Environmental soil physics

Prof. Ayman Suleiman

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Introduction

Humans are dependent on soil and to a certain extent “good” soils are dependent upon the care given to them and the uses made of them by humans.

In this course, we will examine the physical properties of soil and show how these properties can be used to gain a greater understanding of soil behavior in agricultural and environmental applications.

Soil physics: is the study of the **physical properties of soil** and the relation of these soil physical properties to agricultural, environmental, and engineering uses.

Soil physics is quantitative and mathematical in nature and is primarily concerned with the fundamental properties of soils.

Soil physical properties

Five soil physical properties have been identified as important to the maintenance of the physical fertility of soil**:**

- ***Soil texture**: particle size distribution.
- *__Soil temperature and heat__: processes include heat transport and thermal regimes.
- **Soil aeration**: exchange of gases such as oxygen, and carbon dioxide by plant roots and soil microorganisms with atmosphere.
- **Soil water**: processes include retention, infiltration, runoff, redistribution within the profile, evaporation, transpiration and plant response, groundwater, drainage. Irrigation and transport of contaminates.
- **Soil structure**: particle and pore arrangement, compaction, consistency, and tilth.

- Spatial and temporal influences:
  All of the soil physical properties vary with space and time.
- Mathematically this relation can be written as:
  SP= ƒ(x, y, z, t).
- Its important to distinguish between the average rate of change, $\Delta\theta v / \Delta t$, and the instantaneous rate of change, $d\theta v/dt$.
- Quantification of the rates of change of mass and energy in the soil and the direction of these changes are extremely important concepts to the greater understanding of physical behavior of soil.

Areas of important soil physical research:

- **1- transport studies** , including
- - Movement and retention of water, soluble salts and pesticides in and through soil profiles.
- - Extraction and movement of plants nutrient and pesticides to seed and plant roots.
- - Movement of oxygen in the soil.
- - Transfer of heat in soil.
- **2- Tillage of the soil,** including minimum tillage and no- till.
- **3- Influences of soil compaction,** including tillage pans fragipans, and crusts.
- **4- Evapotranspiration,** including drought effects on plants and soil-plant water relation.
- **5- Soils and water conservation,** including reduction in water sediment transport off-site.
- **6- Irrigation and water quality studies.**
- **7- Spatial and temporal variability of soil properties.**

Areas of important soil physical research:

- **8-** Development of **new techniques** to measure soil water such as **remote sensing** and **TDR.**
- **9- Use of GIS** to summarize, model**,** and display soil properties.
- **10-** Computer simulation modeling of dynamic processes in soil science.

Relation of soil physics to physics

- Soil physics concerns itself with the physical – chemical, and physical –biological relationships among the solid, liquid, and gaseous phase of soil as they are affected by temperature, pressure, and energy.
- Physics has been called the most basic of the natural sciences, it seeks to establish mathematical laws to explain and predict behavior of mass and energy .Physics often uses mathematics as its language.

History of soil physics:

- Philip (1974), Gardner (1977), and wartzendruber (1977) indicate that the physical aspects of the soil have been manipulated since the early use of irrigation and tillage. These basic activities are estimated to have been practiced for at least 6000 years.
- Development of physics itself occurred primarily during the latter half of the 17th century.

Chapter 2

Physical Dimension and Units

- Mechanical quantities frequently used in the sciences and engineering include parameters such as:
- Energy , force , area , velocity acceleration.
- Work , power , volume , pressure momentum.

- <u>All of them can be expressed in soil physics in term of four dimensional quantities:</u>
- Length mass time temperature.

- The dimension of a physical quantity is composed of a numerical value (a number) and its unit.

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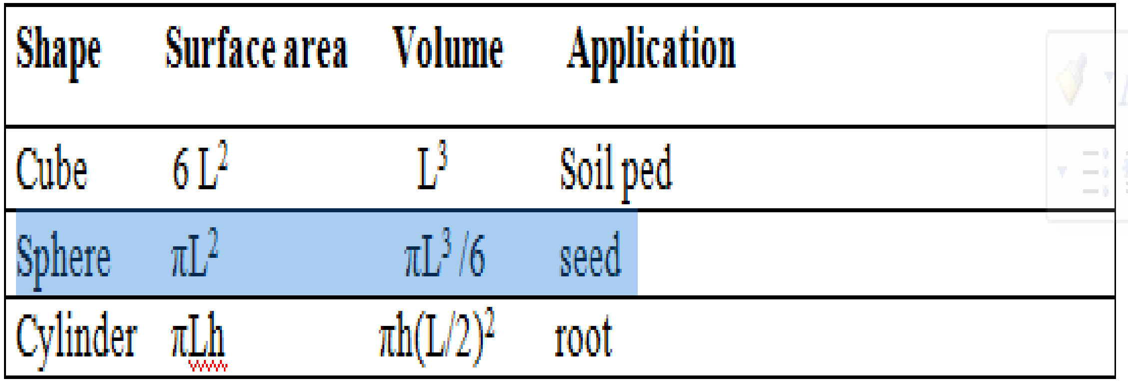
- **Length:**
- The meter is a measure of distance. A meter is the distance that light travels in 1/299,793 second.
- Kilometer = km = 1000m.
- Centimeter = cm = 0.01m = 10^{-2} m.
- Millimeter = mm = 0.001m = 10^{-3}m.
- Micron = µm = 10^{-4}cm = 10^{-6} m.
- Angstrom = A^{o} = 10^{8-} cm = 10^{-10} m
- The derived dimensions of length are *area and volume*,

- **Key relationships for area are:**
- 1 hectare (ha) = 10^4 m^2 and 1 acre = 4046 m^2.
- **Note**: in Jordan the area of land is expressed as dunum and = 1000m^2.
- **Key relationships and for volume,**
- **1000 liters = 1 m^3 .**
- The process of measuring length, area, and volume is called **mensuration**.

Example:

- Determine the surface area and volume of each of the following shapes: cube, sphere, and right circular cylinder. Assume that the height is h; the length of the side is L. And if needed, the radius is L/2.
- **Answer:**



Shape	Surface area	Volume	Application
Cube	$6\,L^2$	L^3	Soil ped
Sphere	πL^2	$\pi L^3/6$	seed
Cylinder	πLh	$\pi h(L/2)^2$	root

Mass and weight

- The base unit for mass is the kilogram: 1 kg is the mass of a Pt-Ir (metal cylinder kept near Paris, France). Mass is a measure of quantity of matter in an entity.
- We don’t measure mass directly, however; we measure its relationship to its weight.

- **Weight**: is the force of attraction from the center of the earth.
- **Example:**
- Calculate the average bulk density of the earth and interpret your answer. Assume that the earth has amass of $6*10^{27}$ g, a radius of $6.4*10^{8}$ cm, and a spherical shape.

- Bulk density =ρb= $6*10^{27}/(4/3)\pi(6.4*10^{8})^{3}$ = 5.5g/cm^{3} = 5.5 Mg/m3 = 5500kg/m^{3}.

From these calculations, we conclude the following:

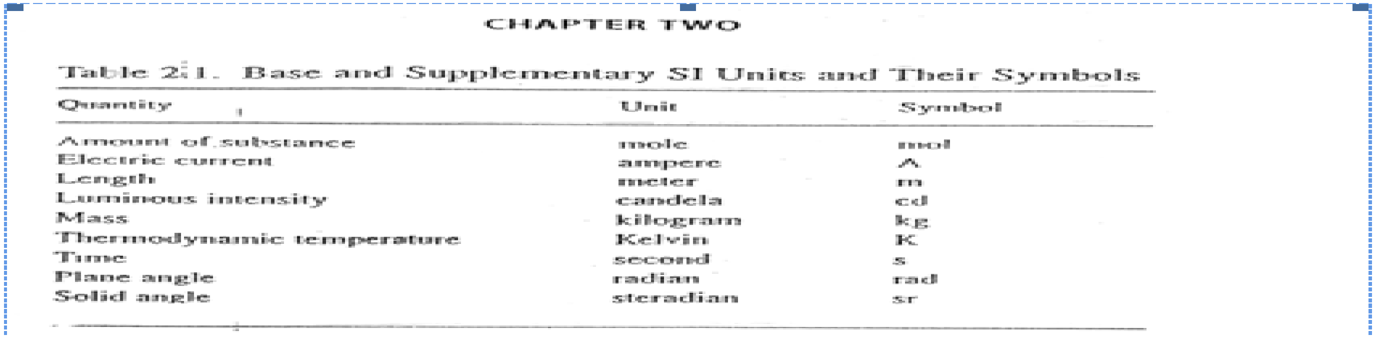
- The average bulk density of the earth is 5.5 times as dense as water the base measure. Note: The bulk density at the soil surface ranges from 1000 to 1800 kg
- The earth has a very dense core.
- This bulk density value also has implications for porosity calculations. The depth of water in the earth is about 3 km on the average and as deep as 8 km in some places.

- **Time:**
- The fundamental invariable unit of time is the second, which is 1/86.400 of the mean solar day.
- **Temperature:**
- The measure of how hot or cold a body is with respect to a standard object.
- Temperature may be measured in Kelvin (K).
- The three-temperature scale frequently used in the United States are shown in figure 2.1
- **Figure: 2.1 temperature scales frequently used in this textbook.**
-
- **$^{o}C = (5/9)\ (^{o}F - 32)$.**
- **$^{o}F = [(9/5)\ (^{o}C)] + 32$.**
- **$K = {}^{o}C + 273.15 = [(5/9)\ (^{o}F - 32)] + 273$**

Systems of units:

- The MKS (meter – kilogram- second) systems an absolute system: that is the definition of these units doses not depend on local variations in the earth’s gravitational field.
- The international system of units is essentially the MKS system.

Systems of units



CHAPTER TWO

Table 2.1. Base and Supplementary SI Units and Their Symbols

Quantity	Unit	Symbol
Amount of substance	mole	mol
Electric current	ampere	A
Length	meter	m
Luminous intensity	candela	cd
Mass	kilogram	kg
Thermodynamic temperature	Kelvin	K
Time	second	s
Plane angle	radian	rad
Solid angle	steradian	sr

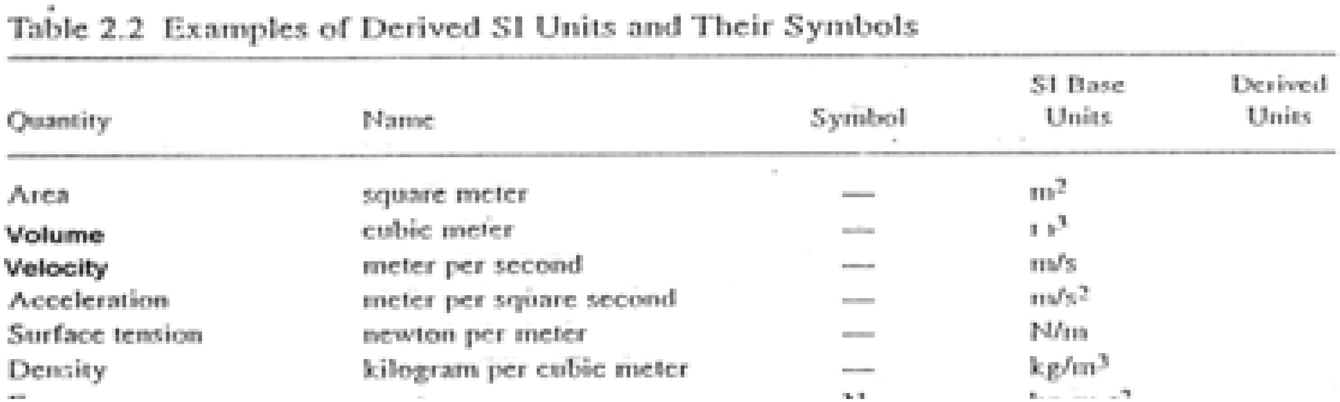


Table 2.2 Examples of Derived SI Units and Their Symbols

Quantity	Name	Symbol	SI Base Units	Derived Units
Area	square meter	—	m^2	
Volume	cubic meter	—	m^3	
Velocity	meter per second	—	m/s	
Acceleration	meter per square second	—	m/s^2	
Surface tension	newton per meter	—	N/m	
Density	kilogram per cubic meter	—	kg/m^3	

Physical Dimension and Units

- **Volume and Mass Relationships in Soils:**
- The soil is Avery complex system. Its behavior depends on magnitude and interaction among the three phases: solid, gas, and liquid.
- **<u>1- Solid phase:</u>**
- Under field conditions, the solid phase occupies from 30 to 60% of the total soil volume .the solid phase occupies approximately 50% of the soil by volume.

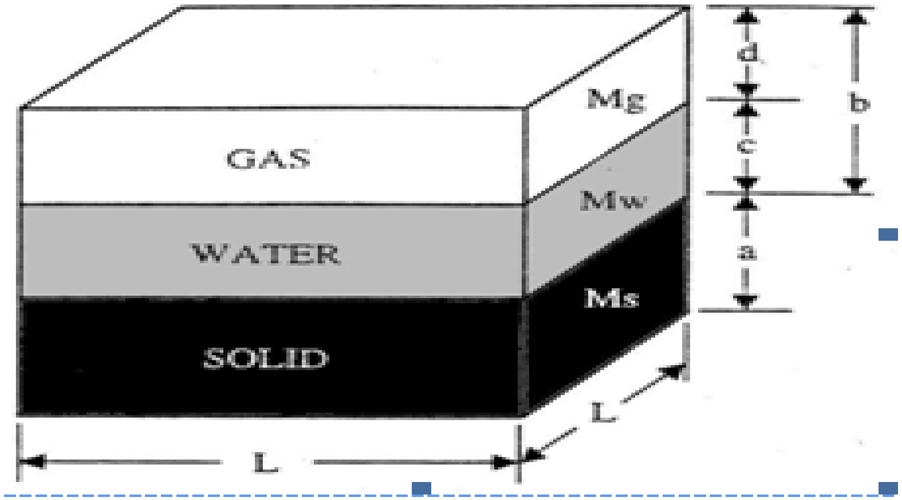


Fig 2.2 Three phases of a unit volume of soil.

	Weight (%)	volume (%)
In organic	98	49
Organic	2	2

- **2- Liquid phase:**
- The liquid phase is dynamic rather than stable because its volume usually varies between 0 to 50 % in the soil.
- Ionic composition in the soil solution includes inorganic ions, such as Na, K, Ca, Mg, Cl, NO3, SO4, HCO3, and various amount of organic ions and molecules.
- **3- Gaseous phase:**
- The gaseous phase is also a dynamic and not stable.
- **Volume /Mass Relationships of Soil Physical Quantities:**
- Mt = Ms + Mw + Mg = total mass of soil.
- Vt = Vs + Vw + Vg = total volume of soil.

- **Example:**
- An individual spherical soil particle has diameter of 2 mm. What is the mass?
- Mass = volume * density
- Volume = $(4/3)\ \pi r^3 = (4/3)\ \pi\ (.001)^3 = 4.189 * 10^{-9}\ m^3$
- Soil Particle density = 2650 kg/m^3.
- therefore assuming the volume of the soil particle can be approximeted by sphere . the mass of the particle of soil is $1.11*10^{-2}$ g

Dry bulk density:

- Dry bulk density of soil, ρ_b, can be calculated as
- ρ_b = Ms /Vt = Ms /Vt = Ms /L^3.
- The dry bulk density of most mineral soils varies from 1000 to 1800 kg /m^3. it may also varies due to tillage, compaction, swelling, aggregation, freezing and thawing, and many other factors. Therefore, it is not considered a static property at or near the soil surface.

Particle density :

- The particle density of ρ_P , can be calculated as :
- ρ_P = Ms / Vs = Ms/ aL^2 = kg solids / m^3 solids
- Particle densities of mineral soils vary between 2600 and 2700 kg / m^3 and for most mineral soils average 2650 kg / m^3. Particle density is considered to be a static property of a soil having a given mineralogical composition.

Total Porosity:

- The symbol for total porosity is f :
- $f = bL^2/ Vt = (Vw + Vg) / V_t = m^3$ voids / m^3 soil.
- For mineral soils, its value generally is between 0.3 and 0.6 m^3/ m^3. For coarse – textured soils, values of tend to be less than for the fine – textured soils, even though the average size of the pores are larger in the coarse – textured soils.

Aeration porosity:

- The symbol for aeration porosity is f_a
- $f_a = = d L^2 /L^3 = m^3$air / m^3soil.
- **Void Ratio:**
- The symbol for void ratio is e:
- $e = (V_g + V_w) / V_s = m^3$ voids / m^3solid.

Water content:

Soil water content by weight, $\theta_W = M_w / M_s$ = kg water / kg solids

% Water (by weight) = θ_W *100 = (wt. (wet soil) – wt.(dry soil) *100)/wt. (dry soil)

= [wt. (water lost)/ wt. (dry soil)]*100

It is frequently necessary to make a conversion from water content on dry – weight basis to water content on wet – weight basis. The equation for this conversion is:

$\theta_{WW} = \theta_{dw/\ 91+}\ \theta_{dw)}$

where θ_{dw} and θ_{WW} are the soil water content on dry and wet basis, respectively.

$\theta_V = V_w / V_t$ = m^3 water /m^3 soil.

$\boldsymbol{\theta_{V} = \theta_{W} (\rho_b/\rho_W)}$

- **Water volume ratio**

θ_r = VW/VS

- **Equivalent Depth Of Water Per Depth Of Soil.**

The quantity D_{eq} is

Depth of water = $D_{eg} = Vw/L^2 = c\ L^2/\ L^2 = \theta_V \Delta z$.

$D_{eq\ =}\ \theta_V(z)\ dz \approx (\theta_{vi}\ \Delta_{Zi})$.

- **Relative Saturation Ratio:**

The relative saturation ratio $\theta_{S,}$ is calculated as:

$\theta_{S\ =}\ V_W\ /\ (V_W + V_g)_{\ =}\ m^3$ water $/m^3$voids.

Relationship among soil physical quantity:

- From basic mass – volume definition given above, relationship between several of the physical quantities have been derived. The derivations of several of these quantities are given below:
-
- 1- The relation between the void ratio and porosity is:
- $e = f/(1-f)$ or $f = e/(1+e)$

The textbook	another method
e= $\frac{(Vg+Vw)}{Vs}$, multiply by V_t and divide by V_t, then	e= $\frac{(Vg+Vw)}{Vs}$
$e = \frac{(Vg+Vw)}{Vs} * \frac{(Vt)}{Vt} = \frac{(Vg+Vw)}{Vt} * \frac{(Vt)}{Vs}$	e= $\frac{Vp}{Vs} = \frac{Vp}{Vt-Vp} = \frac{Vp}{Vp(\frac{Vt}{Vp}-1)}$
$= f*(\frac{(Vt)}{Vt-(Vw+Vg)}) = f*(\frac{(Vt)}{Vt(1-\frac{(Vw+Vg)}{Vt})})$ $= f*(\frac{1}{1-f}) = f/(1-f)$	$= \frac{1}{(\frac{1}{f}-1)} = \frac{1}{(\frac{1-f}{f})} = f/(1-f)$

- The relation between volumetric water content, saturation ratio, and total porosity is:
- $\theta_S = \theta_\upsilon / f$
- 3- The relationship between total porosity, bulk density, and particle density is $f = 1-(\rho_b/\rho_p)$

method 1	method 2
$f = \frac{(Vw + Vg)}{Vt} = \frac{(Vt - Vs)}{Vt}$	$f = \frac{(Vp)}{Vt} = \frac{(Vt - Vp)}{Vt}$ $f = (1 - \frac{Vp}{Vt}]$
$= [1 - \frac{Vs}{Vt}] = [1 - (\frac{Vs}{Vt} * \frac{Ms}{Ms})]$ $= [1 - (\frac{Ms}{Vt} * \frac{Vs}{Ms})]$ $= 1 - (\rho_b/\rho_p)$	$\rho_b = Ms/Vt$ which means $Vt = Ms/\rho_b$ $\rho_p = Ms/Vp$ which means $Vp = Ms/\rho_p$ $f = (1 - \frac{Ms/\rho p}{Ms/\rho b}] = 1 - (\rho_b/\rho_p)$

- -The relationship between volumetric water content, water content by weight, bulk density and water density is:

$$\theta_{\upsilon} = \theta_{w}(\rho_{b}/\rho_{w})$$

$$= (Vw/Vt)(Mw/Mw)(Ms/Ms)$$

$$= (Mw/Ms)(Ms/Vt)(Vw/Mw)$$

$$= \theta_w(\rho_b/\rho_w)$$

$$\theta\upsilon = \frac{Vw}{Vt} = \frac{Mw/\rho w}{Ms/\rho b}$$

$$= \theta w \frac{\rho b}{\rho w}$$

$$f_a = \frac{Vg}{Vt} = \frac{Vg + Vw - Vw}{Vt} = \frac{(Vg + Vw)}{Vt} - \frac{Vw}{Vt}$$

$$= f - \theta_\upsilon$$

- The relationship between water volume ratios, volumetric water content and void ratio is:
- **Prove: $\theta_r = \theta_\upsilon(1+e)$**
- θ_r= VW/VS multiply by and divide by V_t then,
- = (VW/VS)* (VT/VT)

$$= \frac{Vw}{Vt} * \frac{Vt}{Vs} = \theta_\upsilon * \frac{(Vs + Vg + Vw)}{Vs} = \theta_\upsilon *\left(\frac{Vs}{Vs} + \frac{(Vg + Vw)}{Vs}\right) =$$

$$=\theta_\upsilon\,(1+e)$$

Calculation of soil physical quantities (continued):

- Assume that the effective root zone for soybeans in the field was 45 cm. A soil sample was taken from the field 48h after the field was uniformly saturated and allowed to drain. The following information was known or was obtained from soil sample:
- Soil sample volume = 700 cm^3.
- **Wet soil weight = 1150 g**
- **Dry soil weight = 875 g**
- **Particle density of soil = 2.60 g/cm^3.**
- **Permanent wilting point = 0.13 cm^3/cm^3.**

- Based on the information given, determine the values of the quantities listed below for the soil sample. The answers are given in the parentheses:
- Water content by weight (0.314 kg / kg)
-
- Bulk density (1.25 Mg / m^3)
-
- Water content by volume (0.393 m^3 / m^3)

- Volume of water (275 cm^3)

- Total porosity (0.519 m^3/m^3)

- Volume of soil solids (336.7 cm^3)
- Volume of soil pores (363.3 cm^3)
- The sum of macro – and mesoporosity after 48 h of drainage (0.126m^3/m^3)
- Micro porosity (0.393 m^3/m^3)
- Equivalent depth of water (cm) of soil water at field capacity in effective root zone (17.7cm)
-
- Plant available water content after 48 h of drainage by volume (0.263 m^3/m^3)
- Available water content in cm of water after 48 h of drainage (11.8 cm)
- If the water content in the root zone must be maintained at 30 % by volume, how many centimeters of water must be removed from the profile? (4.2 cm)

- If the initial water content is 20% by volume, how many centimeters of water must be added to the soil so that the water content in the soil can be brought to saturation? (14.36 cm)

- If the initial water content is 23% by volume and 5 cm of water are uniformly added to the root zone, the water content in the root zone will be increased to what percentage by volume? (34.1 %)

- What is the maximum retentive capacity of this soil? (0.519 m^3/m^3)

- What is the permanent wilting point in percentage by weight? (10.4%)

- If a soil sample is taken 24h after the field was saturated and the sample had a water content by weight of 37.5 %, what is the aeration porosity of the soil at this time? (0.05m^3/m^3).

Soil Texture

- The particle size distribution of a soil determines to a large extent its physical, chemical, and microbiological behavior.
- **Amount of solid phase:**
- 1- The volume fraction of solid phase , S , is calculated from the ratio of the bulk density and particle density.
- **$S = \rho_b/\rho_p = m^3$ solids / m^3 soil**
- 2- The total volume of soil solids, Vs , in a given soil volume is:
- **$Vs = S \times V_t = m^3$ solids**
- 3- The total mass of the solid phase, Ms, in a given volume of soil is:
- $Ms = \rho_b \times V_t = S \times V_t \times \rho_p =$ kg solids

Example

- Calculate the mass of oven–dry soil having a bulk density of 1400 kg/m^3, depth of 15 cm, and an area of 1 ha. What is the influences of bulk density on mass of soil in this area?
- Ms = 1400 $kg/m^3 \times 10000\ m^2 \times 0.15m = 2.1\times10^6 kg$
- or Ms = (1400/2650) × 10000 $m^2 \times 0.15m \times 26500 = 2.1\times 10^6$ kg.
- As ρ_b increases, the mass of solid – phase increases.

General Review of Soil Separates:

- The term soil texture is an expression of the predominant size range, of the particles. Quantitatively, soil texture refers to **the relative proportions of the various sizes of particles in a given soil**.
- Soil texture is considered a permanent, natural attribute of the soil. Therefore, it is a static soil physical property.
- There is no natural classification of particle size:
- Several classification systems have been developed around the world.

Table 3.1: The International Soil Science Society (ISSS) Classification of Soil particles

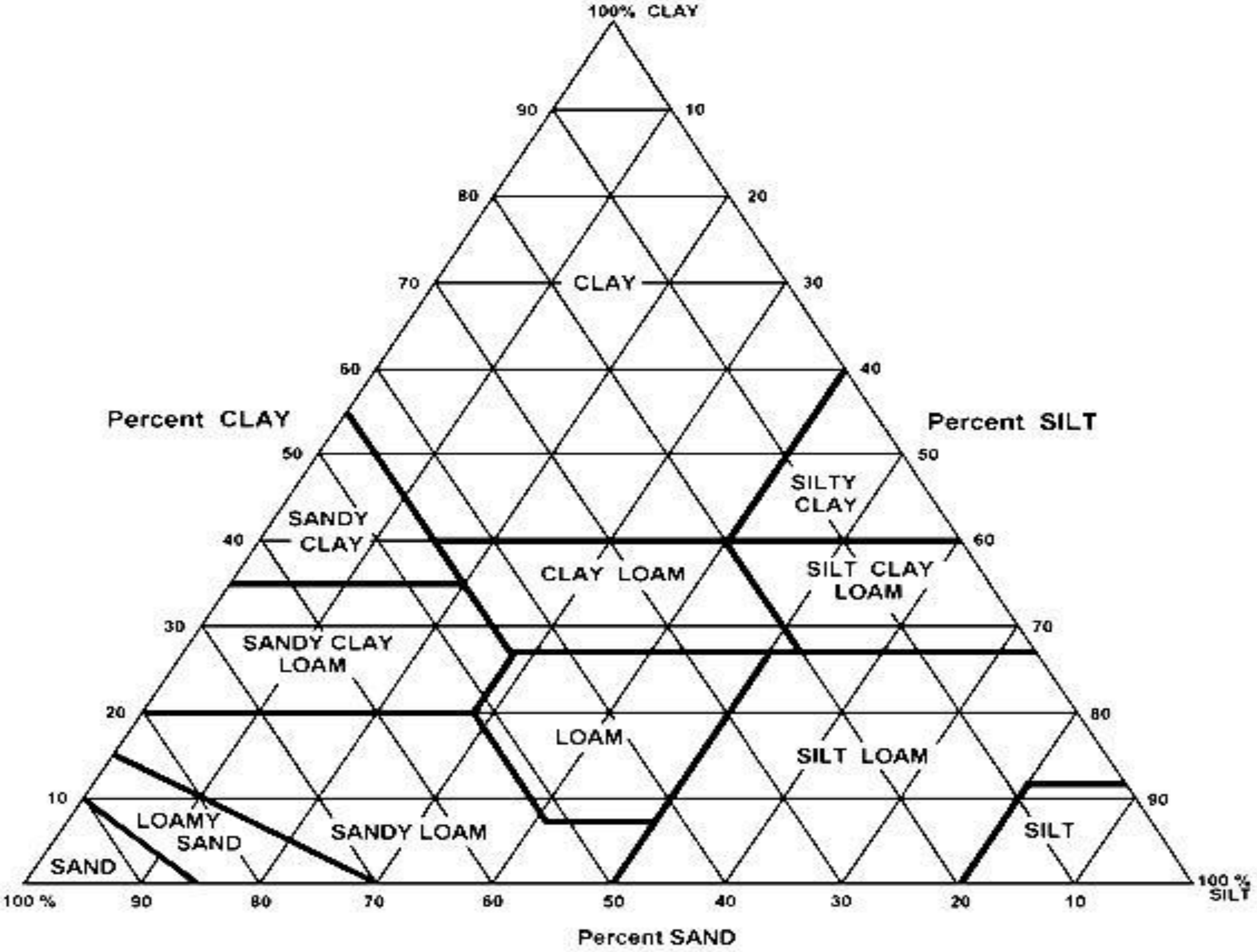
Separate	Particle diameter	
	mm	µm
Coarse sand	2.0-0.2	2000-200
Fine sand	0.2-0.02	200-20
Silt	0.02-0.002	20-2
Clay	<0.002	< 2

Table 3.2: The USDA Classification System of Soil Particle

Separate	**Particle Diameter**	
	mm	**Mm**
Very coarse sand	2.0-1.0	2000-1000
Coarse sand	1.0-0.5	1000-500
Medium sand	0.5-0.25	500-250
Fine sand	0.25-0.10	250-100
Very fine sand	0.10-0.05	100-50
Silt	0.05-0.002	50-2
Clay	<0.002	<2

Textural Classes of Soil Particles:

- Most soils and sediments are mixtures of particles of various sizes. The USDA, textural triangle has 12 classes, which were defined for their practical value in agriculture


100% CLAY
Percent CLAY
Percent SILT
Percent SAND
90
80
70
60
50
40
30
20
10
100 %
100 % SILT
CLAY
SILTY CLAY
SANDY CLAY
CLAY LOAM
SILT CLAY LOAM
SANDY CLAY LOAM
LOAM
SILT LOAM
LOAMY SAND
SANDY LOAM
SAND
SILT

- When the coordinate occurs exactly on the line between two textural classes, it is customary to use the class associated with the finer particle size.
- Loam whenever all three major size fractions occur in sizable proportions.
-
- A general grouping of the 12 basic classes of soil texture is as follows:

Coarse –textured soils:

- -Sands and loamy sands.
- - Sandy loams, including fine sandy loams.
- The sand may be characterized as very fine, coarse, and very coarse.

Medium – textured soils:

- - Loamy soils: loam, silt loam, silt.
- - Moderately heavy soils: clay loam, sandy clay loam, and silty clay loam.

Fine –textured soils:

- - Clay and silty clays.
- - Sandy clays.

- specific definitions of the soil textural classes and their physical caharcterization are presented in the table (3.4).

Importance of textural triangle:

- 1-Indication of specific surface, that is, surface area per unit mass or volume. The smaller the particle is, the greater the specific surface.

