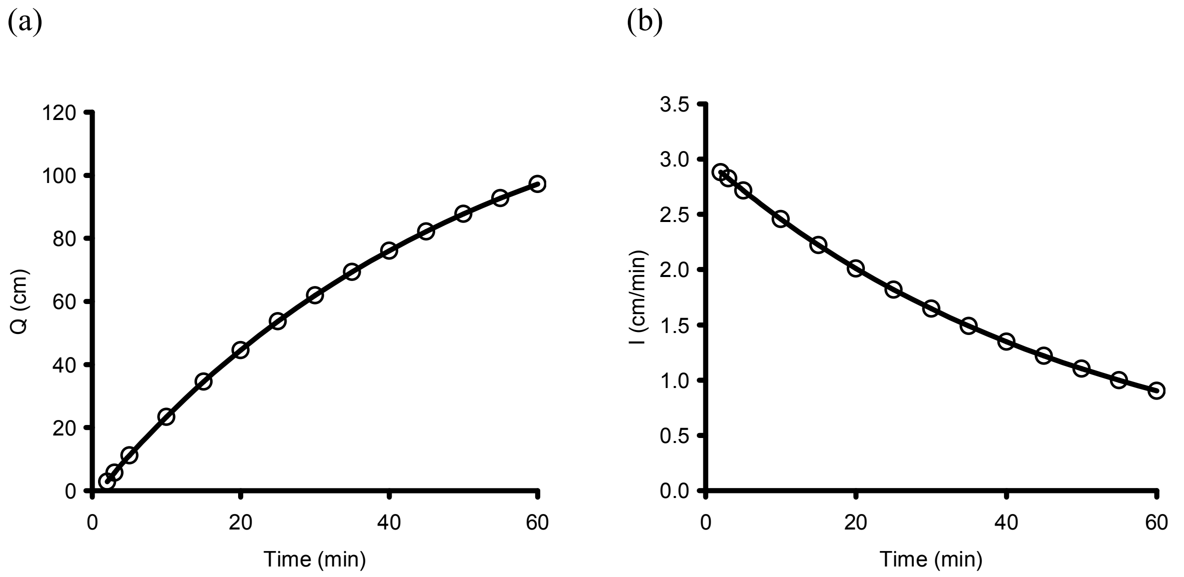


Figure 10. The change in: (a) total infiltration (Q) with time, and (b) infiltration rate (I) with time

