1.1. Different states of water in soil

The water contained within a soil mass is divided as follows (Fig.01):

- **Constitution water**: water that is part of the chemical composition of the soil grains;
- **Bound or adsorbed water**: the water in the soil is subjected to an electrical field near the surface of the grains;
- **Interstitial water**: capillary water and free water.

Fig.01: State of water in soil

Interstitial water occurs in the form of free water when the soil is saturated and immersed in a groundwater table. This water is subject to the laws of hydraulic flow. Interstitial water exists as capillary water above the groundwater table. Capillary water is in equilibrium, on the one hand under the action of gravity and, on the other hand under the action of tension forces that develop at the water–air interface.

1.2. Properties of Free Water:

1.2.1. Linear Flow – Darcy's Law (1856):

*Let us consider a soil cylinder with a cross-sectional area **s** (Fig.2), and assume that a flow occurs from **M** to **N**. Let **q** be the discharge through the cross-section **S**. By definition, the velocity of the water is:*

$$v = \frac{q}{S}$$

It is an **apparent velocity** since water flows only through the pores, and the actual available cross-section is reduced to ***(n·S)***, where ***(n = porosity)***. Moreover, the pores are not straight, and the water follows many tortuous paths.

Which implies that the true normal velocity, denoted v_n

$$v_n = \frac{q}{n \cdot S}$$

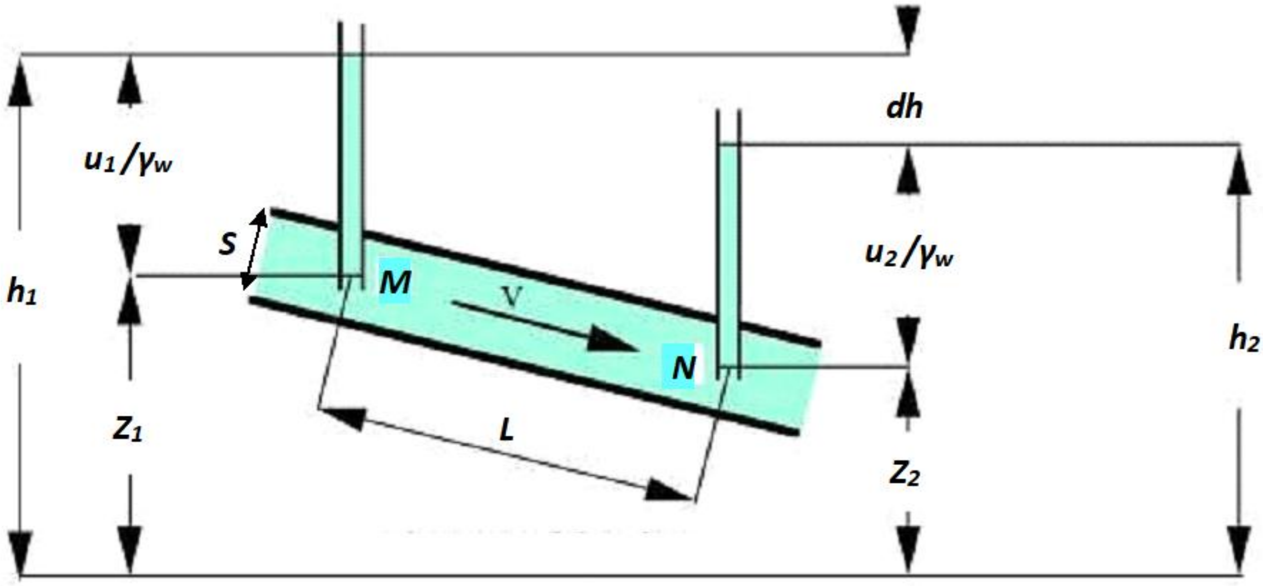

Fig.02. Linear flow

Hydraulic head (Charge hydraulique):

In hydrodynamics, the hydraulic head (**h_1**) at a point (**M**) is defined, according to **Bernoulli's theorem**, by the following expression:

$$h1 = Z_1 + \frac{u_1}{\gamma_w} + \frac{v^2}{2g}$$

Where:

