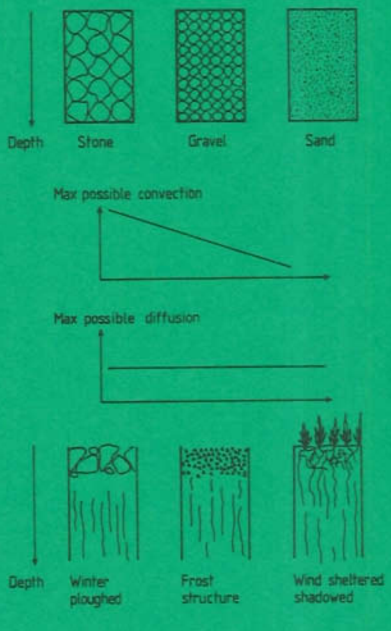
SVERIGES
LANTBRUKSUNIVERSITET

SOIL AIR

Volume and Gas Exchange Mechanisms

Peter Edling



Institutionen för markvetenskap
Avdelningen för lantbrukets hydroteknik

Swedish University of Agricultural Sciences
Department of Soil Sciences
Division of Agricultural Hydrotechnics

Rapport 151
Report

Uppsala 1986

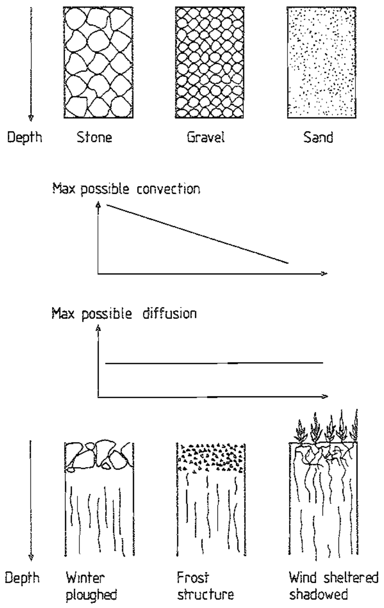
ISSN 0348-1816
ISBN 91-576-2764-9

SVERIGES
LANTBRUKSUNIVERSITET

SOIL AIR

Volume and Gas Exchange Mechanisms

Peter Edling



Institutionen för markvetenskap
Avdelningen för lantbrukets hydroteknik

Swedish University of Agricultural Sciences
Department of Soil Sciences
Division of Agricultural Hydrotechnics

Rapport 151
Report

Uppsala 1986

ISSN 0348-1816
ISBN 91-576-2764-9

Tryck: Sveriges lantbruksuniversitet, Uppsala 1986

Distributor: Sveriges lantbruksuniversitet
Institutionen för markvetenskap

Peter Edling, Valla gård, Husby-Oppunda,
640 33 Bettna

Soil air, volume and gas exchange mechanisms

Akademisk avhandling som för avläggande av agronomie doktorsexamen kommer att offentligt försvaras i Sal L, Undervisningshuset, Sveriges Lantbruksuniversitet, Uppsala, fredagen den 31 oktober 1986 kl. 10.15.

Abstract

This study concerns soil air as a part of the three phase system, solid material, water and air which forms the soil.

An idea of soil air, volume, composition, ecological significance and exchange mechanisms, is formed from the literature. An apparatus for experiments with gas diffusion through naturally layered moist soils is described. The possible replenishment of soil air by diffusion and mass flow is studied. A study on gas diffusion and mass flow on a number of dry and wet laboratory prepared soils is carried out, followed by a study of a number of well described naturally layered soils.

From the information acquired so far, the possible significance of the experimental results on the composition of soil air and its ecological effects is discussed.

The final conclusions are in agreement to those of Lundegårdh (1927), soil aeration status must be regarded on exchange possibilities on the one side and production and consumption of its components on the other, a highly dynamic system.

Key words: Soils, Soil pore system, Soil air, Root environment, Diffusion, Air permeability, Gas exchange in soils

ISSN 0348-1816
ISBN 91-576-2764-9
Tryckort: Uppsala
Tryckår: 1986

CONTENTS

1. INTRODUCTION

Since its foundation in 1932 the Department of Agricultural Hydrotechnics has investigated the soil as a water-binding and -storing system in theoretical, laboratory and field research. Basic research has been concentrated on static properties of the triphasic system Solids - Water - Air and interest has been focussed on the Water factor. However, water and air compete for room in the soil pore space. When water is removed, air enters (and vice versa). This means that the degree of water saturation in the pore space determines the volume available to the air phase.

