18
Soil Air and Aeration

18.1 AIR

Earth is surrounded by a gaseous envelope of air about 80 km thick called the atmosphere. The origin of Earth's atmosphere is still a subject of speculation. One theory seems fairly certain that some five billion years ago when Earth was formed, it was extremely hot and did not have an atmosphere. It is generally accepted that the first atmosphere, created when Earth cooled down, consisted of helium (He), hydrogen (H_2), ammonia (NH_3), and methane (CH_4). Assuming that five billion years ago volcanoes emitted similar gasses as in the modern era, Earth's second atmosphere probably consisted of water vapor (H_2O), carbon dioxide (CO_2), and nitrogen (N_2), because these gasses are emitted from Earth's interior by a process known as "outgassing." With colonization by plants, which absorb CO_2 and emit O_2 during photosynthesis, the atmosphere eventually contained a large concentration of O_2, which now constitutes one-fifth of its volume.

In fact, the envelope of air is a mixture of many discrete gases. Each gas has a distinct physical and chemical property. The atmosphere com prises two types of gases: those whose concentration remains essentially constant or permanent (by percent), and those that are variable and have changing concentrations over a finite period of time. Among the permanent gases, nitrogen (78.1%) and oxygen (20.9%) constitute about 99% of the atmosphere. Other permanent gases are argon (Ar, 0.9%), neon (Ne, 0.002%), helium (He, 0.0005%), krypton (Kr, 0.0001%), and hydrogen (H_2, 0.00005%). The variable gases are water vapor (H_2O, 0 to 4%), carbon dioxide (CO_2, 0.037%), methane (CH_4, 0.0002%), ozone (O_3, 0.000004%), and nitrous oxide (N_2O, 0.00009%) (www.met.fsu.edu/explores/atmcomp.html). A brief description on some of these gases is given in the following sections.

18.1.1 Nitrogen

Nitrogen gas (N_2) is composed of molecules of two nitrogen atoms, and occupies 78.1% of Earth's atmosphere. It is colorless, odorless, and tasteless. The atomic weight of N_2 is 14. Nitrogen is a principal nutrient. The low content of nitrogen in most soils exists in stark contrast to its abundance in the air. This is because gaseous N_2 molecules have very strong bonds, which make the gas chemically stable, but unusable by most biological

organisms. Some species of bacteria absorb N_2 from the air and convert it to ammonium, which can be used by plants. This process is called "biological nitrogen fixation" and is the principal natural means by which atmospheric nitrogen is added to the soil by nitrogen-fixing bacteria living in nodules on the plant roots. An example of a leguminous nitrogen-fixing crop is soybean *(Glycine max)*.

18.1.2 Oxygen

Oxygen gas (O_2) is composed of molecules of two oxygen atoms, and occupies 20.9% of Earth's atmosphere by volume. It is colorless, odorless, and tasteless, and constitutes 86% of the oceans and 60% of the human body. It is the third most abundant element found in the Sun. The atomic weight of oxygen is 16. Almost all plants and animals require oxygen for respiration to maintain life. Oxygen is flammable, reactive, and oxidizes most elements. A chemical reaction in which an oxide is formed is known as "oxidation." The rate at which oxidation occurs varies with the element with which oxygen is reacting, (e.g., burning involves a rapid oxidation, whereas rust, or iron oxide, forms slowly). Carbon in fossil fuels, for example, can be quickly oxidized to carbon monoxide (CO) and carbon dioxide (CO_2), with a considerable amount of heat being given off. Within the stratosphere (the second major layer of the atmosphere, which occupies the region of the atmosphere from about 12 to 50 km above Earth), O_2 molecules combine with free oxygen atoms to form ozone (O_3). It absorbs ultraviolet (UV) radiation from the Sun.

18.1.3 Trace Gases

Oxygen and nitrogen together constitute about 99% of the atmosphere, and the remaining 1 % is made up of trace gases whose concentrations are very small. The most abundant of the trace gases is the noble gas argon (atomic weight=39.9). Noble gases, which also include neon (20.2), helium (4), krypton (83.8), and xenon (131.3), are very inert and do not generally involve any chemical transformation within the atmosphere. Hydrogen (1.008) is also present in trace quantities in the atmosphere. Although low in concentrations, the important trace gases in Earth's atmosphere are the so-called "greenhouse gases." These greenhouse gases include carbon dioxide (44), methane (16), nitrous oxide (44), water vapor (18), ozone (48), and sulfur hexafluoride (SF_6, 146.1). These gases allow sunlight, which is radiated in the visible and ultraviolet spectra, to enter the atmosphere unimpeded, but prevent most of the outgoing infrared radiation from the surface and lower atmosphere from escaping into outer space. The greenhouse gases absorb reflected infrared radiations (heat), thus trapping the heat in the atmosphere. Thus, these gases keep Earth warm through the so-called natural "greenhouse effect," which has raised Earth's temperature from $-18°C$ to $15°C$, an increase of $33°C$. (Refer to the footnote on p. 532.)

   Variable greenhouse gases, can be divided into two categories: (i) those that occur naturally in the atmosphere (e.g., water vapor, CO_2, CH_4, and N_2O) and (ii) those that result from human activities (e.g., chlorofluorocarbons (CFCs), hydrofluorocarbons (HFCs), perfluorocarbons (PFCs), and sulfur hexafluoride). Human activities can also enhance the concentration of naturally occurring greenhouse gases. Each greenhouse gas differs in its ability to absorb heat in the atmosphere, and HFCs and PFCs are the most

heat-absorbent. The atmospheric lifetime of CH_4, a greenhouse gas 21 times more effective than CO_2 in trapping its long-wave radiation, is approximately ten years. Methane (CH_4) can trap 21 times more long wave radiation per molecule than CO_2, and N_2O can absorb 310 times more long wave radiation per molecule than CO_2 (IPCC, 2001). Methane, in contrast to CO_2 and other greenhouse gases, has the unique property of being partly converted to H_2O by cosmic radiation in the mesosphere.

The global mean surface air temperature has increased between approximately 0.3 and 0.6°C during twentieth century (IPCC, 2001). Globally, sea level has risen 10–20 cm over the past century. Worldwide precipitation over land has increased by about one percent. The frequency of extreme rainfall events has increased throughout much of the

TABLE 18.1 Concentration of Some of the Atmospheric Gases in 1 cm^3 Volume

Gas	Formula	Volume (gmol−1)	Concentration (% vol.)	Molar mass (gmol^{-1})	Concentration (g cm^{-3})
Nitrogen	N_2	22.4	78	28	9.75×10^{-4a}
Oxygen	O_2	22.4	21	32	3.0×10^{-4}
Carbon dioxide	CO_2	22.4	0.033	44	6.0×10^{-6}
Methane	CH_4	22.4	0.0002	18	1.6×10^{-9}

[a]$\frac{289}{mol} \cdot \frac{78}{100} \cdot \frac{1L}{10^3 cm^3} \cdot \frac{mol}{24.4L} = 9.75 \times 10^{-4}$ g cm^{-3}.

United States (IPCC, 2001). Some of the sinks, which absorb CO_2, are oceans, soils, and trees. Each year those sinks absorb hundreds of billions of tons of carbon in the form of CO_2. Concentration of trace/greenhouse gases in the atmosphere is also highly variable over time and space. Gaseous concentration is expressed on the basis of density or gL^{-1}, and can be calculated using Avogadro's law (see the footnote to Table 18.1).

Avogadro's law (1811) states, "Identical volumes of any gas at a standard identical temperature and pressure contain the equal number of molecules regardless of their chemical nature and physical properties." This number, known as "Avogadro's number" (N′), is 6.023×10^{23}. It is the number of molecules of any gas present in a volume of 22.41 L and is the same for a very light gas (e.g., H_2) as for a heavy gas (e.g., CO_2 or Bromine, Br). Avogadro's number is now considered to be the number of atoms present in 12 grams of the carbon-12 isotope (one mole of carbon is 12 g).

The concentration of atmospheric gases in a 1 cm^3 volume, can be calculated from the fact that a gram molecular weight of a gas occupies 22.4L of volume at standard temperature and pressure (STP). Thus, the concentration of O_2 in the atmosphere is $3 \times 10^{-4} gcm^{-3}$. Similarly, the atmospheric concentration of other gases can be computed (Tables 18.1).

18.2 SOIL AIR

Soil air refers to air in the soil. It is located in the air porosity, whose volume is inversely proportional to that of the soil water $(f_a \propto \theta^{-1})$. Thus, as the volume of soil water (θ) increases, that of soil air (f_a) decreases, and vice versa. A compacted soil or an undrained soil has smaller amounts of soil air than a well-structured and drained soil. In a well-structured soil the soil air content is higher with soil air occupying most of the large or macropores. In general, soil air content (f_a) and water content (θ) are nearly equal at field moisture capacity for well-structured soils. The increase in bulk density (ρ_b) decreases the total porosity (f_t) and for given water content (θ) decreases the soil air content (f_a). Soil air content is also affected by drainage conditions in the field as poor or improper drainage increases the water content of soil thus lowering the air content. Composition of soil air is highly variable and depends on numerous factors (e.g., soil structure, bulk density, drainage conditions). In a well-aerated soil, the oxygen content of soil air is similar to that of the atmosphere because the consumed O_2 is readily replaced and CO_2 generated is readily removed from the soil-air system. In soils with restricted exchange, soil air differs from atmospheric air in several respects. The CO_2 concentration in soil air is much higher and O_2 concentration much lower than atmospheric air. Soil air is also relatively moister than atmospheric air, and it contains numerous trace gases (e.g., H_2S). The composition of soil air varies greatly from place to place in the soil, as plants consume some gases and microbial processes release others (Tables 18.2 and 18.3). The amount and composition of soil air is determined by the water content of soil unless the soil is very dry. The O_2 content in a well-aerated soil is higher than that of a poorly aerated soil. The latter has higher concentrations of CO_2, CH_4, and N_2O than atmospheric air. As the depth of soil profile increases, the concentration of CO_2 increases with a corres-ponding decrease in O_2 concentration; however, the sum of these two

TABLE 18.2 Measured O_2 and CO_2 Content in Soil Air (% by Volume) at Two Depths

Soil management	O_2 (%)		CO_2 (%)	
	15 cm	46 cm	15 cm	46 cm
Arable land manured	20.52	20.33	0.34	0.50
Arable land unmanured	20.32	20.35	0.34	0.45
Grassland	18.44	17.87	1.46	1.64

Source: Modified from Russel and Appleyard, 1915.