- **h1**: the Hydraulic head,
- **Z_1**: the elevation (altitude) of point (M) with respect to a horizontal reference plane,
- **u_1**: the interstitial water pressure at point (M),
- **v**: the apparent velocity of water.
- **g**: the gravitational acceleration (m/s^2)

In soils, water velocities are low (< 10 cm/s), and the kinetic energy term $\frac{v^2}{2g}$ is completely negligible; therefore, the formula reduces to :

$$h1 = Z_1 + \frac{u_1}{\gamma_w}$$

Since the hydraulic head at point **M** is always $\mathbf{h_1}$, let us denote by $\mathbf{h_2}$ the hydraulic head at point **N**. According to **Bernoulli's theorem**:

- **If $h_1 = h_2$**: there is no flow, and the groundwater table is in equilibrium.
- **If $h_1 > h_2$**: flow occurs from **M** to **N**, and the head loss ($\mathbf{h_1 - h_2}$) corresponds to the energy lost due to **friction (le frottement)**. The head difference is both the driving force and the consequence of the flow.

The **hydraulic gradient (le gradient hydraulique)** ***i*** is defined as:

$$i = \frac{h1 - h2}{L} = \frac{dh}{L}$$

Darcy's law, which governs flow phenomena in soils, is expressed by the formula:

$$v = k.i$$

Where:

- v : flow velocity (seepage velocity) (m/s or cm/s),
- k : soil permeability coefficient (m/s or cm/s),
- i : hydraulic gradient (dimensionless).

The **discharge (le débit)** (**q**) through a cross-section (**S**) is given by:

$$q = v.S$$

Hence:

$$q = k.i.S$$

1.3. Laboratory Measurement of the Coefficient of Permeability:

1.3.1. Test Conditions:

The coefficient of permeability of a saturated soil is a soil property that mainly depends on its grain size distribution, nature, structure and void ratio.

The finer the soil, the smaller the pores, the greater the friction and head losses, and therefore the lower the coefficient of permeability. Clays are often considered impermeable because the flow rates through them are negligible, as their permeability is very low.

The more compact the soil, the lower its porosity. Since the space available for water flow is reduced, the soil becomes less permeable.

Two laboratory methods are used (direct applications of **Darcy's law**):

- the constant head test for highly permeable soils,
- the variable head test for low-permeability soils.

1.3.2. Constant Head Permeameter (Perméamètre à charge constante) :

A permeameter **(Fig.03)** consists of a watertight enclosure (cylindrical mold) in which a soil sample of cross-sectional area **S** and length **L** is placed. The two ends of the sample are connected to two tubes through porous stones.

In the constant head permeameter, the head difference (**h**) between the two faces of the sample is kept constant by means of an overflow device.

The test consists of measuring the quantity of water (**Q**) that flows through the sample during a time interval (**t**). According to **Darcy's law**, we have:

$$\boldsymbol{Q = q.t = k.i.S.t = k.\frac{h}{L}.S.t}$$

Hence:

$$\boldsymbol{k = \frac{Q.L}{h.S.t}}$$

Which implies:

$$k = \frac{q.L}{h.S}$$

Where:

- **Q** = quantity of water (units: m^3, cm^3, …)
- **q** = water discharge (units: m^3/s, cm^3/s, …)