- 2- It is a guide to the value of the land

- 3- Land use capability and methods of soil management

- 4- Soil texture affects the movement and retention of water, solutes, heat, and air.

- **Important**: The use of the words “light” and “heavy” to describe soils originated from difference in the amount of work needed to cultivate soils, not from the density of the particles.
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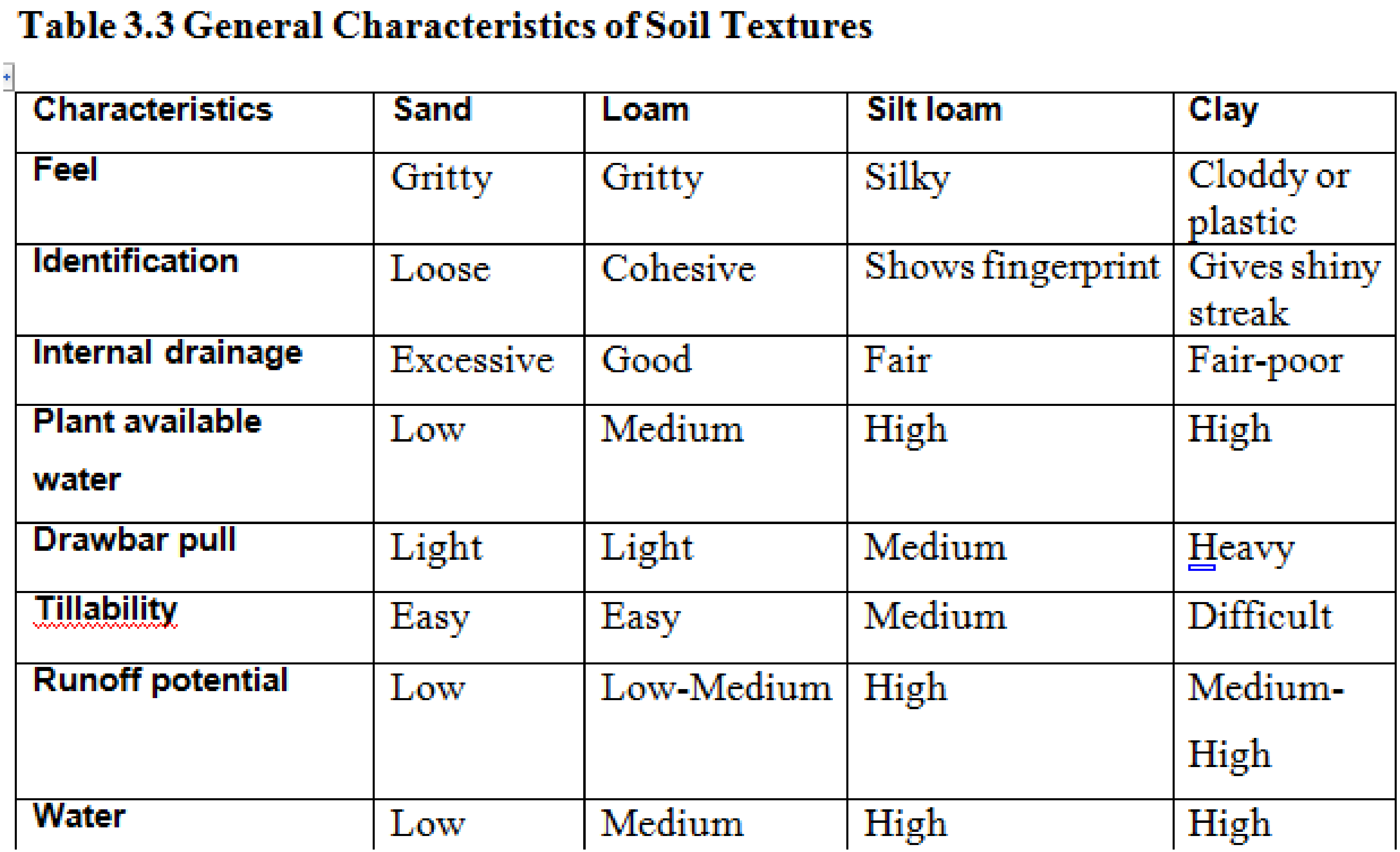


Table 3.3 General Characteristics of Soil Textures

Characteristics	Sand	Loam	Silt loam	Clay
Feel	Gritty	Gritty	Silky	Cloddy or plastic
Identification	Loose	Cohesive	Shows fingerprint	Gives shiny streak
Internal drainage	Excessive	Good	Fair	Fair-poor
Plant available water	Low	Medium	High	High
Drawbar pull	Light	Light	Medium	Heavy
Tillability	Easy	Easy	Medium	Difficult
Runoff potential	Low	Low-Medium	High	Medium-High
Water	Low	Medium	High	High

In soil survey reports, percentages are based on oven –dry mineral soil particles passing through a 2 mm sieve.

Soil Texture

- **3.4 Summation Curves and Uniformity Index:**
- The distribution of particle sizes in a soil can be represented by computing summation curves:
- 1-Compute values of percentage composition (%p).
- % p = (weight of fraction / total weight of washed soil) ×100
- 2- Sum values of % p over size ranges beginning with the percentage of particles, by weight, smaller than the upper size limit.
- The shapes of summation curves also give information on the grading pattern (fig 3.4). For example, a well –graded soils has a flattened and smooth summation curve (fig 3.4a) and poorly graded soils has a step-like summation curve (fig 3.4 c)

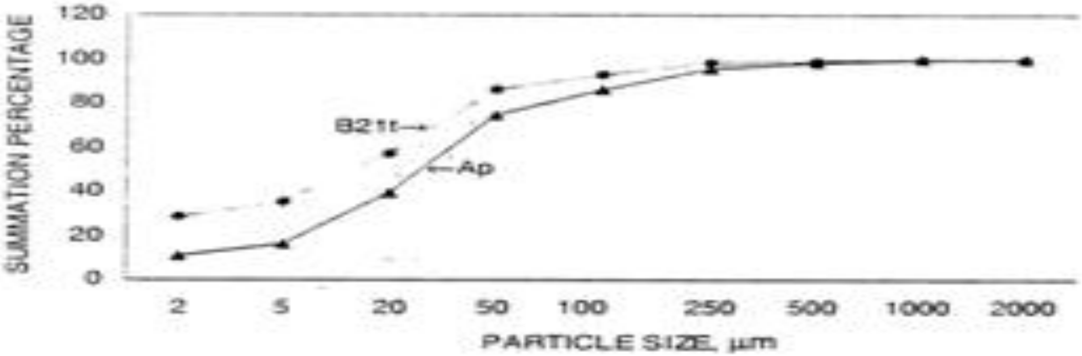
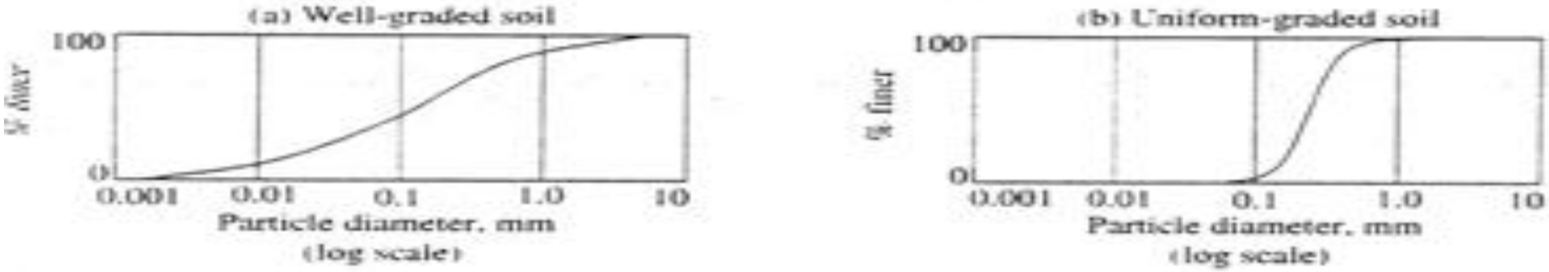


Figure 3.3. Summation curves for two horizons of a Peridge soil.


(a) Well-graded soil

(b) Uniform-graded soil

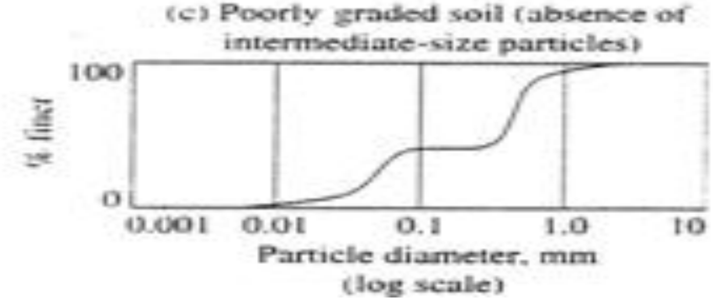

(c) Poorly graded soil (absence of intermediate-size particles)

Figure 3.4. Typical shapes of summation curves of soil.

- Another method used to represent soil particle size distribution is uniformity index
- **The uniformity index , Iu , is mostly used for coarse – textured soils in engineering**
- **Iu = D_{60}/D_{10}**
- Where D60 and D10 are the diameters of soil particle at summation percentages 60 and 10, respectively.
- Example, uniformity index of the A_P horizon of the peridge soil is:
- Iu= 0.043/0.002 = 21.5
- The Iu for B_{21t} = 0.025/????, can not be calculated because of much clay.
- Soils with the same particle size have Iu values of 1.0 ,. Some well- graded sandy soils have Iu values greater than 1000.

3.5 Determination of particle size Distribution:

- The determination of the relative distribution of the size groups of soil particle is called a **mechanical analysis** and is one of the most common physical analyses. Success of the analysis depends on two factors:
- Preparation of the sample to completely disperse all aggregates.
- Accurate fraction of the sample into various separates.

Preparation of the sample :

- The main objective is to obtain maximum dispersion and to maintain this dispersion while performing the particle size analysis.
- This can be accomplished by:
- Removal of the binding (cementing) agents such as organic matter and Fe and Al oxides.
- Mechanical rehydration of clay particle and
- Physical and chemical dispersion of the particles.
- Dispersion may be achieved by mechanical and chemical means.

Mechanical methods include:

1- Shaking in water.

2- Stirring.

3- Rolling with a wooden roller.

4- Agitating with air jets in specially constructed apparatus and

5- Boiling with water.

Chemical methods :

1- Hydrogen peroxides H_2O_2, which is used to destroy the organic binding agents.

2- Sodium acetate, which is used to remove the carbonate cements.

3- Na – dithionate in citrate solution (chelating agent), which makes the iron cements soluble.

Raising the electro kinetic potential until particles repel each other will keep colloidal particles in suspension. This can be done effectively by replacing the adsorbed polyvalent cations (particularly divalent and trivalent cations) with Na or Li.

Some of chemicals that can be used are:

1- Sodium hexametaphosphate, $Na(PO_3)_2$ ~ (Calgon).

2- Sodium hydroxides, Na(OH) .

3- Sodium oxalate , NaC_2O_4 and

4- Sodium carbonate, $NaCO_3$.

- **Sieving:**
- Sieving is the preferred method for separating the coarser fraction of soils (sand)
- After wet sieving, sand should be dried and sieved again because part of the silt fraction will be caught on the 0.05 mm screen.
- **Sedimentation (Stoke’s Law):**
- Derivation of the sedimentation Equation:
- The principle behind the sedimentation of various sizes of the soil particles is known as stokes’ law of sedimentation. This law can be simply derived as follows:
- The particle falls in a liquid media without acceleration but at a constant velocity known as the ***terminal velocity***, *or settling velocity.*

- Since the acceleration is zero, we can equate the vertical upward and downward forces and obtain:
- **Buoyancy Force + Viscous Drag Force = gravity Force**
- B + D = G
- **Viscous Drag, D** = $(2 \pi r)*\eta*3\nu = 6\pi r\eta\nu$.
- **G (gravity Force**) = ma = mg = Volume*density*g = $(4/3) \pi r^3\rho_p g$
- **B $_{\textbf{Buoyancy}}$** = weight of the liquid displaced = volume * density of liquid* gravity = $(4/3)*\pi r^3\rho_l g$
- r: the particle radius (m) ,
- ρ_p : the particle density (kg/m^3) , and
- ρ_1 : the liquid density (kg/m^3) .
- Therefore, we write
- $\mathbf{(4/3) *\pi r^3\rho_1 g +6\pi r\eta\upsilon = (4/3)*\pi r^3 g\rho_p}$.

- solving for the terminal velocity gives:

$$\upsilon = (4/3)\ \pi r^3 g(\rho_p-\rho_1)/6\pi r\eta$$

$$\upsilon = \frac{2gr^2(\rho_p - \rho_l)}{9\eta}$$

- Values of r differ according to USDA or ISSS classification.
- The magnitude of η and ρ_1 is a function of liquid and its temperature value for water, given in the following table.

Physical Properties of Liquid Water

Temperature (°C)	Density (kg/m³)	Specific heat (J/kg deg)	Latent heat (vaporization) (J/kg)	Surface tension (kg/sec²)	Thermal conductivity (J/m sec deg)	Dynamic Viscosity (kg/m sec)	Kinematic viscosity (m²/sec)
−10	0.99794×10^3	4.27×10^3	2.53×10^6	—	—	—	—
−5	0.99918×10^3	4.23×10^3	2.51×10^6	7.64×10^{-2}	—	—	—
0	0.99987×10^3	4.22×10^3	2.50×10^6	7.56×10^{-2}	0.561	0.1787×10^{-2}	1.79×10^{-6}
4	1.00000×10^3	4.21×10^3	2.49×10^6	7.5×10^{-2}	0.570	0.1567×10^{-2}	1.57×10^{-6}
5	0.99999×10^3	4.207×10^3	2.49×10^6	7.48×10^{-2}	0.574	0.1519×10^{-2}	1.52×10^{-6}
10	0.99973×10^3	4.194×10^3	2.48×10^6	7.42×10^{-2}	0.587	0.1307×10^{-2}	1.31×10^{-6}
15	0.99913×10^3	4.19×10^3	2.47×10^6	7.34×10^{-2}	0.595	0.1139×10^{-2}	1.14×10^{-6}
20	0.99823×10^3	4.186×10^3	2.46×10^6	7.27×10^{-2}	0.603	0.1002×10^{-2}	1.007×10^{-6}
25	0.99708×10^3	4.18×10^3	2.44×10^6	7.19×10^{-2}	0.612	0.0890×10^{-2}	0.897×10^{-6}
30	0.99568×10^3	4.18×10^3	2.43×10^6	7.11×10^{-2}	0.620	0.0798×10^{-2}	0.804×10^{-6}
35	0.99406×10^3	4.18×10^3	2.42×10^6	7.03×10^{-2}	0.629	0.0719×10^{-2}	0.733×10^{-6}
40	0.99225×10^3	4.18×10^3	2.41×10^6	6.95×10^{-2}	0.633	0.0653×10^{-2}	0.661×10^{-6}
45	0.99024×10^3	4.18×10^3	2.40×10^6	6.87×10^{-2}	0.641	0.0596×10^{-2}	0.609×10^{-6}
50	0.98807×10^3	4.186×10^3	2.38×10^6	6.79×10^{-2}	0.645	0.0547×10^{-2}	0.556×10^{-6}

- Other formulas for Stoke’s law
- In terms of diameter: **$\upsilon = gd^2 (\rho_p - p_1)/18\eta$**
- 1- At 20ºC and $\rho_p=2.65 Mg/m^3$ **$\upsilon = 8983d^2$**
- 2- We can calculate the time needed for the particle to fall through a distance L in the cylinder to be:
- **$t = 18 L \eta/gd^2 (\rho_p - p_l)$**
- 3- The diameter of a particle can also be calculated from the sedimentation equation as:
- **$d = [18L\eta/tg (\rho_p - \rho_1)]^{0.5}$**

** The use of stokes law for measurement of particle sizes is dependent on certain simplifying assumption:

1- The particles must be large in comparison to liquid molecules. Occurs for all soil particles with d greater than $1*10^{-6}$ m.

2- The volume of the liquid must be great in comparison to the size of the particles.

3- Particles must be rigid, smooth, and spherical. This is why the term equivalent diameter is used

4- There must be no slipping between the particles and liquid.

5- The terminal velocity must not exceed a certain critical value.

- Therefore, the sedimentation equation is not applicable to soil particles having a diameter larger than very fine sand(i.e. roughly 80 μm)
- In summary, Stokes law of sedimentation has certain limitations:
- 1- Shape of particles.
- 2- Temperature.
- 3- Density of the particles.

Techniques of mechanical analysis based on stokes law

- There are two primary methods used to quantify particle size of soils by sedimentation.
- **1- Pipette sampling method:**
- This method is the most accurate in determination particle size distribution and is a direct sampling of density of the solution. The amount of solid in sample is found by evaporating the water by oven-drying, then weighing the residue. The ratio of the weight, W, of particles present in that volume (20 ml) at a time t , divided by initial weight , W_0 , of a particle is equal to its percentage. Usually the silt and clay are sampled while sand is estimated by the weight difference.
- **2- Hydrometer method:**
- Less accurate than the pipette sampling method, but it is a rapid method in which the density of suspension is measured with a special hydrometer. The hydrometer method also depends upon stokes law.

- The concentration **(g/l), C,** of the soil in the suspension can be calculated from the following equation:
- $\boldsymbol{\rho_s = \rho_w + C[1-(\rho_w/\rho_p)]}$
-
- ρ_s: the density of the suspension (**reading of the hydrometer**).
- The stem of the hydrometer is read in density units(g/L on a scale to 60) or in percentages (100 C/C_0) when initial concentration , C_0 , is equal to 0.04Mg/m^3 and ρ_p is 2.65 Mg/m^3
- Some precaution should be kept in mined:
- 1- It is a calibrated method.
- 2- It gives an average reading of the suspension over the size distribution curve.
- 3- Settling of the soil particles on the shoulders of the hydrometer introduces uncertainty in the results.

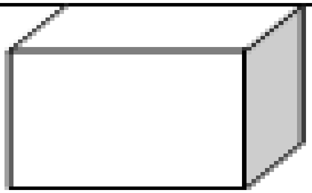
3.6 Specific surface

- The large amount of surface per unit mass is a characteristic property of all dispersed systems. The extent of the surface of a dispersed system is usually expressed in the terms of *specific surface*
- This term is defined as the amount of surface area per unit mass or volume.
- s_m = surface area/ mass (m^2/kg)
- s_v = surface area / volume (m^2/m^3)
- $\mathbf{s_m . \rho_p = s_v}$

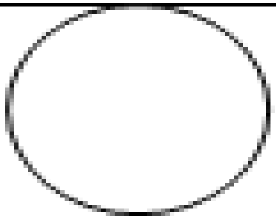
Relation of surface area to particle size and shape:

Specific surface varies with the shape of the particles as following:

CUBE **SPHERE**



$S_v = 6L^2 / L^3 = 6/L$



$S_v = 4\pi r^2/(4/3)\pi r^3 = 3/r = 6/d$

Thus, it can be seen that the specific surface increases directly as the size of the cube decreases.

Plate – shaped **Plate – shaped disk:**

r_p = radius of the particle

z_p = thickness of the particle

Volume = $\pi r^2 z_p$.

$S_v = 2(Lw+wd+Ld)/Lwd$

usually $w = L$, therefore $S_v = 2/d+2/L+2/w$

$= 2/d+4/w. \approx 2/d$

The term d is very small compared with w or L in the Cube, therefore platy shape has a very high specific

Surface area $= 2\pi r_p^2 + 2\pi r_p z_p$.

<u>**thus**</u> ,

$s_v = 2\pi r_p(r_p+z_p)/\pi r_p^2 z_p = 2/z_p + 2/r_p$

- knowledge of the particle size distribution can allow us to approximate the specific surface by summation equation.specific surface on a mass basis is
-

$$S_m = \frac{6}{\rho_p}\sum_{i=0}^{i=n} C_i(d_i^2 / d_i^3) = \frac{6}{\rho_p}\sum_{i=0}^{i=n}(C_i / d_i)$$

Where,

- C_j: the mass fraction of the particles
- d_i : the average dimeter (m)
- ρ_p: the average particle density (kg/m^3) and
- n: the number of soil separates.

Important:

- Many of the physical and chemical properties of the soil are associated with surface activity. Large particles present a significantly smaller area for the retention of water, mineral elements, and organics than do small particles.
- The amount of contact per unit surface also changes with shape. Each sphere can make six – point contacts when placed in a close – packed arrangement. A disk , or plate , will make intimate contact with about ¾ of its flat surface area
- The results of comparisons of various particle shapes show that:

1- A sphere has the smallest surface area per unit volume

2- A cube and a sphere are of the same specific surface.

3- If a sphere is deformed into a rod or into a plate , the surface area increases , with the plate-like particles exhibiting the greatest surface area, (example : clay).

Ch4: Soil Structure

- **Soil Structure:** the size, shape, and arrangement of aggregates and pores of the soil. It is sometimes called soil architecture.
- **Aggregate**: a group of primary soil particles that cohere to each other more strongly than to other surrounding soil particles.
- **Ped:** a unit of soil structure formed by natural processes, in contrast to a clod, which is formed artificially (disturbance like plowing).
- Structure has considerable effect on: Water movement, heat transfer, aeration, strength and erodability.
- Soil structure is not a plant growth factor, but it influences practically all factors that affect plant growth in soil.

4.1 Classification of Soil Structure:

- These classification systems are based on the ***shape and arrangement (type)***, the size (Class), and the **distinctness and durability (grade)** of the soil particles and pores.
- **a- Structure Type(Shape) :**
- **Plate-like:**
- Horizontal plates.
- Founded in the surface layers of virgin soils but may characterize the lower horizons as well.
- The platy types are often inherited from soil parent materials, especially those deposited by water or ice.
- Platy structure can also be created by compaction of silt and clay soils by heavy machinery.
- *Reduce the redistribution of water and air and penetration of roots*.

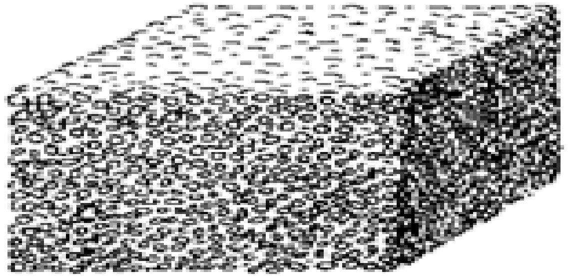
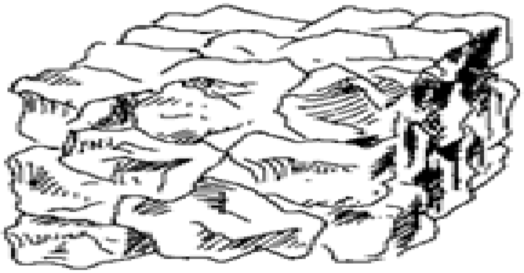
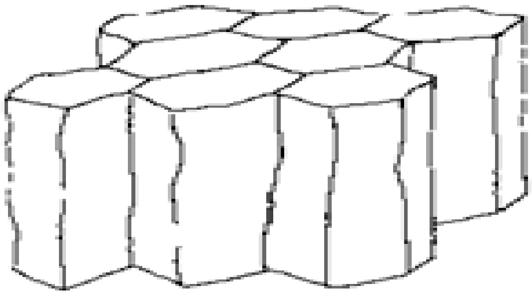
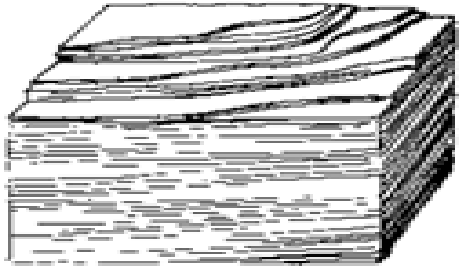
4.1 Classification of Soil Structure:

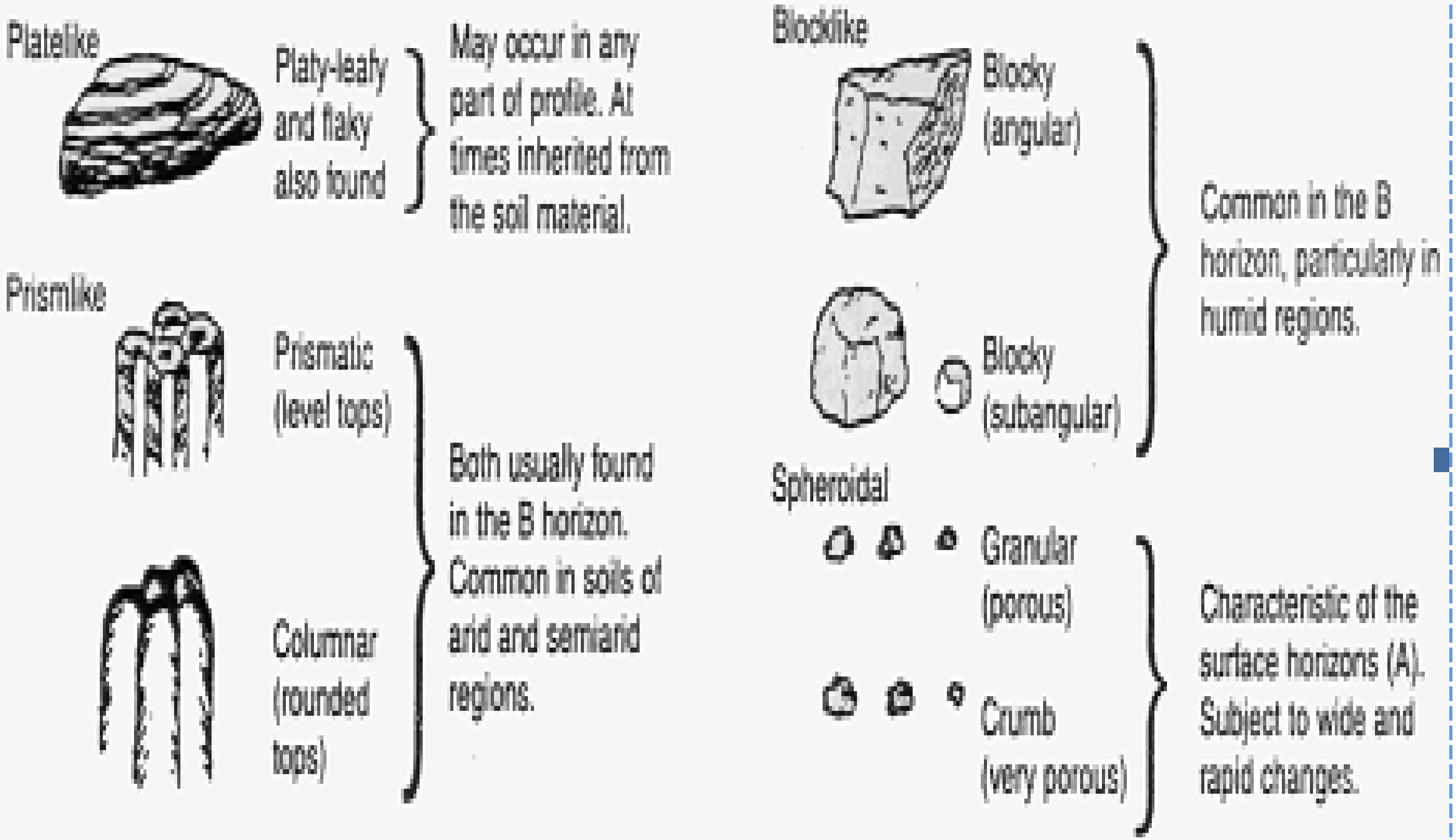
- **B. Prism-like:**
- Vertically oriented aggregates. Diameter of 15 cm or more.
- They commonly occur in B-horizons in arid and semi arid regions; they also occur in some poorly drained soils in humid regions and often are associated with the presence of semctitic clays.
- When the tops of prisms are rounded, it is called **columnar.** Common in subsoil’s high in sodium (i.e., natric horizons).
- Prismatic structure is moderately permeable: columnar structures are slowly permeable.

4.1 Classification of Soil Structure:

- **C. <u>Spheroidal:</u>**
- Curved, irregular shape
- They usually lie loosely and are separated from each other.
- When the granules are especially porous, they are called **crumbs.**
- Granular and crumb structures are characteristics of many mineral soils in the *A-horizon*, particularly those relatively high in organic matter.
- *Prominent in soils in grassland and pasture and in soils that have active earthworm populations*
- The peds are between 1 to 10 mm in size and rounded in form.

4.1 Classification of Soil Structure:

Granular	**Blocky**
	
Columnar	**Platey (Massive)**
	



May occur in any part of profile. At times inherited from the soil material.

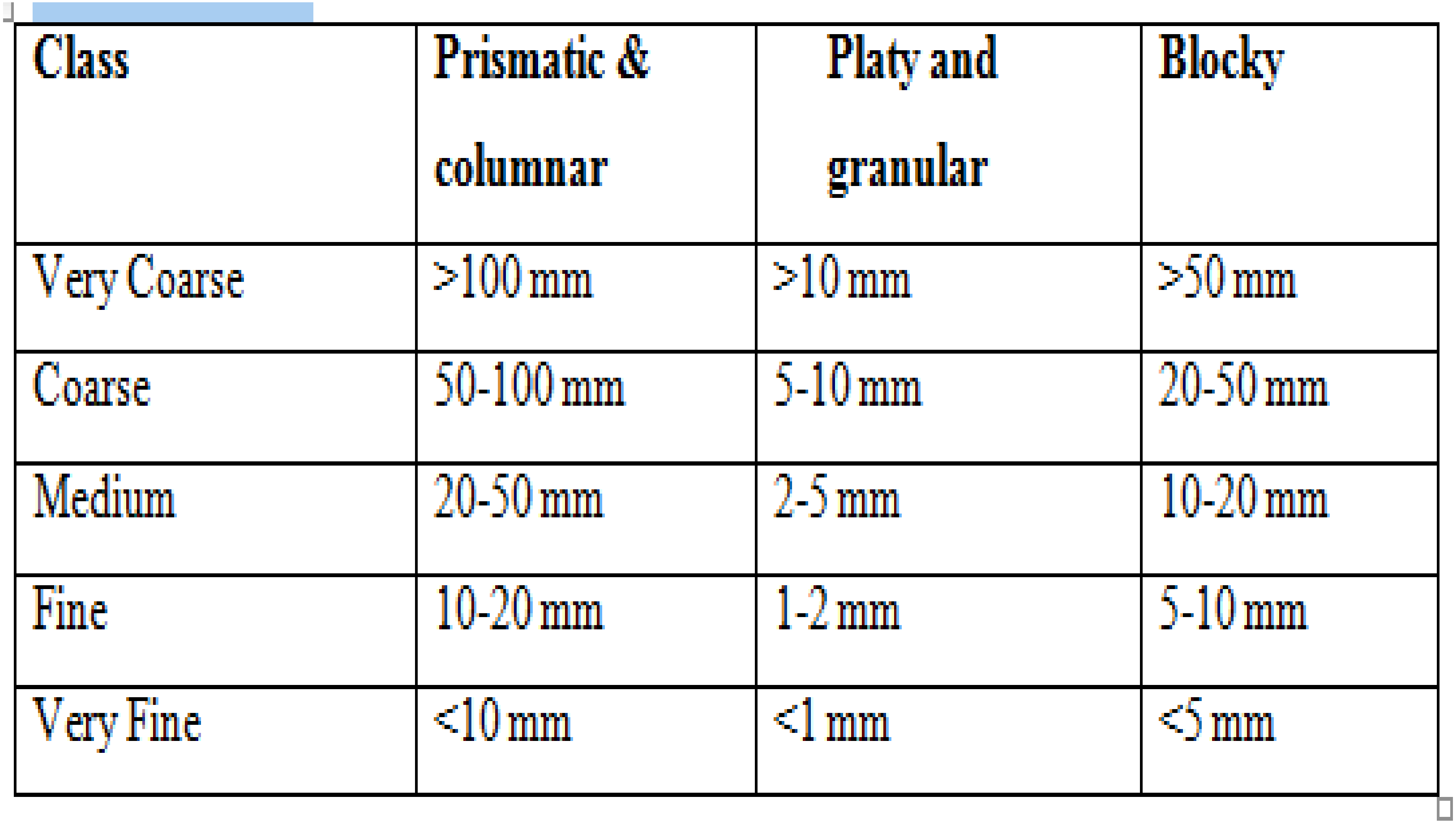
Both usually found in the B horizon. Common in soils of arid and semiarid regions.