TABLE 18.3 Measured O_2 and CO_2 Content (% by Volume) in Soil Air Collected During Summer and Winter

Cropping systems		O_2 (%)	N_2 (%)	CO_2 (%)
Arable land manured and cropped	Summer	20.74	79.03	0.23
	Winter	20.31	79.32	0.37
Arable land unmanured and cropped	Summer	20.82	78.99	0.19
	Winter	20.42	79.37	0.21

Source: Modified from Russel and Appleyard, 1915.
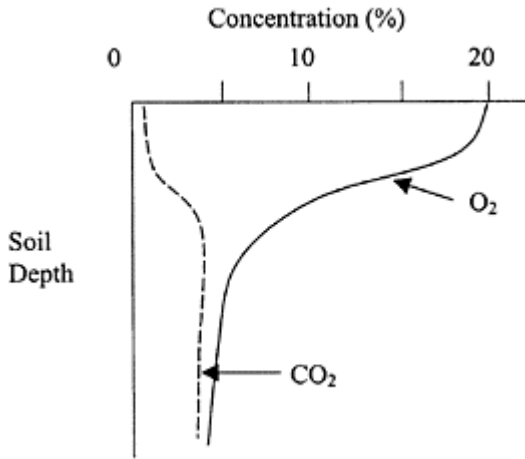
FIGURE 18.1 Schematic of variation of concentrations of O_2 and CO_2 in soil air with depth.

TABLE 18.4 CO_2, O_2 and N_2 Contents in Soil Air for Well-Drained Treatments with Constant Water Table Depths

Water table position	Date	CO_2 (%)	O_2 (%)	N_2 (%)
No water table	2 July	1.2	17	78
	30 July	2.0	18.5	76
	17 August	0.3	17.8	76.5
1 5 cm depth	2 July	6.8	15	75.8
	30 July	8.5	11	81
	16 August	8	7	80.2
30 cm depth	2 July	2.2	16.5	77
	30 July	6.2	11.5	73.5
	16 August	3	17	77

Source: Modified from Lal and Taylor, 1969.

concentrations never exceeds 21% (Fig. 18.1). A soil is considered healthy if the air filled pore spaces are about 50% of the total porosity, and composition of soil air is similar to that of atmospheric air. The reduced soil aeration results from excess water in the soil profile, which may be due to the poor drainage, a shallow groundwater table, soil compaction, swelling clays, or decomposition of organic matter by microorganisms with low O_2 replenishment. As the water table falls below the root zone, the CO_2 concentration in soil air decreases with a corresponding increase in O_2 (Table 18.4). Air permeability of soil, tillage practices (Table 18.5), soil

TABLE 18.5 Soil CO_2 Concentration Data for No-Till (NT) and Moldboard Plow (MB) Plots for Early (21 July), Mid (24 August), and Late (1 October) Season, 1998

| | Average CO2 concentration (ppm) | | | | | |
| | no-till | | | moldboard plow | | |
Depth (cm)	early	mid	late	early	mid	late
5	2000	3000	1000			
10	8000	6000	2000			
20	28000	23000	4000			
30	34000	24000	5000	20000	9000	3000
50	36000	28000	9000	25000	18000	8000
70	35000	27000	13000	27000	16000	10000

Source: Modified from Reicosky et al., 2002.

TABLE 18.6 O_2 Consumption and CO_2 Release for a Cropped and Bare Soil in January (Soil Temperature 3°C) and July (Soil Temperature 17°C)

| | Cropped ($gm^{-2}d^{-1}$) | | Bare ($gm^{-2}d^{-1}$) | |
	January	July	January	July
0_2	2	24	0.7	12
CO_2	3	35	1.2	16

Source: Modified from Curry, 1970.

temperature, and microbial activities (Table 18.6) also affect concentration of CO_2 in soil air. Soil management practices, which improve soil structure, also improve soil aeration. These include no-till, residue mulch, application of manures, conversion of cropland to pasture, etc.

18.3 SOIL AERATION

Soil aeration, the process of the exchange of air (O_2 and CO_2) between soil (or plant roots and soil microorganisms) and the atmosphere is important to plant growth because it maintains O_2 concentration in the root zone at the level needed for root and microbial

respiration. Soil aeration is a vital process for controlling the twin processes of respiration and photosynthesis. Plant roots absorb O_2 and release CO_2 during respiration. The O_2 in soil air also governs the chemical reactions, which provide the necessary conditions for oxidation of reduced elements (Fe^{+2}, Mn^{+2}), which may otherwise be toxic to plant growth. Respiration involves the oxidation of organic compounds (such as glucose), and can be represented as follows:

$$C_6H_{12}O_6 + 6O_2 \xrightleftharpoons[\text{Photosynthesis}]{\text{Respiration}} 6CO_2 + 6H_2O + \text{energy} \tag{18.1}$$
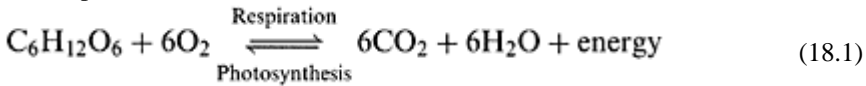
In photosynthesis, the above reaction is reversed (right to left). The total energy is 2883 kJ and biologically useful energy is 1270 kJ. The respiration process increases the concentration of CO_2 in the soil pores and at the same time reduces the O_2 concentration, which creates a concentration gradient, and O_2 flows in the soil profile through the process of diffusion and pushes the CO_2 out of the soil. The rate of O_2 diffusion into the soil profile is proportional to the aeration porosity. The aeration porosity has been defined as the pore space filled with air when the soil sample is placed on a porous plate and equilibrated at 50 cm of suction (Φm). The air circulation in and out of soil matrix also moderates the temperature of the soil. In addition to plant growth, soil air composition alters production and emission of trace gases (e.g., CH_4 and N_2O).

18.4 OXYGEN DEFICIENCY AND PLANT GROWTH

The influence of soil air on plant growth is a complex process and can be grouped into direct and indirect effects. The direct influences are related to the physiological effects of O_2 and CO_2 while the indirect influences affect the biological and chemical transformations in the soil. A decrease in soil O_2 concentration results in a decrease in aerobic microbial population and at the same time an increase in anaerobic microbial population, which is responsible for the changes in soil respiration, enzyme activity, and oxidation-reduction or redox potential. Among physiological influences, most of the effects are solely caused by the lack of O_2 for metabolic activities. The O_2 deficiency restricts the root respiration, growth of plant, water, and nutrient uptake, and changes root metabolism toward fermentation. The reliable index of O_2 availability to plant roots is termed the oxygen diffusion rate (ODR; Glinski and Stepniewski, 1985). The diffusion coefficient of O_2 increases with temperature as a result of decrease in O_2 solubility (Letey et al., 1961). After a certain value of ODR, the seedling emergence remains almost a constant, below this value the seedling emergence declines very rapidly with decrease in ODR. The limiting and critical values of ODR for some crops are presented in Table 18.7. At a critical value of ODR (20×10^{-8} g O2 cm^{-2} min^{-1}) (Stolzy and Latey, 1964), the emergence falls to zero, i.e., no germination of seedling takes place. The

TABLE 18.7 Limiting and Critical Values of ODR
for Some Crops

Crop	ODR ($\mu g \, m^{-2} \, s^{-1}$)	
	Limiting	Critical
Barley	25	8
Oats	30	12
Beans	33	12
Wheat	40	8
Flax	40	13
Maize	40	16
Tomato	40	25
Sugar beet	50	13
Rye	50	12

Source: Modified from Glinski and Stepniewski, 1985.

deficiency of O_2 results in restricted root respiration, which has adverse influences on plant growth, and nutrient and water uptake. The deficiency of O_2 for root metabolism also leads to increase in ethanol (C_2H_5OH) concentration, which decreases the emergence of seedlings. The adjustment of stomata aperture regulates the transpiration, heat balance, photosynthe-sis, and respiration in plants (Glinski and Stepniewski, 1985). The factors affecting stomata aperture are the partial pressures of CO_2, light, water stress, and temperature. The O_2 deficiency to roots results in stomata closure (Sojka and Stolzy, 1980). The wilting thus caused, despite inundation, is called "scalding."

18.5 OXYGEN DEFICIENCY AND SOIL PROPERTIES

Increase in the degree of saturation reduces O_2 content in the soil air. This scenario is very common in undrained or poorly drained soils, where waterlogging or inundation results in O_2 deficiency in soil. The high water content alters soil structural and water transmission properties such as airfilled porosity at a given suction, air permeability, saturated hydraulic conductivity, infiltration characteristic, and compressive strength (Hundal et al., 1976). Soil bulk density may be higher in undrained than drained soil (Table 18.8). The saturated hydraulic conductivity, air-filled porosity at 1 bar (100 kPa), and soil strength may increase with drainage or lowering of the water table (Table 18.8). Soil organic carbon concentration also decreases with drainage or lowering of the water table (Table 18.9). Increase in soil water content also decreases soil temperature (see Table 17.11 in Chapter 17)

TABLE 18.8 Effect of Drainage on Soil Physical Property

	Undrained		Drained	
Property	0–15 cm	15–30 cm	0–15 cm	15–30 cm
ρ_b(g cm$^{-3)}$	1.29	1.36	1.22	1.32
w (%)	30.4	29.2	30.1	29.6
K$_s$ (cm h^{-1})	0.1	0.08	2	0.8
P$_a$	9	7	15	10
UCS (kg cm^{-2})	2.5	3.0	1.8	2.2

[a]Where ρ_b is bulk density; w is gravimetric moisture content at 1 bar (%); K$_s$ is saturated hydraulic conductivity; P$_a$ is air filled porosity at 0.5 bar; UCS is unconfined compressive strength.
Source: Modified from Hundal et al., 1976.