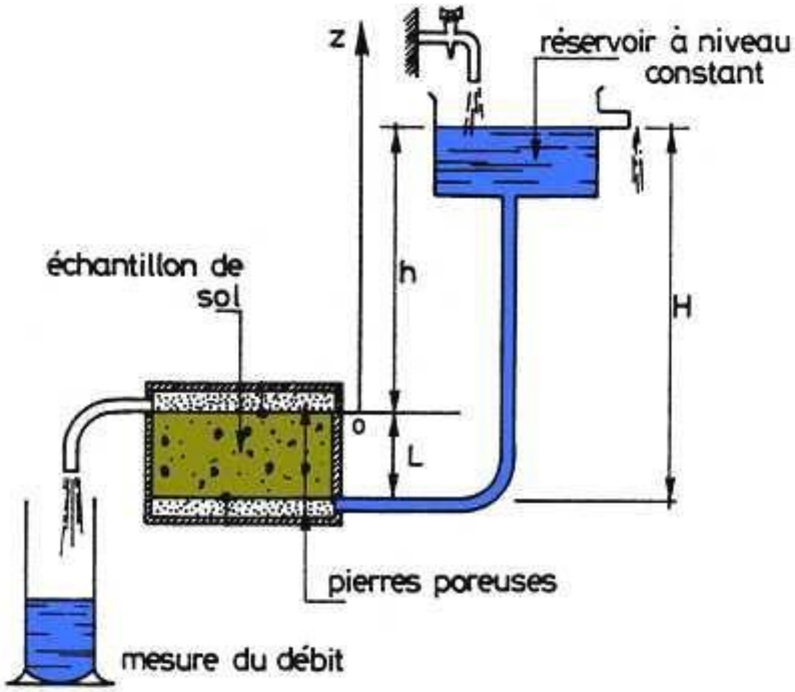

Fig.03: Principle of the Constant Head Permeameter

1.3.3. Variable Head Permeameter

In the variable head permeameter, tube (1) in Fig.4 is filled with water. The test consists of measuring the drop in water level as a function of time. Let **s** be the cross-sectional area of this tube.

During a time interval (**dt**), the quantity of water flowing out is:

$$Q = -s.dh$$

But it is also:

$$Q = q.dt = v.S.dt = k.i.S.dt$$

Since the hydraulic gradient (**i**) at time (**t**) is equal to **h/L**, we have:

$$Q = k.S.\frac{h}{L}.dt$$

By equating the two previous expressions of **Q**, we obtain:

$$k.dt = -\frac{s\,L}{S}\frac{dh}{h}$$

From which the following formulas are derived:

$$k = -\frac{s\,L}{S}\frac{In(h0/h1)}{t1 - t0}$$

Where:

- h_0: head difference at time t_0,
- h_1: head difference at time t_1,
- **ln** (h_0/h_1): natural logarithm of h_0/h_1.

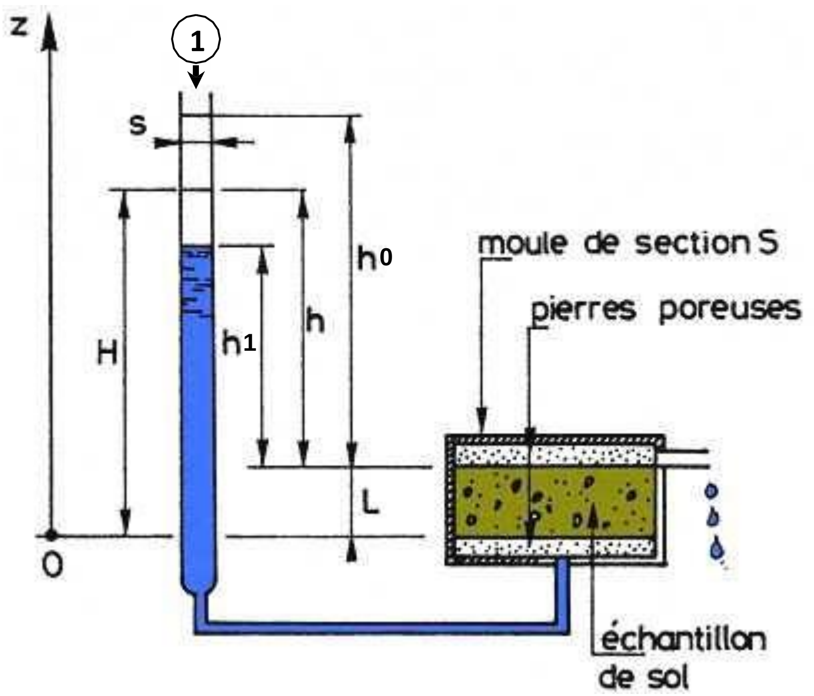

Fig.04.Principle of the Variable Head Permeameter

1.4. Order of Magnitude of the Coefficient of Permeability of Soils

The following table provides an indicative order of magnitude of the coefficient of permeability of soils.