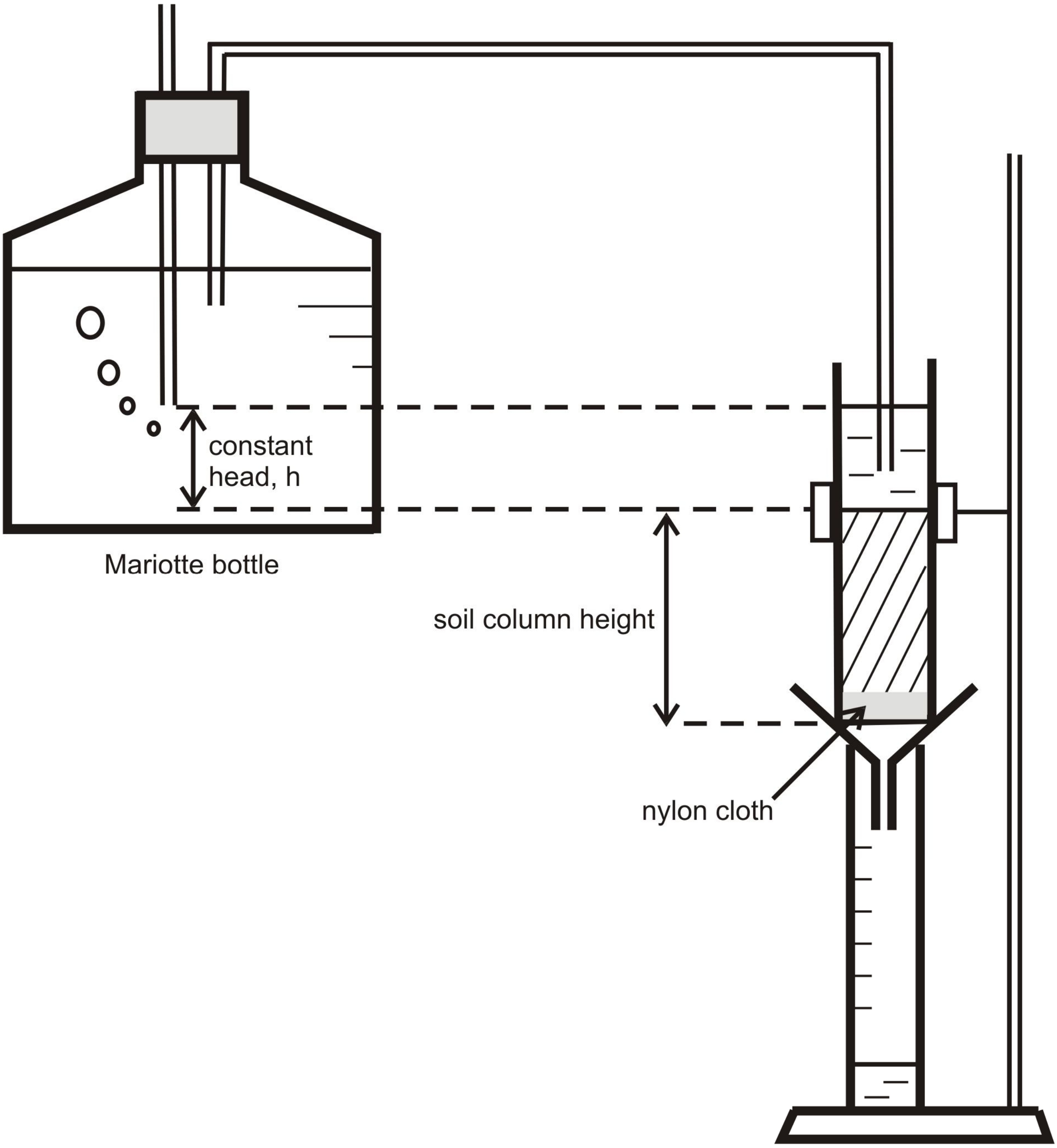


Figure 11. Scheme of system to measure saturated hydraulic conductivity

Figure 12. Measurement of saturated hydraulic conductivity

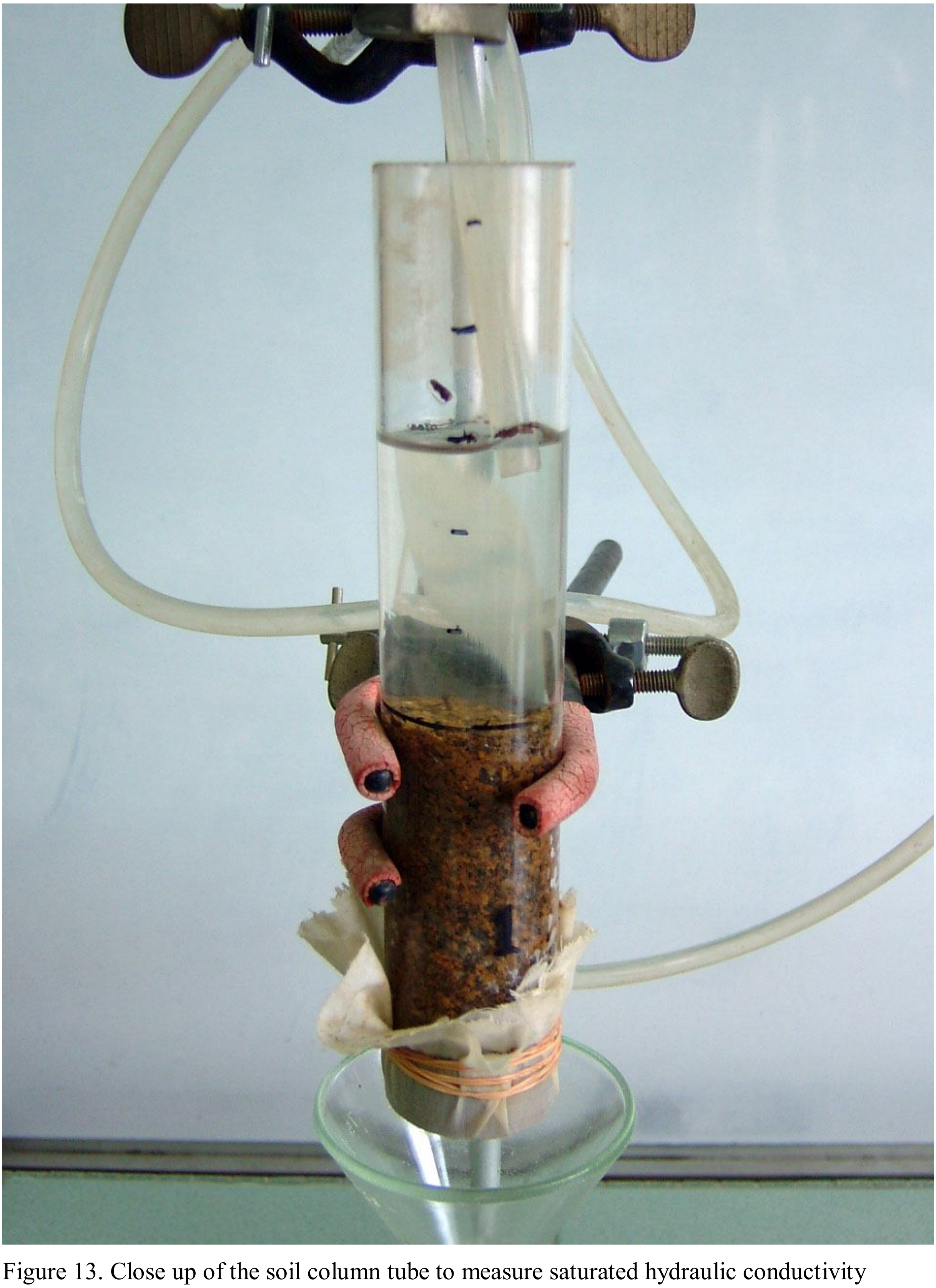