TABLE 18.9 Effect of Water Table Depth on SOC (Mgm^{-3})

		Depth (cm)	
Treatment	Sample number	8 to 16	16 to 24
Drained	7	2.37	2.3
	4	2.53	2.37
Undrained	7	2.58	2.34
	4	2.62	2.59

Source: Modified from Sullivan et al., 1997.

, which depending upon the prevalent climate of area, can increase the intensity of hot/cold, and freeze/thaw cycles, thereby causing a change in soil aggregation and overall structural properties.
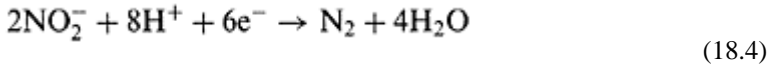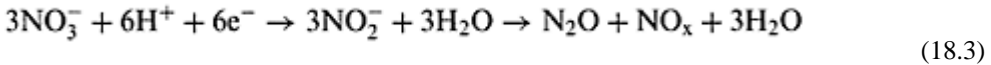
18.6 SOIL RESPIRATION

Soil respiration is the amount of oxygen consumption or CO2 evolution in the soil. The rate of soil respiration varies with space and time and depends upon soil water content, soil type, plant cover, and agriculture measures and amendments. Soil respiration can be measured both under field and laboratory conditions using various types of respirators or respimeters. The respiratory coefficient, which provides useful information on soil aeration, is the ratio of the volume of CO_2 produced to the volume of O_2 consumed. For a well-aerated soil, the respiratory quotient is equal to one. The aerobic or anaerobic conditions of soil can be checked as follows (Monteith et al., 1964):
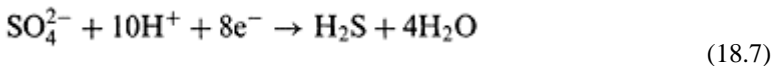
$$R = R_0 Q^{T/10}$$

(18.2)

where R is the flux at $T°C$ and R_0 at $0°C$ and Q is equal to 3 (Monteith et al., 1964). The concentration of O_2 consumed and CO_2 released in a cropped and bare soil is also presented in Table 18.6 as an example.

18.7 OXIDATION REDUCTION PROCESS IN SOIL

The chemical and biochemical reactions, which occur in soil under anaerobic condition, are dentrification and reduction of manganese (Mn), iron (Fe), and sulfate (SO_4). Nitrate (NO_3) is reduced to nitrite (NO_2), then to nitrous oxide (N_2O), and eventually to elemental nitrogen (N_2). The process of decrease in nitrate content with time in a flooded or saturated soil is known as denitrification. The rate of denitrification depends on soil saturation, pH, and temperature. Denitrification is an anaerobic process and an indicator of the absence of O_2 in at least a part of soil volume. The end products in a denitrification process are gaseous (N_2O, NO, and N_2) (Ponnamperuma, 1972).

$$3NO_3^- + 6H^+ + 6e^- \rightarrow 3NO_2^- + 3H_2O \rightarrow N_2O + NO_x + 3H_2O$$

(18.3)

$$2NO_2^- + 8H^+ + 6e^- \rightarrow N_2 + 4H_2O$$

(18.4)

Manganese reduces from a manganic (Mn^{+4}) to magnous (Mn^{+2}) state, iron from a ferric (Fe^{+3}) to ferrous (Fe^{+2}) state, and sulfate (SO_4) to hydrogen sulfide (H_2S).

$$Mn^{4+} + 2e^- \rightarrow Mn^{2+}$$

(18.5)

$$Fe^{3+} + e^- \rightarrow Fe^{2+}$$

(18.6)

$$SO_4^{2-} + 10H^+ + 8e^- \rightarrow H_2S + 4H_2O$$

(18.7)

Some of the toxic substances produced during anaerobic conditions are (H_2S), ethylene (C_2H_4), and acetic ($C_2H_4O_2$), butyric ($C_4H_8O_2$) and phenolic (C_6H_5OH) acids.

The process of production of CH_4 is known as methanogenesis. Methanogenic bacteria generate CH_4 biologically, largely from acetate (CH_3COOH) dissimilation and CO_2 reduction. The methanogens are capable of obtaining energy for growth by converting CO_2 and molecular hydrogen into CH_4 and H_2O.

$$CO_2 + 4H_2 \rightarrow CH_4 + 2H_2O$$

(18.8)

Some methanogenic bacteria are also capable of transforming acetate into CH_4 and CO_2.

$$CH_3COOH \rightarrow CO_2 + CH_4 + 35.6 \text{ kJmol}^{-1}$$

(18.9)
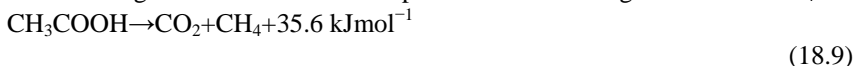
The H_2 is a product of anaerobic degradation of organic matter. The H_2 with acetate is one of the most important intermediates in the methanogenic degradation of organic matter and serves as a substrate for methanoge-nic process (Conrad, 1999). Methanogenesis is a major pathway for organic matter decay in sediments. The factors controlling methanogenesis are temperature, concentration of other electron acceptors, water table position, substrate (e.g., H_2) availability, and oxygen supply (Boon and Mitchell, 1995; Grunfeld and Brix, 1999). As temperature increases, water table in the root zone rises and other electron acceptors (e.g., NO_3, Fe_3, SO_4) reduce, methanogenesis increases (Kluber and Conrad, 1998). Methanogenesis occurs in flooded soils, as well as in soils at low water content incubated under anaerobic condition (Boon and Mitchell, 1995). Rice fields are estimated to contribute $100\pm50Tg$ yr^{-1} of the greenhouse gas CH_4 (Kluber and Conrad, 1998). Production of CH_4 occurs during fermentation process by anaerobic bacteria. In flooded soils, CH_4 appears from several days to weeks after flooding. The organic matter amendment stimulates CH_4 formation in alkaline soils, whereas it is suppressed in acid soils (Glinski and Stepniewski, 1985).

$$C_6H_{12}O_6 \rightarrow 2CH_3COOH+CO_2+CH_4+346.8 \text{ kJmol}^{-1}$$

(18.10)
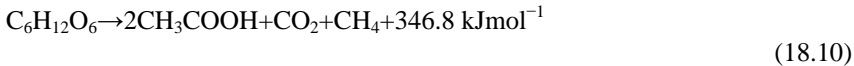
The electron transfer is the primary source of energy needed by microorganisms for various processes. Glucose releases electron upon oxidation as follows

$$C_6H_{12}O_6 = 2CH_3COCOOH+4H^++4e^-$$

(18.11)

In anaerobic conditions O_2, NO_3, H^+, and high valency iron and manganese accept electrons and are reduced to H_2O, N_2, H_2, lower valency Fe^{+2}, and Mn^{+2}, respectively.

$$0_2+4H^++4e^- \rightarrow 2H_2O$$

(18.12)

$$2H^++2e^- \rightarrow H_2$$

(18.13)

This tendency of a substance to accept or donate electrons is measured in terms of the oxidation-reduction potential, commonly known as the oxidation-reduction potential or "redox potential." It is defined as "the potential in volts required in an electric cell to produce oxidation at the anode and reduction at the cathode." The redox potential is a relative term and is measured relative to a standard hydrogen electrode also known as reference electrode whose potential is assumed to be zero. The potential has an inverse relationship with the rate of reduction of substances. The redox potential of soil is closely linked to the availability of O_2, especially at low O_2 levels, and can identify the changes in availability of O_2. The redox potential can be represented as follows:

$$E_h = E_0 - \frac{RT}{nF} \ln\left(\frac{Ox}{Red}\right)$$

(18.14)

where E_h is the potential difference between the reference electrode and inert platinum (Pt) electrode, E_0 is the potential of reference electrode, R is the gas constant, T is

absolute temperature, n is the number of electrons transferred in the reaction, F is Faraday's constant, "Ox" is activity of oxidized specie, and "Red" is the activity of reduced specie. From Eq. (18.14), it is clear that the E_h is proportional to the natural log of reduced and oxidized products. In well-drained soils, a sufficient amount of O_2 is available, therefore, they can be called oxidized systems. The typical E_h values for oxidized systems are > 400 mV. The O_2 disappears at about 300 mV, NO_3^- is removed between 200 and 300 mV, and $Fe_3^+, Mn_4^+,$ and SO_4^- are reduced sequentially with decrease in Eh value (Poonamperuma, 1972; Scott, 2000).

18.8 FLOW OF AIR IN SOIL

The gaseous exchange between soil and atmosphere occurs by two processes: convection and diffusion. The convective flow of air in soil occurs as a result of the total pressure difference between the soil air and

TABLE 18.10 Increase in CO_2 Content for Calcareous Silty Clay Loam (SCL) and Sandy Loam (SL) Near Field Capacity Under Tensions for Short Period of Time

Soil	Duration minute	Tension (kPa)	CO_2(%)
SCL	35	35	6.6
SCL	40	29	8.7
SCL	67	41	4.8
SL	18	39	17.4
SL	20	28	4.6

Source: Modified from Boynton and Reuther, 1938.

outer atmosphere. The pressure difference is caused as a result of O_2 consumption by plant roots, CO_2 production in the soil (Table 18.10), change in the barometric pressure in the atmosphere, soil temperature, moisture content, or water table depth of soil due to evaporation, drainage, or water supply by rainfall or irrigation, etc. Various studies have pointed out that convection of air in soil is predominant for shallow depths and in soils with large pores (Rolston, 1986). The convective flow of air in the soil is similar to water flow and is proportional to the pressure gradient across the flow domain. However, since air is compressible, the density and viscosity are also the functions of pressure and temperature. Unlike water flow, gravity is not important for airflow. Air is not attracted to mineral particles and occupies the larger pores. Using Darcy's law for water flow [refer to Eq. (12.3)] the convective flux (q_a) for laminar airflow is given as follows

$$q_a \propto -\nabla P$$

(18.15)

where is the three-dimensional gradient of soil air pressure. If the permeability of air-filled pore space is k_a, and viscosity of soil air is ηa, then one-dimensional convective flow can be given as follows (Hillel, 1998)

$$q_a = -\frac{k_a}{\eta_a}\left(\frac{dP}{dx}\right)$$

(18.16)