Hydrotechnical procedures such as drainage, irrigation and soil improvement, which are aimed at controlling soil water, also influence soil air. Furthermore, many of the ecological effects of controlling water are direct results of changes in the soil air phase. A closer study of soil air is therefore of direct interest in hydrotechnical research.

Soil air must be considered as a dynamic entity despite the fact that changes in its composition do not necessarily involve volume changes. Even if the air content of a soil remains constant, changes occur continuously in the air phase - water vapour and carbon dioxide escape to the atmosphere while oxygen enters the soil air. The aim of this investigation was to determine the quantities of gases which can be transported through soil under various conditions. Results of experimental work to determine possible gas transport in soils form the central part of this report.

The background to experimental work is provided by a preceding section of the report where general facts from current literature are expressed in the form of basic models.

Experiments were carried out to determine the rate at which gas can diffuse and flow through soils in disturbed and natural conditions. Investigations on soils in natural conditions were concentrated on higher moisture content values. At such moisture contents, evaporation losses are greatest and transport of ecologically active gases such as oxygen and carbon dioxide is most restricted. Soils with these moisture conditions have received least attention in the literature so it was hoped that this investigation could remedy this deficit and also illustrate aspects of oxygen supply to plant roots and movements of water vapour in soil.

Presentation of experimental results in this report is followed by a consideration of soil ventilation based on experimental experiences and theoretical deductions.

This report is a revised version of the author´s licentiate thesis from 1973. For support, help and encouragement in the efforts to create this report, I want to thank Professors Janne Eriksson and Waldemar Johansson, Mrs Mary McAfee, Mrs Barbro Hultqvist, Mr Jan Lindström, Mr Hans Johansson, Mr Peter Wigren, my wife Hélène and my family and finally all my friends and colleagues in the farming community of Södermanland. For inspiration over the years I want to thank my former chief Professor Sigvard Andersson.

2. MODEL REPRESENTATION OF SOIL AIR - ITS ROLE AND CHANGES IN ITS COMPOSITION

2.1 Volume relationships in soil

The soil consists of three phases - solid, liquid and gas. The solid phase, which consists of mineral particles and organic matter, forms the soil skeleton and this provides the floor, walls and roof of the pore space. Thus the pore space can be a simple or complicated system of canals and rooms, depending on the skeletal formation. The water and air phases compete for space in the pores and their relative proportions are determined by the degree of saturation of the soil. Division of a soil into its three phases can be shown in a volume diagram (Fig. 1).

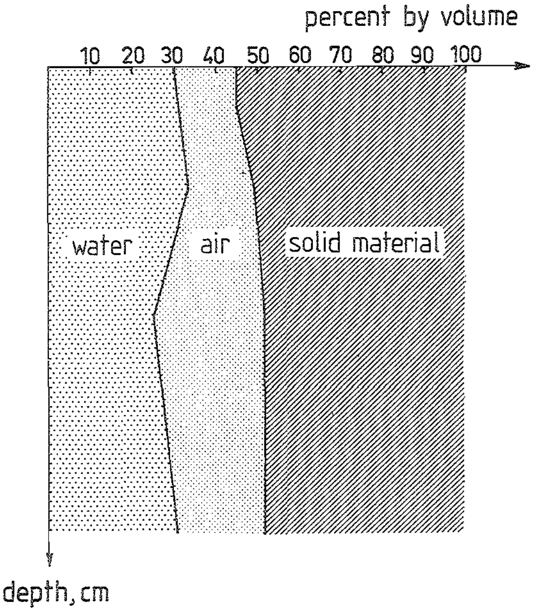


Figure 1. Volume diagram.

In a model, the pore system of the soil can be represented by a number of regular, parallel capillaries of various diameter. In any given capillary with diameter d and in contact with a watertable, the height of water rise in that capillary is determined under ideal conditions by the formula for capillary rise $h_c = 0.3/d$. Saturation level for a particular horizon in the profile is determined by the pore size distribution, watertable depth, evaporation and infiltration effects. A possible model presentation of the soil pore system is shown in Fig. 2.