Figure 13. Close up of the soil column tube to measure saturated hydraulic conductivity

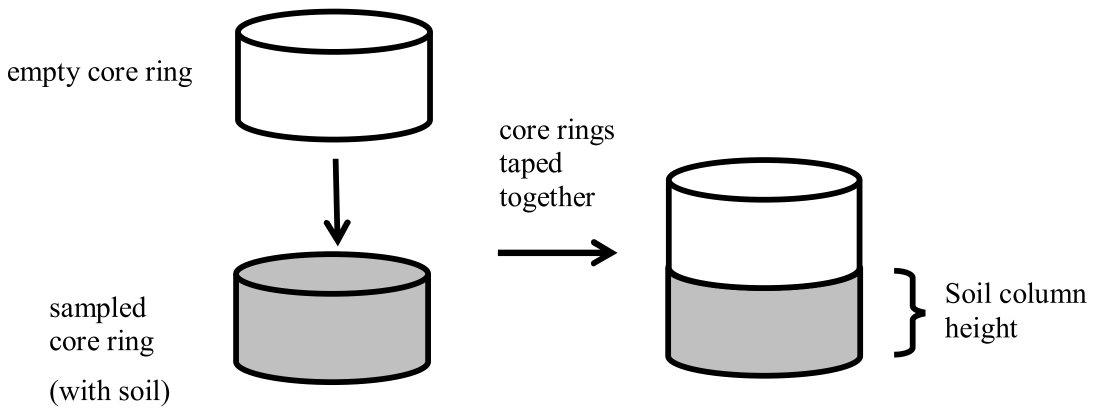


Figure 14. Using core rings for the measurement of saturated (or unsaturated, *see next chapter*) hydraulic conductivity of undisturbed soils