If the density of soil air is ρ_a, then air flux (q_a) expressed as mass flow per unit area per unit time is

$$q_a = -\frac{k_a\rho_a}{\eta_a}\left(\frac{dP}{dx}\right)$$

If soil air is assumed to be an ideal gas at pressure, P, occupying a volume, V, then the ideal gas equation for soil air can be written as

$PV = nRT$

(18.18)

where n is number of moles of gas, R is the universal gas constant per mole, and T is absolute temperature. Substituting the density, $\rho_a = M/V$, and $M = nm$, in Eq. (18.18) where m is the molecular weight, and after rearranging, the following relationship for density is obtained.

$$\rho_a = \frac{m}{RT}P$$

(18.19)

For a one-dimensional compressible fluid, the rate of change of pressure with respect to time is equal to the rate of change of mass flux with respect to length of fluid mass and can be expressed as

$$\frac{\partial \rho}{\partial t} = -\frac{\partial q_a}{\partial x}$$

Substituting Eqs. (18.17) and (18.19) into Eq. (18.20) results in

$$\frac{m}{RT}\frac{\partial P}{\partial t} = \frac{\partial}{\partial x}\left(\frac{\rho_a k_a}{\eta_a}\frac{\partial P}{\partial x}\right)$$

(18.21)

For small pressure differences, $\rho_a k_a/\eta_a$ can be assumed a constant (Hillel, 1998).

$$\frac{\partial P}{\partial t} = \alpha\frac{\partial^2 P}{\partial x^2}$$

(18.22)

where $a = (RT\rho_a k_a)/m\eta_a$. The above equation is an approximate equation for the transient-state convective flow of air in soil. Convective flow rarely meets more than 10% of the

O_2 demand of plant roots (see Example). Thus, diffusion is the more important mechanism of soil aeration (Russell, 1952).

18.9 FICK'S LAW AND GASEOUS DIFFUSION IN SOIL

The gaseous transport of O_2 and CO_2 in the soil occurs both in the gaseous and liquid phases. The process of diffusion (random thermal molecular movement from high to low concentration; also refer to Chapter 16) maintains the air exchange between soil and surrounding atmosphere, whereas the supply of O_2 and removal of CO_2 from the plant roots or live tissues takes place by diffusion through water films. According to Fick's law, the mass rate of transfer of a diffusing gas through a unit area of bulk soil is proportional to the concentration gradient measured normal to the surface through which diffusion is taking place. If D is the diffusion rate ($cm^2\ s^{-1}$), C is the concentration of diffusing substances ($g\ cm^{-3}$), q_x is the rate of transfer of mass per unit area ($g cm^{-2}s^{-1}$), and x is the distance of diffusion (cm), the diffusion of gases in both phases can be represented by the following one-dimensional equation

$$q_x = -D\frac{\partial C}{\partial x}$$ (18.23)

The three-dimensional diffusion of gases according to Fick's law is represented as follows:

$$q_x + q_y + q_z = -\left(D_x\frac{\partial C}{\partial x} + D_y\frac{\partial C}{\partial y} + D_z\frac{\partial C}{\partial z}\right)$$ (18.24)

where q_x, q_y and q_z are the rate of transfer of mass per unit area, and D_x, D_y, and D_z are gaseous diffusivity, in x, y, and z directions. The partial differential equation of diffusion can be derived, similar to Laplace's equation, by equating the difference between the inflow and outflow of a diffusing substance in a volume element to the change in concentration with time.

$$\frac{\partial C}{\partial t}\Delta x\Delta y\Delta z = -\left(\frac{\partial q_x}{\partial x} + \frac{\partial q_y}{\partial y} + \frac{\partial q_z}{\partial z}\right)\Delta x\Delta y\Delta z$$ (18.25)

or

$$\frac{\partial C}{\partial t} = -\left(\frac{\partial q_x}{\partial x} + \frac{\partial q_y}{\partial y} + \frac{\partial q_z}{\partial z}\right)$$ (18.26)

From Eqs. (18.24) and (18.26), assuming the diffusion coefficient is independent of direction, the differential equation for three-dimensional gas flow is obtained as follows:

$$\frac{\partial C}{\partial t} = D\left(\frac{\partial C^2}{\partial x^2} + \frac{\partial C^2}{\partial y^2} + \frac{\partial C^2}{\partial z^2}\right) = \nabla D$$ (18.27)

TABLE 18.11 Diffusion Coefficient of Some
Gases Under Standard Pressure and Temperature

Gas	Density (kgm^{-3})	Diffusion coefficient ($m^2 s^{-1}$)	
		In air	In water
O_2	1.429	1.78×10^{-5}	2.6×10^{-9}
CO_2	1.977	1.39×10^{-5}	1.91×10^{-9}
N_2	1.251	1.8×10^{-5}	1.9×10^{-9}
H_2	0.08	6.34×10^{-5}	5.85×10^{-9}
Water vapor	0.768	2.39×10^{-5}	
NH_3	0.771	1.98×10^{-5}	2.0×10^{-9}
N_2O	1.978	1.43×10^{-5}	
C_2H_4	1.261	1.37×10^{-5}	

Source: Data from Weast et al., 1989.

The one-dimensional form of gaseous diffusion in a porous medium is given by

$$\frac{\partial C}{\partial t} = D \frac{\partial C^2}{\partial x^2} \qquad (18.28)$$

which is similar to Eq. (16.17) when mass flow (second term on the right hand side) is zero. D varies inversely with the molecular weight of gas and is a direct function of temperature and pressure of the gaseous medium. Under standard pressure and temperature, the D in soil air is 10,000 times greater than in soil water (Table 18.11). Under normal atmospheric pressure and 25°C, the D ranges from 0.05 and 0.28 cm^2 s^{-1}; the value depends on the volume of phase available for diffusion. The D is not affected by the shape of solid surfaces or by the particle size or pore size distribution of soil solids because mean free path of diffusing molecule is generally much smaller than the width of the pores.

   Considering the diffusive path in the air phase of soil, the diffusion coefficient in soil D_s is much smaller than in air D_a. The ratio Ds/Da is known as relative diffusion coefficient. The D_s and D_a are related by some function of air-filled porosity (f_a), which are presented in Table 18.12. The tortuosity coefficient of 0.66 (Table 18.12) (Penman, 1940) suggests that straight-line paths are only 66% of total average path of diffusion in soil. Van Bavel (1952) suggested the value of coefficient to be 0.61 rather than 0.66. The advantage of using the dimensionless coefficient or ratio is that the effects of state variables such as pressure, temperature, and type of gas are cancelled.

TABLE 18.12 Models of D_s/D_a as a Function of
Volumetric Air Content

Relationship	Reference
$\dfrac{D_s}{D_a} = \kappa f_a^2$	(κ is a constant), Buckingham (1904)
$\dfrac{D_s}{D_a} = \dfrac{f_a}{[k - f_a(k - 1)]}$	Burger (1919)
$\dfrac{D_s}{D_a} = 0.66 f_a$	Penman (1940)
$\dfrac{D_s}{D_a} = 0.61 f_a$	Van Bavel (1952)
$\dfrac{D_s}{D_a} = f_a^{1.5}$	Marshall (1959)
$\dfrac{D_s}{D_a} = \alpha f_a^\beta$	Currie (1960)
$\dfrac{D_s}{D_a} = \dfrac{f_a^{10/3}}{\phi^2}$	Millington (1959), Millington and Quirk (1961)
$\dfrac{D_s}{D_a} = -0.12 + 0.9 f_a$	Wesseling and Van Wijk (1957)
$\dfrac{D_s}{D_a} = -0.1 + 0.9 f_a$	Wesseling (1962)

The O_2 and CO_2 can diffuse both in gaseous and aqueous systems, a diffusion constant K_a can be defined, which separates the contribution from these two phases. The diffusion constant in air (K_a) is given as follows:

$K_a = f_a D_a$

$$(18.29)$$

And diffusion constant in water K_w is

$K_w = \theta D_w$

$$(18.30)$$

The ratio of Eqs. (18.29) and (18.30) after rearrangement yields

$$K_W = \frac{\theta}{f_a} \frac{D_W}{D_a} K_a = a_b \frac{D_W}{D_a} K_a \qquad (18.31)$$

where a_b is Bunsen's solubility coefficient.

18.10 SOURCES AND SINKS OF GASES IN SOIL

The continuity equation states that the rate of change of concentration of a diffusing gas equals the rate of change of flux with distance. Mathematically, it is expressed as follows:

$$\frac{\partial C}{\partial t} = -\frac{\partial q_x}{\partial x}$$

(102)

Equation (18.32) implies that a diffusing substance follows the law of conservation of matter. However, during the transport of CO_2 and O_2 through the soil system, the plant roots or anaerobic activities along diffusional path absorb O_2 and release CO_2. Considering S_g to be a source and sink term in time and space, Eq. (18.32) is modified as follows (Hillel, 1998; Scott, 2000):

$$\frac{\partial C}{\partial t} = -\frac{\partial q_x}{\partial x} \pm S_g(x, t)$$

(18.33)

Substituting Eq. (18.23) into Eq. (18.33) and assuming D constant in diffusional path yields

$$\frac{\partial C}{\partial t} = -\frac{\partial}{\partial x}\left(-D\frac{\partial C}{\partial x}\right) \pm S_g(x, t)$$

(18.34)

$$\frac{\partial C}{\partial t} = D\frac{\partial^2 C}{\partial x^2} \pm S_g(x, t)$$

(18.35)

After a rainstorm or irrigation, the larger pores drain quickly and smaller pores or intraaggregate micropores drain slowly. The gaseous diffusion also takes place rather rapidly from interaggregate macropores. The plant roots are also confined to larger pores between aggregates but do not penetrate them. Therefore, larger pores remain well aerated whereas micropores remain anaerobic.