Common in the B horizon, particularly in humid regions.

Characteristic of the surface horizons (A). Subject to wide and rapid changes.

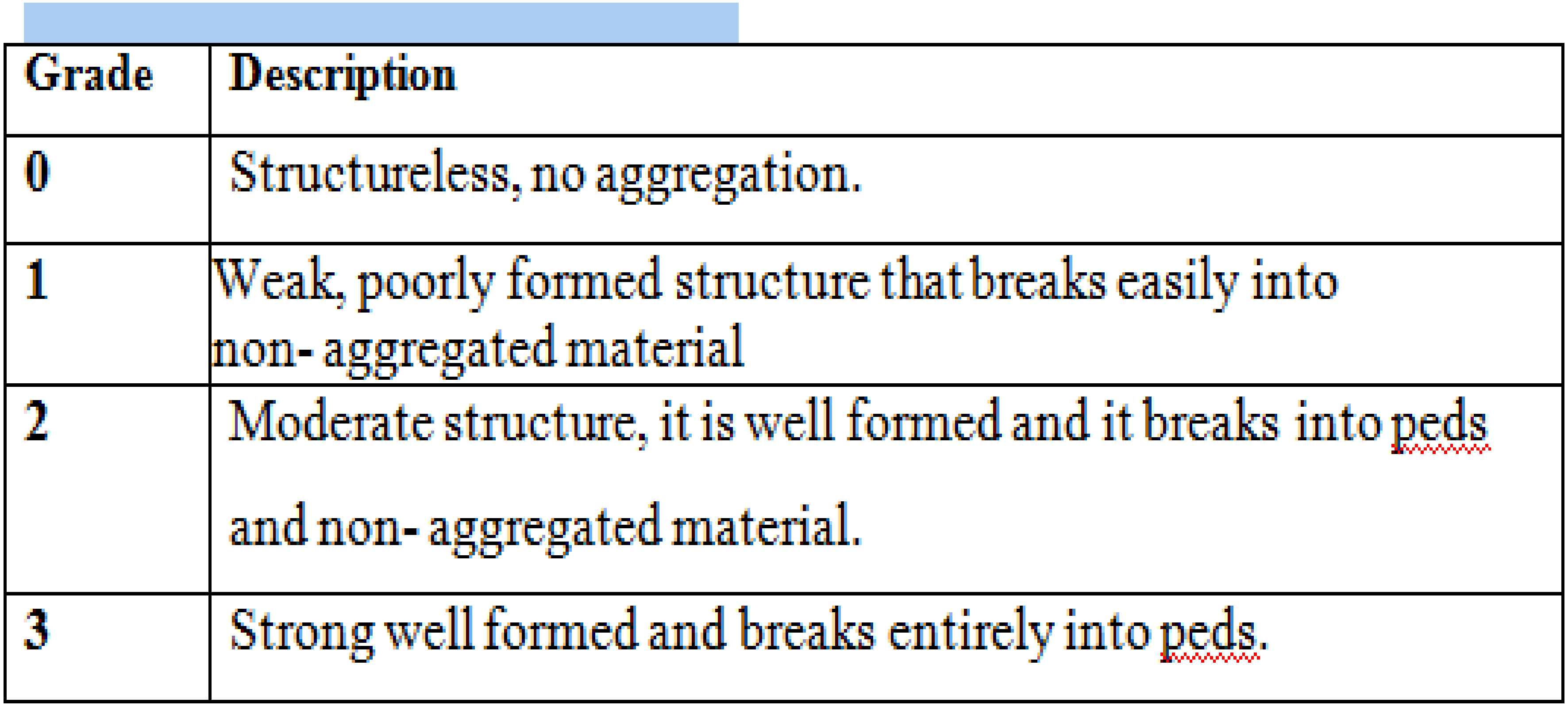
Types of soil structural units (peds) and their accompanying subtypes.

b- Class (size):



Class	Prismatic & columnar	Platy and granular	Blocky
Very Coarse	>100 mm	>10 mm	>50 mm
Coarse	50-100 mm	5-10 mm	20-50 mm
Medium	20-50 mm	2-5 mm	10-20 mm
Fine	10-20 mm	1-2 mm	5-10 mm
Very Fine	<10 mm	<1 mm	<5 mm

C- Grade (durability and stability):

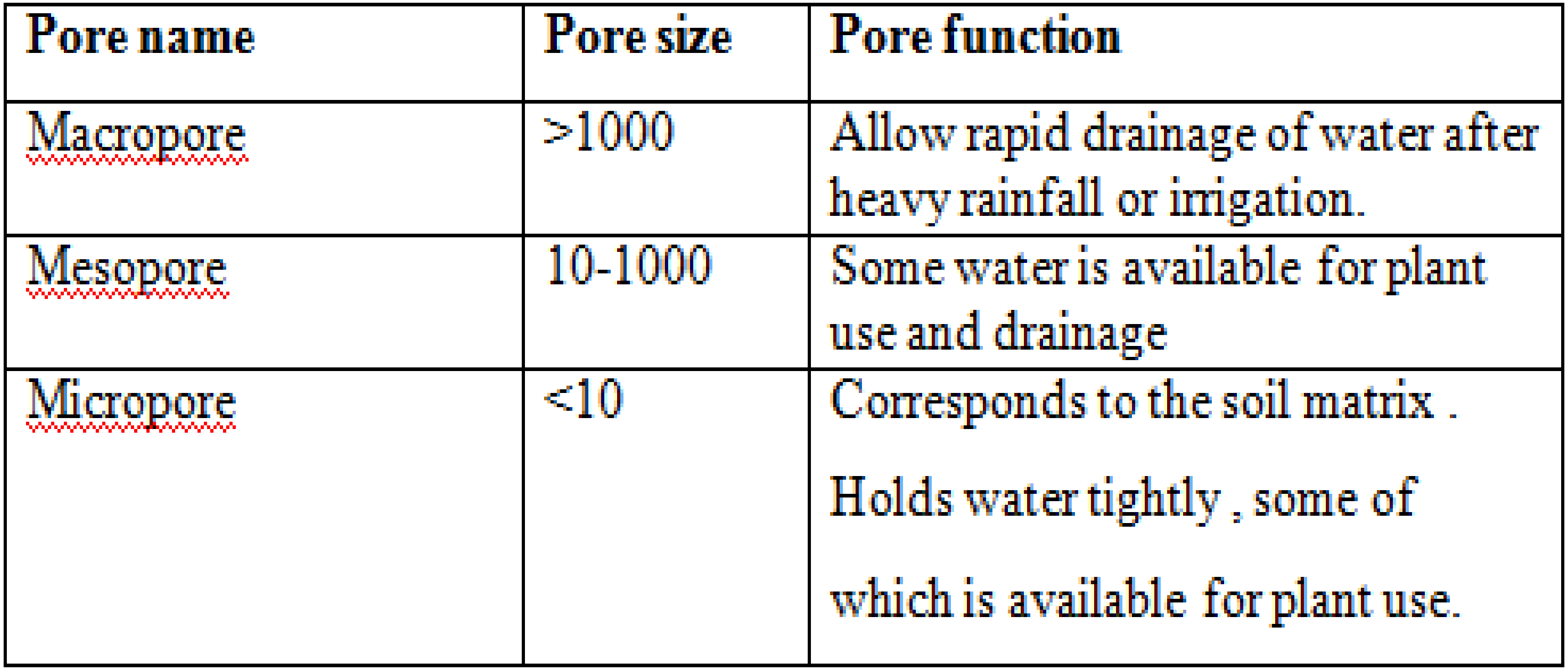


Grade	Description
0	Structureless, no aggregation.
1	Weak, poorly formed structure that breaks easily into non- aggregated material
2	Moderate structure, it is well formed and it breaks into peds and non- aggregated material.
3	Strong well formed and breaks entirely into peds.

Soil structure is expressed in order of Grade, Size, Shape.
<u>Example</u>: moderate, fine, sub-angular blocky structure”.

4.2 Soil Pores:

- Soil pores influence the ability of soils to support plants, animal, and microbes.
- Luxmoore (1981) arbitrarily classified pore size into three groups as following:



Pore name	**Pore size**	**Pore function**
Macropore	>1000	Allow rapid drainage of water after heavy rainfall or irrigation.
Mesopore	10-1000	Some water is available for plant use and drainage
Micropore	<10	Corresponds to the soil matrix . Holds water tightly , some of which is available for plant use.

4.3 Genesis of Soil Structure:

- The size, quantity and stability of aggregates recovered from soils reflect an environmental conditioning that includes factors that enhance the aggregation of soil particles. These factors and processes can be summarized as following:

1. physical processes:

- Wetting and Drying, cause swelling and shrinkage stresses that result in the weakening of the larger clods and development of shrinkage cracks in clods and in fresh sediments. Furthermore , water flowing between separated surfaces can deposit clay. Surface modified by deposition of clay are called *cutans*.
- Freezing and Thawing: affect aggregation by the expansion of water as it changes to ice within the profile. Under field conditions, freezing is usually not rapid, which allows considerable movement of water upward toward the lower soil temperatures and also toward any ice formed as water potential in contact with it decreases. Cracks enlarge by expansion on freezing. Continuous freezing may result in platy structure.

- **2. Chemical processes:**
- Soil chemical proceeses important to soil structure formation involve :
  - Clay and exchangeable ions.
  - Inorganic cements.
  - Organic compound and cementing agents.
  - Ions such as Ca, Mg, and K ions have a flocculating effect on clay, while H and Na ions have a dispersion effect.
  - Sesquioxides form irreversible and slowly reversible colloids and help to form water stable aggregates.
  - Fats, waxes, lignin, proteins, resins, and certain other organic compounds have direct stabilizing effect.
  - Organic compounds are effective in inducing aggregation and in stabilizing structure by forming irreversible or slowly reversible cementing colloids.
  - Soil amendments of *Sewage sludge* are believed to have the effect of organic compounds and improve soil structure. However, studies are still going on these amendments (Restrictions on the use of sludge were indicated in Jordanian Standards that recommend the use of sludge for forestry and in arid areas).
  - Long C-chain compounds and polymers might be used to bind clay particles and enhance aggregation. These are called *soil-conditioners*.

3. Biological processes:

1. Roots:

- Extract water and cause shrinkage and cleavage planes.
- They leave behind organic residues which enhance aggregation and root channels when they die.

2. Soil animals move material, create burrows.
3. Microorganisms decompose plant and animal residues, leaving humus.

- The other biological agent that creates structural pores and channels is the earth worm. In many studies, an absence earth worm was considered as environmental indicator for soil pollution and degradation.

4.4 Evaluation of Soil Structure and aggregate Stability:

- There are no theoretical methods developed so far to quantitatively express soil structure. The structure of soil is characterized indirectly by soil physical and morphological techniques.
- **1- Soil Physical Techniques:**
- **Total porosity and bulk density**
- Total porosity in a soil varies depending on: the texture and organic–matter content, the depth of the profile, soil management.
- **b. Pore size distribution**
- Pore size distribution is strongly influenced by aggregation of soil particles as well as by soil texture. The size distribution of soil pore is determined from the water retention curve (also known as characteristic curve). The equivalent radius of the largest pore that will be filled with water is a function of the soil water pressure through the capillarity equation. Mathematically, this is express as
- $\mathbf{r = (2\sigma \cos\theta)/(\rho_w g h) = 0.15 / h \qquad (h < 0)}$
- Where **r** is the equivalent radius of the pore (cm), **σ** is the surface tension (kg / s^2)
- = (Dyne/cm), **θ** is the contact angle between the water and the pore wall (usually assumed to be zero), $\boldsymbol{\rho_w}$ is the density of water (kg /m^3), **g** is the gravitational acceleration (m/ s^2), and **h** is the soil water suction (cm of water).

- Macropores can retain water only at low negative soil water pressure.
- In applying equation (**2**), the following assumptions are made:
- The soil is rigid and does not change its volume with changes in the soil water content.
- The pores are tubes of circular cross section
-
- **Important:** *Derivation of this equation and its implication will be discussed in more details in the chapters of soil water.*
-
- **c. Effective porosity, f_e**
- Associated with each size class of pores can be calculated from equation (2) and Poiseuille's equation by
- $$f_e = (8\,\eta I_m)/(g\rho_w r^2)$$
- Where **η** is the viscosity of water (kg / m* s) , **I_m** is the macropore flow (m/ s) , and **r** is the minimum radius of size class . Values of **I_m** were calculated as difference between infiltration rates at saturation and –2 cm pressure. This calculation results is upper bound estimates of the macroporosity associated with size classes.

- The number of effective pores, N per unit area for each size class is given by:
- **$N = f_e/(\pi r^2)$**
- **Assumptions:**
- Laminar flow of water.
- Smooth, cylindrical pores.
- Pore radius determined by capillary theory.
-
- **d. Aggregate Analysis**
- One of the most widely used techniques for assessing soil structure is the analysis of the size distribution and water stability of aggregates.
- It is a measure of the resistance of aggregates to breakdown when subjected to destructive force.
- **Mean weight diameter (MWD)** = $\Sigma (w_i d_i) / w$
- w_i: is the weight of each soil content at i sieve size,
- d_i: is the mean diameter of i sieve ,
- w: is the total weight of sample (or 100 percentage weight).
- This is done by dry sieving to measure the resistance for erosion by wind, or wet sieving to measure the resistance of aggregates to the water crusting force

d- Infiltration Rate:

- Infiltration rate of water can be expressed mathematically by the empirical power equation:
- $\boldsymbol{i = at^{-b}}$
- Where ***i*** is the infiltration rate (mm/h) , the volume of water infiltrating a unit soil surface area per unit time t (h) , and the parameters **a** and **b** are statistically fitted to the data that reflect soil properties.
- \- The above equation can be applied to a short time infiltration experiments.
- The condition of soil surface has a significant influence on the infiltration rate of the soil and therefore, soil structure might be evaluated from this physical property of soil.

2- Soil Morphological Techniques:

- Soil profile descriptions typically include a field evaluation of soil structure that includes sizes, shapes, and degrees of development of pores and of natural aggregates such as the peds. Mostly qualitative in nature, One morphological techniques used for soil structure evaluation is to develop and examine soil thin sections.

Ch5. Fate and Transport of Mass and Energy

- **5.1 Introduction**
- Transport of water, heat, oxygen, and solutes such as fertilizers, pesticides, heavy metals, and salts has practical significance for both agriculture and environment. The description, measurement, prediction, and possibly management of mass and energy transport in soil are important concerns in soil physics. We are primarily concerned with the prediction of the ***mass***, ***temperature***, and ***velocity*** variations (or distributions) within the medium such as soil.
- We use two sets of equations:

1. Conservation or balance equations.

2. Rate equations or flux laws.

The approach taken is that the transport phenomena of mass and energy are <u>analogous</u>: that is, there are similarities in mathematical form and in physical mechanisms.

- **Conservation of mass and energy:**
- We begin our discussions by defining a system, then we derive the balance equation of a unit volume of soil .

Definition of a system:

- ***A soil system*** is defined as any arbitrary portion or whole of the universe as specified by scientist. The remainder of the universe is called the surrounding. Therefore, a system has distinct boundaries.
- a three – dimensional component idealizing a system can be represented as shown in figure 5.1 .

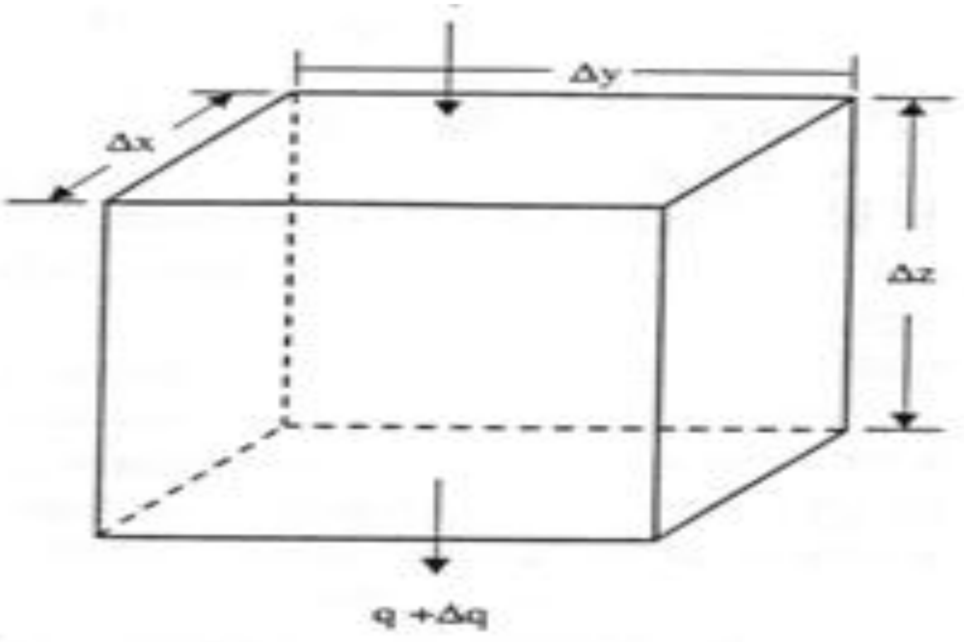
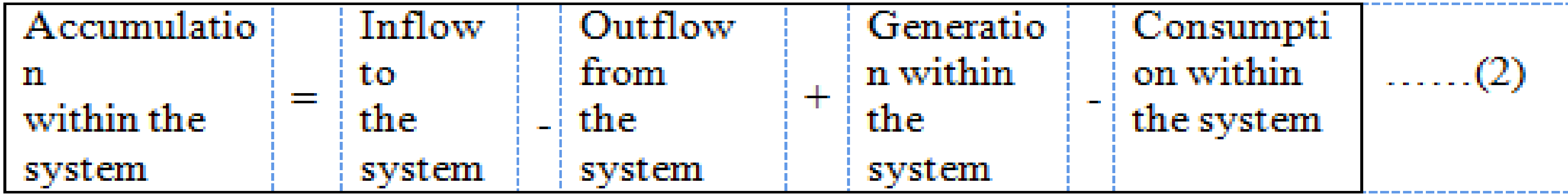


Figure 5.1. Schematic diagram of a unit volume element of soil.

- A **closed system** is one in which there is no such transfer of mass and/or energy during the time interval of interest.
- An **open system** is one in which material is transferred across the system boundary
- Most agricultural and environmental systems are open.
- There are two other ways to categorized open systems:
- **<u>1- A steady – state system</u>**
- Occurs when there is no change in the storage within the system.
- The flow rate is constant: **Rate in – Rate out = 0**…………………………………….(1)
- **<u>2- A transient – state system</u>**<u>:</u>
- Occurs when there is storage and/ or consumption within the system.
- Most transport systems in the field environment are transient – state system

5.2 Balance equation

- The generalized law of the conservation of mass is expressed as a balance equation.



Accumulation within the system	=	Inflow to the system	-	Outflow from the system	+	Generation within the system	-	Consumption within the system	……(2)

Generations refers to all production within the system. •

•Consumption refers to all degradation.

- Equation (2) can be expressed in near the soil sueither <u>integral</u> or <u>differential</u> form.

- Equation (2) reduces to different equations as following:

•No generation (or consumption) of mass and energy within the system:

Accumulation = Inflow – Outflow …………………………..(3)

•No accumulation: **Inflow = Outflow** ……………………………….(4)

•If there is no flow in and out of the system, (enclosed isolated systems).

Accumulation = generation - consumption…………………(5)

- Most of phase conservations occur at or near the soil surface.
- Applications of the principle of mass and energy conservation to the inflows and outflows in the soil profile enable a balance equation to be written for the soil and/or component of soil.
- **Example: part of mass balance calculations (page 92).**

Depth Interval (cm)	Bulk Density (g/cm^3)	Concentration (µg /g)	Amount (µg)
0-5	1.24	1.7	10.54
5-10	1.32	1.9	12.54
10-15	1.35	2.3	15.53
15-20	1.4	1.8	12.60
20-25	1.45	1.0	7.25
25-30	1.47	0.1	0.74

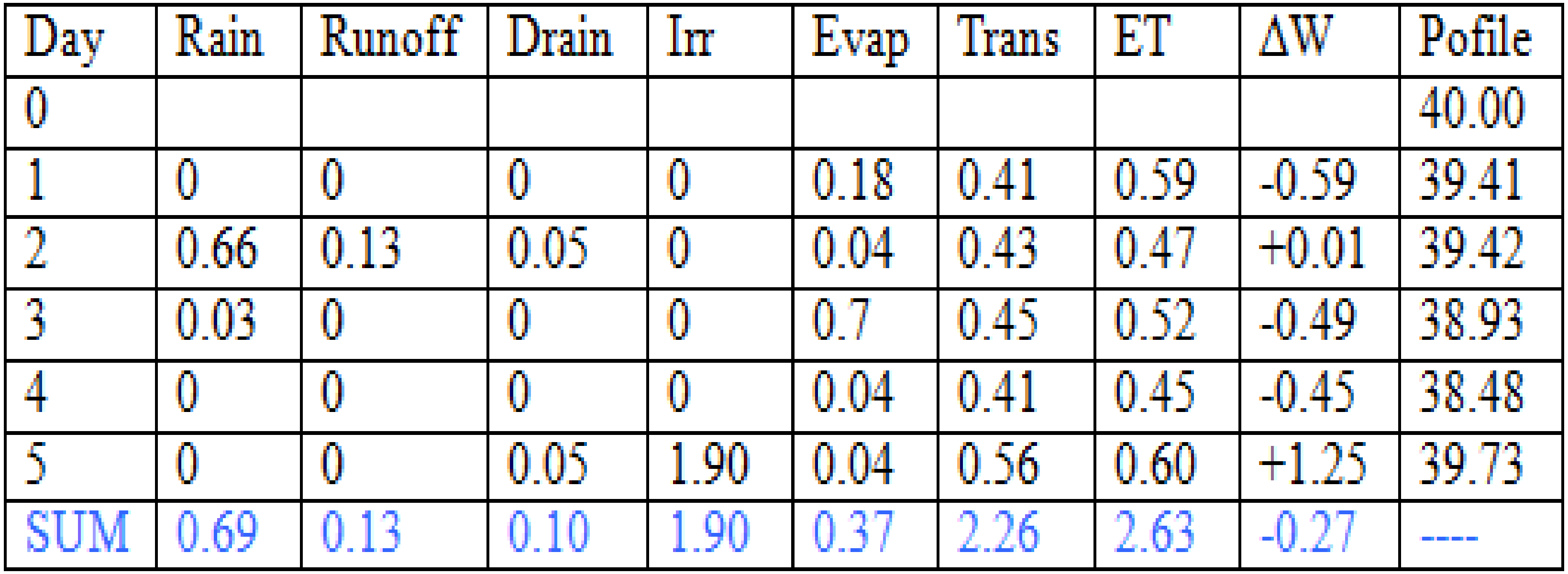
- The total amount of pesticides in the 0-30 cm depth interval is 59.20 μg/ cm².this amount can be determined from:
- Total = 5 or = 59.2 μg/ cm² ……………………………………….(6)
- **5.2.1 Water balance:**
- The water balance equation can be written by considering the possible fate of the precipitation received on a certain time. The equation is concerned with a volume of soil contained by imaginary surface. The depth of the soil volume can be chosen arbitrarily
- $\mathbf{P+I \pm R - ET \pm D = \Delta W}$………………(7)

- **Where:**
- **P**: is the precipitation recived in the area (of any size) for which the balance is being considered
- **I**: is the irrigation
- **R**: is the net surface runoff
- **ET**: is the evaporation
- **D**: is the net drainage .
- **Δw**: is the change in storage of water .
- The input is P and I the outputs are R, ET, and D

- Over a short time period, however, changes in storage will be significant, so the input and output will usually not be equal and the system will be at a transiet state.
- The system boundaries (i.e the volume of soil and the time interval) are defined. For example, if water balance is to be calculated for soybeans then the depth of the root zone as well as the area of the field should be stated.
- To calculate rates of processes, the duration between the measurments must known.

Example: water balance calculations (page 94)

Suppose that we obtaiened the following data for daily water balance (quantities are in centimeters).



Day	Rain	Runoff	Drain	Irr	Evap	Trans	ET	ΔW	Pofile
0									40.00
1	0	0	0	0	0.18	0.41	0.59	-0.59	39.41
2	0.66	0.13	0.05	0	0.04	0.43	0.47	+0.01	39.42
3	0.03	0	0	0	0.7	0.45	0.52	-0.49	38.93
4	0	0	0	0	0.04	0.41	0.45	-0.45	38.48
5	0	0	0.05	1.90	0.04	0.56	0.60	+1.25	39.73
SUM	0.69	0.13	0.10	1.90	0.37	2.26	2.63	-0.27	----

Substituting the sum of water balance categories into equation (6) results in

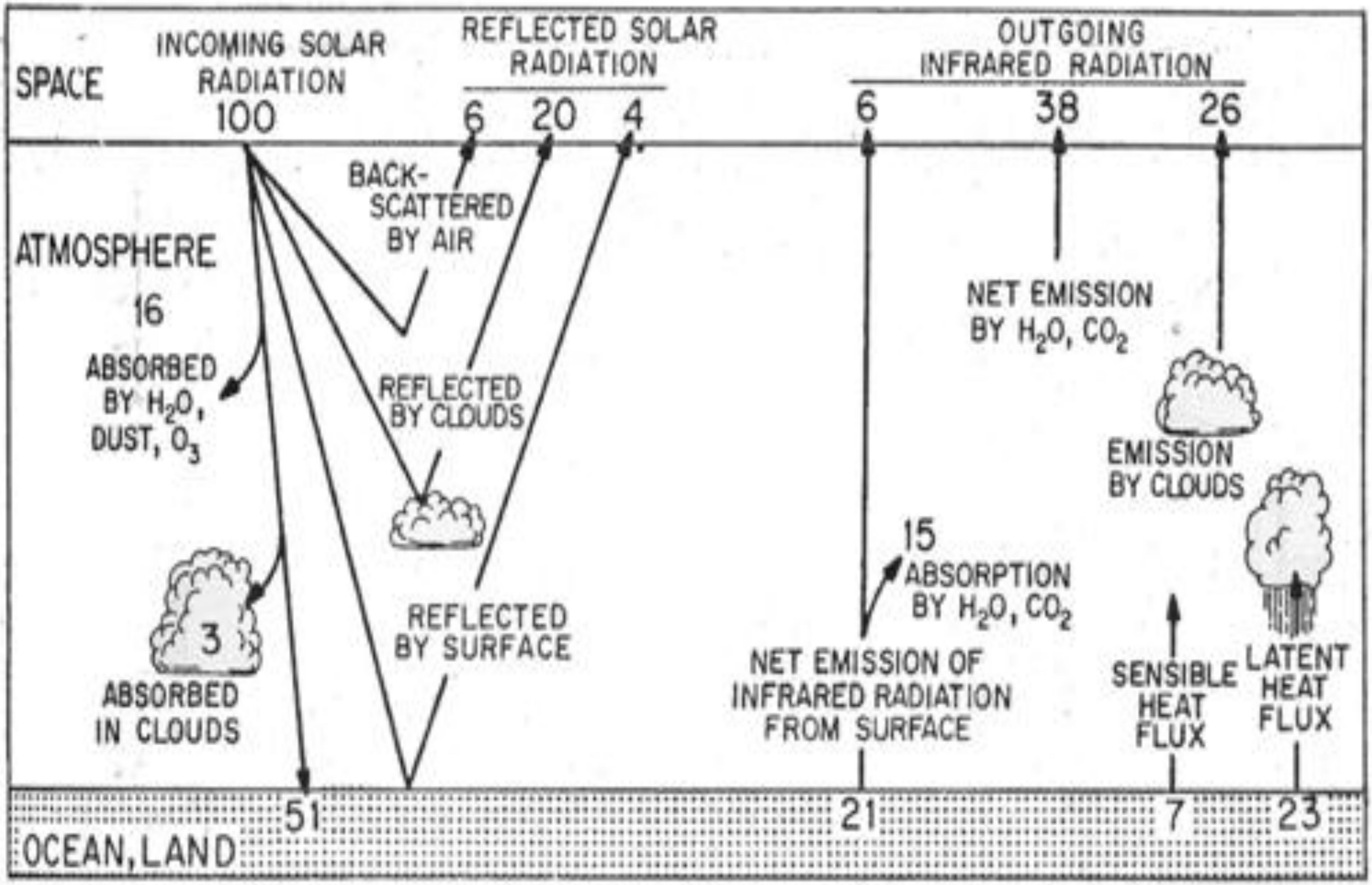
$$0.69 - (0.13+0.10) + 1.90 – 2.63 = - 0.27 \text{ cm}$$

$$= 39.73 – 40.00 \text{ cm.}$$

5.2.2 Energy balance:

- Many of the important soil physical properties are directly influenced by heat energy received at the surface. The energy balance equation is also a detailed statement of the law of conservation of energy (figure 5.3), mathematically it is stated as:
- $\mathbf{R_n = H+LE+M-G}$

- **<u>Where:</u>**
- **Rn**: is the net radiation flux density ($MJ/m^2.d$)
- **H:** is the energy flux density utilized in heating the air (sometimes called the sensible heat)
- **LE** :is the energy utilized in evaporation where **L** is the latent heat of vaporization (2.45 MJ/kg) and is the conversion factor needed to convert E to energy units and **E** is the amount of water evaporated from the soil ad transpired by crops (usually called the evapotranspiration).
- Note: **ET = LE/Lρ$_w$**
- **M**: is the energy flux density utilized in miscellaneous energy terms such as photosynthesis and respiration.
- **G: is** the energy utilized in heating the soil, water, and vegetation ($MJ/m^2.d$).
- Solar radiation provides almost all of the energy received at the surface of the earth, **the solar constant radiation reaching at the top of the atmosphere = 1375 Wm^{-2}.**


SPACE
INCOMING SOLAR RADIATION
100
REFLECTED SOLAR RADIATION
6
20
4
OUTGOING INFRARED RADIATION
6
38
26
BACK-SCATTERED BY AIR
ATMOSPHERE
16
ABSORBED BY H_2O, DUST, O_3
REFLECTED BY CLOUDS
NET EMISSION BY H_2O, CO_2
EMISSION BY CLOUDS
15
ABSORPTION BY H_2O, CO_2
3
ABSORBED IN CLOUDS
REFLECTED BY SURFACE
NET EMISSION OF INFRARED RADIATION FROM SURFACE
SENSIBLE HEAT FLUX
LATENT HEAT FLUX
51
21
7
23
OCEAN, LAND

- **The annual mean global energy balance for the earth-atmosphere system. (Numbers are given as percentages of the globally averaged solar irradiance upon the top of the atmosphere).**
- All radiation arriving at the soil surface is given a positive sign and all radiation reflected from or emitted by the surface is given a negative sign.
- Evaporation from the soil surface is considered positive, and condensation is considered negative.
- In the flow of soil heat below the soil surface (either with G or H). The upward z direction is taken as positive and downward z is taken as negative.
- **The units of energy flux density are expressed as W/m^2 or $J/m^2.s$.**

- **Example:**
- Calculation of the energy balance:
- The measured components of the energy budget were Rn = 9.0 MJ/m^2 *d, H= 1.2 MJ/m^2 *d, and G = - 0.4 MJ/m^2 *d . Find the energy used in evapotranspiration. LE and calculate ET, the evapotranspiration rate.
-
- LE = R_n - H+G = 9.0-1.2 +0.4 = 7 .4 MJ/m^2 *d
- LE /R_n = 7.4/9.0 = 0.82
- ET = LE/L.ρ_w = 7.4 MJ/m^2 *d /(2.45 MJ/kg * 1000 kg water /m^3)
- = 0.003 m/day
- = 0.3 cm/day = 3 mm/day
-
- **Net radiation** is the difference between total upward and total downward radiation fluxes and is a measure of the energy available at the soil surface.