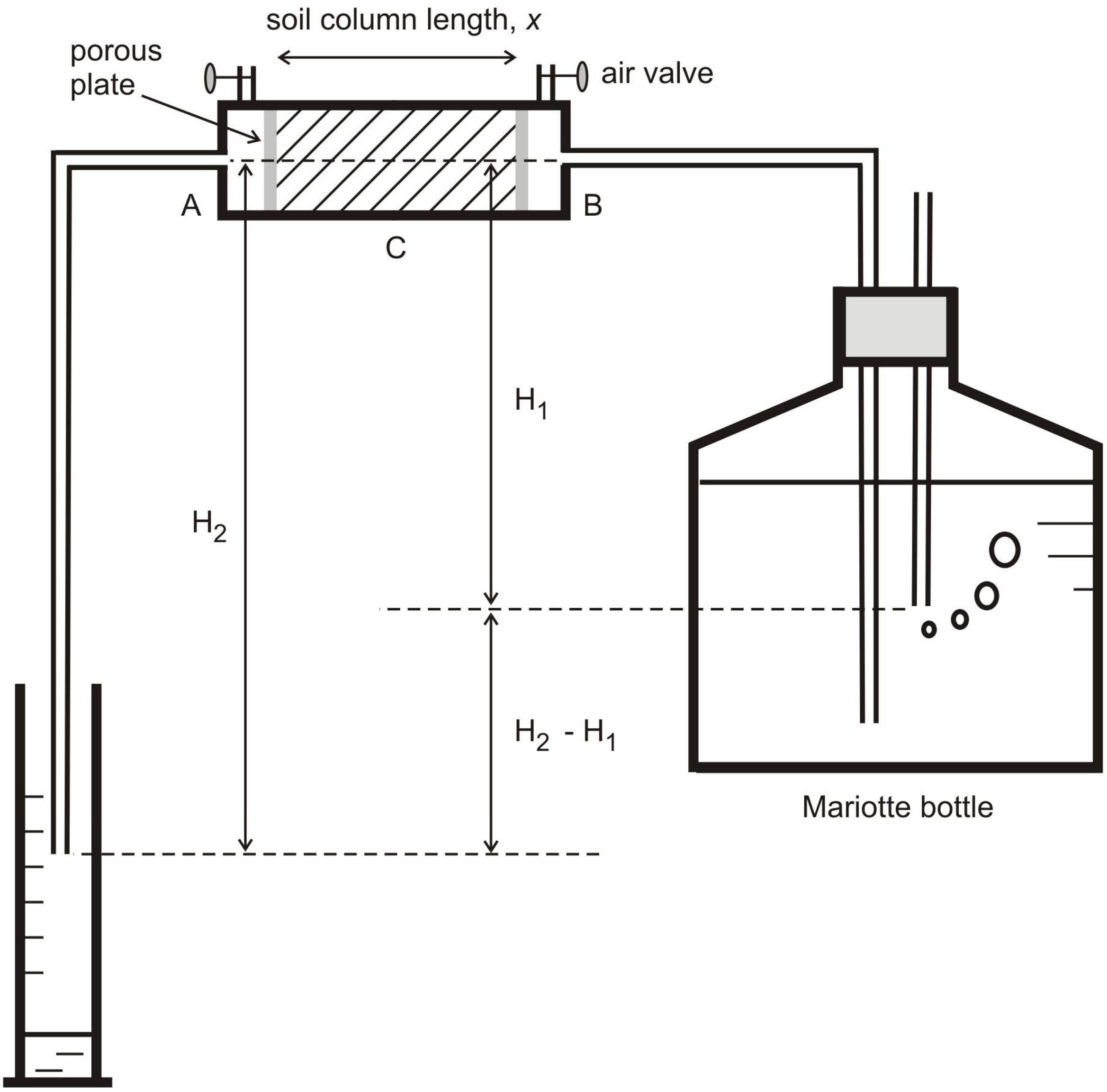


Figure 15. Scheme of system to measure unsaturated hydraulic conductivity

Figure 16. Measurement of unsaturated hydraulic conductivity