18.11 MEASUREMENT OF SOIL AERATION

Measurement of soil aeration involves assessing: (i) fractional pore space, (ii) composition of soil air, and (iii) rate of diffusion of O_2 from atmosphere into the soil. The aeration is measured by measuring the air-filled porosity at a standard value of soil suction or soil water content. This is done by collecting a core sample from a soil at field capacity (normally 24 to 48 h after a deep wetting or at soil water suction of about 50 cm of water) and measuring air-filled space with an air pycnometer. Alternately, first obtaining total porosity from bulk density (ρb) and particle density (ρs), and subtracting the water content of core can calculate the air-filled porosity (fa). Measurement of the relative concentration of O_2, CO_2, and other gases in the soil air provides important information on the aeration and soil structure. Depletion of O_2 level content in soil air is a good indicator of the restricted gas exchange in the soil matrix. This method, although static, is better than the measurement of air volume alone. However, it requires extraction

of a sample that is large enough to provide a measurement but at the same time small enough to be representative. Another drawback of this method is soil disturbance and contamination or mixing of air from the atmosphere. The repeated measurements of O_2 or CO_2 concentrations in soil air without extracting a sample can also be obtained by the electrode methods (McIntyre and Philip, 1964; Phene, 1986). The measurement of depletion of O_2 or increase in CO_2 can be made both in situ or in a laboratory by gas chromatography technique, which provides reliable measurements. The method allows rapid and precise measurement of N_2, O_2, Ar, CO_2, CH_4, Ne, H_2, CO, NO, C_2H_4, and C_2H_6 by employing a wide range of methods, detectors, and column packing (Blackmer and Bremner, 1977). The in situ method for measuring O_2 and CO_2 are based on detecting the thermal conductivity by paramagnetic oxygen analyzer and potable carbon analyzer (van Bavel, 1965), respectively.

An early approach to measure aeration involved the determination of the fractional air space or air filled porosity (f_a) at a standardized value of soil wetness. This was measured by either taking a core sample from the field two days after a deep wetting, or saturating the core sample with water and then subjecting it to a suction of 50 cm. All the pores with an effective diameter greater than 0.06 mm (r =0.147/50 cm) are drained of water. The air space as a fraction of porosity now can be determined with an air pycnometer (Page, 1948; Vomocil, 1965). Alternatively, the air space can be determined by the difference of porosity and volumetric wetness $(f_a=f_t-\theta)$. However, these two methods are not adequate as considerable uncertainties exist in the measurement and aeration dynamics remains almost untouched.
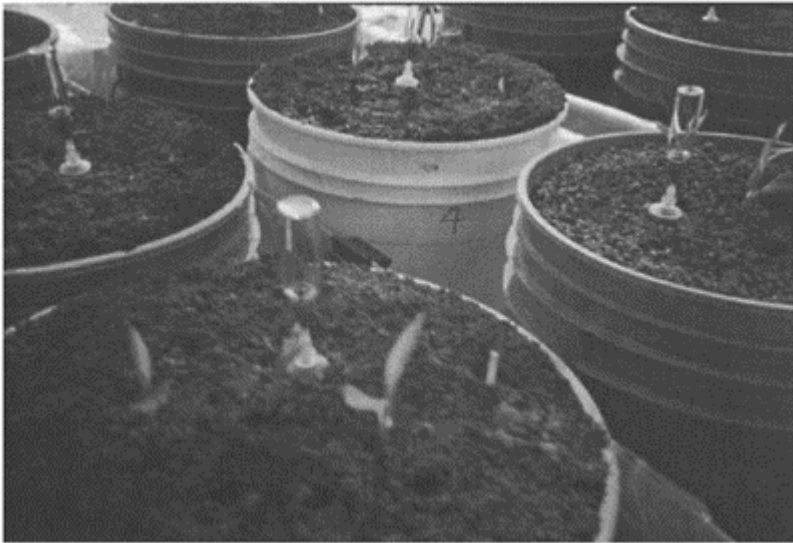

FIGURE 18.2 Soil air diffusion tube installed in a greenhouse water table management experiment. Similar

diffusion tube is used under field
conditions. Soil air sample is taken
from the tube using a syringe.

The other traditional method involves the determination of the composition of soil air
(Fig. 18.2). This method, although again static, is better than the measurement of air
volume alone. The depletion of O_2 content in soil air can be a good indicator of the
restricted gas exchange in the soil matrix and between soil and the atmosphere (Fig.
18.3). Still, the main concern here is how to extract a sample that is large enough to
provide a measurement but at the same time small enough to represent the sample point
and to avoid disturbances and mixing of soil air or contamination from the atmosphere.
The gas chromatography technique can provide reliable measurements. An alternative
method, which permits repeated measurements of oxygen concentrations in soil air
without extracting a sample, is based on the use of membrane-covered electrodes
(McIntyre and Philip, 1964).

Soil aeration can be characterized by the oxygen diffusion rate in the soil or ODR,
(Erickson and van Doren, 1960). The method is based on the hypothesis that the moisture
films around plants roots and organism limit the rate of O_2 diffusion. The ODR can be
measured by a platinum electrometer under a constant electric potential (Lemon and
Erickson, 1955). Once the O_2 present near electrode surface is depleted further depletion
is a function of O_2 diffusion to electrode surface or current. The electric current (I, A) is
proportional to the rate of O_2 flux at the electrode

FIGURE 18.3 A static chamber is
used under field conditions to assess
the gaseous emission over a short
period of 10 to 15 minutes. (Waterman
Farm, Columbus, OH, 1998.)

surface and can be expressed as follows:

$$ODR = \frac{60MI}{nFA}$$

(18.36)

where ODR is oxygen diffusion rate (g m^{-2}s^{-1}), M is the molar mass of oxygen (32g mol^{-1}), F is the Faraday's constant (96,500 coulombs equiv.$^{-1}$ mol^{-1}), A is the electrode surface area (m^2), and n is equal to four (equiv. mol^{-1}) and is the number of electrons required to reduce one molecule of O_2. The ODR values in soils vary from 0 to 200 µg m^{-2} s^{-1} and increase with suction and air-filled porosity of soil (Glinski and Stepniewski, 1985). The ODR method is satisfactory in soils having higher aeration and is less effective for poorly drained or flooded soils. The methods of soil aeration measurement are listed in Table 18.13. Another approach of characterizing soil aeration is to measure the air permeability.

18.12 AIR PERMEABILITY

The gaseous exchange between soil and the atmosphere and the transport of gases within the soil are complex phenomena. Characterizing soil aeration by measuring content and composition of soil air are inadequate because they do not take into account the process dynamics, directions, and rate of change. Air permeability of soils has been recognized as an important

TABLE 18.13 Methods of Measurement of Soil Aeration

Method	Reference
Air pycnometer	Page (1948), Vomocil (1965)
Membrane covered electrodes	McIntyre and Philip (1964)
Gas chromatography	Bremner and Blackmer (1982)
Closed chamber	Matthias et al. (1980)
Flow through chamber	Denmead (1979)

parameter for soil aeration and contaminant remediation techniques and is fundamental to our understanding of environmental problems in the vadose zone. The vadose zone comprises the region between the land surface and underlying groundwater aquifers varying in depth and composition. It is the geologic zone through which water, solutes, nutrients, and/or contaminants travel prior to reaching groundwater. In agricultural research, knowledge of air-filled pores, pore size distribution, tortuosity, air permeability, and their variation along the cross section or depth is important to describe aeration, structure, and compaction of the soil. Precise impact of these parameters on crop yield is not known. In general, poor structure, low air-filled porosity, and water permeability adversely affect crop yield (Moore and Attenborough, 1992).

Air permeability of porous media, including soils, is governed by the convective transport of air through the media under a pressure gradient. The gaseous flow as a consequence of the pressure head difference is often reported as the mass flow of gas. The other mechanism of gas transport is the diffusion, which occurs due to the change in concentration gradients or the partial pressures of the components of the gaseous mix. If the concentration and pressure gradients exist concurrently, both these processes can occur simultaneously. The mass flow of gas is important when differences in pressure are due to the change in barometric pressure, temperature, or soil water content. However, diffusion is considered the primary mechanism.

In general, a soil matrix consists of a mixture of fluid and gaseous phases. Since viscosity of air is small compared to that of water, soil air remains at most phases in the soil matrix at or near atmospheric pressure. A small pressure gradient is sufficient for soil air to move into or out of the soil system. As a result it has a negligible effect on flow of water and therefore most water transport analysis ignores the simultaneous movement of soil air. The negligible influence due to the low-pressure gradients in soil air is generally, but not necessarily always, true. In case of border irrigation, effects of air compression ahead of the wetting front during infiltration of water into the soil can occur (Dixon and Linden, 1972; Morel-Seytoux and Khanji, 1974). During drainage, air entry through the restrictions within the soil pore space causes surge of water in the draining soil columns (Corey and Brooks, 1975). Airflow through soils is essentially nondestructive and air permeability is sensitive to the changes in soil structure (Corey, 1986). Air permeability can be used as a soil quality indicator to characterize the changes in soil structure resulting from different soil management practices (Ball et al., 1988).

Air permeability is a function of pore characteristic and several soil hydrological properties, which are often more difficult to measure. Air permeability at −100 cm soil suction is a potential indicator for providing information about changes and differences of soil structure (Kirkham et al., 1958). Air permeability is related to air-filled macroporosity at different water contents to identify the changes in soil structure and soil water dynamics by soil management practices and biological activities (Blackwell et al., 1990), which are useful for studying the remediation of contaminated soils by modeling the soil-vapor extraction system (Moldrup et al., 1998). Tortuosity expresses a structural condition of soil and can be used as an index of soil structure (Moldrup et al., 2001). Soil structure has a strong influence on air permeability, and convective transport of air takes place through the larger pore networks in well-structured soils. The flow pattern in well-structured soils can be different for the air and water flow because of the differences in geometries and tortuosities of the two mediums. The saturated hydraulic conductivity is strongly correlated to air permeability at −100 cm of suction (Loll et al., 1999). The relationship between air permeability and saturated hydraulic conductivity in undisturbed soil media can be developed using pore scale network models (Fisher and Celia, 1999).