- Net radiation can also be expressed mathematically as:
- $\mathbf{R_n = R_{swbal} + R_{Iwbal} = R_{sg}(1 - P) + R_{Iwbal}}$
- Where:
- **R_{swbal}** : is the net shortwave radiation
- **R_{Iwbal}**: is the net longwave radiation (amount of absorbed minus the amount emitted)
- **R_{sg}**: is the global shortwave radiation arriving at the surface (direct beam plus scattered) measured with pyranometer
- **P**: is the albedo (fraction of global radiation reflected by the surface) measured with two pyranometers (one facing up and one facing down).
- The albedo for most soil and plant conditions in the field varies from 0.10 to 0.30 and for full canopy conditions averages about 0.23.
- Albedo tends to be lower for wet soil than for dry soil.

- ***The quantity Rsg(1- P):   is the shortwave radiation*** .
- When water does not limit transpiration, most of the energy in the form of R_n is utilized as energy in evapotranspiration, ET.
- Under humid conditions, the ratio LE/ Rn is approximately 0.8.
- When bare soil is wet, most of the energy goes into evaporation.
- As the soil dries, water becomes less available for ET, and more of the energy goes into heating the soil.

5.3 The continuity equation for one dimensional flow of water in the soil without extraction:

- Transport of mass and energy always obeys a balance equation, and mathematical derivations of the fundamental transport laws must reflect the balance, or conservation laws. One of the most important uses of a balance equation is for the equation of continuity. The derivation of the equation of continuity for one dimensional flow of water without extraction follows.

1-Assume that the volume element of soil can be represented as shown in figure 5.1 and that there are no consumptive and generative process operating.

2-Then in a given unit of time, the volume of water entering the volume element = $\mathbf{q\Delta x\Delta y}$, which is the flow rate (m^3/ s) per unit area.

3-The parameter **q**: is the flow rate of water per unit area and is known as the flux density with units of $m^3/m^2.s$.

4-The terms **Δx/ Δy**, and **Δz** each represent the length of a side of soil volume in the x, y and z directions, respectively.

5-The volume of water leaving the volume element = **$(q+\Delta q)\,\Delta x\Delta y$**.

-
- Therefore, according to the conservation equation (2), the volume of water accumulated in the volume element during Δt equals
- **Inflow – Outflow = net accumulation.**
- **$q\Delta x\Delta y\Delta t - (q + \Delta q)\Delta x\Delta y\Delta t = - \Delta q\Delta x\Delta y\Delta t$**
- The volumetric water content of the volume element is θv. Therefore,
- **$\Delta \theta_v$ = change in water content /volume of soil**
- **$= - (\Delta q\Delta x\Delta y\Delta t)/\Delta x\Delta y\Delta z$ or $\Delta\theta_v/\Delta t = - \Delta q/ \Delta z$**
- Taking the limit of the left – hand terms as Δt approaches zero and limit of the right hand terms as Δz approaches zero and recognizing that several independent variables are uncontrolled gives:
- **$\partial\theta v /\partial t = -\partial q/\partial z$**
- A similar procedure can be used for the derivation of the energy balance equation.

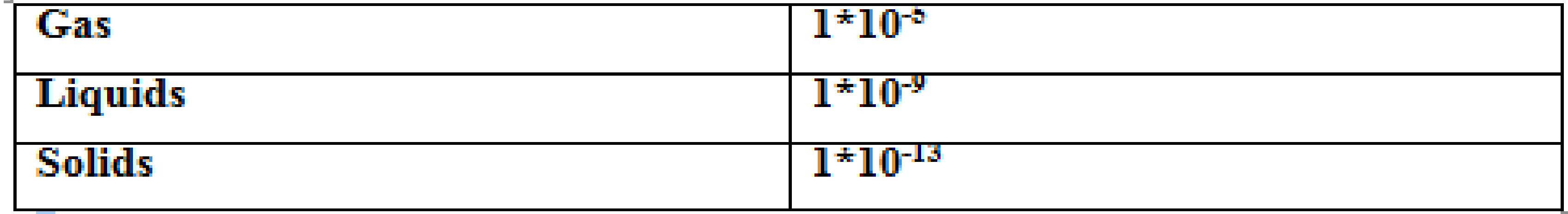
5.4 General Transport Mechanisms:

- Transport can occur by two mechanisms: **Diffusion and Convection**.
- In the case of heat (energy), three basics mechanisms are usually identified:
- **Conduction, Convection, and Radiation**.
- Convection, on other hand, is a mixture of conduction and transport of a fluid
- We will begin with a discussion of transport of mass under steady – state conditions.
- **Concentration, flux density, and influences of soil phase:**
- The concentration of a solute in soil is defined as a measure of the amount of that solute (Q) in a specific volume (m^3)
- **Mass per unit volume**
- The total solute concentration in soil can be defined with units of mass of solute per soil volume (kg / m^3), the total solute concentration C_t is:
- $\mathbf{C_t = \rho b C_s + \theta_v C_l + f_a C_g}$
- **Where**:
- **C_s**: is the solid – phase solute concentration (kg / kg of solid).
- **C_l**: is the solute concentration in water ($kg / m^3$of soil solution).
- **$C_{g:}$** is the solute concentration in the gaseous phase ($kg / m^3$of soil air)
- **-** The parameters ρ_b, θ_v, and f_a are the bulk density, water content by volume, and aeration porosity, respectively.

- **The mass flux**, **j** is defined as the amount of substances passing through a plane perpendicular to the direction z during a specified time interval, divided by the magnitude of the time interval, mathematically, the mass flux, j, is
- **j= Q/t**
- **Where**:
- **Q:** is the amount of substance,
- The **SI** units of mass flux are kilograms or moles per second
- Quite often we also see the flux density reported.
- **The flux density, J**, is the flux divided by the cross – sectional area and mathematically is:
- **J =Q/At**
- Where, **A**: is the cross sectional area of the plane perpendicular to the direction of flow. The units of mass flux density are kg/m^2.s or **moles per meter squared per seconds.**
- ****** The flux density is the preferred term in expressing the amount of mass transferred when the shapes of the boundaries are changing or are unknown.

- The transport of the same solute in soil is sum of the flux density in each phase, mathematically, this is
- $\mathbf{J_t = J_s + J_w + J_a}$
- Where:
- $\mathbf{J_{t:}}$ is the total flux density (in kilograms per square meter per second) .
- $\mathbf{J_{s,}}$ $\mathbf{J_{w\,,}}$ and $\mathbf{J_a}$ are the flux densities in the solid, aqueous , and gaseous phases , respectively . For almost all strongly sorbed solutes, transport in the solid or adsorbed phases is negligible and for practical purposes, J_s can be assumed to be zero.
- **5.4.1 Mass transport :**
- Examples include the diffusion of oxygen and nutrients to plant roots and convective movement of nitrates, chlorides, and pesticides within and below the soil profile.

- **Mass transport can occur by two physical mechanisms:**
- **1- Molecular diffusion:**
- Occurs as a result of concentration gradients
- **Table 5.1. Molecular diffusivities as a function of phase (m^2 / s)**



Gas	**$1*10^{-5}$**
Liquids	**$1*10^{-9}$**
Solids	**$1*10^{-13}$**

- Diffusion transport results from the movement of solute molecules from regions of higher concentration to regions of lower concentration, mathematically; this transport mechanism can be represented by Fick’s first law of diffusion as:
- **j = - D.A.ƏC/Əz**
- **<u>Where</u>**:
- **j** :is the solute flux and is the quantity of solute per unit time (kg / s) .
- **A**: is sectional area (m^2).
- **C** is the solute concentration (kg/m^3)
- **z:** is the spatial coordinate (m)
- **D:** is the transport coefficient, whose value depending on the nature of the solute and that of the diffusing medium. The parameter D is also known as the diffusivity or molecular diffusion coefficient and has dimensions of length squared per unit time (m^2 /s) .

- **The quantity ƏC/Əz** is known as the concentration gradient (kg/m^4)
- Diffusion results from concentration inequalities and occurs in the direction of decreasing concentration.
- It is a steady- state mass transport equation.
- Under conditions where the concentration gradient is I, transport by diffusion is 10.000 times faster in the gaseous phase than in aqueous phase, which in turn is 10.000 times faster than in the solid phase.
- **Example (page100)**
- A soluble, nonvolatile, pesticide (Kill All), was broadcast-applied to a given soil surface. After two weeks, the concentration of the pesticide at the 5 cm soil depth was 5 μg/L, while the concentration at the 10 cm depth was 15 μg/L. If the average soil water content was 0.25 cm^3/cm^3, the soil water flux was zero, and the diffusion coefficient for this pesticide in this soil was 4×10^{-4} cm^2/h, calculate or determine the following:

- 1- The concentration of Kill All in the soil at both depths.
- C_1 = (5 μg/L)(0.25 cm^3/cm^3)(1 L/1000 cm^3)= 0.00125 μg/cm^3 soil
- C_2 = (15 μg/L)(0.25 cm^3/cm^3)(1 L/1000 cm^3)= 0.00375 μg/cm^3 soil
- 2- The average concentration gradient in this horizon:
- $\boldsymbol{\partial C/\partial z}$ = (0.00375 – 0.00125)/[(-10cm)-(-5cm)] = -0.0005 μg/cm^4
- 3- The rate of diffusive movement in μg/cm^2.h and the diffusion rate in kg/ha.day
- $J_d = (4\times10^{-4}\ cm^2/h)(\ -0.0005\ \mu g/cm^4) = 2\times10^{-7}\ \mu g/cm^2.h$
- $J_d = (2\times10^{-7}\ \mu g/cm^2.h)(10^{-9}\ kg/\ \mu g)(10^4 cm^2/1m^2)(10^4\ m^2/1\ ha)(24\ h/day) = 4.8\times10^{-7}\ kg/ha.day$
-
- 4- The direction of transport of this herbicide in the profile
- Since the final answer is positive, the direction of flow is upward.

- **Notes on Fick's Law**
- Use the law as it is, if the final answer is:
- -ve: this means that $C_2 > C_1$, direction is from C_2 to C_1.
- +ve: this means that $C_2 < C_1$, direction is from C_1 to C_2.

- **2- Convection:**
- Occurs by bulk – fluid motion. In soils, the fluids are water and air. i.e mass flow.
- Convection transport is sometimes called ***advection*** and depends on the velocity of the moving fluid and the concentration of solute in the fluid.

- Mathematically, the flux density due to the convection can be expressed by:
- $J_c = q_z C_l$ (16)
- Where,
- **q_z:** is the flux density of water ($m^3$of water / m^2 .s of soil) moving through the soil in the z direction.
-
- **Note:** **q_z = pore water velocity * θ_v**
-
- **C_l:** is the concentration of solute in water (kg/ m^3) .
- **The ratio q_z/θ_v:** is known as the pore water velocity, v_z (m/s) and this is known as the unsaturated soil hydraulic conductivity where rate is function of soil water content.
- The system or phase should be well defined. Examples of these systems include the solution phase, the aqueous phase, the gaseous phase and the solid phase or entire soil.

- **Example (page101):**
- After 2 days of a heavy rain, the average concentration of Kill All in the AP horizon is 10µg /L. The rate of downward water movement through the horizon has a pore water velocity of 0.15 cm^3 / cm^2.h and the water content is 0.40 cm^3 /cm^3. What is the transport rate by convective flow of Kill All in kilograms per hectare per day? What is the direction of convective flow?
- $J_c = q_z C_I$ = (-0.15 cm^3/cm^2 .h) (10µg/L) (L/1000 cm^3) (0.40 cm^3/cm^3)
- (10^8 cm^2 /ha) (24 h/d) (1 kg/10^9µ g)
- = $-1.44*10^{-3}$ kg/ha.day

- Since the sign of q_z is negative, J_c is negative, and the direction of flows is downward.
- In soil, equation (16) does not accurately represent the total convection since the flow pathways are tortuous (i.e. not straight), and the velocity of water is faster in center of pores than at the edges and is faster through the larger pores. Therefore, $\mathbf{J_h}$, hydrodynamic dispersion, is usually used:
- $\mathbf{J_c = q_z C_I + J_h = q_z C_I - D_h \partial z C_I}$
- **Ch5. Fate and Transport of Mass and Energy**

5.4.1 Mass transport :

- **2- Heat (Energy) Transport:**
- Heat transfer (or heat) is energy in transit due to differences in temperature .we will consider three heat transfer processes: conduction, convection, and radiation.
- **1-conduction:**
- The transfer of energy from the more energetic to the less energetic particles and molecules in porous material due to the interactions between the particles and between the molecules.
- Conduction is the primary heat – energy transfer in the soil. It also acts to transfer energy up and down the stems of plants.
- Conduction is accomplished by the transmission of translational, rotational, and vibrational energy
- Conduction of energy in one dimension can be described by the steady – state forms of Fouriers law:
- $h = -k A (\partial T/\partial z)$ (18)

- **Where:**
- **h**: is the heat flux in J/s or W
- **ƏT/Əz**: is the temperature gradient in degrees per unit length.
- **The ratio ƏT/Əz**: is the slope of the temperature – distance relationship
- **k:** Thermal conductivity J/ m .s .K . = W/m.K.
- **Thermal conductivity** is defined as: quantity of heat transport through soil per unit temperature gradient. Also, it is the proportionality constant between the flux density h/A, and the temperature gradient **ƏT/Əz**.

Note 1: For most agronomic uses in soils, the value of k can be considered to be independent of temperature.

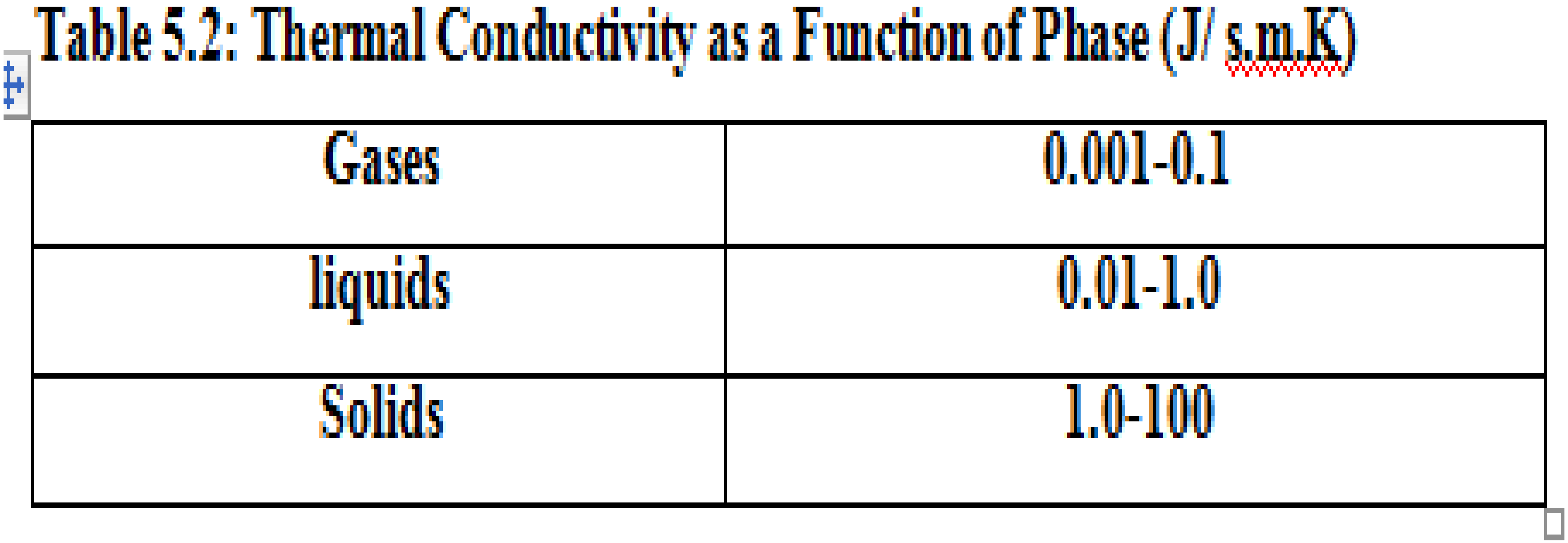


Table 5.2: Thermal Conductivity as a Function of Phase (J/ s.m.K)

Gases	0.001-0.1
liquids	0.01-1.0
Solids	1.0-100

Note 2:

- The heat flux density, H, is obtained by dividing the heat flux by the cross-sectional area. The units of H are W/m^{2}

- **Example:**
- Consider a 1-cm thick soil crust having a temperature at the surface of 35 ºC and at the bottom of 25ºC. If the thermal conductivity through the crust is 1.1 W/ m.K . Calculate the flux density of heat conducted through the soil crust. In which direction is the heat conducted? In incremental form.
- **Fourier's equation is:**
- **$H = - k (\Delta T/\Delta z)$** $= -1.1[(308 – 298) / 0.01] = -1100$ W / m^2
- Since the highest temperature occurs at the soil surface, heat is conducted down through the crust.

2-Convection:

- When a fluid such as water is at rest or in motion with a surface at a temperature that is different from the fluid, energy flows I the direction of the lower temperature. The energy flux leaving or entering the surface can be calculated from **Newton's law of cooling** as:
- $\mathbf{h = C_v\, Av(T_S - T_o)}$ **(19)**

Where:

- **T_s:** the temperature of the soil in contact with the fluid (K) .
- **T_o:** is the temperature of the fluid far away from the surface so that no influence of the surface is evident (k).
- **A:** cross sectional area (m^2)
- **v:** is the velocity of the fluid (m/s)
- **C_v**: is the volumetric heat capacity (J /m^3 .k)
- **h :** is the heat flux (J/s or W)

1- Convection of heat energy is extremely important in the atmosphere, where air is warmed or cooled as air masses come in contact.

2- Convection of heat energy is lesser importance in soils but may occur during the infiltration and redistribution of water from rainfall and irrigation.

- **3- Radiation:**
- All forms of matter emit thermal radiation; the emission may be attributed to changes in electron configurations of the constituent atoms or molecules. The energy of the radiation field is transported by electromagnetic waves (or photons).
- The energy travels as discrete packets called photons or quanta whose energy content depends on their wavelengths or frequencies.
- We can use the Stefan- Boltzmann law to obtain:
- $\mathbf{h = (\varepsilon\sigma T^4).A} \qquad \mathbf{(20)}$
- **where :**
- **σ :** is Stefan- Boltzmann constant = $5.675*10^{-8}$ $W/m^2.k^4$
- **ε**: emissivity of object = es/eb
- This equation gives the maximum energy flux that can leave a surface of area A at absolute temperature T.

- **Analogies between One – Dimensional Transport Phenomena:**
- We have used the same mathematical models to develop the steady – state flux laws for mass and energy in one dimension , this laws can be written as :
- **Ficks law :**
- $j = -DA(\partial C/\partial z)$ **or** $J = -D(\partial C/\partial z)$
- **Fouriers law :**
- $h = -KA(\partial T/\partial z)$ **or** $H = -K(\partial T/\partial z)$
- The transport rate equations take the same fundamentals form of :
- Flux density = (flow rate)/ area = Q /At.
- = - transport coefficient × potential gradient.

Where :

- **Q**: is the quantity.
- **A**: is the cross –sectional area.
- **t**: is the time
- **D** and **k** are the molecular transport properties.
- **C** and **T** are potentials.

- Although these equations are similar, they are not completely analogous because the transport coefficients have different units .
- In order to make these terms compatible, we define the thermal diffusivity as:
- $\alpha = (k / \rho_b) C_w = k / C_v$

where:

- **α:** is the thermal diffusivity in units of length squared per unit time (m^2/s)
- **k:** is the thermal conductivity (J/ m.s.K)
- **$\rho_{b:}$** is the bulk density (kg/m^3)
- **$C_{w:}$** is the gravimetric heat capacity (J/kg.K)
- **$C_{v:}$** is the volumetric heat capacity (J/ m^3.K)
- **ρ_b** and **C_w** normally are assumed to be constant within a horizon .

The various quantities in mass and energy transport are summarized in table 5.3

Table 5.3 Analogous Terms in the One – Dimensional Flux Laws:

Property	Mass	SI Units	Energy	SI Units
Flux	j	kg/s	h	J/s or Watt
Flux density	J	$kg/m^2.s$	H	$J/m^2.s$ or $W.m^2$
Transport property	D	m^2/s	k	$J/m^2.s.K$
Potential gradient	$\partial C/\partial z$	kg/m^4	$\partial T/\partial z$	K/m
Diffusivity	D	m^2/s	$\alpha = k/(\rho_b C_w)$	m^2/s
Concentration	C	kg/m^3	$\rho_b C_w T$	J/m^3
Gradient concentration	$\partial C/\partial z$	kg/m^4	$\partial(\rho_b C_w T)/\partial z$	J/m^4

Notes:

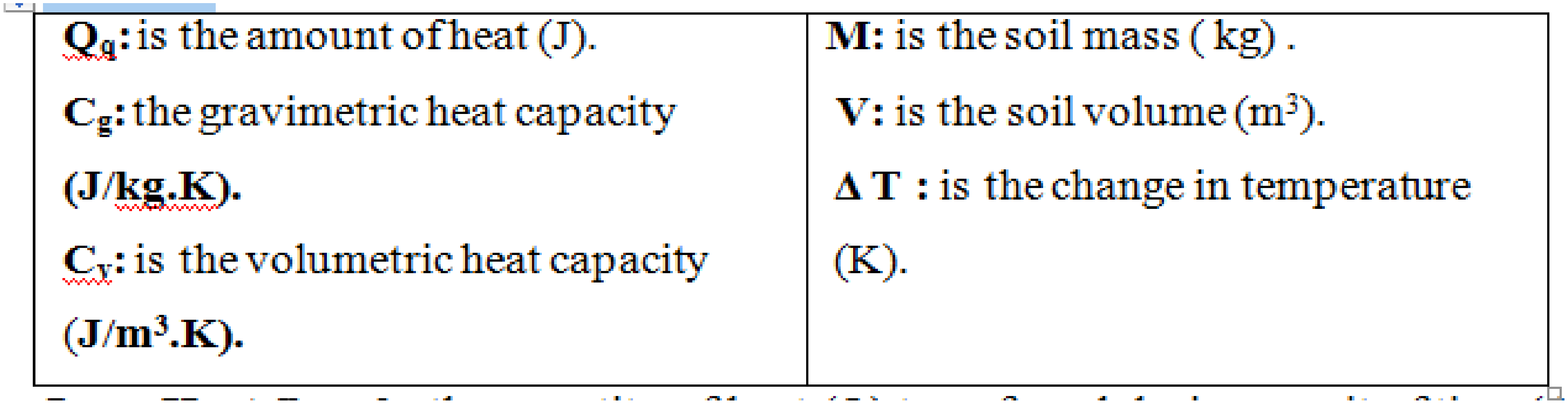
- The concentration term for energy is derived from **$\rho_b.C_{w.}T$ = kg/m^3 (J/kg.K) (K) = J/ m^3** = energy/ volume. This analogous with concentration of mass/volume.
- The flux laws can be written as flux density = -diffusivity * concentration gradient and also flux density = - driving force / resistance, where the driving force = concentration differences = C2 –C1. Resistance = distance / (transport coefficient * area). e.g. . $\Delta z/\Delta A$.

Ch6. Soil Temperature

- Temperature is one of the most important factors affecting the rates of physical, chemical, and biological processes in soil.
- **6.1 Classes of Soil Temperature (Soil temperature Regimes, page 109):**
- These classes are based on the mean annual soil temperature and given below:
- **Pergelic:** the soil has a mean annual temperature lower than0ºC.
- **Cryic**: the mean annual soil temperature is higher than 0ºC but lower than 8ºC.
- **Frigid:** this class of soils has warmer temperate in summer than those in Cryic regime.
- **Mesic**: the mean annual soil temperature is 8ºC or higher but lower than 15ºC, and the difference is greater than 5ºC at 0.5 m.
- **Thermic**: the mean annual soil temperature is 15ºC or higher but lower than 22ºC
- **Hyperthermic**: the mean soil temperature is 22º C or higher, and the difference between the mean summer and mean winter soil temperature is greater than 5º C at 0.5 m.
- When difference is less than 5ºC, the term is called iso (isomesic, isothermic…)

- **6.2 Thermal Concepts:**
- **Heat**, **Qq**: the kinetic energy of the random motion of the particles of which the soil is composed.
- **Temperature**: it’s the ability to transfer heat. It is a measure of the intensity of heat.
- **Heat capacity**: is the amount of temperature change in the soil in response to the absorption or release of heat.
- **Specific heat of materials**: the ratio of the heat capacity of the material to that of water and is dimensionless. **Gravimetric heat capacity**, $\mathbf{C_g}$: the amount of heat required to raise the temperature of 1 kg of soil by 1 degree Kelvin. In the SI system, the units are J/kg.K.
- **Volumetric heat capacity,** $\mathbf{C_v}$: the amount of heat required to raise the temperature of 1 m^3 of soil by 1 degree Kelvin. In the SI system, the units are $J/m^3.K$.

- **Relationships among thermal concepts:**
- $C_g \rho_b = C_v$
- $Q_q = C_g M \Delta T$
- $Q_q = C_v V \Delta T$
- **Where:**



Q_q: is the amount of heat (J).	**M:** is the soil mass (kg) .
C_g: the gravimetric heat capacity **(J/kg.K).**	**V:** is the soil volume (m^3).
C_v: is the volumetric heat capacity **(J/m^3.K).**	**Δ T :** is the change in temperature (K).

- **Heat flux, h**: the quantity of heat (Q) transferred during a unit of time (t), the SI units of heat flux are J/s.
- **Heat flux density**, **H**: the quantity of heat(Q) transferred during a unit of time (t) across a unit surface area(A) perpendicular to the direction of heat flow.The SI units of heat flux density are $J/m^2.s$ orW/m^2.
- **Thermal conductivity, k:** the ability of the substance to transfer heat from molecule to molecule. It's defined as the quantity of heat(J)transmitted a distance of 1m through a substance per unit cross section. Per unit of temperature gradient per second. The units of k are J/m.s.K= W/m.K
- **Thermal Diffusivity, α** is an expression of the rate at which a substance heats up as the result of a thermal gradient. It is the rate change of temperature with time and is proportional to the thermal conductivity and inversely proportional to heat capacity on a volume basis .The units of α are m^2/s.
- **Note:** most of these concepts were tackled in chapter 5, and the heat transfer mechanisms were discussed in details.

- <u>**6.2 Factors Affecting Soil Temperature:**</u>
- <u>**1-Environmental Factors:**</u>
  - **Solar Radiation:**
- The daily radiation flux density reaching the top of the atmosphere ranges between1350 and 1400 W/m^2 (solar constant).
- <u>The radiation flux that reaches the soil surface depends on:</u>

1- The angle at which the soil faces the sun.

The angle at which the soil faces the sun varies with latitude, season, time of day, steepness and direction of slope (aspect). In the northern, hemisphere, southeastern slopes are warmest during the early summer, southern slopes are warmest at midsummer, and southwestern slopes are warmest during the fall. Correspondingly, a west – facing slope is normally somewhat warmer than an east-facing slope.

2- Insulation:

- The earth is insulated by air, water vapor, clouds, dust, smog, snow, plants and mulch.
- In Temperate Zone, between 4.18 and 31.4 MJ/m^2 are received at the earth's surface per day. Approximately 24.3 $MJ/m^2$are required to evaporate a layer of water 1cm thick.
- ***Note***: 1 calorie = 4.184 J.