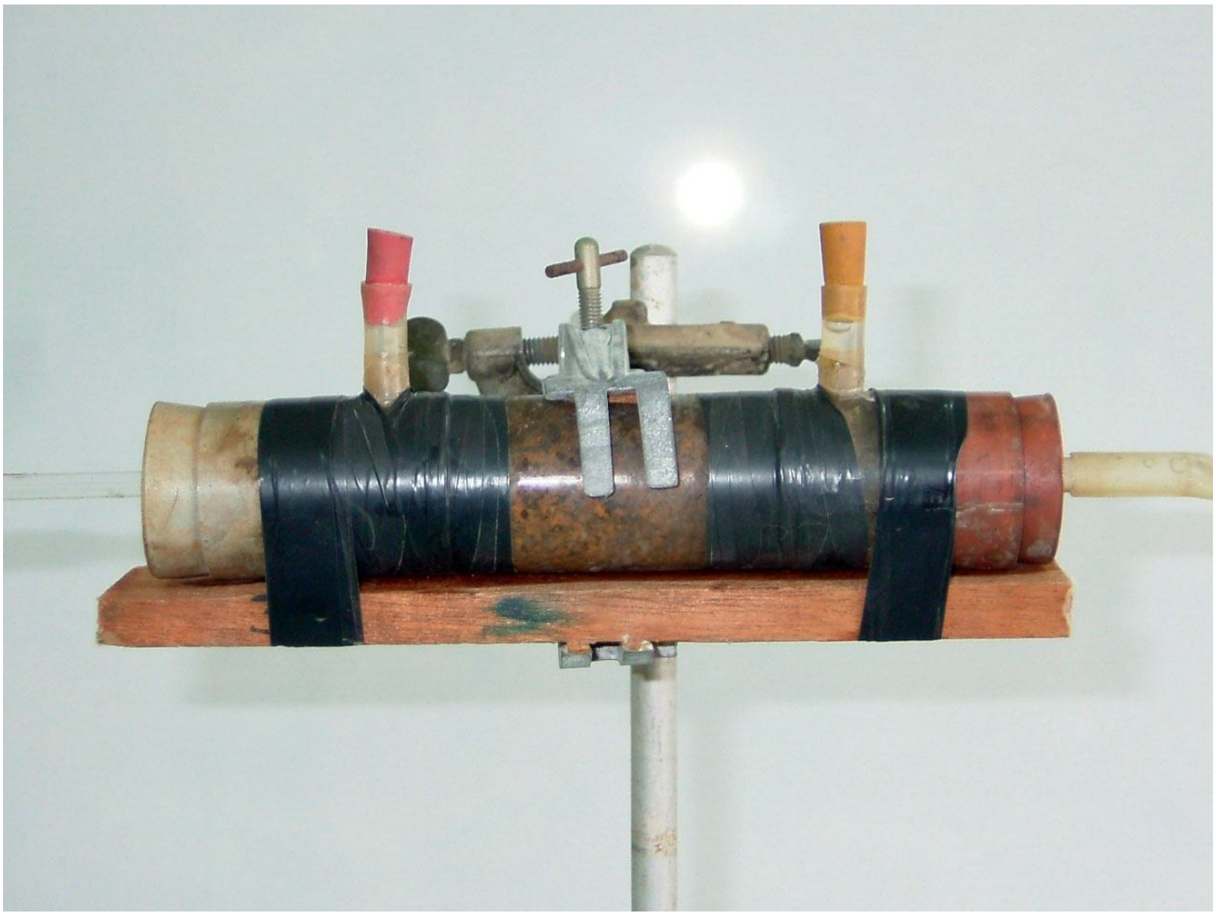


Figure 17. Close up of the soil column tube to measure unsaturated hydraulic conductivity