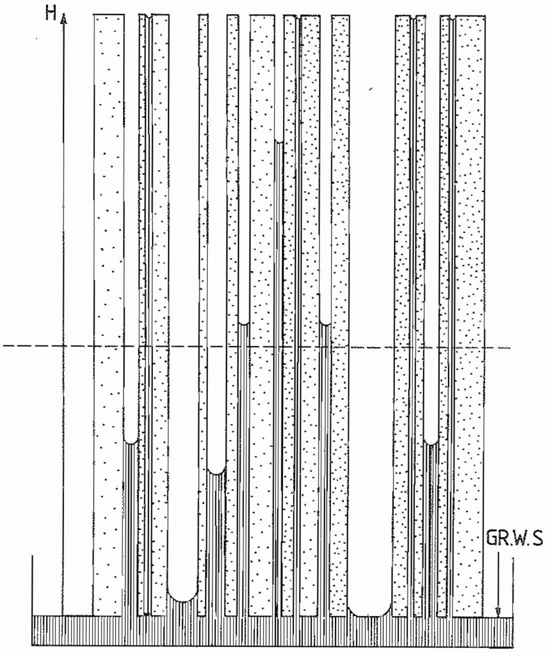


Figure 2. Model of a simple capillary system

In an actual soil profile, the situation is infinitely more complicated than the model shown, even though water and air content of the soil is determined by the same factors of pore size distribution, watertable depth and changes in the soil moisture store. Additions to and subtractions from the soil moisture store are functions of climate, topography and vegetation. Fig. 3 shows a simplified diagram of the proportions of solids, water and air at different levels over the watertable in a natural soil.

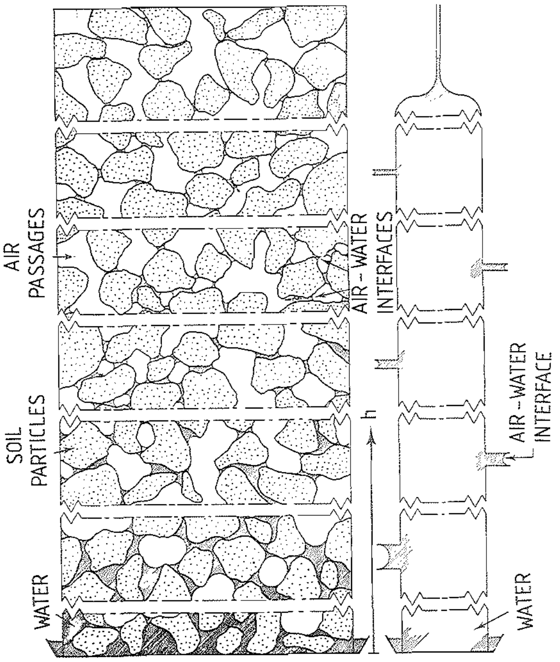


Figure 3. Model of a coarse-textured natural soil. After Gardner & Chatelain (1947).

Differences in pore size distribution while all other conditions remain constant lead to differences in the proportions of water and air in a soil. An example of this is shown in Fig. 4 where the wh_t - relationship (Andersson, 1962; Andersson & Wiklert, 1970) or the soil moisture characteristic is plotted for a heavy clay and a silt soil. Assuming that homogenous profiles of these materials are in contact with a 1 m deep watertable, tension at any point in the profiles is proportional to the height of that point over the watertable. The wh_t - relationship shows the proportions of air and water in the profiles to be different at any particular tension. When height over the watertable increases from 0 to 1 m, moisture content of the clay soil decreases continuously and its air content increases by a corresponding amount while the moisture content of the silt remains unchanged. Only if the watertable falls below 10 m depth will the upper part of this silt profile drain in any significant amount.

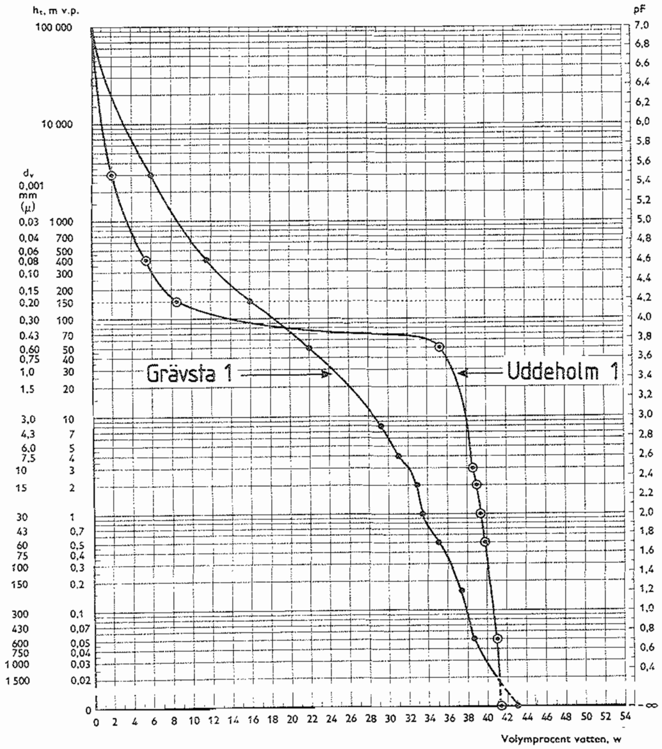


Figure 4. wh_t-relationships or moisture-characteristics for a heavy clay, Grävsta, and a silt soil, Uddeholm.