18.12.1 Governing Principles

According to Darcy's law for laminar flow, velocity of a given fluid is proportional to the pressure difference and inversely proportional to the length of flow path (Kirkham, 1946). Therefore, Darcy's law is applicable for the airflow through soils. The pore sizes and macropores or cracks greatly contribute to airflow in a soil. According to Poiseuille's

equation, air flow through a single pore varies as the fourth power of the pore radius $(Q_a \alpha r^4)$ According to Darcy's law, air permeability (k_a) can be defined by the following relationship:

$$q_a = -\frac{k_a}{\eta_a} * \frac{dp}{dx} \tag{18.37}$$

where q_a is the volume flux per unit area $(L^3 L^{-2} T^{-1})$; ηa is the dynamic viscosity of air $(ML^{-1} T^{-1})$; p is the pressure of air $(ML^{-1} T^{-2})$; and x is the distance in direction of flow (L). Air permeability of soil samples is calculated by modifying Eq. (18.37) as follows:

$$k_a = \frac{Q_a \eta_a L}{A \rho_a g \Delta H_a} \tag{18.38}$$

where Q_a is the volumetric flow rate $(cm^3\ s^{-1})$, ΔH_a is difference in pressure head (cm), ρ_a is density of air (gcm^{-3}), g is acceleration due to gravity $(cm\ s^{-2})$, A is the cross-sectional area (cm^2) and L is the length of the sample (cm). Note the dimensions of air permeability coefficient, k_a as L^2, which are similar to the intrinsic permeability of soil and therefore k_a is also referred to as intrinsic permeability of air (Reeve, 1953). The cross-sectional area and the length of soil sample are replaced by a shape factor to measure in situ air permeability.

18.12.2 Air Permeability Measurement Methods

Wyckoff and Botson made air permeability measurements as early as 1936 by forcing a mixture of water and air through long tubes of unconsolidated sands. The experiments were repeated with different flow velocities and water and air permeability were measured simultaneously. The air permeability methods can be broadly divided into steady state and unsteady state methods. The steady state methods are based on establishing a steady airflow rate at the inlet through the soil sample and measuring the flow rate and pressure head difference across the sample. The transient methods are generally quicker, easy to use, and require less volume of air for the experiment (Smith et al., 1998). Another approach to air permeability measurement is known as the acoustic technique. The following sections describe these methods in more detail, and the merits and demerits of each method are presented in Table 18.14.

Steady State Methods

Steady state methods can also be described as constant pressure gradient methods or constant flux methods (Grover, 1955). In this method, air is pressed across a core sample at a constant pressure above the atmospheric pressure (Fig. 18.4). The flow rate of air at the inlet end of the core is measured over a given time interval. The constant pressure gradient method is suitable for highly permeable samples and those at higher water content.

TABLE **18.14** Merits and Demerits of Air
Permeability Measurement Methods

Steady state laboratory method		Transient method		Acoustic method
Pressure gradient	Flux method	Core method	Field method	Field method
Constant pressure gradient	Constant flux	Drop in pressure in air tank	Drop in pressure in air tank	Reflection and transmission of audio frequency
Easy and simple	Constant flux and gradients are difficult to attain	Rapid and easy	Practical, rapid economical and easy	Rapid but requires skilled labor
Suitable for highly permeable soils	Suitable for less permeable soil	Suitable for both	Suitable for both	Suitable for homogeneous soils
Does not alter water content	Water content is altered	Does not alter water content	Does not alter water content	Does not alter water content
Disadvantage of air flow between soil and core	Disadvantage of air flow between soil and core Soil shrinkage	Disadvantage of air flow between soil and core	—	—
Well developed	Well developed	Well developed	Well developed	Under development
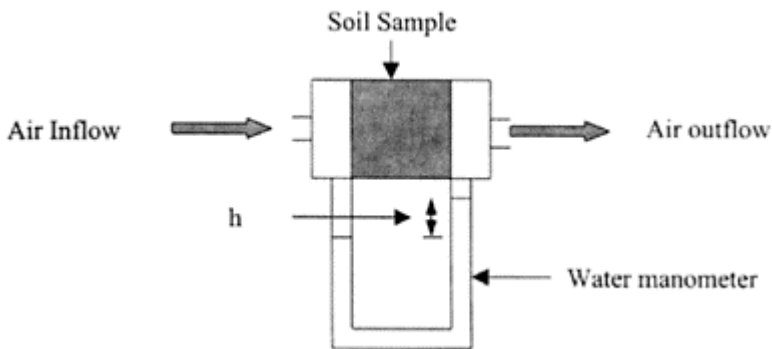
FIGURE 18.4 Schematic of the
apparatus of measure air permeability
(k_a).

Because the air pressure gradient is small for these soils, the measurements can be made
easily without altering the water content or liquid phase of the samples significantly. The

air permeability of soil using a constant pressure gradient method for air pressure less than 0.2 m of water can be calculated by Eq. (18.38).

The pressure difference across the core sample can be measured for an applied constant air flux at one end of the column (Blackwell et al., 1990). This method is relatively simple and requires a gas cylinder with a flow meter. The core sample can be placed inside a chamber and is connected to a water manometer for measuring the pressure difference across the sample, and the pressure gradient may be as high as 0.2 m of water. The air permeability can be calculated by the constant flux method as follows (Kirkham, 1946):

$$k_a = \frac{Q_a \eta_a L}{A \rho_a g \Delta H_a} \left[1 - \frac{\Delta H_a}{2H_i} \right]^{-1} \qquad (18.39)$$

If $\Delta H_a \ll 0.2\,\text{m}$, the term in parentheses approaches unity and Eq. (18.39) reduces to Eq. (18.38). The constant flux method is simple and straightfor-ward and is useful for soils of low permeability. However, it has two basic disadvantages: (i) the dry air changes the water content of core sample, and thus the air permeability and (ii) the constant air flux and gradient are difficult to achieve.

The conservation of moisture in the core sample is an important requirement for air permeability measurement. The water content of the core sample is a function of the pressure difference between both air and water; the pressure gradient between these two competing fluids should be equal in magnitude and direction. One of the methods is the stationary liquid method (Brooks and Corey, 1964), in which air flows upward and through the sample in response to a pressure gradient equal to that in a static liquid. The method was originally developed for measuring the air permeability of porous rocks. The details of this method are given by Corey (1986).

Another method uses both fluids flowing with equal pressure gradient in any direction. Ceramic porous plates are kept both at the inlet and outlet end of the system for controlling the inflow and outflow of water through the sample. The soil water pressure and piezometric head difference is measured directly by a pressure transducer connected with piezometer rings. The procedure allows the water to keep on flowing while permeability measurements are made. The inflow and outflow can be adjusted to obtain a steady state flow condition. A more accurate assessment can be made by simultaneously measuring the change of water content during permeability measurement mounting the soil sample on a scale and calculating the change in water content of soil sample by change in weight (Brooks and Corey, 1966). The merits and demerits of this method are presented in Table 18.14.

Transient Methods

Transient methods are more practical, quicker, less expensive, and easier to use (Table 18.14). They can be employed for air permeability measurements both in a lab on soil cores and in situ in the fields. They also require less volume of air to pass through the soil core or soil volume for a given permeability determination. The duration of the tests is shorter and soil desaturation or drying is less as compared to a steady state method. Kirkham (1946) first proposed the transient method as an in situ field method. In the

transient method one end of a soil core is connected to a close pressurized air tank and the other end is kept open (Fig. 18.5). The rate of drop of air pressure in the tank is measured as air flows out through the other end and is used to calculate the air permeability of soil (Kirkham,
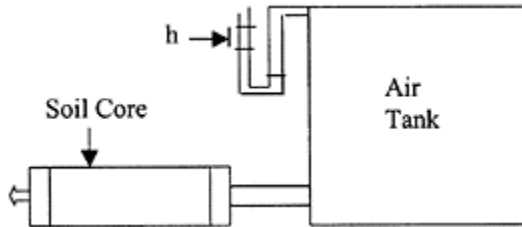

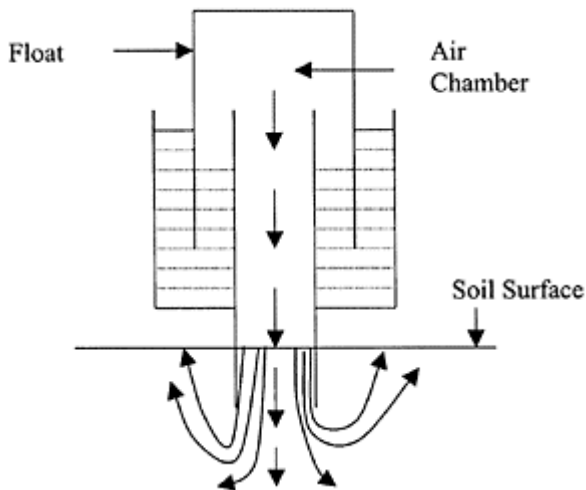**FIGURE 18.5** Schematic of an air permeability apparatus on soil cores for a transient method.

FIGURE 18.6 Schematic of an in situ air permeability apparatus for a transient method.

1946; Smith and Mullins, 1991; Stephens, 1996; Smith et al., 1997; Smith et al., 1998). Schematics of an in situ permeameter are given in Fig. 18.6 (Kirkham, 1946; Grover, 1955). Iversen et al. (2000) further modified the in situ air permeameter and replaced the float by a gas cylinder and a water manometer. Their apparatus is suitable for soil permeability measurements in situ, on site and in the laboratory on soil cores.

The in situ methods of air permeability measurements are preferred over the lab methods and have the advantage of assessing scale effects on permeability and anisotropy (Garbesi et al., 1996). However, these methods involve insertion of steel or plexiglass containers in the ground. Some of the disadvantages of these methods are the unknown sample length, disturbance of natural soil condition, and lack of or low reproducibility. Traditionally, air permeability calculations are made for isothermal conditions. Smith et al. (1997) showed that the traditional isothermal method is inaccurate because the pressure in the air tank cools the air inside. As a result, rate of change of air pressure in the tank does not represent the true mass flux out of the air tank and through the soil sample. Smith and colleagues proposed mathematical expressions for compensating the temperature effects for known temperature changes in the soil tank.