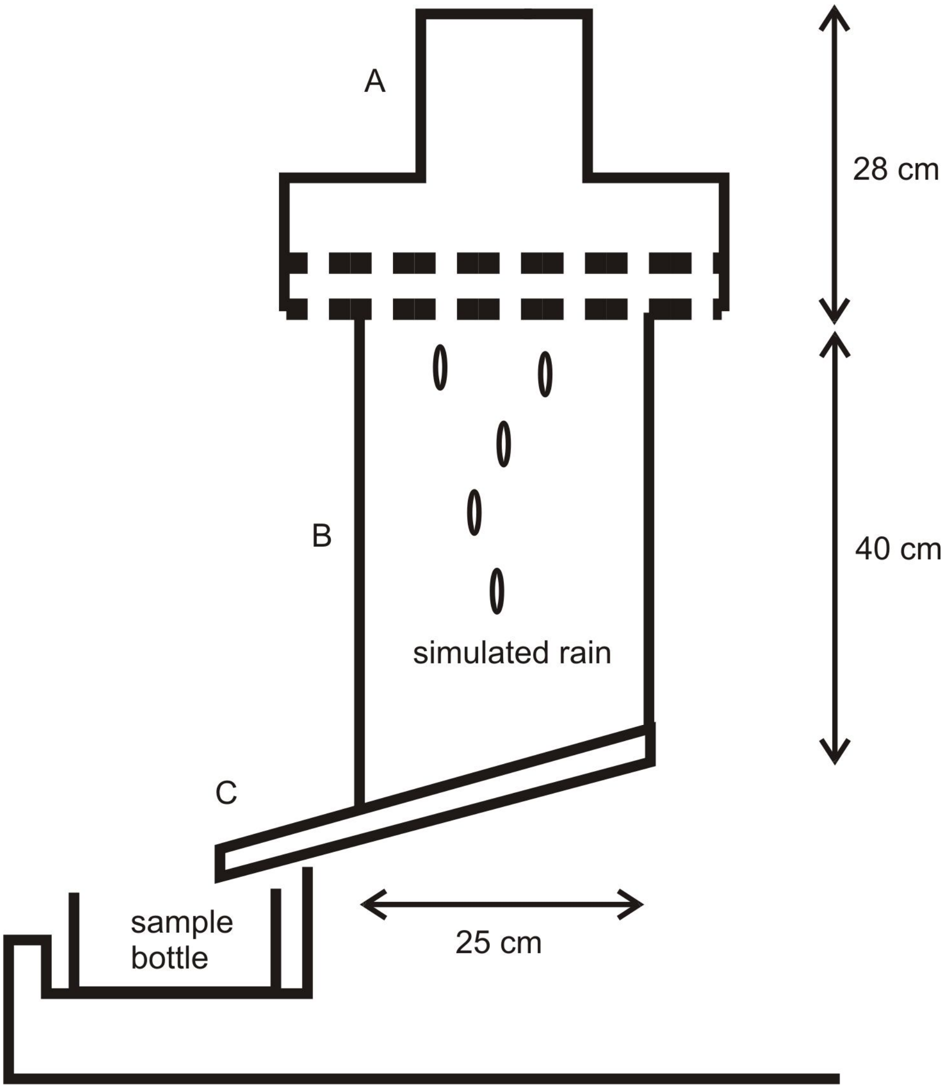


Figure 18. Vertical cross section of rainfall simulator, where A, B and C are the sprinkler system, sprinkler support, and a stainless steel frame with attached gutter, respectively.

Figure 19. Rainfall simulator apparatus