  1 cal/ cm^2 .min = 697.93 W/ m^2.
- **Radiation from the sky :**
- Radiation from the sky contributes a relatively large amount of heat to the soil in areas where the suns rays penetrate the earth's atmosphere at oblique angles.
- **Conduction of Heat from the Atmosphere :**

  It can have a substantial effect on soil temperature only by contact.
- **Condensation:**
- Condensation is an exothermic process. Whenever water vapor condenses from the atmosphere, it heats the soil because heat is transferred from the water to the soil. Under such conditions, increases of 5ºC and more in the soil temperature have been noted .in similar way the freezing of water releases heat from the water to the soil.

- **Evaporation:**
- Evaporation is an endothermic process. The greater the rate of evaporation, the more the soil is cooled. The energy required to change from liquid to vapor phase depends on the temperature and approximately 2.45 MJ/ kg at 20º C, this energy is known as the latent heat of vaporization.
- **Rainfall and Irrigation:**
- Depending on their temperature, rainfall and irrigation water can cool or warm the soil.
- **Insulation:**
- The soil can be insulated from the environmental temperature factors by plant cover, mulches, Snow, clouds, and fogs.
- During the summer, insulated soil is cooler than soil that is directly exposed to the elements, whereas in the winter, the temperature situation is reversed.
- **Vegetation:**
- Transpiration of water, reflection of incident radiation (back radiation), and energy used for photosynthesis by plants tend to decrease the temperature of the microclimate and indirectly of the soil. a plant cover serves as insulation .

- Most of the energy that reaches the earth in daytime is used as energy for evapotranspiration or radiated back to the atmosphere. Only a small portion perhaps as much as 10% is used to heat the soil.
- **2- Soil factor:**
- **2-1 Mineralogical Composition:**
- The mineralogical composition affects the heat capacity, thermal conductivity, and therefore the thermal diffusivity. These relationships are given in table 6.1-6.3.
- **2-2 Soil Water Content:**
- An increase in soil water content in dry range, from oven-dry to air-dry, increases the thermal conductivity of a soil, but only very little. As soon as the water starts to form bridges from particle to particle, the thermal conductivity increases markedly.
- Relationships examples are shown in figures 6.1 and6.2.

- **2-3 Soil Color:**
- Light-colored soils reflect more than darker soils. Dark-colored soils have high organic matter contents and soil water contents. Soils will not warm up as rapidly in spring as will well-drained, light-colored soils.

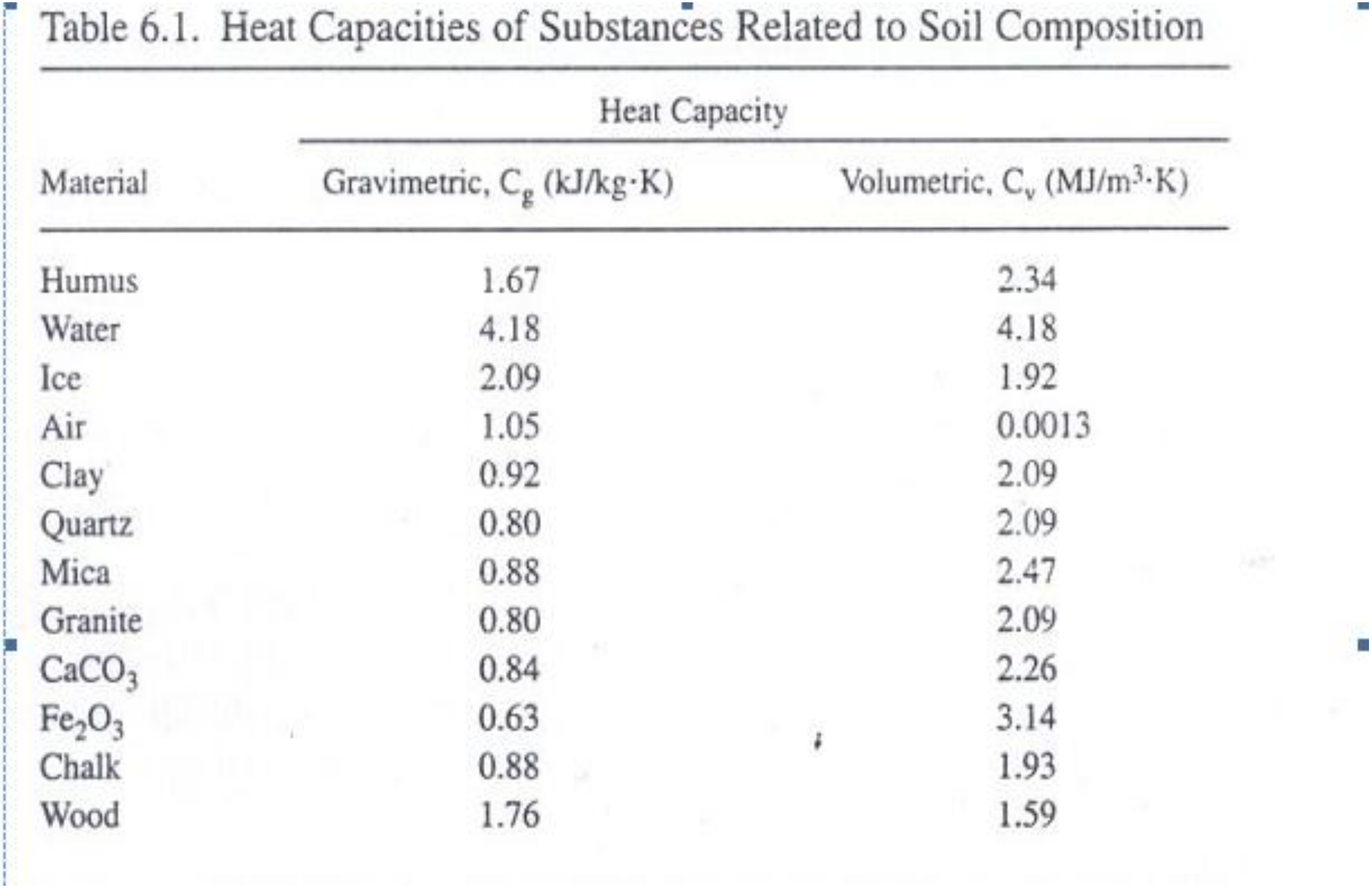


Table 6.1. Heat Capacities of Substances Related to Soil Composition

	Heat Capacity	
Material	Gravimetric, C_g (kJ/kg·K)	Volumetric, C_v (MJ/m^3·K)
Humus	1.67	2.34
Water	4.18	4.18
Ice	2.09	1.92
Air	1.05	0.0013
Clay	0.92	2.09
Quartz	0.80	2.09
Mica	0.88	2.47
Granite	0.80	2.09
$CaCO_3$	0.84	2.26
Fe_2O_3	0.63	3.14
Chalk	0.88	1.93
Wood	1.76	1.59

Table 6.3. Thermal Properties of Soil Constituents at 20° C and 1 atm

Material	Density, ρ (Mg/m^3)	Gravimetric Heat Capacity, C_g (kJ/kg·K)	Volumetric Heat Capacity, C_v (MJ/m^3·K)	Thermal Conductivity, k (J/m·s·K)	Thermal Diffusivity, α (m^2/s)	
Quartz	2.65	0.732	1.92	8.368	43×10^{-4}	$\rightarrow 10^{-5}$?
Many soil minerals	2.65	0.732	1.92	2.930	15×10^{-4}	$\rightarrow 10^{-5}$
Organic matter	1.3	1.925	2.51	0.251	1×10^{-4}	$\rightarrow 10^{-5}$
Water	1.0	4.184	4.18	0.594	1.4×10^{-4}	$\rightarrow 10^{-5}$
Air	0.0012	1.004	0.00121	0.026	2.1×10^{-5}	

Sources: Adapted from van Wijk and DeVries 1963, p. 105.

Note: Both k and α are low for organic matter and air compared to soil minerals.

6.3 Soil Heat Capacity

The volumetric heat capacity can be written as:

$\mathbf{C_v = SC_s+\theta_v C_w+f_a C_a}$……………………..(6)

Where:

- **C_v:** is the volumetric heat capacity of the soil (J/m³. K).
- **S:** is the volume fraction of the soil solids (m³/m³), that is $\rho_b/\rho_{p.}$
- **θ_v:** is the volume fraction of the soil water(m³/m³).also called the volumetric soil water content.
- **f_a**: is the volume fraction of the soil air(m³/m³).also called the aeration porosity.
- **C_s:** is the volumetric heat capacity of the soil solid (J/m³.K).
- **C_w:** is the volumetric heat capacity of the soil water (J/m³.K).
- **C_a:** is the volumetric heat capacity of the soil air (J/m³.K).

We note the volumetric heat capacity of solid fraction ($J/m^3.K$) can be written as

$\mathbf{SC_s = X_m C_m + X_o C_o}$……………………….(7)

<u>Where:</u>

- $\mathbf{X_m}$**:** is the volume fraction of mineral content.
- $\mathbf{X_o}$**:** is the volume fraction of the organic matter component.
- $\mathbf{C_m}$**:** is the volumetric heat capacity of mineral component.
- $\mathbf{C_o}$**:** is the volumetric heat capacity of the organic component.

1- For most mineral soils, the air and organic contributions to the heat capacity are assumed negligible because of their low magnitudes (0.01 to 0.03).

2- Using the information in table 6.1.we see that:

- $\mathbf{C_a}$ = 1.30 kJ/m³.K and $\mathbf{C_o}$ **=** 2.34 MJ/m³.K
- $\mathbf{C_v = X_m C_m + \theta_v C_w}$

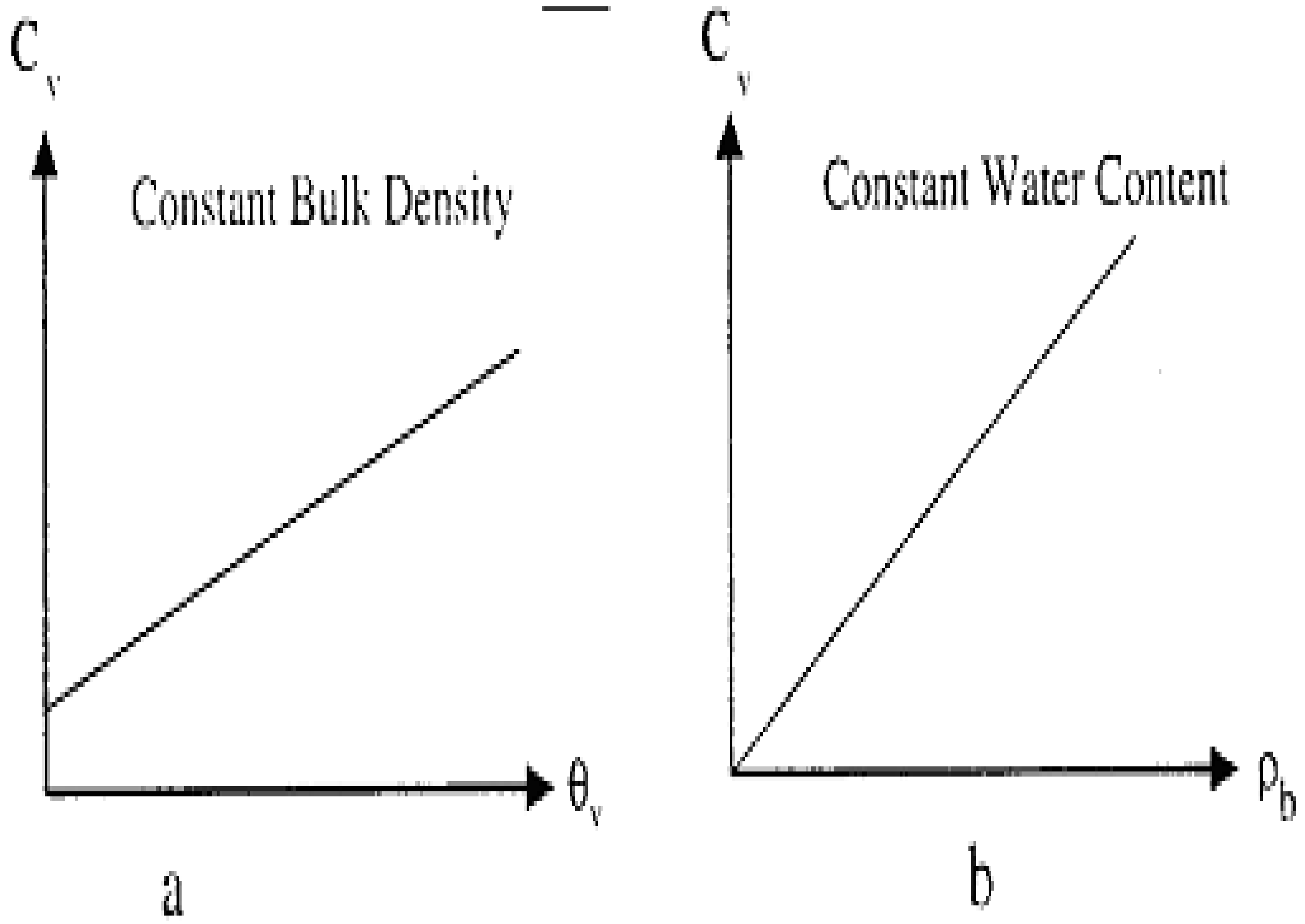
$$\mathbf{= \rho_b C_{gm} + \theta_w C_w \rho_b =}$$

$$\mathbf{= \rho_b [C_{gm} + \theta_w C_w]}$$

$$\mathbf{Cv = \rho_b [837 + (4180.\theta_w)] \ldots\ldots (9)}$$

Where:

- **Cv:** volumetric heat capacity of soil (J/m^3.K), $\boldsymbol{\rho_b}$**:** soil bulk density in kg/m^3.
- **837** is the gravimetric heat capacity of solid phase (J/kg.K).
- **4180** is the gravimetric heat capacity of water (J/kg.K).
- *C_v is directly dependent on the bulk density and soil water content*



Relationship between the volumetric heat capacity and water content (a) and bulk density (b).

- **<u>Example:</u>**
- If the gravimetric heat capacity of oven-dry mineral soil is near 837J/kg. K, what is the volumetric heat capacity of 1m^{3}of oven-dry soil consisting of half solids and half pore space?
- $\mathbf{C_v = SC_s+\theta_v C_w+f_a C_a \ \ = S\rho_p C_{ms}}$
- Since there is no water in the soil and the contribution from the soil air is negligible, we have:
- $\mathbf{C_v = SC_s = S\rho_p C_{ms}}$
- $= (0.5m^3/m^3)\ (2650kg/m^3)(837J/kg.K)$
- $= 1.1*10^6\ J/m^3K$
- $= 1.1\ MJ/m^3K$
- **<u>The same soil filled with water would have heat capacity of</u>**
- $\mathbf{C_v = SC_s+\theta_v C_w}$
- $= 1.1*10^6+ [(0.5m^3/m^3)\ (4.18*10^6J/m^3.K)]$
- $= 3.2*10^6\ J/m^3.K$
- $= 3.2\ MJ/m^3.K$

- If only one-half of the pore space was filled with water, its heat capacity would be:
- **$C_v = SC_s + \theta_v C_w$**
- $C_v = 1.1*10^6 + [(0.25 m^3/m^3)(4.18*10^6 J/m^3.K)] + [(0.25 m^3/m^3)(1.21*10^{-3} J/m^3.K)]$
- $= 2.15$ MJ/m^3.K
- This example shows that Cv would be expected to range between 1 and 3 MJ/m3.K (a threefold variation due to water content).equation (8) and the results of these problems clearly show that soil water content and bulk density directly affect the magnitude of the heat capacity.

- **6.4 Thermal conductivity of soil:**
- Heat flow through porous materials such as soil is more complicated than flow through solids because of varying combinations of phases and densities of soil components. The heat flux amounts to 10-30% of the net radiation that is received diurnally for a bare soil. Heat entering the soil is nearly matched by the heat leaving the soil. Conduction of heat occurs by transmission of thermal energy of motion from one particle to another. Transport of heat by conduction is the most important mechanism in the soil. The amount of heat transfer depends on soil thermal conductivity, k, and temperature gradient, as described by **Fourier's law:**
- $H_s = -k\, ƏT/Əz$
- **Where:**
- **H_s**: the heat flux density ($J/m^2.s$ or W/m^2)
- **ƏT/Əz**: is the temperature gradient (K/m)
- **k**: the thermal conductivity ($J/m^2.s.K$ or W/m.K).

- Fourier’s law is sufficient to describe heat conduction under steady- state conditions; temperature and flux density remain constant in the time. Soil heat flow is also assumed to be unidirectional, that is, only in the vertical direction and no heat is generated or converted into other forms and soil medium is assumed to be isotropic.

- The magnitude of k at complete dryness is low. At low soil water content, the soil particles are covered by thin adsorbed water layers. An increase in soil water content causes the contact areas to expand, which results in an increase in K. As a result, k can be described as the steady-state thermal transport property characteristics for each soil medium within certain physical restrictions. We let:
- $\mathbf{k=\alpha C_v}$
- Where $\boldsymbol{\alpha}$ is the thermal diffusivity.

- Values of k vary with:
- Soil phase: solid>liquid>air
- Textural composition: sand>loam>clay>peat
- Bulk density: directly related (see figure6.1)
- Water status: directly related (see figure6.1)
- Structure: similar to textural composition since aggregation of primary particles results in larger secondary particles
- Surface contact between particles.
- **6.5 Methods of calculation of soil heat flux density:**
- **1- Calorimetry:**
- Calorimetry utilizes the law of conservation of energy to give:
- $H_{si} - H_{so} = \Delta H_s$ (18)
- **Where :**
- **H_{si}**: the soil heat flux density going into the layer ($J/m^2.s$)
- **ΔH_{so}**: the soil heat flux density going out of the layer ($J/m^2.s$),and
- **ΔH_s** is the rate of change in the heat content per unit area ($J/m^2.s$).values of ΔH_s can be determined as follows. the amount of heat needed to change the temperature of a given soil volume is Qq with SI units of J.

- **$Qq = C_vV\Delta T$ (19)**
- Dividing by cross- sectional area and time, we obtain
- **$Qq/A\Delta t = \Delta H_s = C_vV\Delta T/(A\Delta t) = C_v(\Delta T/\Delta t)\ \Delta z$**
- Or
- **$\Delta H_s = C_v(\Delta T/\Delta t)\ \Delta z$ (20)**
- <u>Where:</u>
- **C_v:** the volumetric heat capacity of the layer ($J/m^3.K$)
- **Δz**: the soil profile thickness (m)
- **$\Delta T/\Delta t$:** the change in temperature with change in time (K/s)

- **Example (page 124):**
- A soil was found to have a net heat flux density, $\Delta H_{s,}$ of –2.13MJ/ m^2.d. If the average volumetric water content was 0.25m^3/m^3 and the bulk density was 1300 kg/m^3, calculate the daily average temperature increase that occurs in 1 d in the top meter of the soil assuming that the heat is evenly distributed in the profile.
- **$\Delta H_s = Cv. (\Delta T/\Delta t)\ \Delta = (C_v)(1.0)(\Delta T/\ \Delta t)$**
- $\Delta T = \Delta H_s / C_v = 2.13/[(1300*837)+(0.25*4.18*10^6)]$
- $= 2.13/2.13 = 1.0°C$
- The sign of ΔH_s is negative, indicating that heat is entering the soil. However, from a soil energy balance perspective, the heat flux density H_s is positive because it is entering the soil volume. Thus, it is an influx. This is a different sign conversion than that used with the soil heat flow. In general, uniform redistribution of heat throughout the profile does not occur at the field soils.
-

- **2- Temperature gradient method:**
- Fourier's law: $\mathbf{H_s = -k\partial T/\partial z = -k(\Delta T/\Delta z) = -k[(T_2-T_1)/(z_2-z_1)}$
- **Example (page 125):**
- The temperature of a DeWitt silt loam was measured at two depths twice in one day. The data are presented at the table below. If the thermal conductivity is 1.76 J/m.s.K, calculate the heat flux density and direction of heat flow at the two measurement times. Remember we assume that z is positive in the upward direction and the reference level is at the soil surface.

Sampling time	Temperature (in K)	
	Surface	0.10 m depth
06:30	294	303
16:30	327	309

- <u>At 06:30 hours.</u>
- H_s = - 1.76[(303-294)/(-0.1-0)] = -1.76(9/-0.1) =158.4 J/m^{2}s
- Since upward flow is considered to be positive, a positive H indicates upward flow of heat and a negative H_s indicates downward flow of heat. At this time of the day, soil heat flow is toward the soil surface.
- <u>At 16:30 hours.</u>
- H_s=-1.76[(309-327)(0.1-0)] = -316.8 J/m^{2}s
- Thus, at this late-afternoon time, the flow of heat is vertically downward into the profile.

6.6 Experimental Estimation of Soil Temperature

- **1- Liquid in –Glass Thermometer.**
- Probably the most common instrument for measuring soil temperature. A good contact is required between the reservoir and the soil of which the temperature is to be measured.
- **2- Bimetallic Thermometer:**
- Two strips of metal that have different thermal expansion coefficients and firmly welded together.
- **3- Resistance Thermometer:**
- Electrical methods are based on change of resistance with temperature or on the thermoelectric effect.

- **4- Thermocouple:**
- The thermocouple is one of the most widely used instruments for measuring soil temperature. When two wires of different metals are welded together at two points and the welds are kept at different temperatures, an electrical potential difference between the welds I is established, and current flows throughout the circuit formed by two wires. This effect is called the thermoelectric, or Seebeck effect.
- **5- Remote Sensing Thermometer:**
- According to Stefan-Boltzmann equation:
-
- $\mathbf{R_1 = e\delta T^4}$
- Or $\mathbf{h = (\varepsilon\sigma T^4).A}$ equation (20) in chapter 5.
-
- **Where:**
- **h**: is the heat flux in J/s or W
- **σ**: Stefan- Boltzmann constant = $5.675*10^{-8}$ W/m^2.k to units of h
- **ε**: emissivity of object = es/ eb
- This equation gives the maximum energy flux that can leave a surface of area A at absolute temperature T.

Ch7: Soil Aeration

- **Soil aeration**: is the process by which gases consumed or produced in the soil are exchanged for gases in the atmosphere.
- Soil aeration controls the soil status of two life-sustaining processes: **respiration** and **photosynthesis**.
- Respiration involves the oxidation of organic compounds and can be represented by the general reaction in which glucose is oxidized:
- $C_6H_{12}O_6+6O_2 \rightarrow 6CO_2+6H_2O$ **(1)**
- In photosynthesis, the reaction is reversed.
- Oxygen is the most abundant atom in the earth's crust (53.77%), only about 0.01% of all oxygen is molecular oxygen in the atmosphere.
- The most important source of O_2 for land organisms is the atmosphere, where the concentration is buffered at about 21%

7.1 Gaseous composition of the soil atmosphere:

- The chemical composition of the soil atmosphere depends on conditions such as water content, temperature, porosity, microbial activity, and structure. A comparison of the compositions of the soil atmosphere is poorly drained silt loam is given in Table 7.1.
- ***Volume percentages of CO_2 and O_2 in the soil gaseous phase a t two sampling times and five depths.***

Sample at soil depth (cm)	Winter		Summer	
	CO_2	O_2	CO_2	O_2
30	1.2	19.4	2.0	19.8
61	2.4	11.6	3.1	19.1
91	6.6	3.5	5.2	17.5
122	9.6	0.7	9.1	14.5
152	10.4	2.4	11.7	12.4

- Soil O_2 and CO_2 concentrations are inversely related. Concentrations of CO_2 increases with depth while O_2 concentration decreases with depth.
- Trace gases such as methane, nitrous oxides, hydrogen sulfide, and ammonia, are also present in the soil.
- Soil structural units largely determine how air moves into the soil profile.
- **Soil gases can exist in three states:**
- Free state, filling the empty soil pores or voids.
- Dissolved in water phase.
- Adsorbed on solid phase.

- **Volume fraction of soil air:**
- Since the solid phase is essentially constant over time, the fractions of water and air are dynamic. And their ratios on a volume basis are inversely related.
- **$f_a = f - \theta_v$ (2)**
- **$f_a = 1 - \theta_v - S$ (3)**
- **Where:**
- **f_a** is the volume fraction of air(aeration porosity)
- **θ_v** is the volumetric water content.
- **S** is the volume fraction of the solid phase (ρ_b/ρ_p)
- **7.2 Computations of gas concentration:**
- The percentage of the total volume that would be occupied by each gas in its pure state called the partial pressure of that gas.
- The ideal gas law can be used to convert gas partial pressure to gas concentration.

- **7.2.1 On amass basis:**

$C= (M/RT)P$……………………………(4)

- **Where:**
- **M** is the molar mass (kg/mol), **R** is the gas constant (8.314 J/mol.K)
- **T** is the absolute temperature (K), **C** is the gas concentration (kg/m^3 of the soil)
- **P** is the partial pressure of the gas (Pa)
- **C** is the gas concentration (kg gas/m^3 of soil)
- **Notes:**

1- The partial pressure of the gas is the volume fraction (0.21 for O_2; 0.79 for N_2) multiplied by atmospheric pressure ($1.013*10^5$ Pa at the mean sea level)

2- Two useful ***conversions*** to remember are

- J = N.m
- Pa = N/m^2. Thus, Pa=J/m^3

3- unit of C is $kg/m^3 = [(kg/mol)/[(J/mol.K)*K]*Pa = kg/J* J/m^3= kg/ m^3$

Example (page 144):

- Determine the gaseous concentration of $O_{2,}$ $N_{2,}$ and CO_2 at a standard temperature (25ºC) and pressure (101.3 kPa).
- $C(O_2) = [0.032/(8.314*298)](0.21*1.0113*10^5) = 0.275\ kg/m^3$
- $C(N_2) = [0.028/(8.314*298)](0.79*1.013*10^5) = 0.904\ kg/m^3$
- $C(CO_2)=[0.044/(8.314*298)](0.003*1.0113*10^5)= 0.00054\ kg/m^3$.

7.2.2 On molar basis

- Equation (4) becomes:
- **$C= (1/RT)P$ (5)**

Where:

- The units of C are in moles per cubic meter.
- Gas concentration in the soil, C, is related the concentrations of gas in the soil air C_a, and in the soil solution, C_1.
- **$C=C_a(f_a+a_b\theta_v)$ (6)**

Where:

- **f_a** is the aeration porosity (m^3/m^3)
- **a_b** is Bunsens coefficient of solubility (m^3/m^3). This represents the solubility of the gas per unit volume of water at at the given temperature and normal pressure (101.3 Pa)
- **θ_v** is the volumetric soil water content (m^3/m^3)

$a_b = a_w/a_a$ (7)

<u>Where:</u>

- **a_w** is the solubility coefficient in water (mol/ m^3.Pa)
- **a_a** is the solubility coefficient in air (mol/ m^3.Pa)
- Values of **a_b** for O_2 and CO_2 decreases with increases in temperature and are given in table7.2
- The solubility coefficient of the gas in water is determined from Henrys law as:

$a_w=C_w/p$ (8)

<u>Where:</u>

- **C_w** is the concentration of gas in water
- **P** is the partial pressure
- For air the relationship is similar and given as:

$a_a=C_a/p$ (9)

- Application of equation (6) to soils shows that:
- ***The contribution of the gaseous phase to the concentration of oxygen in soil is low. Therefore, changes in the term f_aC_a are closely related to the soil gas concentration***

7.3 Mechanisms of soil gas exchange:

7.31. Convective flow

- There are four conditions to be considered in evaluating convective flow of gases in soil:
  - Changes in soil temperature.
  - Changes in the volume of air space due to changes in soil water content.
  - Changes in barometric pressure in the atmosphere.
  - Wind gusts over the soil surface.
    - Convective flow of soil air results in a mass of air moving with a zone of higher total gas pressure to a zone of lower total gas pressure. Convective flow of air is similar to the convective flow of water in that the flow of both fluids is proportional to a pressure gradient.
    - The process is dissimilar in that water is relatively incompressible in comparison to air, which is highly compressible. Thus, air density and viscosity are strongly dependent on pressure and temperature.

- In one dimension (z), the volume convective flux density of air is:
- **J_v = -(k/η) ӘP/Әz = -KӘP/Әz**…………………………(10)
- Where:
- **J_v** is the volume convective flux density of air (m^3/m2.s or m/s), **k** is the intrinsic permeability of air –filled pore space (m^2), **η** is the viscosity of soil air (Pa.s), **P** is the total air pressure (Pa) and **K** is the air conductivity (m^2/Pa.s). The quantity **ӘP/Әz** is the total air pressure gradient (Pa/m)
- Assuming the total pressure gradient is constant, equation (10) indicates that the increase in K result in the flux density of air. The intrinsic permeability, k, depends only on the geometry of soil pores and not on the gas.
- The viscosity of water is 50 times greater than air:
- Viscosity of air~20 µPas and Viscosity of air~1.0 mPas.

- Air conductivity, K, has the same units as water conductivity expressed on a volume basis ($m^2/Pa.s$) and depends on the aeration porosity, size, continuity, and geometry of gas-filled soil pores. It is highly dependent on the volumetric water content of the soil.
- If gas flux density is expressed in terms of mass/(area*time), the convective transport equation is
- $\mathbf{J_m = -(Ck/\eta)\ \partial P/\partial z}$ ……………………………….(11)

<u>Where</u>:

- **J_m** is the mass convective flux($kg/m^2.s$)
- **C** is the concentration of soil air (kg/m^3).

<u>7.3.2 Diffusion</u>

- Transport of gas by diffusion occurs as a result of the gradient of partial pressure (or concentrations of the various constituents in soil air).
- Depends on both soil and gas properties.

1- The differences in gas concentration between two regions.

2- The length of diffusion bath.

3- The diffusion coefficient of the gas,

4- The shape and continuity of the air-filled pores.

- For steady-state conditions, the diffusive transport can be described by Fick's law and in one direction is governed by:
- **$J_d = -D (ƏC_a/Əz)$**

<u>Where:</u>

- **J_d** is the diffusive flux density ($kg/m^2.s$)
- **D** is the apparent gas diffusion coefficient in the soil (m^2/s)
- **C_a** is the gas concentration ($kg/m^3$of soil air).
- **z** is the distance(m).
- If partial pressure, p, is used instead of concentration of the diffusing component, the above equation becomes
- **$J_d = -(D/b)\ Əp/Əz$**

<u>Where:</u>

- b: is the ratio of the partial pressure to the concentration ($Pa.m^3/kg$)
- The quantity D/b is the inverse of the diffusion impedance (delay) and may be designed as the diffusion constant K. Diffusion of gases in the soil due to the partial pressure gradient of one component is always accompanied by a counter diffusion of other gas components.