In addition to providing a basis for discussion of soil moisture storage and permeability properties (Andersson, 1962 & 1969), the wh_t - relationship is also a starting point for studies of soil air. But it is not sufficient to consider air volume alone, the status of this air must also be evaluated by its changing composition. This report expands a hypothesis which was first put forward by Lundegårdh (1927), that soil air must be considered on the basis of utilization and production of its component gases and on its means of replenishment.

2.2 General properties of soil air

In a sterile soil such as a dry gravel heap, the composition of soil air is identical to that of the surrounding atmosphere. The situation is different, however, in a living soil (Romell, 1922). Soil air has generally the following features:

a) its oxygen content is lower and its carbon dioxide and moisture contents are higher than those of the atmosphere. However, the sum of oxygen and carbon dioxide contents is usually 21 % in both atmosphere and soil air,

b) soil air composition varies with time and with depth in the soil profile. The oxygen content generally decreases and the carbon dioxide and moisture contents increase with depth.

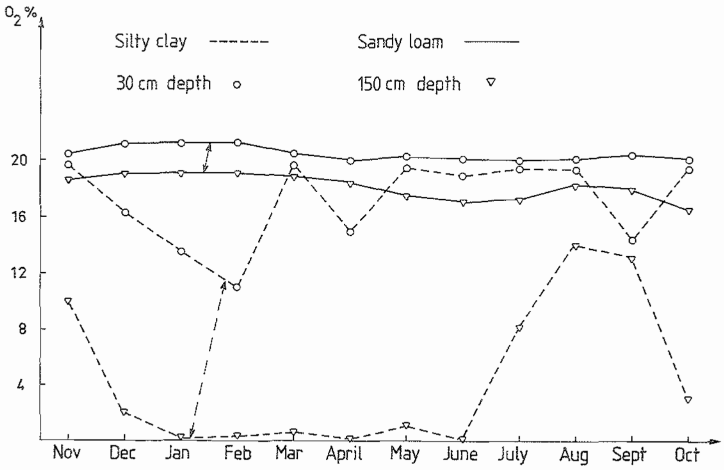


Figure 5. Seasonal variation in oxygen content of soil air at 2 depths for 2 soil types. (After Boynton & Reuther, 1939).

Fig. 5 shows seasonal variations in oxygen content with depth in two American garden soils. Different time and depth functions are obtained under different climatic conditions. An English investigation (Rusell & Appleyard, 1915) found distinct spring and autumn minima and summer and winter maxima for these functions. Oxygen content decreases rapidly after rain or irrigation because the air-conducting passages penetrating the soil are blocked. When this water drains from the larger air canals, leaving them open again, oxygen content returns to normal. A secondary, very gradual decrease in oxygen content may then occur because of increased root and microbe activity combined with delayed ventilation effects (Flühler & Läser, 1975).

In special circumstances, the sum of oxygen and carbon dioxide contents can deviate from 21 %. Such circumstances are caused partly by intense nitrification, partly by rapid replenishment of soil moisture reserves (Russell & Appleyard, 1915).

During intense nitrification oxygen is consumed in the oxidation of ammonia compounds without any release of other gases, so the sum of oxygen and carbon dioxide contents falls below 21 %. During rapid replenishment of soil moisture the different solubilities of oxygen and carbon dioxide in water leads to temporary changes in the CO_2 content so that the sum of oxygen and carbon dioxide contents can fall below 21 %. Gases which are soluble in water form a type of gas store which is not actually part of the soil air phase. Russell & Appleyard (1915), Romell (1922) and Greenwood (1969) talk of a 'second atmosphere'. This 'second atmosphere' has the same relationship to soil air as soil air has to the atmosphere, namely lower oxygen content and higher carbon dioxide content.

It would seem that soil ventilation conditions are more important than seasonal changes in biological activity in determining oxygen content of soil air. Fig. 5 shows for example a case where low oxygen content occurred in a denser soil during winter when biological activity was lowest. This agrees with results described by Stolzy & Flühler (1978).

2.3 Ecological functions of soil air

Soil air is ecologically active in its functions as an oxygen source for plant roots and as a sink for waste products from roots. Soil air also aids drying by its movements through the profile and this can have the ecological effects of drying wet soils or dessicating dry soils.