Acoustic Methods

Sound reflected from a soil surface interferes with the incident sound and causes an interference pattern in the total sound field. If the acoustic properties of porous media are known, interference patterns can be modeled from theory of sound propagation. Attenborough (1985) presented analytic approximations for calculating the sound levels from outdoor sound sources on or near the ground in terms of acoustic properties of soil. For a homogeneous porous media these acoustic properties depend on the air filled porosity of soil surface (Attenborough, 1987). The acoustic techniques involve the measurement of both the reflection and transmission of audio frequency and sound by the soil. Sound reflection measurements give qualitative indications of relative air permeability. Inserting a probe microphone at a given depth and keeping another vertically separated above ground makes sound transmission measurements. The theoretical predictions for homogeneous soils are fitted to the measured reflection and quantitative information on surface air filled porosity, and air permeability is obtained. The acoustic techniques have been validated on a series of trial plots, for a variety of soils and have been found within 10% of those obtained conventionally (Sabatier et al., 1990, Moore and Attenborough, 1992). They have shown a potential for the measurement and monitor-ing of management induced or seasonal changes in soil surface properties (Table 18.14).

18.13 MANAGEMENT OF SOIL AERATION

Poor aeration, due to inundation or compaction, affects crop growth by seedling mortality (Fig. 18.7) and creating poor soil physical conditions (Fig. 18.8). Therefore, management of soil aeration is important, which can be accomplished in a number of ways (e.g., tillage, drainage, mulches, etc.). Tillage is usually defined as the mechanical manipulation of soil to improve soil aeration conditions, control weeds, and incorporate organic matter in the soil, which directly affect crop production. Tillage practices on one hand open the soil and increase the porosity and soil aeration at least temporarily, and on the other hand compact the soil surface, increase bulk density, and reduce aeration for the soil below the plow layer. The use of conservation tillage and/or no-till improves the soil physical conditions and quality (Lal, 1989). A no-till system improves organic carbon

concentration in soil, which in turn improves aggregation and aeration porosity of soil. The aeration porosity and ODR under different tillage practices are generally significantly correlated (Flowers and Lal, 1998). The data in Table 18.15 show that aeration from no-till treatment is much higher than moldboard or chisel treatment for 5 cm and 15 cm depths, however, corresponding ODR values are not necessarily a direct function of soil aeration (Flowers and Lal, 1998). This apparent

FIGURE 18.7 Poorly drained sites are inundated during spring leading to seedling mortality, patchy stand, and low yields. (Courtesy of Dr. N. R. Fausey, USDA Drainage Unit, Columbus, OH.)

FIGURE 18.8 Soil structure is adversely affected in poorly drained soils. (Courtesy of Dr. N. R Fausey, USDA Drainage Unit, Columbus, OH.)

discrepancy may be due to the fact that intraaggregate ODR measure-ments are lower and represent anoxic conditions, whereas interaggregate measurements are higher and reflect higher aeration status between aggregates (McCoy and Cardina, 1997). The crop residue on soil surface

TABLE 18.15 Mean Aeration Porosity (f_a) and ODR at 5 cm Depth from Moldboard Plow (MP), Chisel Plow (CP), and No-Till (NT)

Moldboard plow		Chisel plow		No-till	
AP (cm^3 cm^{-3})	ODR (μgcm^{-2}s^{-1})	fa (cm^3 cm^{-3})	ODR (μg cm^{-2}s^{-1}))	fa (cm^3 cm^{-3})	ODR (μgcm^{-2}s^{-1})
5 cm depth					
0.054	12.2	0.13	12	0.185	12.5
0.06	13.8	0.145	11.8	0.195	12
0.12	22	0.19	23	0.23	17.5
15 cm depth					
0.002	7.5	0.035	10	0.058	8.8

| 0.018 | 5.8 | 0.045 | 5 | 0.072 | 9 |
| 0.07 | 13.5 | 0.095 | 21 | 0.105 | 19 |

ORD = oxygen diffusion rate; f_a = air porosity.
Source: Modified from Flowers and Lal, 1998.

TABLE 18.16 The Dependence of Oxygen Diffusion Rates on Volumetric Water Content (θ)

θ (cm^3cm^{-3})	ODR $(\mu gm^{-2}s^{-1})$	θ (cm^3cm^{-3}) $3cm^{-3})$	ODR $(\mu gm^{-2}s^{-1})$	(cm^3cm^{-3})	ODR $(\mu gm^{-2}s^{-1})$
0.34	86	0.38	68	0.5	25
0.36	74	0.52	28	0.57	20

Source: Modified from Flowers and Lal, 1998.

maintains the soil temperature and soil water content and increases earthworm activities and macropore channel formations, etc., which result in increased porosity and soil aeration. Soil aeration can also be improved by maintaining the soil below saturation levels. This can be achieved by installing surface or tile drainage systems especially in areas where groundwater table is shallow. The soil ODR values are strongly affected by soil water content and fluctuate in response to rainfall (Sojka, 1997; Flowers and Lal, 1998). As the water content increases from 0.38 cm^3 cm^{-3} to 0.5 cm^3 cm^{-3}, a sharp decline in ODR from 68 to 28$\mu gm^{-2}s^{-1}$ (>100%) is observed (Table 18.16) (Flowers and Lal, 1998). The use of different kinds of mulches on soil surface protect the soil from hot and cold cycles, freeze and thaw processes, and help maintain aggregation in soil consequently maintaining higher soil aeration.

18.14 WATER TABLE MANAGEMENT IN POORLY DRAINED SOILS

Water table management is defined as "the management or regulation of amount or volume of soil water in the profile for a healthy environment for plants." Water table management also implies the management of drainage system for maintaining the temperature and aeration in the soil for sustainable agriculture. Water table management consists of: (i) conven-tional subsurface drainage, (ii) controlled drainage, and (iii) subirrigation (http://www.ohioline.ag.osu-state.edu/).

   Water table management in its simplest form is synonymous with conventional drainage. It is practiced commonly in the midwestern United States to remove excess water from the soil profile. A conventional drainage system essentially consists of an outlet (ditch or a stream) for discharging excess water and drainage pipes or tiles, which are made of corrugated plastic tubing, clay, or concrete tile. The excess water in the soil profile enters the perforated drainage pipes installed at certain depth from soil surface and flows out to the open ditch or stream through gravity. The controlled drainage system is

very similar to a conventional drainage system except that in the former, drainage or outflow of water is intercepted by a control device, which manages the water table at any specified level below soil surface. The soil moisture status is better managed in a controlled drainage system. A subirrigation system is a combination of controlled drainage and an irrigation system and is used for drainage as well as
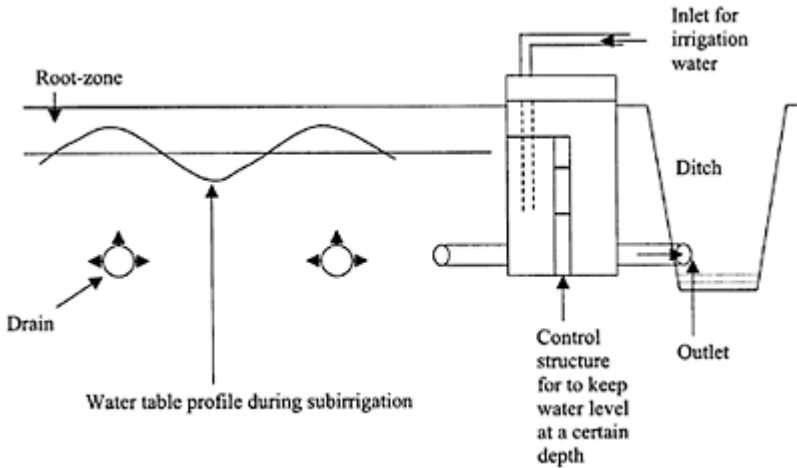

FIGURE 18.9 Schematic of a subirrigation method.

irrigation (Fig. 18.9). The drain spacing for a subirrigation is usually 30 to 50% less than conventional drainage system. Irrigation occurs below the ground surface and water table depth is maintained at a desired depth within the crop root zone.

The lowering of the water table during spring and fall facilitates field operations, and the raising of the water table during growing season by controlled drainage and/or subirrigation provides plants with much needed water. Drainage control strategies are used to improve the quality of surface water by keeping nitrate and other chemicals within the soil profile and not letting them flow out into surface drains. For reducing nitrate concentration in surface waters, three strategies are basically followed: (i) reduce air-filled porosity by maintaining a shallow water table within the root zone, thus creating anaerobic conditions to enhance denitrification; (ii) keep the water table shallow to reduce the volume of outflow from drainage; and (iii) decrease the leaching potential of soil nitrate by decreasing the depth of soil profile through which water infiltrates (Dinnes et al., 2002). Thus, in addition to improving quality of drainage water by reducing leaching of agrochemicals from soil profile, water table management can increase efficiency= crop production in three ways: (i) by retaining more nitrate in soil profile for plant, thus reducing fertilizer costs; (ii) increasing water availability during water demand pe=ds; and (iii) keeping adequate soil air in the profile (Mejia et al., 2000).

Example 18.1

Corn was grown in a 10 ha agricultural field, which has an effective root zone of 75 cm. Assuming the daily rate of soil respiration as $8gO_2m^{-2}$ and transpiration as 5 mm, calculate the fraction of the O_2 requirement supplied by convection. Also calculate the volume of CO_2 drawn from the atmosphere. Note that the air is drawn from the atmosphere immediately by the pressure difference created in the soil by soil moisture extraction, molecular weight of O_2 is 34 and CO_2 is 44, and 22.4 liter=1 mole of gas at STP.