Nature of the Soil	Order of Magnitude of k (m/s)	Degree of Permeability
Medium to coarse gravels	10^{-3} to 10^{-1}	Very high
Fine gravels, sand	10^{-3} to 10^{-5}	Fairly high
Very fine sand, silty sand, loess	10^{-5} to 10^{-7}	Low
Compact silt, silty clay	10^{-7} to 10^{-9}	Very low
Clay	10^{-9} to 10^{-12}	Practically impermeable

Soils are often stratified (layered) and therefore exhibit permeability anisotropy, meaning their permeability varies depending on the direction of flow.

Each soil layer has its own permeability, which influences the overall (equivalent, Fig.5) permeability of the soil mass. This overall permeability is different depending on whether water flows:

- **Parallel to the stratification planes** (horizontal flow), or
- **Perpendicular to the stratification planes** (vertical flow).VCXW

Assuming the layers are horizontal, we define:

- $\mathbf{k_h}$: average horizontal permeability coefficient
- $\mathbf{k_v}$: average vertical permeability coefficient

In general, permeability is much higher **along the layers** than **across them**.

Let a discharge flow through a soil layer of unit width, with:

- $\mathbf{k_1, k_2, \ldots, k_n}$: the coefficients of permeability,
- $\mathbf{H_1, H_2, \ldots, H_n}$: the thicknesses of the different layers,
- $\mathbf{H = H_1 + H_2 + \ldots + H_n}$: the total thickness,
- **h**: the total head loss,
- $\mathbf{k_v}$: the average coefficient of permeability perpendicular to the stratification planes,
- $\mathbf{k_h}$: the average coefficient of permeability parallel to the stratification planes.

➢ **If the flow is perpendicular to the stratification planes (vertical flow),** the principle of continuity requires that the discharge, and therefore the flow velocity, is the same in each layer (i.e., identical across the different layers) since the flow is steady.

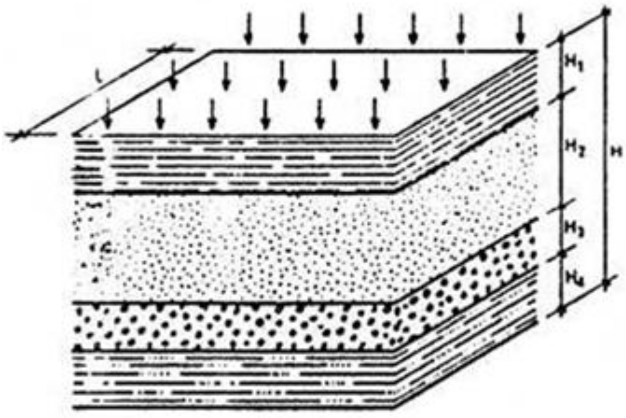

Fig.05. Flow Perpendicular to the Stratification Planes

$$Kv = \frac{H}{\frac{H1}{K1} + \frac{H2}{K2} + \cdots + \frac{Hn}{kn}}$$

Or in the form:

$$\frac{H}{Kv} = \sum_{i=1}^{i=n} \frac{Hi}{Ki}$$

- **If the flow is parallel to the stratification planes (horizontal flow, Fig.6)**, the total discharge is the sum of the discharges flowing through each layer for a slice of unit width (L = 1 m) and hydraulic gradient **i**.

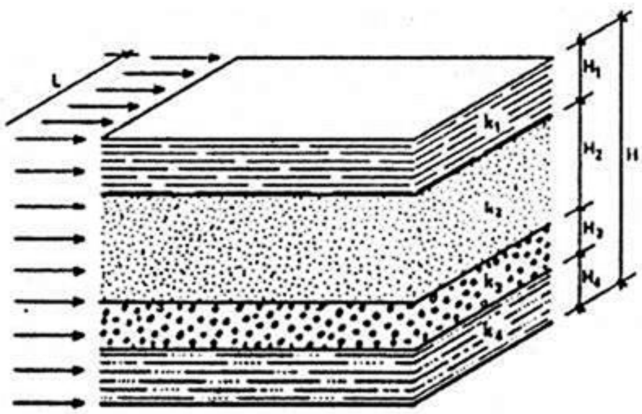

Fig.6: Flow Parallel to the Stratification Planes.

- **The average horizontal permeability coefficient:**

$$Kh = \frac{H1.K1 + H2.K2 + \cdots + Hn.kn}{H}$$

Or in the form :

$$Kh.H = \sum_{i=1}^{i=n} Ki.Hi$$