Oxygen derived from soil air by roots and microbes is used in aerobic respiration for breakdown of energy compounds and release of energy. Energy compounds are produced by photosynthesis or obtained from outside sources. Aerobic respiration releases energy according to the formula $C_6H_{12}0_6 + 6\ H_20 + 6\ 0_2 \rightarrow 6\ CO_2 + 12\ H_20 + 690$ cal.

If oxygen is not available, plant structures and microbes can release energy by anaerobic respiration or fermentation. Breakdown of a glucose molecule by this process follows the formula $C_6H_{12}0_6 \rightarrow 2\ CO_2 + 2\ C_2H_5OH + 30$ cal. Energy exchange is low and as a result of continued anaerobic respiration plant-toxic substances such as aldehydes, alcohol etc. accumulate in the soil. These lead to reduced growth of the higher plants and aerobic bacterias (See Gliński & Stępniewski, 1983).

The change from aerobic to anaerobic respiration occurs gradually as oxygen content decreases. There is, however, a minimum oxygen content of soil air below which the higher plants cannot survive. This is called the 'lower critical boundary' (Cannon, 1925). The upper critical boundary occurs when no increase in plant growth is produced by a further increase in oxygen content of soil air. This boundary is sometimes referred to as the lower optimal boundary. An upper optimal boundary is defined as the point when 'luxury breakdown' of the products of photosynthesis occurs (Andersson & Kemper, 1964).

The effects of carbon dioxide concentration in soil air have been widely discussed. Tacket & Pearson (1964) found that roots could continue to grow at very high CO_2 concentrations as long as oxygen supply was not limiting. Kowalik & Stepnięwski (1979) agree with this finding in principle while noting the existence of conflicting results in the literature.

During purely aerobic respiration the mole ratio of CO_2 produced to 0_2 consumed is approximately 1:1, i.e. the respiratory quotient is 1. A respiratory quotient greater than 1 indicates anaerobic conditions. The mole ratio varies somewhat depending on the compound being broken down - it is 1,3 for organic acids, 1 for carbohydrates, 0,8 for proteins and 0.7 for fats. Thus the respiratory quotient may exceed 1 even under aerobic conditions, for example in combustion of high-oxygen, low-hydrogen compounds such as oxalic acid (Gliński & Stępniewski, 1983).

Possible respiration rate is determined by the quantity and viability of the root mass and soil microflora and by soil temperature.

All ecologically active factors including soil air, are directly involved in stimulating root and microbial growth and thus increasing respiration rate.

The relationship between temperature and respiration rate follows the van't Hoff law - reaction rate is doubled or trebled for every 10°C increase in temperature (Jensen, 1960). However, oxygen diffusion rate into tissues gradually restricts rate of aerobic respiration since rate of physical processes such as diffusion is not increased to the same extent by temperature increase (see Section 2.4.1). When oxygen demand becomes greater than supply, root growth is inhibited by lack of oxygen. An example of this is shown in Fig. 6. Optimum temperature, above which internal transport mechanisms of the root become limiting, is lower for the lower O_2 concentration.

It should be noted that while on one side certain processes can continue at very low O_2 concentrations, on the other side very small oxygen deficits can affect root function (Amoore, 1961). The significance of this is apparent in experimental results recorded later in this report.

Root and microbe activity (O_2 consumption and CO_2 production) is in general not uniformly distributed throughout the profile. Activity varies with the number and location of roots and microbes and with temperature changes in the soil. A 'soil activity profile' can be produced for a particular soil by determining biological activity at various depths.

Capacity for biological activity of a soil is 5 - 20 litres O_2 produced per day and m^2 of surface, assuming that respiration is aerobic (Romell, 1922; Currie, 1962; Brown et al., 1965; Greenwood, 1969).

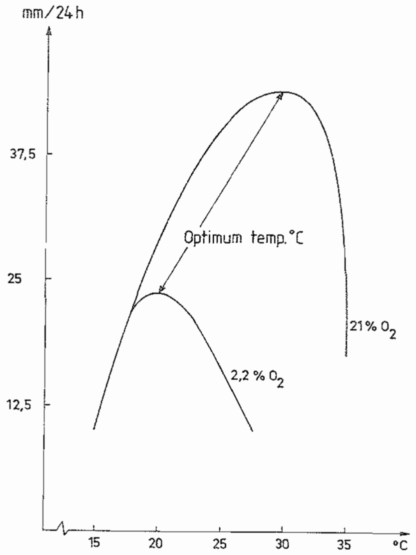


Figure 6. Root growth of Gossypium barbadense (mm/day) as a function of temperature (^{0}C) at two oxygen concentrations around the root. (After Cannon 1925)

When the oxygen necessary for root and microbial growth cannot be supplied, growth of the higher plants is inhibited by direct and indirect causes. Factors acting directly on higher plants have been mentioned previously - toxicity effects of byproducts of anaerobic respiration and decreased energy exchange from the respiration process. The latter effect leads to inhibited uptake of moisture and nutrients, resulting in one of the more acute symptoms of oxygen deficiency, wilting.