Solution
   The volume of water extracted by roots from $1m^2$ area=$1m^2*0.005$ m= $0.005m^3$=5L.
   The same volume of air from the atmosphere will replace the volume of water extracted or removed from the soil. Therefore, volume of air drawn from atmosphere = 5L.
   The O_2 content of air is 21% and CO_2 is 0.037%.
   Volume of CO_2 drawn from atmosphere = (5*0.037)/100 =0.00185L
   Mass of CO_2 drawn from atmosphere=(0.00185/22.4)*44 = 3.63×10^{-3}g

Volume of O_2 drawn from atmosphere=(5*21)/100=1.05L
   Mass of O_2 drawn from atmosphere=(1.05/22.4)*32=1.5g
   Percentage of daily O_2 requirement supplied by convection=(1.5/8)*100= 18.75%

Example 18.2

Determine the gaseous concentration of O_2, N_2, and CO_2 at standard temperatures and pressure (25°C and 101.3 kPa, respectively). The partial pressure for O_2 is 0.21*atm and for N_2 is 0.79*atm, where atm is the atmospheric pressure= 1.013×105 Pa at sea level, the gas constant, R is equal to 8.314 Jmol^{-1} K^{-1}.

Solution
   Concentration can be calculated from the ideal gas law. Writing Eq. (18.18) in terms of partial pressures, the concentration on a mass basis is given as

$$C = \frac{M}{RT}p$$

   where C is concentration (kg m^{-3}), M is molar mass (kg mol^{-1}), R is gas constant, and p is partial pressure of gas (Pa).

$$\text{Concentration of } O_2 = \frac{0.032}{8.314 * 298} 0.21 * 1.013 x 10^5 = 0.275 \, kg \, m^{-3}$$

$$\text{Concentration of } CO_2 = \frac{0.044}{8.314 * 298} 0.0003 * 1.013 x 10^5 = 0.00054 \, kg \, m^{-3}$$

$$\text{Concentration of } N_2 = \frac{0.028}{8.314 * 298} 0.79 * 1.013 x 10^5 = 0.9044 \, kg \, m^{-3}$$

Example 18.3

For a soil of bulk density $1.2 Mgm^{-3}$ and a constant water content of $0.2 m^3 m^{-3}$, calculate the amount of O_2 and CO_2 in the soil for (a) water content retained at $0.2 \, m^3 \, m^{-3}$ (b) if 1 cm depth of water is removed by evaporation or drainage, and (c) if 1 cm depth of water enters the soil matrix. Assume air replaces water instantaneously and concentration of dissolved air does not change.

Solution

(a) The total porosity of soil$=f_t=1-(\rho_b/\rho_s)=1-(1.2/2.65)= 0.55$
The air-filled space in the soil $f_a=f_t-\theta=0.55-0.2=0.35 \, cm^3 \, cm^{-3}$

In a well-aerated soil, air contains 21% of O_2 and 0.05% of CO_2.
Calculating the concentration on 1 m^3 basis
The concentration of $CO_2=(44g/22.4 \, L)*0.05*(1000 \, L/1 \, m^3)=98.2 \, g \, m^3$
The concentration of $O_2=(32g/22.4 \, L)*0.21*(1000 \, L/1 \, m^3)=300g \, m^3$
Since air filled porosity was 35%, assuming 1 m^3 volume of soil, the total volume of air in the soil system$=0.35 \, m^3$
The amount of CO_2 in the soil$=98.2*0.35=34.37$ g
The amount of O_2 in the soil$=300*0.35=105$ g
(b) Total water in the soil profile (1 m^3)$=0.2*1*1*1=0.2 \, m^3$
If 1 cm depth of water is removed$=1/100*1*1*1=0.01 \, m^3$
The total volume of water left in the soil profile after 1 cm of water is removed$=0.19 \, m^3$
Water content of the soil$=0.19 \, m^3 \, m^{-3}$
The air-filled space in the soil $f_a=f_t-\theta=0.55-0.19=0.36 \, cm^3 \, cm^{-3}$
The amount of CO_2 in the soil$=98.2*0.36=35.35$ g
The amount of O_2 in the soil$=300*0.36=108$ g
(c) If 1 cm water is added, the air-fiilled pore space changes to$=0.55-0.21=0.34$
The amount of CO_2 in the soil$=98.2*0.34=33.39$ g
The amount of O_2 in the soil$=300*0.34=102$ g

Example 18.4

If the topsoil of a recently cleared and cultivated tropical forest with a topsoil depth of 35 cm and bulk density of 1.35 g cm^{-3} contains 2.5% of readily decomposable organic

residues having a carbon content of 30%, assuming constant O_2 consumption rate of 0.06 kg m^{-2}d^{-1}, calculate how much carbon is released to the atmosphere in three weeks. Note atomic weight of O_2 is 16 and C is 12.

Solution

Mass of soil in top layer=35*1.35=47.25 g cm^{-2}

The mass of decomposable organic matter=0.025*47.25=1.18 g cm−

The mass of carbon in top layer=0.3*1.18=0.35 g cm^{-2}

The daily Carbon release rate (from CO_2)=atomic weight of C/atomic weight of O_2=12/32

O_2 consumption rate=0.375*0.06=0.0225 kg m^{-2}

Total mass of C released in 21 days=21*0.0225=0.4725 kg m^{-2}

Daily organic matter decomposed=0.0225/0.25=0.09 kg m^{-2}

Total organic matter decomposed in three weeks=0.09*21=1.89 kg m^{-2}

Example 18.5

40 cm deep homogeneous soil matrix is at uniform water content of 25%. If the density of soil air as 1.275 kg m^{-3}, and k/η ratio equal to 60 μm2 s^{-1}, calculate the convective flow through soil if the pressure head difference is 10 cm.

Solution

The convective airflow through a unit cross section (q_v) can be calculated from Eq. (18.16).

q_v=60×10^{-6}*0.1/0.4=1.5×10^{-6}m^{2}s^{-1}

The convective flow in terms of mass=q_v*ρ_{sa}=1.5×10^{-6}*1.275=1.91× 10^{-6}kg m^{-1}s^{-1}.

Example 18.6

Assuming a homogeneous soil profile having a bulk density of 1.48 gcm^{-3}, and water content of 30%, calculate (a) air filled porosity and (b) effective diffusion coefficient (D_S), and diffusion rate using at least three models from Table 18.12 when O_2 concentration diminishes linearly from 21% at soil surface to 12% at 80 cm. The bulk air diffusion coefficient (D0) is given as 0.185 cm^2 s^{-1}, particle density of soil as 2.65 g cm^{-3} and concentration of O_2 in atmosphere as 0.0003 g cm^{-3}).

Solution

(a) The porosity of soil=(1−(bulk density/particle density))=1−(1.48/2.65)=0.44

Therefore, air filled porosity=0.44−0.3=0.11.

(b) The concentration of O_2 in atmosphere reduces to 0.12×32/22.4=0.00017 g cm^{-3}. We chose the following three models and calculated effective diffusion coefficient (D_s) and the steady state one-dimensional diffusive flux from Fick's law [Eq. 18.16)].

Change in O_2 concentration=0.0003−0.00017=0.00013 g cm^{-3}
Change in elevation=80−0=80 cm

Models	D_s (cm^2 s^{-1})	q_x (g cm^2 s^{-1})
Penman (1940)	0.0137	2.23×10^{-8}
Marshall (1959)	0.0067	1.09×10^{-8}
Millington and Quirk (1961)	0.0033	0.54×10^{-8}

PROBLEMS

1. Under normal standard pressure and temperature conditions (i.e., 101.3 kPa, and 25°C, respectively), calculate the ratio of mobility of O_2 and CO_2 in air and water. Use the diffusion coefficient values for O_2 and CO_2 as given in Table 18.11. The solubility coefficient for O_2 and CO_2 is 0.0333 and 0.942, respectively.

2. Calculate the amount of oxygen contained in the top 20 cm soil profile, if the soil volume is composed of 30% air, 25% of which is O_2. Soil consumes $6 \text{gm}^{-2} \text{d}^{-1}$ of O_2. Assuming no replenishment, if the consumption rate is constant, how many days will this stored O_2 last?

3. A homogeneous soil profile has a porosity of 0.48, and moisture content of 32%. The O_2 concentration diminishes linearly from 21% at soil surface (concentration of O_2 in atmosphere as 0.0003 gcm^{-3}) to 1/3 at a depth of 60 cm. Calculate the diffusion rate using any five models from Table 18.12. The bulk air diffusion coefficient (D0) is given as $0.185 \text{cm}^2\text{s}^{-1}$.

4. A static gas chamber has a diameter of 15 cm and is 15 cm high. Concentration of CO_2 in the chamber is increased by 200 ppmv in 15 minutes. Calculate efflux of CO_2 from soil air to the atmosphere.

5. Assume an effective rooting depth of 50 cm in corn at the initial tasselling stage of growth. The oxygen demand is $10 \text{gm}^2\text{d}^{-1}$. Calculate the proportion of O_2 supplied by convection (displacement of water by air), if the rate of water intake is 5 mmd^{-1}.

6. While measuring CO_2 flux, a field experiment conducted using a diffusion chamber technique showed that the CO_2 concentration in the chamber was 500 ppm in 10 minutes. If the chamber is 50 cm^2 in a cross-sectional area and 10 cm high, calculate the flux assuming that CO_2 concentration in the atmosphere is 0.03%.

7. For a soil with bulk density of 1.2 gcm^{-3} and gravimetric water content of 0.15, calculate the weight of O_2 and CO_2 (in gm^{-3}) in 1 m^3 of soil.

8. Calculate the effective diffusion coefficient (D_s) for the soil in problem 7, assuming that diffusion coefficient in air is 0.2 cm^2 sec^{-1}.

9. Compute O_2 influx from atmosphere into soil using effective diffusion coefficient calculated in problem 8, when O_2 concentration in 50 cm depth is 59% of that in the atmosphere.

10. The following data on methane emission was obtained from an experiment in Houston, Texas. Review these data and list soil physical properties and processes that will reduce CH_4 emissions.

Treatment	Seasonal methane emission (gm^{-2})
Straw	
No straw	27.4
With straw	35.6
Tillage	
Late tillage	15.8
No-tillage	16.5
Early tillage	14.2
Flooding	
Late flood	15
Normal flood	9.3
Midseason flood	4.9
Multiple aeration	1.2

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