7.4 Characterization of soil aeration:

- **1- Aeration porosity, f_a**
- ***The air-filled porosity*** is expressed as a fraction or percentage of the total soil volume that is occupied by air; values of f_a depend on soil texture, water content, and structure.
- **2- Composition of soil atmosphere:**
- The composition of the gas phase is also a static indicator of soil aeration. Measurement of the composition of soil gases was the earliest method used to characterize soil aeration.
- Measurement techniques have limitations caused by volume of the gas sample required, the volume of soil sampled, and the time required to take the sample.
- **3- Oxygen diffusion rate:**
- The aeration status of the soil can be characterized by oxygen diffusion rate (ODR) in the soil. The (ODR) method has been found to be satisfactory in the higher range of the soil aeration and to be less sensitive in the lower range of soil aeration (poor drained and flooded soils). In (ODR) method, oxygen reduction rate is measured at a cylindrical Pt electrode when a constant electrical potential is applied.
- Many studies have shown that there is a fairly good relationship between ODR and plant response.
  - Values $< 0.20 \mu g/cm^2.min$ often indicate inadequate aeration for most plant species.
  - Values $> 0.40 \mu g/ cm^2.min$ are considered sufficient for optimum growth of most plants.

4- Oxygen-reduction potential:

- Oxygen-reduction potential or redox, is a measure of electron availability (or activity) potential in a chemical or biological system. A relatively inert electrode, usually platinum, is used to measure the potential of all redox couples and gives value of E_h

1- Well-drained soils have characteristics E_h values of grater than 400 mV

2- In some soils with high carbon contents, E_h values as low as 350 mV can be observed.

3- Oxygen disappears at E_h values of approximately 300Mv.

4- NO_3 is removed between 200 and 300 mV, followed sequentially by reduction of Mn^{+4}, Fe^{+3}, and SO_4^{-2}

5- Respiration quotient and respiration rate:

- Aerobic respiration involves the decomposition of complex carbon molecules, the consumption of oxygen, and release of CO_2, water, and energy for cellular growth. Under these conditions, the respiratory quotient, RQ, is defined:
- **RQ = Volume of CO_2 released /Volume of O_2consumed** (usually =1.0)

- However, when respiration is anaerobic, Rq increases since O_2 is no longer consumed but CO_2 continues to be evolved.
- Factors affecting the respiration rate include:
- Soil conditions such as organic matter and water content
- Cultivation and cropping practices
- Atmosphere factors such as air temperature, and
- The presence of biological organisms.
- **6- Reduced chemical forms:**
- The aeration status of a soil is related to its oxidation and reduction status. Therefore, the ionic status of several ions is indicative of soil aeration.

- **Response of plants to poor soil aeration:**
- Several factors affect plant response to poor aeration. These include the age of the plant, nutrition, temperature, light intensity, and the plant species or variety.
- However, when heavy rain coincides with temperatures high enough to maintain rapid respiration rates, the root zone can become anaerobic, a condition that, even if sustained for only a day, has serious effects on the metabolism and growth of sensitive plants,
- The characteristic whole plant symptoms are wilting, yellowing of leaves, and the development of adventitious roots at the base of the stem. Stimulation of the production of ethylene (C_2H_4) may cause growth abnormalities such as leaf epinasty, the downward curvature of leaf axis.

Ch8: Soil Water Principles

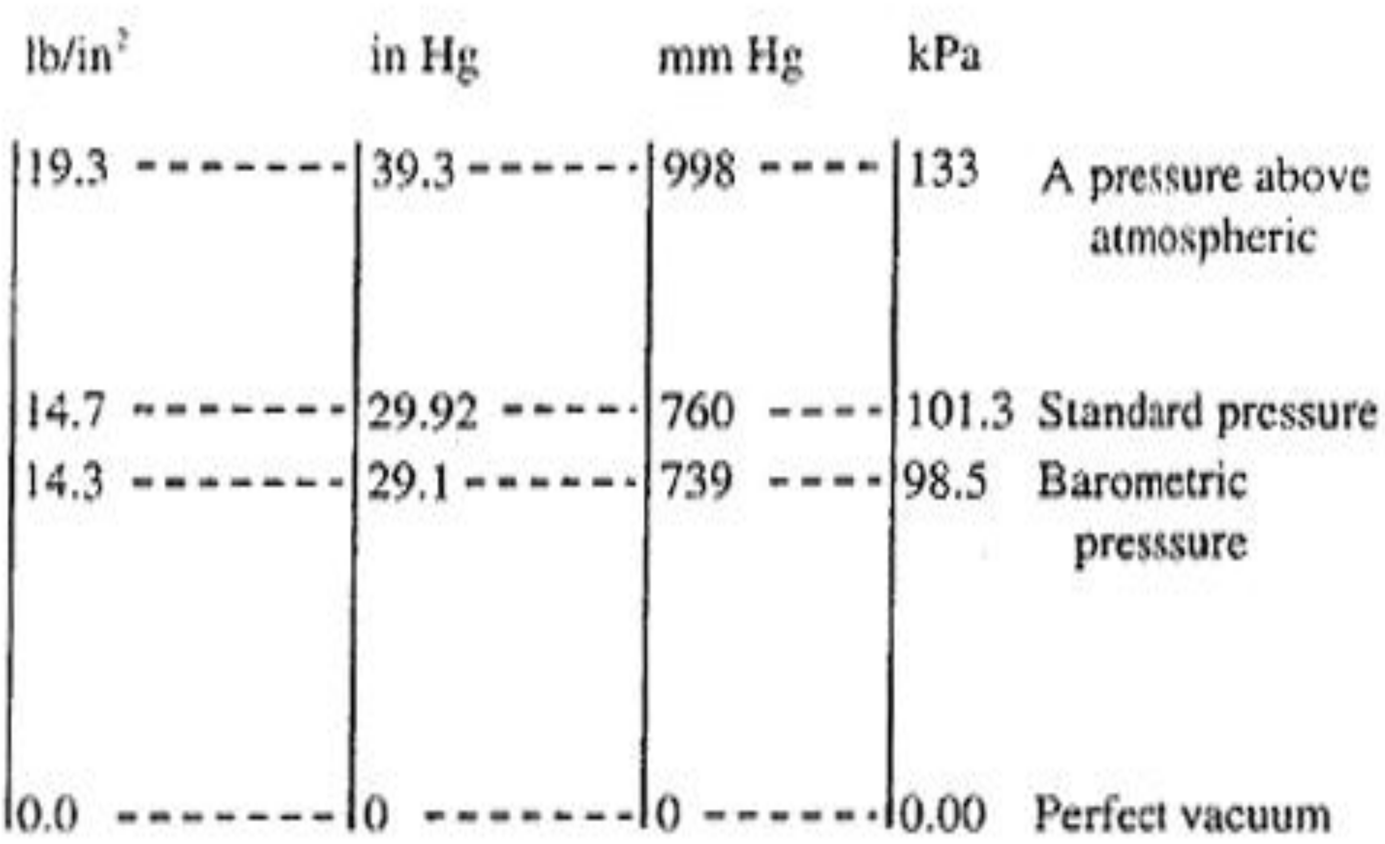
8.1 Review of pressure concepts:

- Pressure is defined as “normal force (F) per unit area (A), F/A. The absolute pressure P (N/m^2 or Pa) at the bottom of the static column of water exerted on the sealing plate is the sum of the pressure exerted by the water column and the atmospheric pressure.
- **$P = F/A = \rho_w g h + p_o$**
- Where
- **F:** the force (N).
- **A**: the cross-sectional area (m^2). **ρ_w:** the density of water (kg/m^3)
- **ρ_w:** the density of water (kg/m^3)
- **g**: the acceleration of gravity (m^2/s)
- **h**: the height of water column (m)
- **p_o**: the atmospheric pressure at the top of the column of water (Pa).

1-Pressures, like temperatures, can be expressed by either absolute or relative scales.

2-Whether relative or absolute, pressure is measured in a pressure-measuring device depends on the nature of the instruments used to make the measurements. Another common pressure-measuring device is the Bourdon gauge, which normally (but not always) reads zero pressure when open to the atmosphere.

3-The four common systems of pressure measurement: pounds per sequare inch (psi or lb/in^2). Inches in mercury (in Hg), mm of Hg (mm Hg), and kilopascals (kPa). Other frequently used systems are feet of water (ft H_2O), atmospheres (atm), and bars (100 kPa =1 bar)



lb/in²	in Hg	mm Hg	kPa	
19.3	39.3	998	133	A pressure above atmospheric
14.7	29.92	760	101.3	Standard pressure
14.3	29.1	739	98.5	Barometric presssure
0.0	0	0	0.00	Perfect vacuum

Figure 8.6. Absolute-pressure comparisons when barometric pressure is 29.1 and 29.92 inches of Hg.

- **Expressed in various units, the <u>standard atmosphere</u> equals to:**
- 33.91 feet of water.
- 14.7 pounds per square inch.
- 29.92 inches of Hg.
- 760 mm of Hg.
- $1.013*10^5$ Pa, or 101.3 kPa.
- 101.3 J/kg

8.2 Expressions of the potential energy per unit quantity of water

1- Energy per unit mass, Ψ_{pm}

- Work =force*distance =energy.
- Work/mass = energy/mass = J/kg or J/mole.

Note: 1 J/kg = 1 kpa

2- Energy per unit volume, Ψ_{pv}

- **Ψ_{pv}**, is the work divided by volume,
- Force*distance/volume = force/area = pressure.
- Work/volume =J/m^3 = Pa.
- Potential enrgy on a volume basis is obtained from potential energy on a mass basis by multipling by the density of water, ρ_w.

3- Energy per unit weight, Ψ_{pw}

- Work per unit weight of water, **Ψ_{pw}**, is known as ***hydraulic head***.
- Work/weight = force * distance/ weight = distance or head.
- $N.m.s^2/kg.m = kg.m.s^2/kg.s^2$ =m

- In addition, hydraulic head is defined as:
- Pressure/weight =(force/area)/(density*gravity)=head
- $(kg.m/m^2.s^2)/(kg.m/m^3.s^2)=m.$
- This energy /weight method of expressing potential energy is simpler and more convenient because it only has units of length. Therefore, it is used mostly to characterize the state of soil water in the terms of the total hydraulic head, the gravitational head, and the matric head.

Useful conversions (page 175):

- $\boldsymbol{\Psi_{pw} = \Psi_{pm}/g = \Psi_{pv}/(\rho_w g)}$
- $\boldsymbol{\Psi_{pv} = \Psi_{pm}\rho_w}$
- $\boldsymbol{\Psi_{pm} = \Psi_{pw}g = \Psi_{pv}/\rho_w}$

- **Example (page 175):**
- Convert the potential energy/mass of 100 J/kg to its equivalent in volume (Pa) and weight (m) units.
- On a volume basis, we use $\boldsymbol{\Psi_{pv} = \Psi_{pm}\rho_w}$
- (100 J/kg)(1000 kg/m^3) = 100, 000 J/m^3
- = 100, 000 Pa = 100 kPa
- On a weight basis, we use $\boldsymbol{\Psi_{pw} = \Psi_{pm}/g}$
- (100 J/kg)(9.8 m/s^2) = 10.2 m= 1020 cm.

Table 8.2. Useful Conversions of Soil Water Potential Units

Expression	J/kg	Pa	bar	psi	cm H_2O	cm Hg
Energy/mass						
1 J/kg	1	10^3	10^{-2}	0.145	10.2	7.5
Energy/volume						
1 Pa	10^{-3}	1	10^{-5}	1.45×10^{-4}	0.0102	0.0075
1 bar	10^2	10^5	1	14.5	1020	75
1 psi	6.89	6890	0.0689	1	70.2	51.2
1 atm	101.3	1.013×10^5	1.013	14.7	1033	76
Energy/weight						
1 cm H_2O	0.0981	98.1	9.81×10^{-4}	0.0142	1	0.735

Note: To convert from the unit of the expression, multiply by the factor to obtain the dimensions of the other factor.

8.3 Total soil water potential:

- **Definition:**

The amount of work that must be done per unit of a specified quantity of pure water in order to transport reversibly and isothermally an infinitesimal quantity of water from pool of pure water, at a specified elevation and at atmospheric pressure, to the soil water at the point under consideration.

- The total soil water potential energy can also be defined mathematically as:
- $\boldsymbol{\Psi_t = \Psi_g+\Psi_m+\Psi_p+\Psi_o}$

Where

- $\boldsymbol{\Psi_t}$**:** the total soil water potential energy.
- $\boldsymbol{\Psi_g}$: the gravititional potential energy.
- $\boldsymbol{\Psi_m}$**:** the matric potential due to capillary pressure and other soil and water surface forces.
- $\boldsymbol{\Psi_p}$: the pressure poential.
- $\boldsymbol{\Psi_o}$**:** the osmotic potential due to salts.

- **Component water potential**
- **<u>8.3.1 Gravitational potential energy:</u>**
- Gravitational force = weight = mass* acceleration
- = **mg**
- Thus the potential energy per unit mass is:
- $\boldsymbol{\Psi g_m = gz}$
- Equation (12) forms the definition of gravitational potential of soil water when expressed on a mass basis. It shows that Ψg is linearly related to changes in elevation and is independent of the chemical or pressure conditions of soil water: that is. It is independent of soil properties.

- In terms of gravitational potential energy per unit weight, equation (11) becomes
- **$\Psi g_w = z$**
- **<u>Where:</u>**
- **z**: the distance measured in the vertical direction from some arbitary chosen reference level. Once again, this equation shows that Ψ g depends only on the position of the point relative to the reference level, and not on any soil property or characteristic.
-

- **Example:**
- Assume that a piezometer (an open-ended cylindrical tube) has been inserted into a soil profile and that this tube contains water at the 0.90m depth (figure 8.7). On a mass, volume and weight basis, compute the gravitational potential at the water table assuming that the reference level is (1) located at the soil surface, (2) at the depth of water table, and (3) at a drain that is located at a depth of 1.2m, respectively. The solution is given in the table below for three positions and three bases of expression.

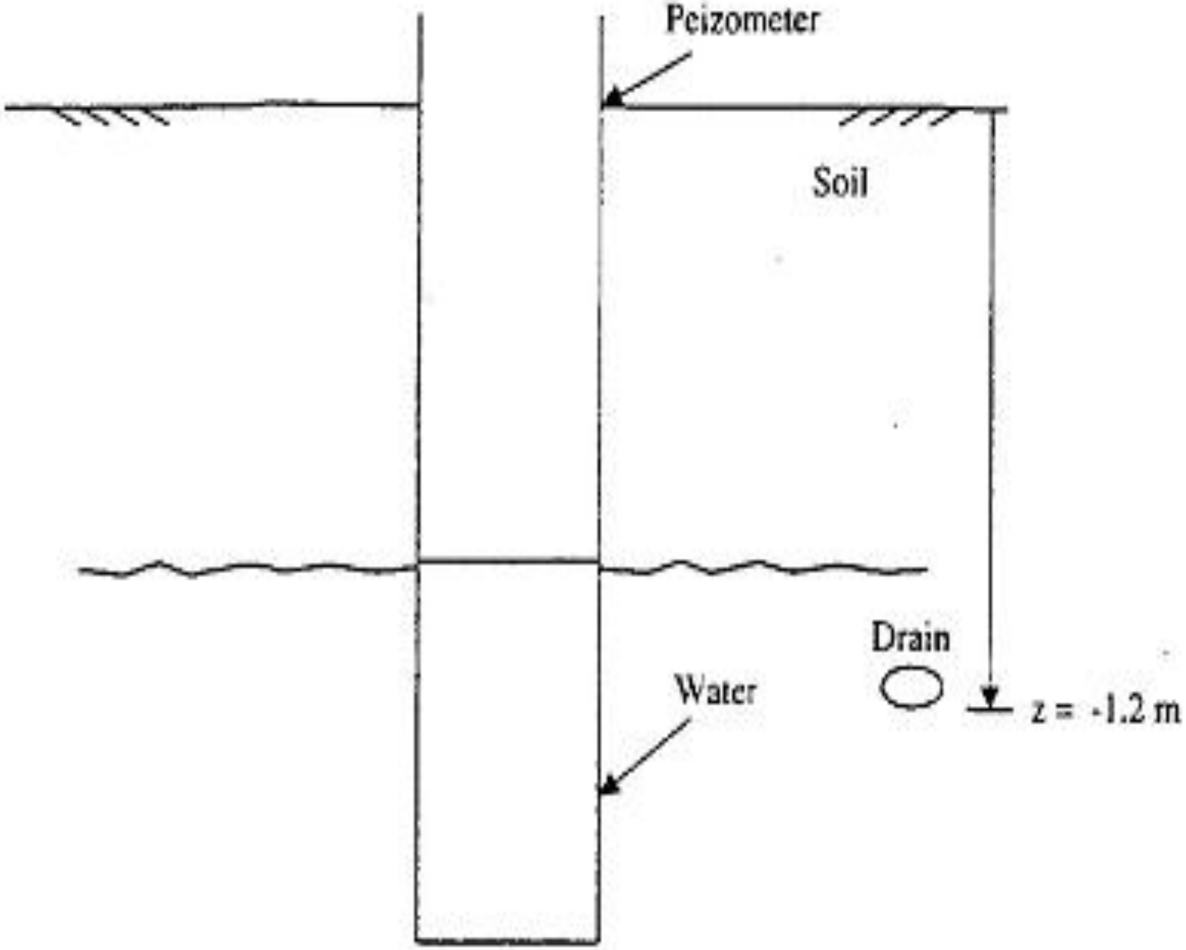


Figure 8.7. Schematic of a piezometer inserted into a soil profile containing a drain at the 1.2 m depth.

	Basis of expression		
Position	Mass (J/kg)	Volume (Pa)	Weight (meters)
Reference level at the soil surface			
Soil surface	0	0	0
Water table	–8.82	–8,820	–0.9
Drain	–11.76	–11,760	–1.2
Reference level at the water table			
Soil surface	8.82	8,820	0.9
Water table	0	0	0
Drain	–2.94	2,940	–0.3
Reference level at the drain			
Soil surface	11.76	11,760	1.2
Water table	2.94	2,940	0.3
Drain	0	0	0

The results show that varying the position of the reference level changes the absolute value of the gravitational potential.

8.3.2 Pressure potential energy:

- The positive pressure potential that occurs below the groundwater table has a hydrostatic pressure of water per unit mass with reference to the atmospheric pressure of:
- **$P = gh$**

Where:

- **h**: the submergence depth below the free water surface. The potential energy of this water is then
- **$E = PdV$**

Therefore, the pressure potential energy per unit volume of soil water is:

- **$\Psi_p = \rho_w gh.$**

And the potential energy per unit weight is:

- **$\Psi_p = h$**

8.3.3 Osmotic potential energy:

The soil water potential caused by solutes in the soil or plant water system where a semi-permeable membrane is present is designed as the osmotic potential.ψo

- The units of potential osmotic energy based on mass are joules per kilogram (J/kg). And the governing equation is
- $\Psi_o = gh_o$

Where:

- **h_o**: the height of a hanging water column with respect to the point of measurement in the solution.
- The quantity ho is always negative and can be considered as the gauge pressure.
- The unit of potential osmotic energy based on volume is the Pascal (Pa)
- $\Psi o = \rho_w g h_o$

1- The vant Hoff equation can be used to calculate the osmotic potential form:

- **Ψ_o = - RTCs**

Where:

- **R**: the gas constant (8.134J/K.mol)
- **T**: the Kelvin temperature
- **Cs**: the solute concentration (mol/m³)
- **Ψ_o**: the J/kg or J/m^3 of water.

Depending on whether the unit quantity of water is expressed on a mass or a volume basis.

2- The equation developed b the U.S. Salinity Laboratory Staff:

- $\mathbf{\Psi_o = - 0.056\ TDS.}$
- $\mathbf{\Psi_o = - 36\ EC}$

<u>Where:</u>

- **TDS:** Total Dissolved Solids in mg/L or ppm
- **EC**: measured in deci-Siemens per meter (dS/m)
- $\mathbf{\Psi_o}$: the Kilopascals (**kPa**)

3- When ψ_o is expressed as per unit volume of water, the equation:

- $\mathbf{\Psi_o = (RT/V).\ln(e/e_0)}$

<u>Where:</u>

- $\mathbf{\Psi_o}$: the osmotic potential in Kilopascals (**kPa**)
- **V**: the molar volume (the volume of 1 mole of water is 18 cm^3/mol)
- $\mathbf{e_o}$: the vapor pressure of water,
- **e**: the vapor pressure of the solution.

Note that for water, ρw is equal to M/V.

<u>Example (page 190-191):</u>

- Compute the volume of ψ_t of a saturated soil at 20ºC at a point C through which the reference level passes. The water table is 1.2 m above point C. Assume that at the saturation, the volumetric water content of the soil is 0.5 m³/m³ and that 1cm³ of soil at point C has 3*10-4 moles of solute.

We calculate the potentials on volume basis. First, we write

- $\boldsymbol{\Psi_t = \psi_g + \psi_p + \psi_m + \psi_o}$

In this particular case, ψg = 0: since the soil is saturated, ψm =0: and the other potentials are calculated as follows:

- Ψ_o= -RTCs = -8.31 J/K.mol*293K*($3*10^{-4}$ mol/$5*10^{-7}$m³) = -$1.46*10^3$kPa
- ψ_p =1.2*9800 = 11.8 kPa
- ψ_t = -$1.46*10^3$+11.8 kPa = -$1.448*10^3$kPa

<u>Note:</u> the Cs is defined on a soil solution basis and not on a soil basis.

8.3.4 Matric potential energy, Ψ_m

- Storage or retention of water by soils is a result of attractive forces such as adhesion and cohesion between the solid and liquid phases, responsible for the binding of water in the pores or capillaries.

1- Ψ_m can be measured in equilibrium with various vapor pressures. On a mass basis:

- $\Psi_m = (RT/M) \ln(e/e_0)$
- **e** is the vapor pressure of the soil water, e_0 is the vapor pressure of pure water, e/e_0 is the relative vapor pressure or relative humidity and has values less than 1.
- **R** is the gas constant (8.314 J/K.mol), T is the temperature in K and M is mass in kg of a mole of water (0.018015 kg/mol).

Example (page 183):

- Assume that the relative vapor pressure of water in soil at a temperature of 20ºC is equal to 0.85. Calculate the Matric potential of the soil water.

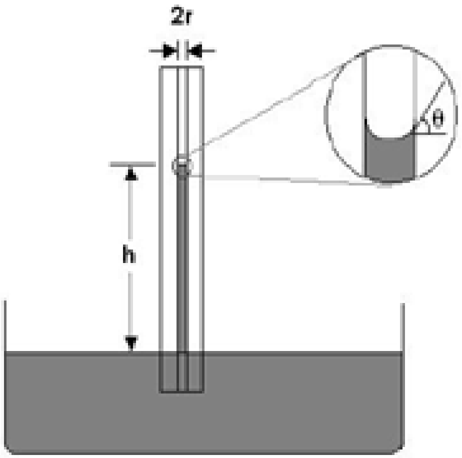
1- Ψ_m= [(8.314*293)/0.018]*ln(0.85) = -21.989J/kg = -21.989kPa = -2244m.

2- If the relative vapor pressure is 0.989, what is the matric potential?

- Ψ_m =[(8.314*293)/0.0181]*ln(0.989) = -1496J/kg = -1496 kPa = -152.7m.

This last calculation shows that the range of intrest in e/e0 from a plant grwth stand point lies between 0.989 and 1.00. a very small range of relative humidity.

2.Estimation of matric potential from the capillary rise equation:



$$F_{up} = (2\pi r)\sigma\cos(\theta)$$

$$F_{down} = mg = \text{volume*density*gravity} = \pi r^2 h \rho_w g$$

$$(2\pi r)\sigma\cos(\theta) = \pi r^2 h \rho_w g$$

or: $$h = 2\sigma\cos(\theta)/(r\rho_w g)$$

Setting P = pressure = $\rho_w gh$. We can write

$$P = 2\sigma\cos(\theta)/r$$

Assuming that the surface is completely wet and the contact angle is zero, equation (23) Becomes:

$$P = 2\sigma/r$$

Where:

Where **r** is the equivalent radius of the pore (cm), **σ** is the surface tension (kg/s^2) = (Dyne/cm), **θ** is the contact angle between the water and the pore wall (usually assumed to be zero), ρ_w is the density of water (kg/m^3), **g** is the gravitational acceleration (m/s^2), and **h** is the soil water suction (cm of water). **h**: the height of water in the capillary tube (m)

Where:

- **P**: the pressure differences between water and the top of the column and the atmosphere.
- In terms of diameter of the capillary table, equation (22) becomes
- **$h=4\sigma \cos(\theta)/(d\rho wg)$**

Where:

- **h**: the equilibrium height of water. Substituting for he properties of water at 20ºC and assuming a contact angle of 0 degree gives
- **$h=0.15/r$**
- **h and r have the units of cm.**

Example page 73 (evaluation of soil structure)

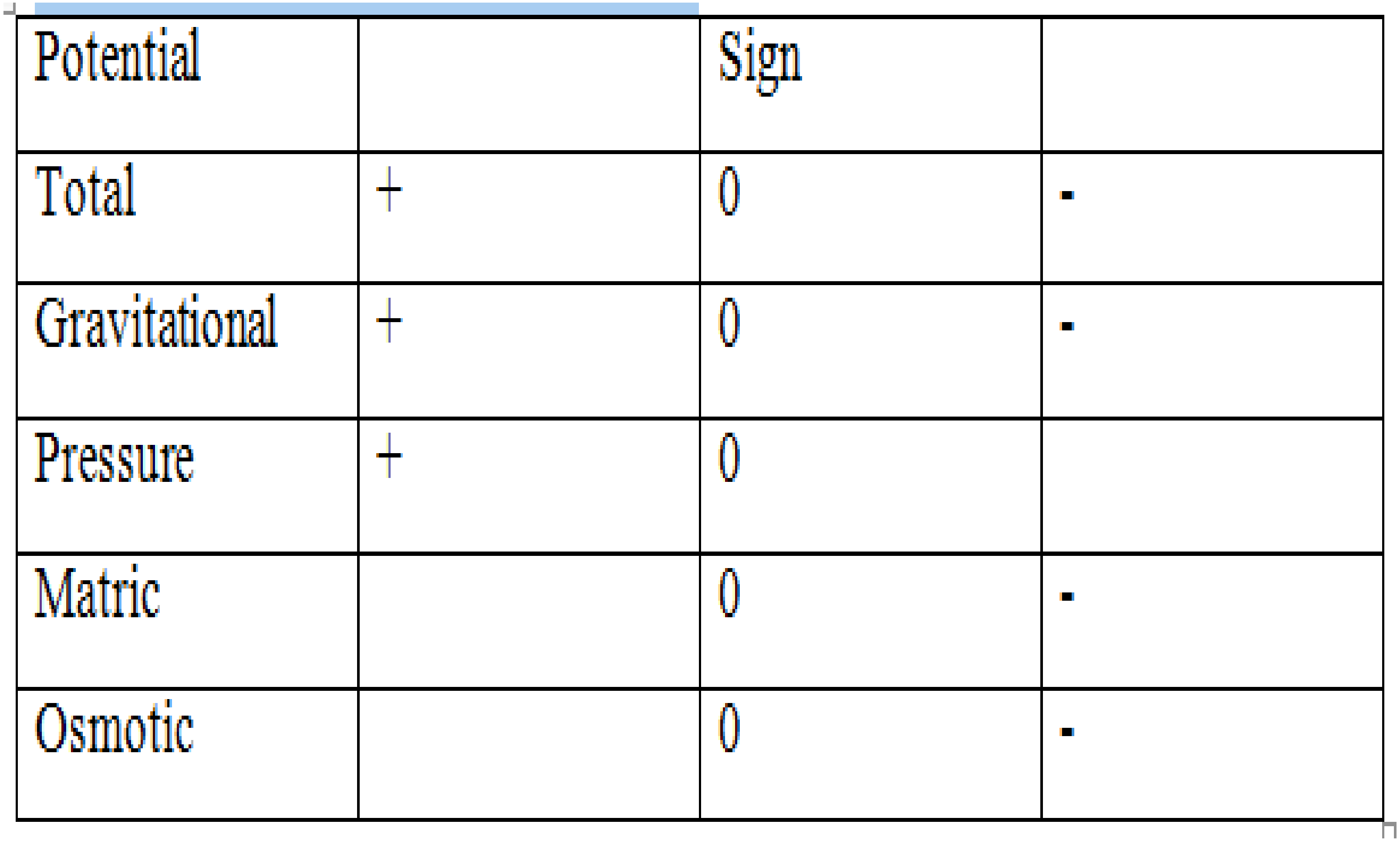
- Show the influence of soil water suction on the equivalent pore radius at suctions of -10 and –100 cm:
- $r = 0.15/10 = 0.015$ cm and $r = 0.15/100 = 0.0015$ cm.