Indirect effects of lack of oxygen on higher plants are a result of changes in the microflora and chemical environment of the soil. They lead in the long run to changes in soil pH, reducing conditions and anaerobic microbial activity. These changes are characterized by occurrence of denitrification (nitrogen losses), absence of nitrification and abnormal concentrations of certain cations (Fe^{2+}, Mn^{2+}) All these factors inhibit growth.

As early as 1663 Schering Rosenhane wrote "...so you must in any case avoid what is wet and boggy, because it is sure and known, that when the roots reach the water or a wet ground, they must die".

Many parameters by which soil aeration can be evaluated as an ecological factor have been proposed, for example:

1. Air content of the soil.
2. Soil air composition.
3. Gas diffusion ratio, relative diffusion.
4. Soil permeability to air.
5. Production and consumption of components of soil air, activity.
6. Adaption of the soil moisture characteristic.
7. Redox potential.
8. Polarographic O_2 gas diffusion, oxygen diffusion rate (ODR).

Systems 7 and 8, which present many technical and methodological problems will not be discussed further in this report.

A problem with system 5, soil activity, is when CO_2 production is used as a measure of soil oxygen status. Since the respiratory quotient may be greater than 1, CO_2 production is not an automatic measure of O_2 consumption though it is an accurate indicator under pure aerobic conditions (Stolzy & Flühler, 1978).

2.4 Air exchange in soil

Soil air must be continually replaced, otherwise all its oxygen would be consumed and replaced by carbon dioxide. Replacement of soil air takes place by way of different transport mechanisms - diffusion, effusion and convection (mass flow). All these obey the well-known proportionality law:

$$q = -k \cdot A \cdot I$$

where q = resultant transport per unit time,
k = transport coefficient,
A = cross-sectional area perpendicular to flow
I = gradient, in pressure or concentration.

This relationship is presented below in the form of Fick´s, Knudsen's and Darcy's laws, which differ in their coefficient and in the way the gradient is calculated.

2.4.1 Diffusion

The molecules in a liquid or gas strive to disperse themselves statistically throughout the available space. This means that the system is continually striving to even out differences in concentration regardless of whether total pressure is constant or fluctuating. This process is known as diffusion and it continues as long as concentration gradients are created in the system. In soil such gradients are created by consumption and production of gases, for example in root respiration, microbiological activities and evaporation.

Diffusion obeys Fick's law, which can be written thus (Stefan, 1890; Krischer, 1963; Bygg, 1972):

$$q = -D \cdot A \cdot I \qquad (1)$$

where D = coefficient of diffusion. Dimensions of D depend on dimensions in which gradient I is measured. If I is concentration gradient, D has dimensions area/time. In this case Fick's law is written

$$q = -D \cdot A \cdot dC/dz \qquad (2)$$

where dC/dz = concentration gradient

D is a function of properties of the diffusing gas, the mixture of gases present, pressure and temperature. This means that a gas in a particular combination of gases has a particular diffusion coefficient. The influence of pressure and temperature is shown in this equation by Kennard (1938):

$$D_1 = D_2 \cdot P_2/P_1 \cdot (T_1/T_2)^n \qquad (3)$$

where D_1, D_2 = diffusion coefficients for a certain substance in a particular combination at total pressure P_1 and P_2 respectively and absolute temperature T_1 and T_2 respectively and n = exponent with value 2.00 for vapours, 1.75 for gases.

For diffusion in bulk gas D can be referred to as D_0. Table 1 shows some values for these.

In current literature and in this report, the fact that D is also a function of composition of the gas mixture is disregarded. During drying of a soil, the diffusion coefficient may change by one or two percent with the change from diffusion in a wet to a progressively drier soil air (Stefan, 1871).

Table 1. Values of diffusion coefficients in bulk gas. D_0 values are obtained from Landolt-Börnstein (1969).

Diffusing gas	Other gas present	Temp., ^0C	D_0, cm^2/s
H_2O	air	0	0.219
H_2O	O_2	35	0.282
H_2O	N_2	34.4	0.256
CO_2	air	20	0.155
CO_2	O_2	20	0.153
CO_2	N_2	20	0.163
O_2	air	0	0.178
O_2	air	10	0.189
O_2	N_2	20	0.22
CS_2	air	20	0.102

It should be noted that the coefficient for diffusion in water is around 10,000 times less than the coefficient for diffusion in air.

In a body of gas free from other phases (bulk gas) the diffusion coefficient is, as previously discussed, D_0. In a system which is partly filled with air (e.g. a soil) flow by diffusion is restricted by the presence of the other phases. The relationship between the coefficient for diffusion through such a system D and coefficient D_0 above is known as relative diffusion, D/D_0.

D/D_0 is a measure of the permeability of a system to diffusion and is a function of air content and air distribution in that system. D/D_0 can be expressed thus:

$$D/D_0 = f\ (f_a, a, b, c) \qquad (4)$$

where f_a = air content of the system, air-filled porosity,
a = real average path of diffusion in soil (tortuosity)
b = frequency of narrow or blocked pores
c = other factors

Equation (4) can also be written:

$$D/D_o = \frac{f_a}{100} \cdot S_D \qquad (5)$$

where S_D = combination of factors a, b, c in (4).

S_D, specific diffusion, is obtained by rearranging (5).

$$S_D = \frac{100}{f_a} \cdot \frac{D}{D_o} \qquad (6)$$

Specific diffusion, S_D, can be said to be "relative diffusion", D/D_o, per unit air in a porous media.

During diffusion, no friction occurs between the air pathways and the transported gas. This means that D is independent of mean pore diameter or pore size distribution. D and therefore D/D_o are determined by the total cross-sectional area of pathway in a system.

In its simplest form, diffusion through a system consisting of uniform, straight pores in the direction of gradient is $D/D_o = f_a/100$. S_D in this case is equal to 1, since pores do not twist and turn or vary in diameter. In natural systems, $S_D < 1$.

The relationship between D/D_o and f_a has been widely investigated. Some representative values from various sources are shown in Table 2. Methods by which these results were obtained are described in Section 3.3.1.

According to these, diffusion is independent of average pore radius but proportional to total pore cross-sectional area. Very fine pores are an exception to this (see Section 2.4.2).

Fig. 7 shows in principle how diffusive flow (mass diffusing per unit area and time, q_D) is dependent on total cross-sectional area of the pore system (ΣA) and independent of pore size distribution or pore radius (r).

Table 2. Relationship between relative diffusion D/D_o, specific diffusion (S_D) and air-filled porosity (f_a, %).

Source	D/D_o	S_D	Remarks
Buckingham (1904)	$(f_a/100)^2$	$f_a/100$	
Penman (1940)	$\frac{f_a \cdot 0.66}{100}$	0.66	
Blake & Page (1948)	$\frac{f_a}{100} \cdot 0,67$	0,67	Two different soils investigated
	$1,3(\frac{f_a}{100} - 0,1)$	$1,3 - \frac{13}{f_a}$	
van Bavel (1952)	$\frac{f_a}{100} \cdot 0,6$	0,6	
Marshall (1960)	$(\frac{f_a}{100})^{\frac{3}{2}}$	$(\frac{f_a}{100})^{\frac{1}{2}}$	Theoretical analysis
Currie (1960)	$\alpha \cdot (\frac{f_a}{100})^{\beta}$	$\alpha \cdot (\frac{f_a}{100})^{\beta - 1}$	

2.4.2 Effusion

When diffusion occurs in very fine canals, collisions between diffusing molecules and the canal walls influence diffusion rate. This effect is known as effusion or Knudsen motion and it is greatest in canals whose radius approaches the mean free path (distance a molecule in random motion travels before colliding with another) of diffusing molecules. After repeated impacts with the canal walls, molecules ricochet forward in thermal equilibrium with molecules of the wall material. Effusion becomes significant at canal radius of approximately 0.1 µ. This value can be compared with the mean free path of air molecules which is around 0.06 µ at 0°C.

Effusion through a straight, cylindrical canal obeys Knudsen's law (Krischer, 1963; Bygg, 1972)

$$q = -\pi \cdot r^3 \cdot \frac{8}{3} \cdot \sqrt{\frac{g}{2 \cdot \pi \cdot R} \cdot \frac{M}{T}} \cdot \frac{dp}{dz} \qquad (1)$$

where q = mass of gas transported per unit time,
r = canal radius.
g = acceleration due to gravity,
R = universal gas constant
M = molecular weight of the effusing gas
T = absolute temperature
dp/dz = gradient, partial or total pressure

If equation 1 is converted to the form $q = -k \cdot A \cdot I$ it can be seen that k_E (coefficient for effusion) is proportional to canal radius and to total cross-sectional area of canals. Very small canal radius means that friction forces along the walls become significant and transport then becomes dependent on canal radius. Fig. 8 indicates how effusive flow is related to canal radius and total cross-sectional area of canals.