The results for water as a function of diameter are:

- h = 0.0148 m r = 0.1 cm
- h = 0.10 bar or 100 cm r = 0.0015 cm
- h = 3 m (FC) r = 0.0005 cm

- h = 150 m (PWP) r = 0.00001 cm

Summary of soil water potential concepts:



Potential		Sign	
Total	+	0	-
Gravitational	+	0	-
Pressure	+	0	
Matric		0	-
Osmotic		0	-

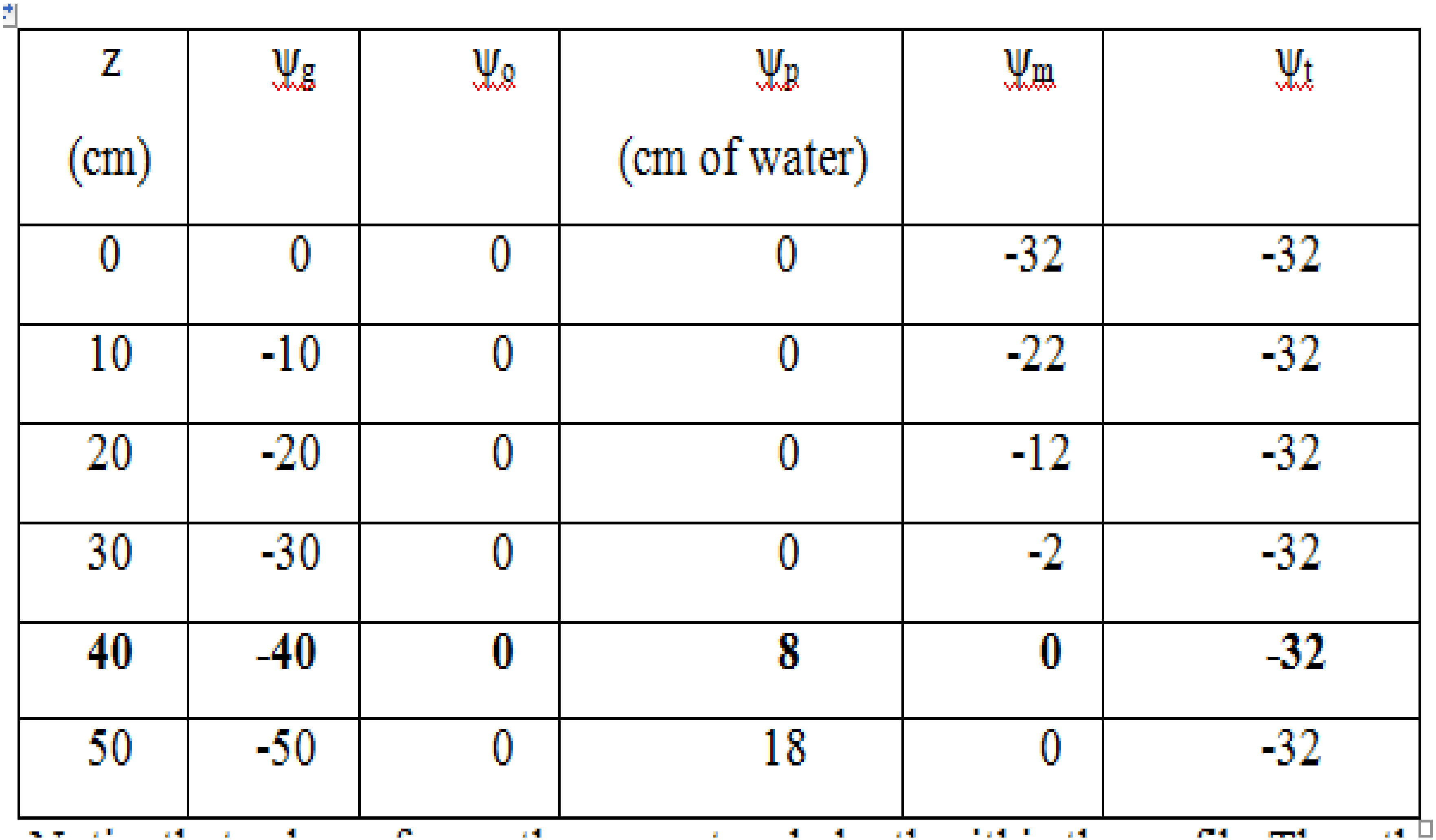
- Soil water moves from a region of higher total potential energy to a region of lower total potential energy.
- Potential Distribution of Soil Water Under Field Conditions:
- The driving force to move this water is the differences in the total potential energy divided by the distance between the points. This ratio is known as the hydraulic gradient
- Water potential can be expressed as
- $\boldsymbol{\Psi_t = \psi_g + \psi_p + \psi_m}$ **(40)**
- This particular combination of potentials is commonly known as the hydraulic potential.

Ch8: Soil Water Principles (Numerical examples on water potential)

Example 1 (Page 193-194)

- A Dewitt silt loam has a perched (suspended) water table above the clayey B horizon. Assuming that the B horizon is at the 40 cm depth, and the height of water ponded above this horizon is 8 cm. Calculate the vertical distributions of the total soil water potential and the component potential energies under ***hydraulic equilibrium*** conditions to the 50 cm depth in 10 cm intervals. Show the effects on the potential energies of moving the reference level from the soil surface to the water table.

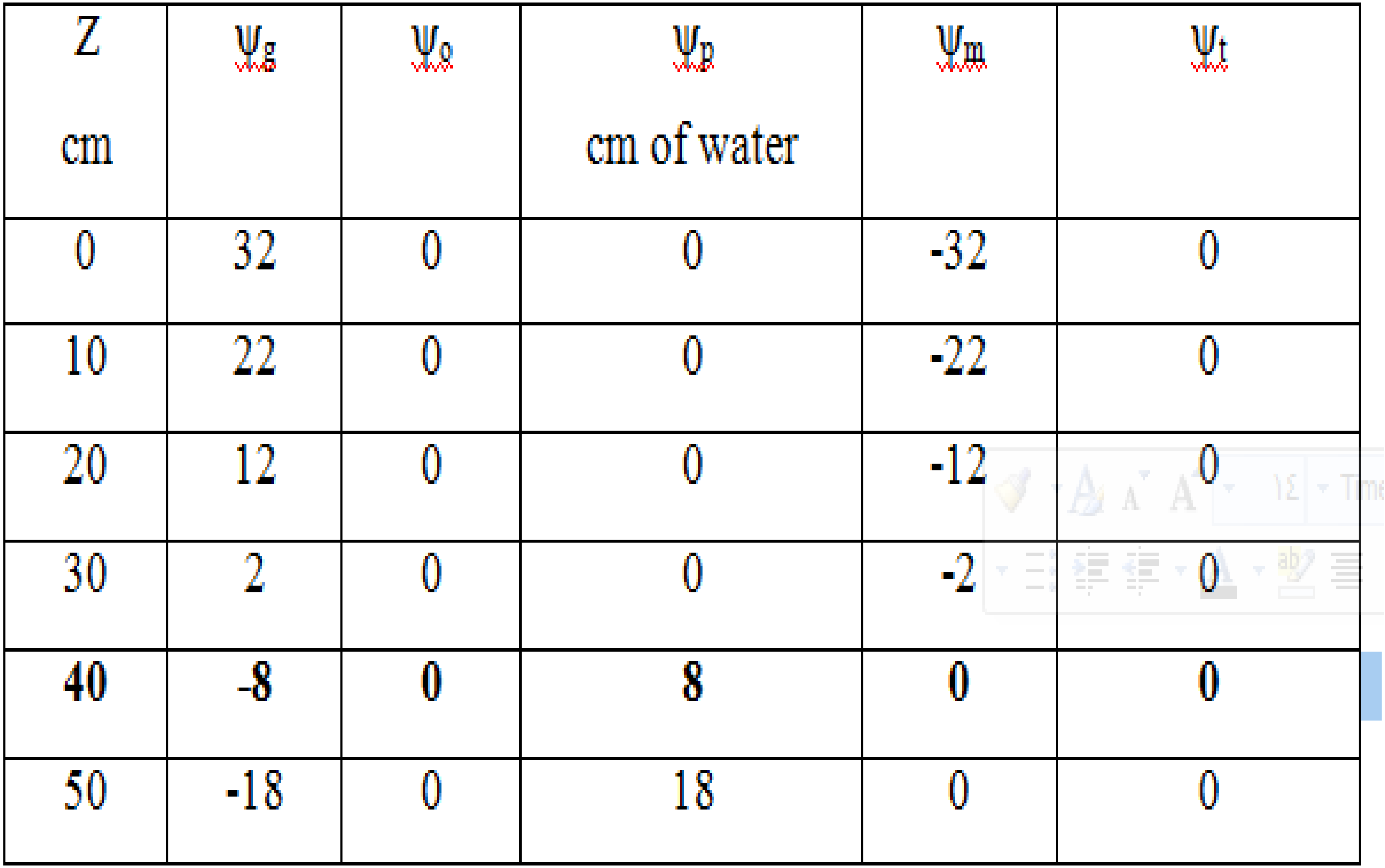
1- <u>The reference level is placed at the soil surface</u>. The potential energy distributions within the profile are tabulated below.



Z (cm)	Ψ_g	Ψ_o	Ψ_p (cm of water)	Ψ_m	Ψ_t
0	0	0	0	-32	-32
10	-10	0	0	-22	-32
20	-20	0	0	-12	-32
30	-30	0	0	-2	-32
40	**-40**	**0**	**8**	**0**	**-32**
50	-50	0	18	0	-32

- Notice that values of ψ_t are the same at each depth within the profile. Thus, the flow of water in this profile is zero and the system is at hydraulic equilibrium.

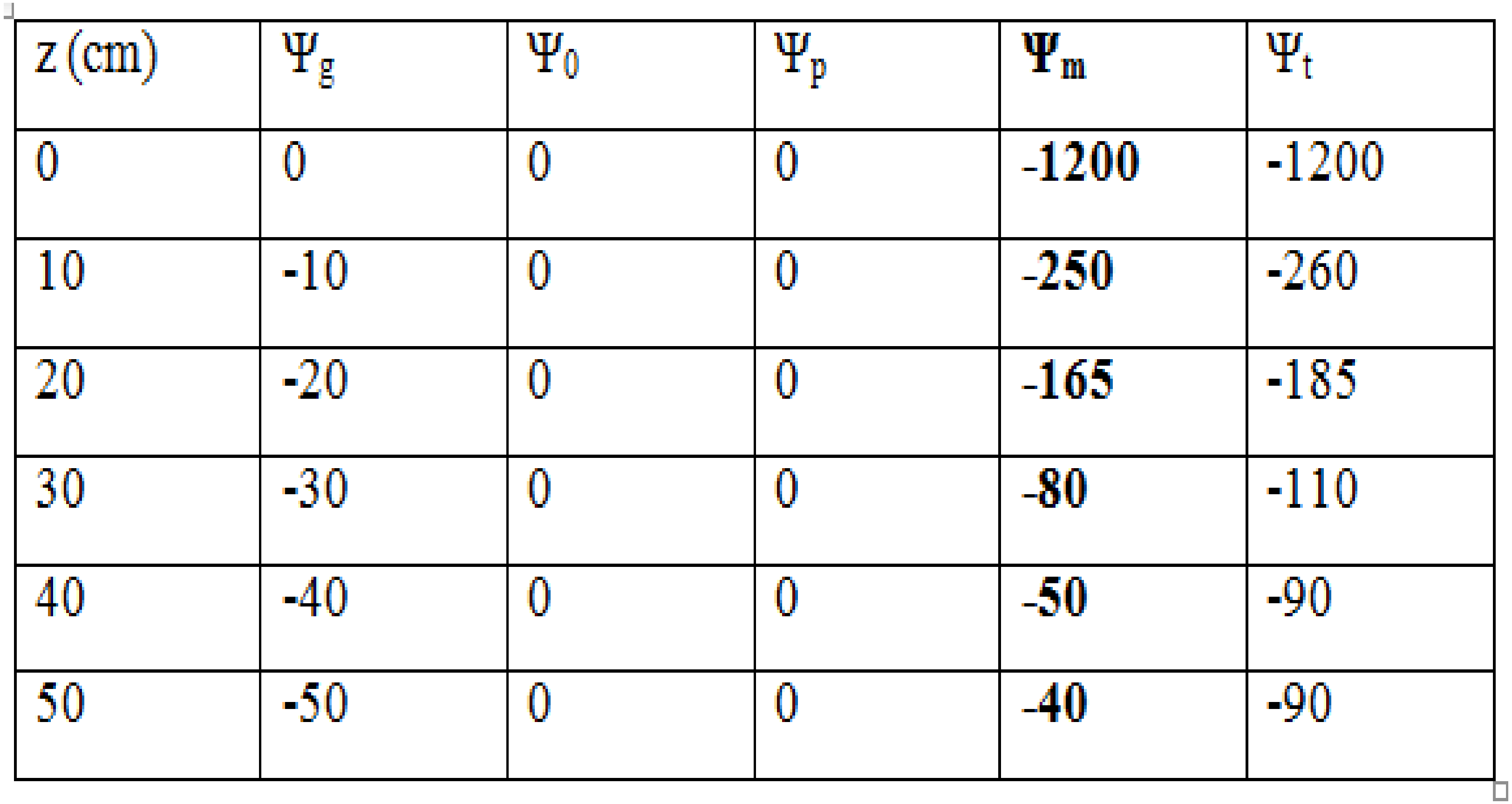
2- The reference level is placed at the water table



Z cm	ψ_g	ψ_o	ψ_p cm of water	ψ_m	ψ_t
0	32	0	0	-32	0
10	22	0	0	-22	0
20	12	0	0	-12	0
30	2	0	0	-2	0
40	**-8**	**0**	**8**	**0**	**0**
50	-18	0	18	0	0

- A comparison of the results in two tables that the magnitudes of ψ_t and one of the component potentials (ψ_g) changed depending on the placement of the reference level. The other potential energies did not change. However, the difference between ψ_t from one location to another did not vary in the profile regardless of the placement of the reference level. Under moist conditions in the field, equilibrium conditions do not occur. Thus, water moves within the profile in various directions and it varies rates. Under this condition, ψ_t is unequal within the soil profile, and non-equilibrium conditions exit

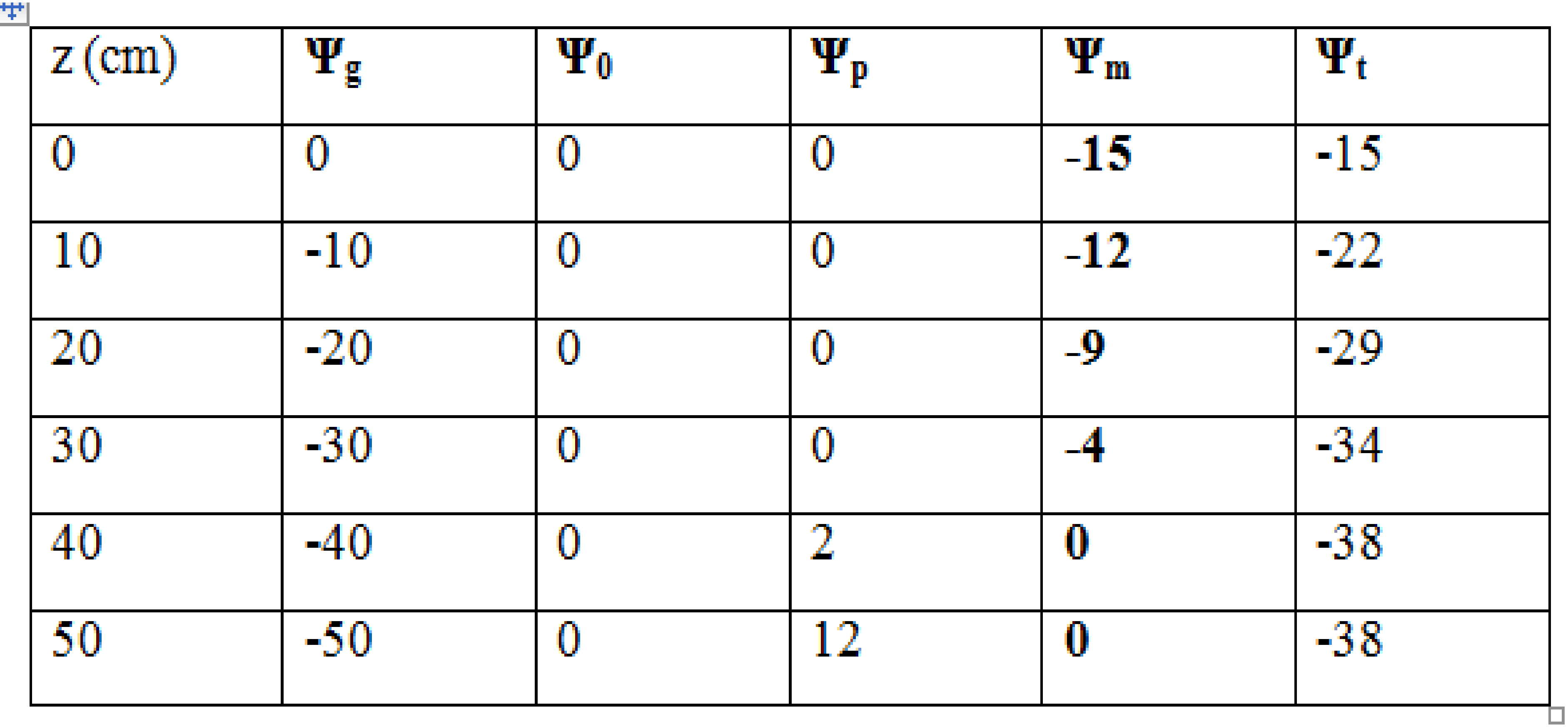
- **Example (page 194- 195):**
- Assume that water is evaporating from ***an unsaturated*** DeWitt silt loam. Given the values of $\boldsymbol{\psi_m}$ in the table, find values of $\boldsymbol{\Psi_t}$ and its component potential to the 50 cm depth in the 10 cm increments. Assume that the reference level is placed at the soil surface.



z (cm)	Ψ_g	Ψ_0	Ψ_p	$\mathbf{\Psi_m}$	Ψ_t
0	0	0	0	**-1200**	-1200
10	-10	0	0	**-250**	-260
20	-20	0	0	**-165**	-185
30	-30	0	0	**-80**	-110
40	-40	0	0	**-50**	-90
50	-50	0	0	**-40**	-90

In this example, values of $\boldsymbol{\psi_t}$ increased from the surface to the 40-cm depth: therefore, soil water was moving toward the soil surface. Hydraulic equilibrium conditions existed between the 40 and 50-cm depths.

- **Example (page 195):**
- Given the values of $\mathbf{\Psi_{m,}}$ assume that 10 cm tile drain, which contained water at a height of 2 cm, was placed on top of B horizon of the DeWitt silt loam and water moves to the drain. The clayey B horizon is located at the 40 cm depth. Find the values of $\boldsymbol{\psi_t}$ and its component potentials to the 50 cm depth in 10 cm increments. Assume that the reference level is at the soil surface.



z (cm)	Ψ_g	Ψ_0	Ψ_p	Ψ_m	Ψ_t
0	0	0	0	-15	-15
10	-10	0	0	-12	-22
20	-20	0	0	-9	-29
30	-30	0	0	-4	-34
40	-40	0	2	0	-38
50	-50	0	12	0	-38

These data show that $\boldsymbol{\Psi}_t$ decreased with depth to the tile drain. And thus, water was moving toward the drain. The height of water pounded in the drain was 2 cm.

8.4 Methods for soil water determination:

1- Water contents methods:

- Methods used to determine soil water content include gravimetric techniques, nuclear techniques such as neutron probe and gamma-ray attenuation, and electromagnetic techniques such as time domain reflectance.
- **Gravimetric techniques:** In situ samples are taken and dried in oven (105 ^oC for 24 hours). The gravimetric soil water content is calculated as the ratio of the water lost to the oven-dry soil weight.
- **Nuclear techniques**: Non-destructive methods rely upon sensor placed in soil with an evaluation unit connected to its cables at the time of measurement. The sensor is inserted each time an observation is desired into the soil by a vertical hole lined with an access tube. A repetitive measurement through a long-term period does not disturb the soil.

- **Gamma-ray attenuation:** This method assumes that scattering and absorption of gamma rays are related to the density of matter in their path and that the density of soil remains relatively constant as water content changes.
- **Electromagnetic techniques**: Electromagnetic techniques include those methods that depend upon the effects of water content on the electrical properties of soil.
- **<u>2- Potential energy methods:</u>**

2.1 Piezometer:

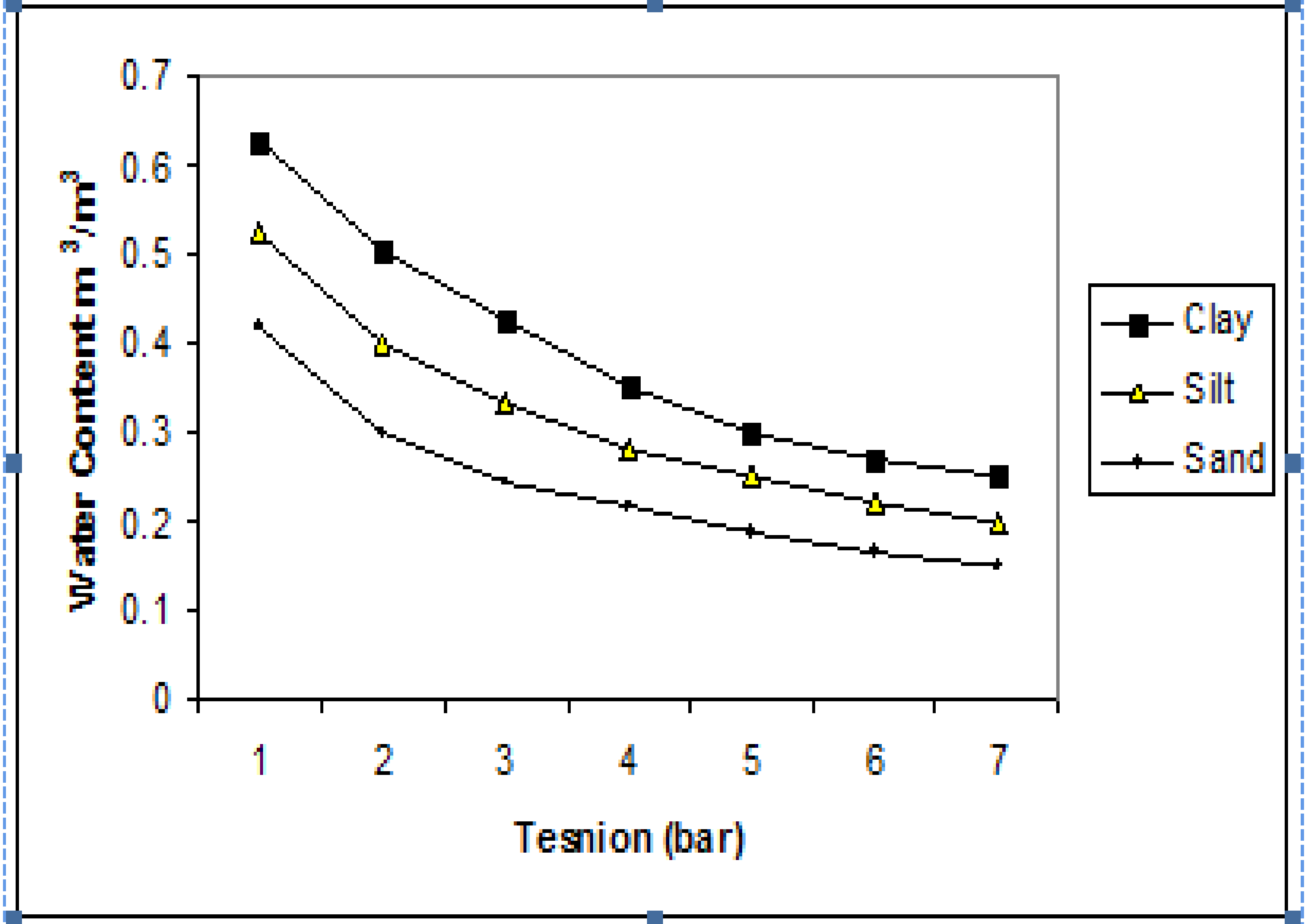
2.2 Tensiometer.

2.3 Psychrometer.

- The potential of an aqueous solution at its reference water level equals the potential of water vapor immediately above the water level.
- Psychrometers have to be extremely accurate.
- The accuracy of the procedure is usually on the order of h = ±100 cm

- **8.5 Soil water retention curve:**

The relationship between water content and matric potential is a fundamental physical property of a soil and historically has been called by at least three terms: soil water characteristic, soil water release curve, and the soil water retention curve.

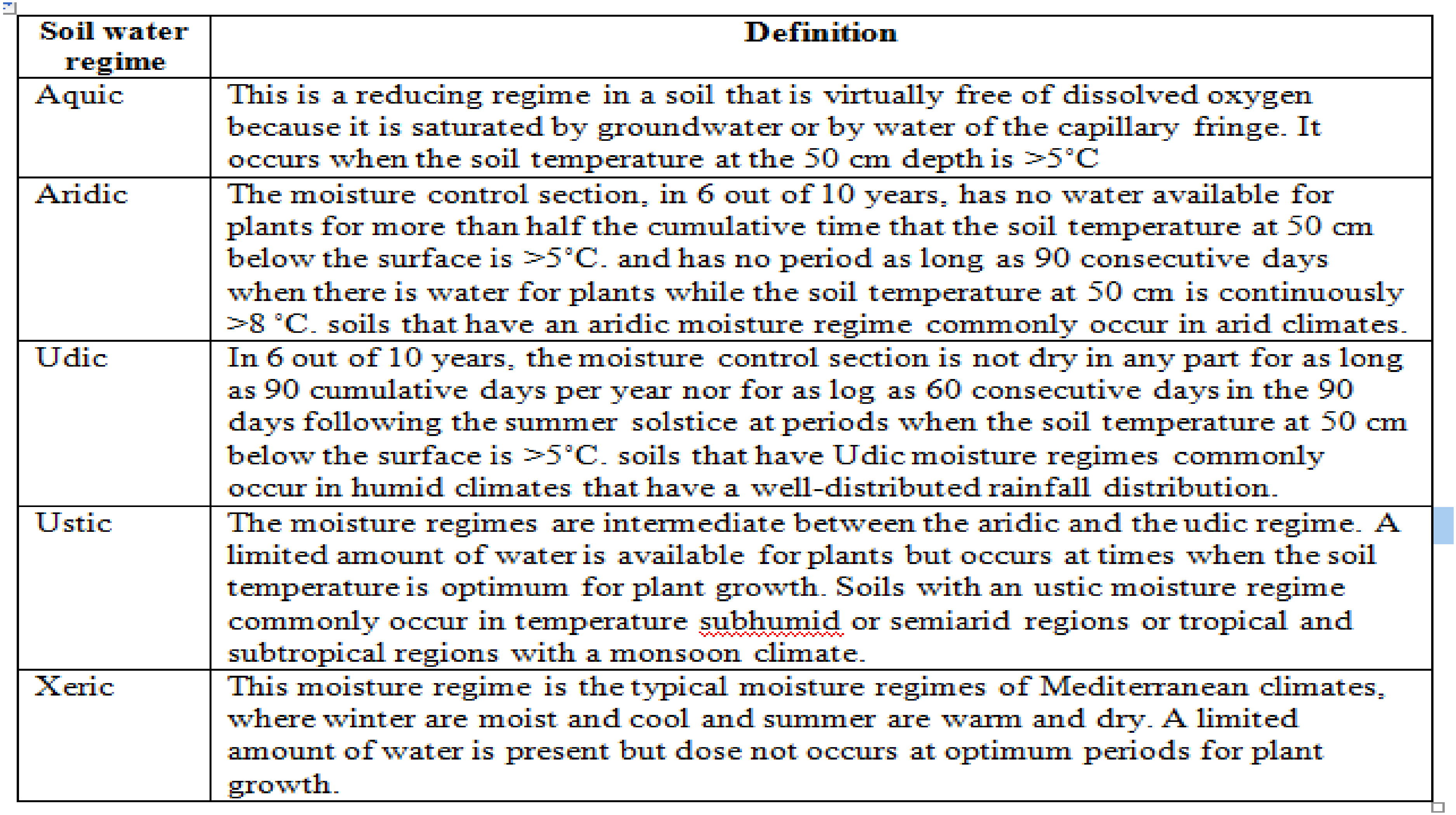


Water Content m 3 /m 3
0.7
0.6
0.5
0.4
0.3
0.2
0.1
0
1
2
3
4
5
6
7
Tesnion (bar)
Clay
Silt
Sand

Water terms in the soil classification system:

Soil water regimes:

Table 8.8. Classes of soil water regimes in soil taxonomy.



Soil water regime	Definition
Aquic	This is a reducing regime in a soil that is virtually free of dissolved oxygen because it is saturated by groundwater or by water of the capillary fringe. It occurs when the soil temperature at the 50 cm depth is >5°C
Aridic	The moisture control section, in 6 out of 10 years, has no water available for plants for more than half the cumulative time that the soil temperature at 50 cm below the surface is >5°C. and has no period as long as 90 consecutive days when there is water for plants while the soil temperature at 50 cm is continuously >8 °C. soils that have an aridic moisture regime commonly occur in arid climates.
Udic	In 6 out of 10 years, the moisture control section is not dry in any part for as long as 90 cumulative days per year nor for as log as 60 consecutive days in the 90 days following the summer solstice at periods when the soil temperature at 50 cm below the surface is >5°C. soils that have Udic moisture regimes commonly occur in humid climates that have a well-distributed rainfall distribution.
Ustic	The moisture regimes are intermediate between the aridic and the udic regime. A limited amount of water is available for plants but occurs at times when the soil temperature is optimum for plant growth. Soils with an ustic moisture regime commonly occur in temperature subhumid or semiarid regions or tropical and subtropical regions with a monsoon climate.
Xeric	This moisture regime is the typical moisture regimes of Mediterranean climates, where winter are moist and cool and summer are warm and dry. A limited amount of water is present but dose not occurs at optimum periods for plant growth.

Chapter 9: Soil water flow

9.1 Introduction

The flow of water in the soil is important for calculations of water balance and for the redistribution of water, solutes, and energy within the soil-plant –atmosphere system.

- **Flow of water in model systems: Tubes and Pipes.**

Flow of water in straight, narrow, cylindrical tubes varies with several properties of water and characteristics of the tube. If we consider only non-turbulent flow conditions, the velocity distribution in the tube is near zero at the walls because of adhesion of water to solid surface. The velocity is maximal on the axis and constant on the cylindrical surfaces that are concentric about the axis. The movement of the fluid is in response to pressure gradient $\Delta P/L$ acting in the axial direction.

The velocity or flux of fluid in a straight tube can be calculated from Poiseulle's law, which is:

- **$Q/t = \pi^4 \Delta P/8\eta L$**
- Where:
- **Q/t**: the volume per unit time known as the flux (m^3/s)
- **η**: the viscosity of the fluid($kg/m.s$)
- **$\Delta P/L$**: the pressure gradient (N/m^3).

Dividing the flux by cross sectional area of tube gives the flux density.

- **$q = Q/At = r^2 \Delta P/8\eta L$**
- The quantitative criterion to determine whether flow of fluid is laminar or turbulent is the Reynolds number (Re). Values of **Re** are dimensionless and for a cylindrical tube can be calculated from:
- **$Re = qd\rho_f/\eta$**

Where:

- **q**: the mean water flux density (m/s)
- **d**: the effective tube or pore diameter (m)
- $\rho_{f:}$ the density of fluid (kg/m^3).
- **η**: the viscosity of the fluid ($kg/m.s$)
- In a straight tube, turbulence occurs when Re is greater than 100-2000. For transport of water in soils, $Re < 1$
- Thus laminar flow of water normally found in the soils.

- **9.2 Darcy’s law**
- Unfortunately, soil pores do not present uniform, smooth tubes but are highly irregular, tortuous, and interconnected.
- Volume discharge from a one-dimensional column of soil is proportional to the column cross-sectional area, the length of the column, and the total hydraulic head loss:
- Q/t is proportional to $A\Delta H/L$
- The ratio $\Delta H/L$ is known as the hydraulic gradient (m/m) and is the driving force for moving water through soil. The specific discharge rate is known as the flux density and for water is symbolized by q with units of $m^3/m^2.s$.

This equation, which as Darcy's law, is written in incremental from as:

- **$q = Q/At = - K (\Delta H/L)$**

Where:

- **K**: the proportionality coefficient known as the hydraulic conductivity (m/s).
- The flow rate of water in the soil pores can also be calculated by dividing the flux density by the volumetric soil water content. Mathematically, this becomes:
- **$q_z / \theta_v = v$** (used in chapter 5, convective flow),
- and is known as the pore water velocity . the magnitude of v will be grater than q since $\theta_v < 1$. The pore water velocity also is an average velocity.
- The relation between the hydraulic conductivity and the properties of fluid media is:
- **$K = k(\rho_f g/\eta)$**

Where:

- **K**: the hydraulic conductivity (m/s).
- **ρ_f**: the fluid density (kg/m^3).
- **g**: the gravitational acceleration (m/s^2).
- **η**: the viscosity (kg/m.s)
- **k**: the intrinsic permeability (m^2)

- \- The ratio of the viscosity to the density is known as the ***kinematic viscosity*** (m^2/s)
- \- The ratio ($\rho_f g/\eta$) characterizes the properties of the fluid and is known as the ***fluidity***.

9.3 Transport of soil water under saturated conditions:

- Transport of water in saturated, homogeneous soil columns:
- For saturated flow of soil water, the total hydraulic head is composed of the sum of the pressure and gravitational heads. Mathematically, this becomes:
- $\mathbf{H = H_g + H_p}$
- $\mathbf{H_g}$: it is determined at any point by the height of the point relative to an arbitrary placed reference level.
- $\mathbf{H_p}$: is determined by the height of water column above that point.
- $\mathbf{q = Q/At = K[(H_i - H_0)/L]}$

Where:

- $\mathbf{H_i}$: the total hydraulic head at the inflow (m)
- $\mathbf{H_0}$: the total hydraulic head at the outflow (m)

General saturated-flow equation :

- **$q = Q/At = -K(\Delta H/\Delta z) = -K[(H_2-H_1)/(z_2-z_1)]$**

Darcy’s law can also be written in differential forms as:

- **$q = -K\delta H/\delta z$**
- **$q_x = - K_x\delta H/\delta x$**
- **$q_y = -K_y\delta H/\delta y$**
- **$q_z = K_z\delta H/\delta z$**

Additional material (Advanced)

- For one –directional flow of water without plant root extraction, the mass conservation equation can be written as:
- **$\delta\theta v/\delta t =-\delta q/\delta z$**
- In saturated soil, $\delta\theta v/\delta t=0$, and if Ksat is assumed to be constant, the equation becomes
- **$Ksat(\delta^2H/\delta z^2) =0$)**

Where Kx, Ky and Kz are the saturated hydraulic conductivities in the x, y, z directions, respectively. When the Ksat have the same magnitude regardless of the direction, then equation (30) reduces to :

- **$\delta^2H/\delta x^2+\delta^2H/\delta y^2+\delta^2H/\delta z^2=\Delta^2H=0$**

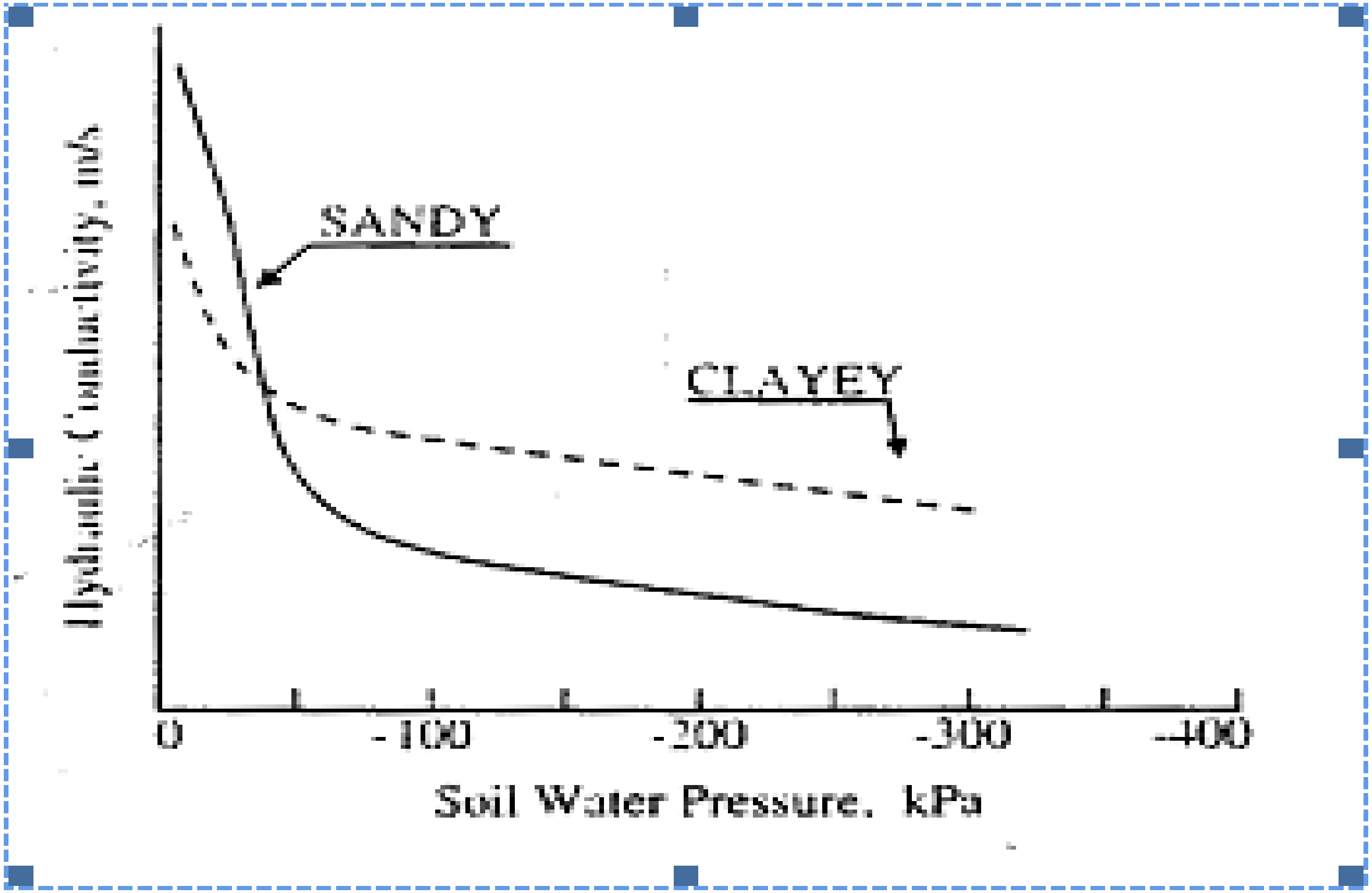
- Which is known as the ***Laplace equation***. This equation can be solved analytically for certain simple boundary conditions. In most cases of interest, however, numerical techniques are needed to solve the equation.

9.4 Transport of soil water under unsaturated conditions

- Most of the processes involving soil water flow in the field occur while the soil is in an unsaturated sate. Unsaturated-flow processes are, in general, much more complicated and difficult to describe quantitatively and mathematically than saturated flow.
- **Darcy’s law for unsaturated flow:**
- For unsaturated flow, Darcy’s law written as:
- $\mathbf{q = -K(\theta_v)\delta H / \delta z}$
- Where, expressed on a weight basis, **q** is the soil water flux density (m/s)
- **H**: is the total soil water potential (m).
- **Z**: the spatial coordinate (m).

K (θ_v): the hydraulic conductivity (m/s), which is dependent on the volumetric soil water content (θ_v) and matric potential (h).

Unsaturated hydraulic conductivity



Soil water diffusivity (advanced):

- To simplify the mathematical and experimental treatment of unsaturated-flow processes it is often advantageous to change the flow equation into a frame analogous to the equations of diffusion of solutes and conduction of heat. With use of the chain rule, we can write the hydraulic gradient as:
- **$\delta h/\delta z = (\delta h/\delta \theta_v)(\delta \theta_v/\delta z)$**
- Where: The specific water capacity is defined as C (θ_v) or C(h) = $\delta \theta_v/\delta h$ and has units of m-1. For horizontal flow of soil water in the y direction.
- **$q= -K(\theta_v)\, \delta H/\delta y = -[K(\theta_v)/C(\theta_v)]\, \delta \theta_v/\delta y$**
- We define the soil water diffusivity as: **$(\theta_v) = K(\theta_v)/C(\theta_v)$**
- Which has SI units of m²/s.the diffusivity from of the soil water flow equation becomes:
- **$q = -D(\theta_v)\, \delta\theta_v/\delta y$**

9.5 Methods of measuring Hydraulic conductivity:

Measurement of K_{sat} take advantages of Darcy’s equation and assume constant temperature.

1- The constant head permeameter

- Recommended for values of Ksat greater than 10-5 m/s. With this method, K_{sat} is determined from
- $\mathbf{K= QL/A\Delta H_t}$
-

2- The falling head permeameter:

- Recommended for soils with lower flow rates or those that have **Ksat** less than 10-5 m/s. This is a transient-state method. The resulting equation for calculation of K_{sat} is
- $\mathbf{K= [aL/A\ (t-t_0)\ \ln(h_0/h)]}$

Where:

- **a**: the cross sectional area of the small tube (i.e., standpipe)
- **L**: the length of the sample
- **A**: the cross-sectional area in the core sample.
- **h0** and **h** are the hydraulic heads at t_0 (the initial time) and t (the final time), respectively.
- A plot of **ln(h0/h)** versus time results in a straight line and the slope is proportional to K.

Other methods:

1- The core method also can be used to measure **$K(\theta_v)$ or K(h)** within a range of 0-20 Kpa. As soil core can be placed is pressure chamber with different pressure between top and bottom (gradient) and flow is measured.

2- Double-tube method and auger hole method.

3- Instantaneous profile method in the field between saturation and 0–30kPa.

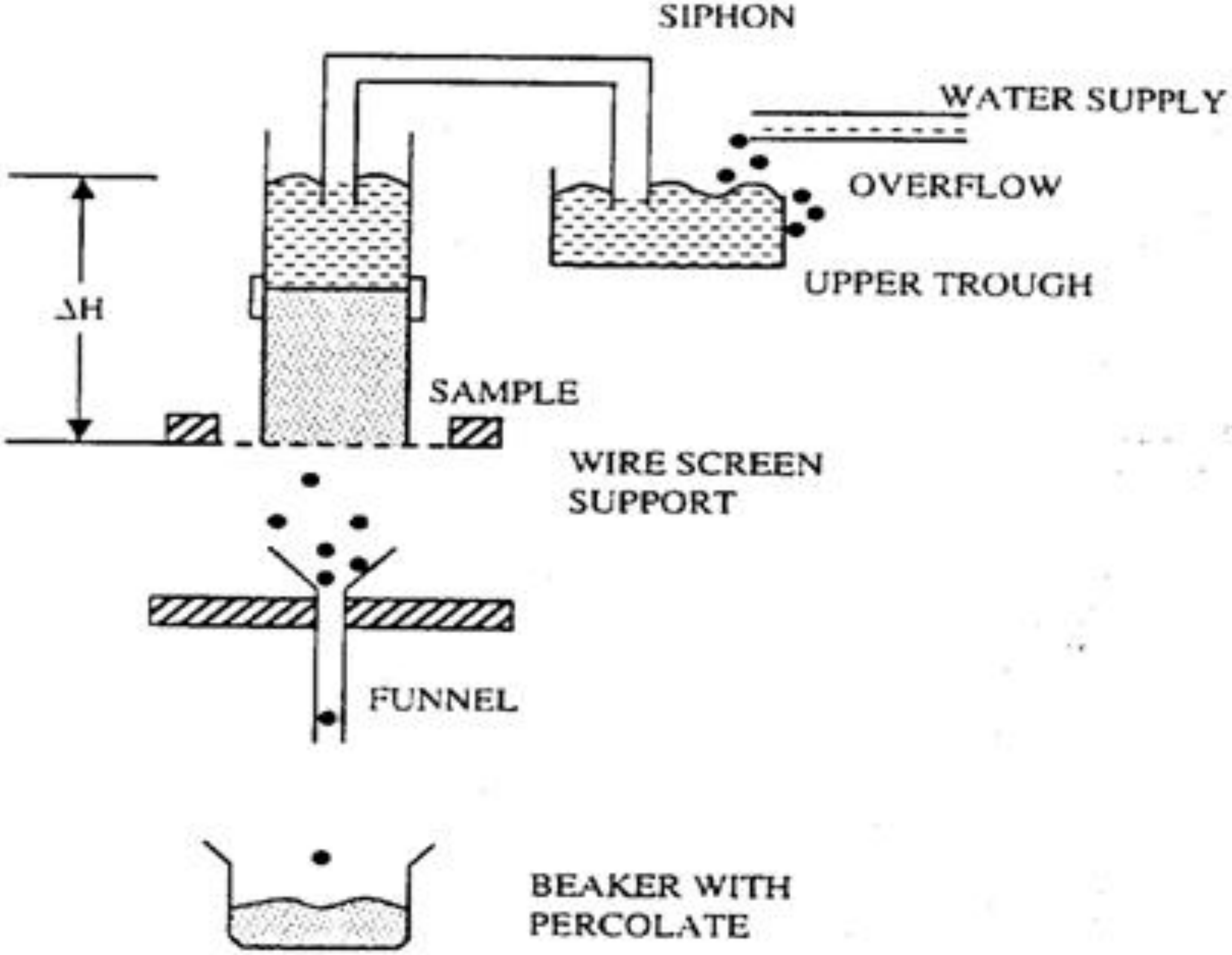


Figure 9.9. Diagram of the constant-head system for measuring saturated hydraulic conductivity. (Redrawn from Klute 1965.)

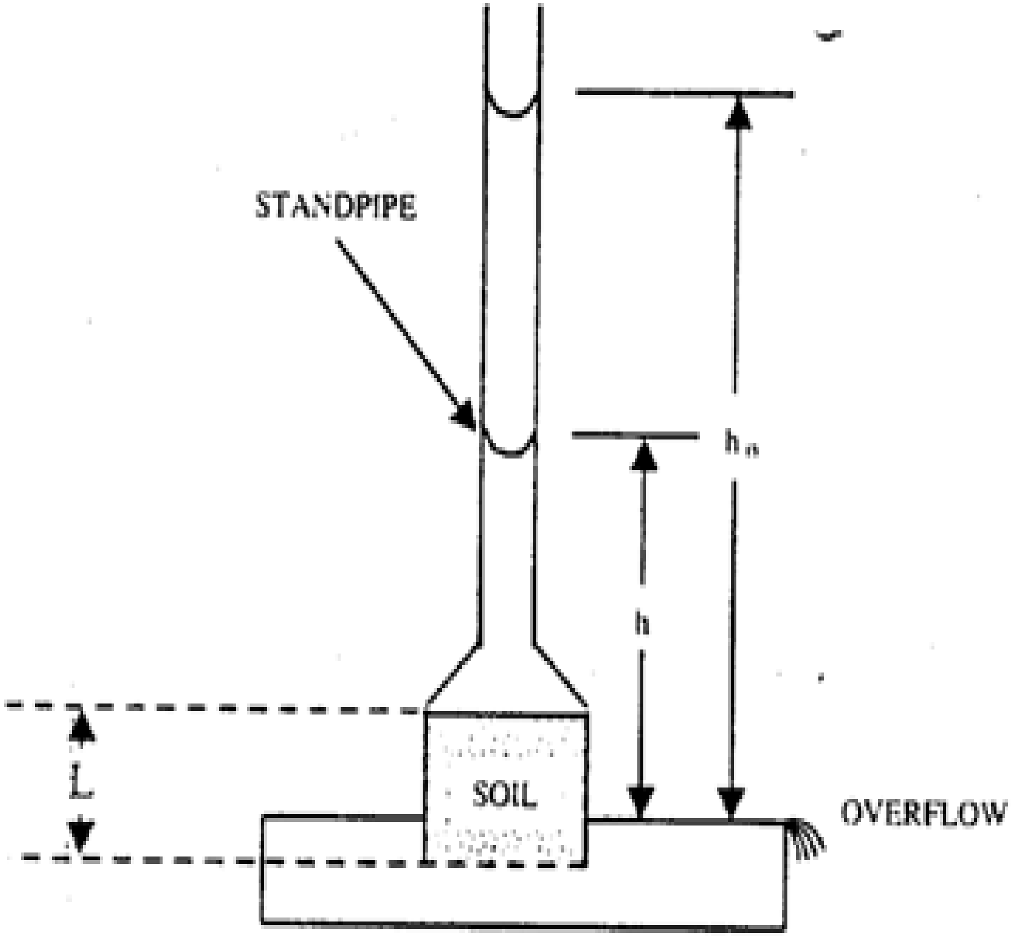


Figure 9.10. Diagram of the falling-head system for measuring saturated hydraulic conductivity. (Redrawn from Klute 1965.)

Chapter 10: Soil water flow processes in the field

Infiltration:

- It is the downward entry of water into the soil from rainfall, irrigation, or snowmelt. The infiltration process is dynamic and complex and varies with several soil, vegetation, and climatic parameters.

Infiltration rate:

- It's the flux density of water passing through the soil surface and flowing into the soil profile.
- **I= Q/At**

Where:

- **Q**: the quantity or volume of water absorbed or infiltrating by unit area A of soil surface per unit time t. the SI units of I are $m^3/ m^2.s$, which results in units such as m/s.

<u>Cumulative infiltration:</u>

- The second important infiltration parameter is the cumulative infiltration, I, which has SI units of length (e.g., m: it also may have units of cm or mm) and represents the equivalent depth of water infiltrated.
- I = (3).
- The second method of calculating (I) is by integrating the change in volumetric soil water content distribution in the profile between the surface (z=0) and the depth to the wetted front ($z=z_f$). Mathematically, this is represented by:
- I= (4)

<u>Where:</u>

- **z_f.:** the depth of the wetting front.
- **$\Delta\theta_v$**: the differences between the initial volumetric water content. $\theta_v(z, t_1)$ and the final volumetric water content, $\theta_v(z, t_2)$. The wetting front is the boundary between the wetted region and the dry region of soil during infiltration.

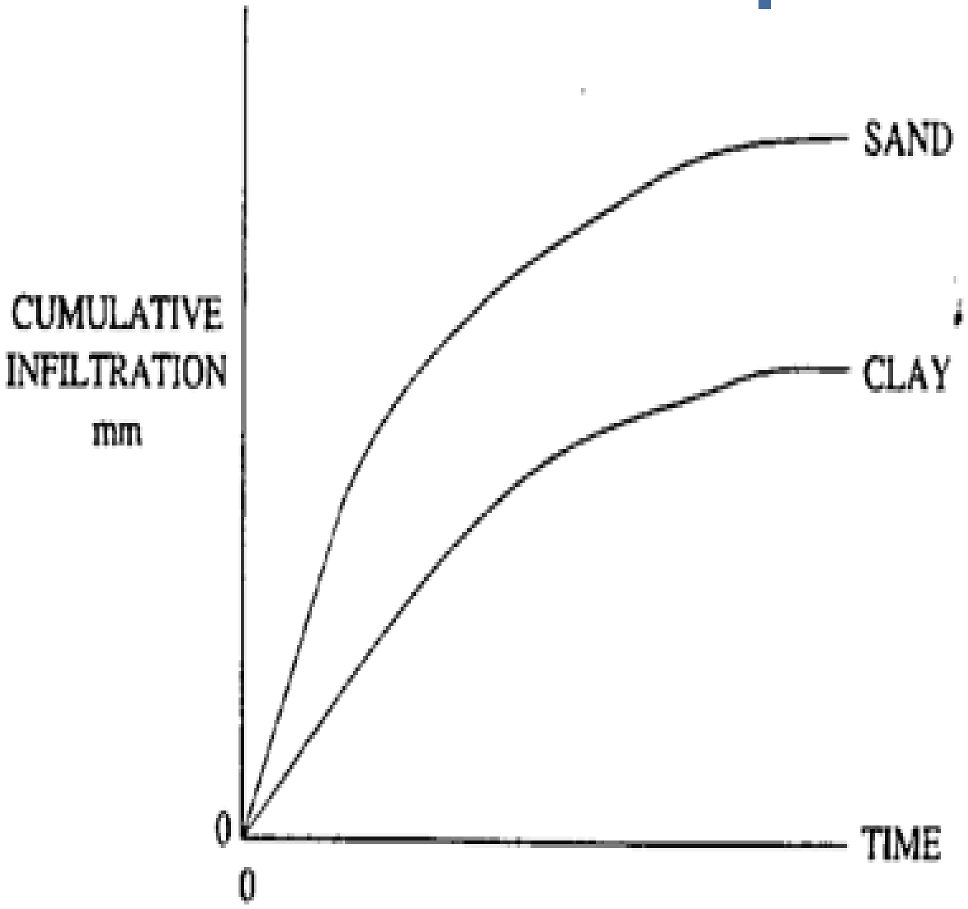


Figure 10.3. Relationship between the cumulative infiltration of water and time for two soil textures.

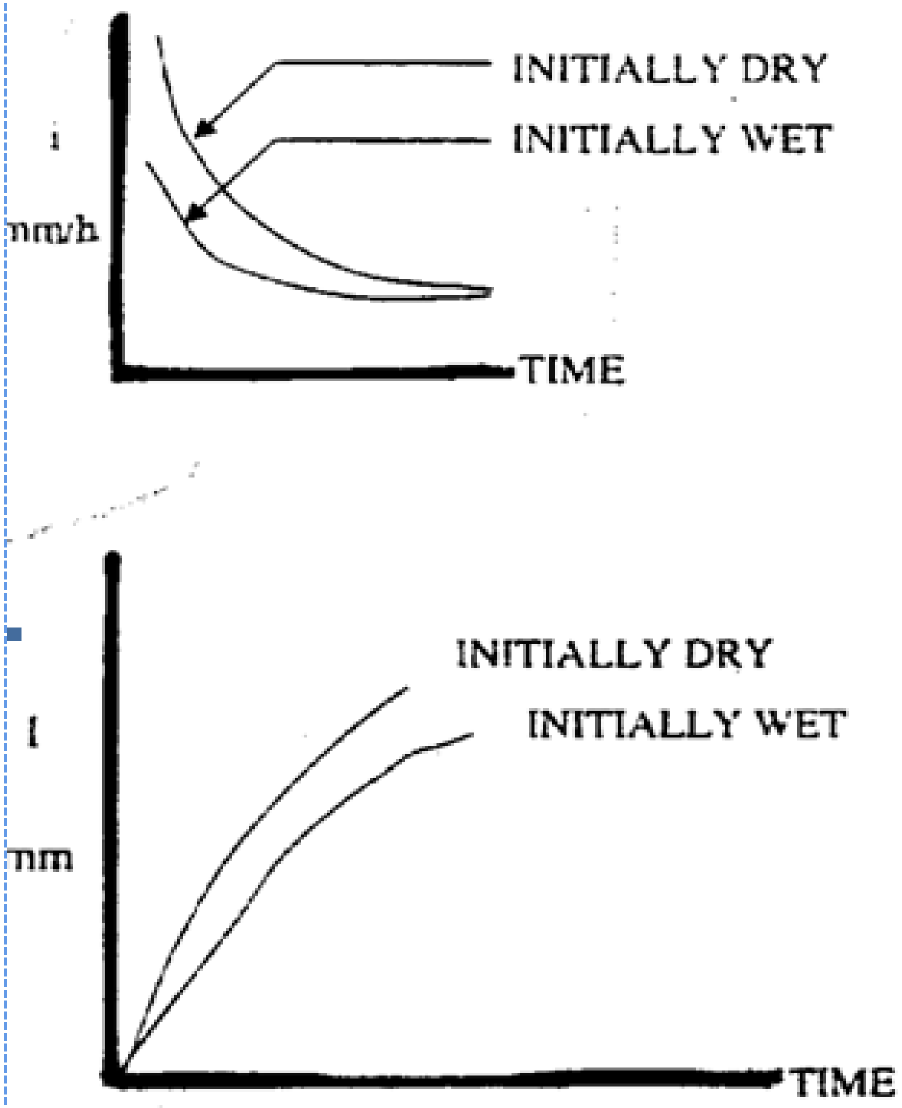


Figure 10.4. Relationships between the infiltration rate, cumulative infiltration, and time under two initial soil water contents at the soil surface. The symbols i and I represent the nfiltration rate and cumulative infiltration, respectively.

Factors that affect infiltration:

- **1. Initial wetness and matric potential:**
- The greater the initial water contents of the soil, the lower the initial infiltration rate and the quicker the attainment of a constant infiltration rate.
- **2. Texture:**
- The saturated hydraulic conductivity is higher in coarse-textured soils than in fine textured soils. Similarly, the infiltration of water is higher in coarse –textural soils than in fine-textural soils.

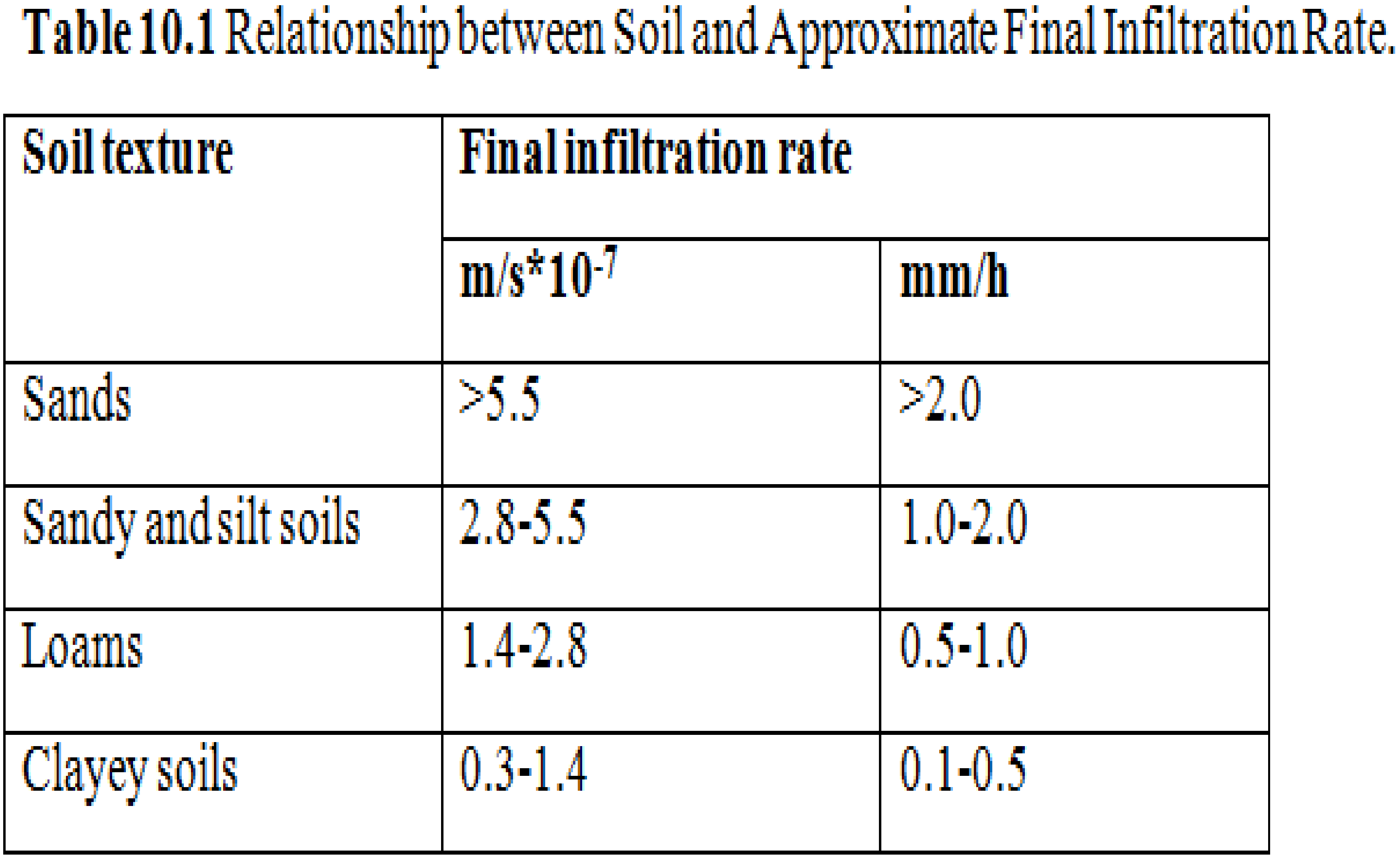


Table 10.1 Relationship between Soil and Approximate Final Infiltration Rate.

Soil texture	**Final infiltration rate**	
	m/s*10^{-7}	**mm/h**
Sands	>5.5	>2.0
Sandy and silt soils	2.8-5.5	1.0-2.0
Loams	1.4-2.8	0.5-1.0
Clayey soils	0.3-1.4	0.1-0.5

- **<u>3. Structure:</u>**
- In general, differences in soil structure and aggregate size influence water movement in the same way as texture.
- **<u>4. Tillage:</u>**
- Tillage such as moldboard plowing and chiseling tends to increase soil porosity at the soil surface by 10-20%, depending on soil texture.
-
- **<u>5. Type of clay :</u>**
- Soils containing montmorillonitic clays crack upon drying, thus providing a high initial infiltration rate. This soil may also swell quickly upon wetting, thereby reducing the infiltration rate.
-
- **<u>6. Vegetative cover:</u>**
- Bare soils tend to have lower infiltration rate than soils protected by a vegetative cover, on bare soils, the impacting raindrops tend to puddle the soil, which results from the decomposition of aggregates, leading to the formation of surface crust. A crust impedes infiltration. The net result is a lowering of infiltration.

7. Rainfall intensity:

- In high intensity rains, the raindrops tend to be large and have more energy when they strike the soil. Thus, high intensity rains are more effective in sealing the soil surface than low –intensity rains.

8. Slope:

- On a steep slope, the precipitated water ends to runoff rapidly and, thus, has less opportunity to infiltrate than on a milder slope.

9. Channel characteristics:

- The shape of the irrigation channel dose not normally affect infiltration significantly, as the entire border is covered with water in border or flood irrigation. The shape dose influences infiltration in furrow irrigation because the hydraulic conditions vary with shape, roughness, slope, and flow rate at any point in the furrow.

10. Water temperature:

- As the temperature of the soil profile increases, the rate of water infiltration also increases. Temperature affects the viscosity and density of water (i.e., the fluidity).
- **11. Air entrapment:**
- The entrapment of air within the profile reduces the infiltration rate.

12. Soil salinity:

- High-saline soils have low infiltration rates. Also, swelling and depression of clay in response to electrolyte composition and concentration in the flowing water change the size of the conducting pores.