2.4.3 Convection (mass flow)

A total pressure gradient in a system results in flow of a substance from a zone of higher pressure to one of lower pressure. The convective flow of gas or water through a porous medium is described by Darcy's law (see Andersson, 1953):

$$q = k \cdot A \cdot I \qquad (1)$$

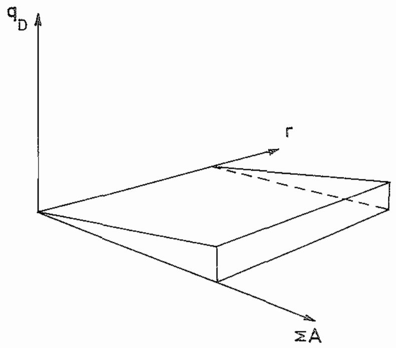
where q = mass of gas flowing per unit time,
k = permeability coefficient
A = cross-sectional area
I = pressure gradient, pressure drop in the flowing mass.

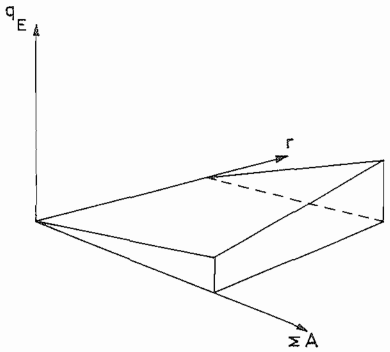
The permeability coefficient is closely dependent on pore size distribution of the system. For a system consisting of one straight, cylindrical tube the Hagen-Poiseuille law can be applied:

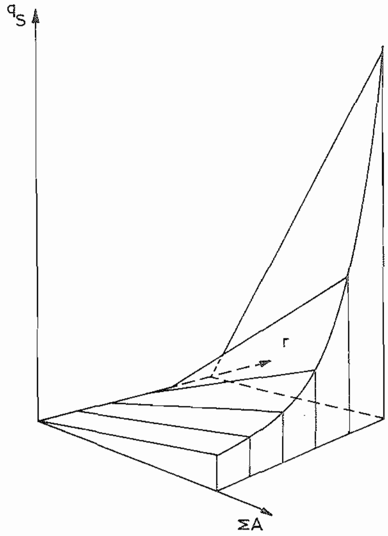
$$q = \frac{\pi \cdot r^4 \cdot \gamma \cdot I}{8 \cdot \eta} \qquad (2)$$

where r = radius of the tube,
γ = weight of the flowing gas or liquid,
η = viscosity of the flowing gas or liquid.

Total cross-sectional area in this case is πr^2 and this can be doubled by increasing the radius of the cylinder by 41.4 % or by introducing a second, identical cylinder. In the former case, the permeability coefficient of the system is quadrupled and in the latter case it is doubled. It becomes obvious that the permeability coefficient is proportional to both total crossectional area of the canals and to the square of their radius (Andersson, 1953). This implies that friction is acting against flow. Fig. 9 expresses graphically the relationship between convective flow, total cross-sectional area of canals and the square of their radius.







Figures 7, 8, 9.
Diffusive flow (qD), effusive flow (qE) and convective flow (q_S) through a pore system as a function of pore radius (r) and total cross-sectional area (ΣA).

The pressure gradient which acts as a driving force for convection of soil air is created mainly by micro- and macroclimatic factors outside the pore space. Such factors include barometric pressure changes in the atmosphere, temperature gradients, water movements in the soil profile and wind gusts over the soil surface (Romell, 1922). Other factors inside pores can also cause convective flow, for example when the respiratory quotient exceeds 1, i.e. production of gas exceeds consumption. The most striking example of this can be seen in marshes when gases formed during anaerobic breakdown of the organic matter bubble up to the surface.

According to Boyle's law, air density changes with changes in air pressure. This means that volume of air in a pore which is in contact with the watertable or the bedrock expands and contracts as air pressure falls and rises. This movement causes some convective flow through the soil surface and direct replacement of air in the upper layers of the profile. This effect is enhanced by the fact that air content of the profile usually decreases with depth (water content normally increases with depth under Swedish conditions).

Diurnal fluctuations in temperature cause flow of soil air partly by expanding and contracting the air, partly by convection when cold air meets warm. The former effect is similar to a pressure change and the entire column of air moves up or down without any mixing. The latter effect, convective flow, is similar to formation of cumulus cloud where warmer air is pressed upwards because of its lighter mass and colder air rushes in to replace it. Typical conditions for this are on a summer evening when the earth surface is warm and the air cool. Russell & Appleyard (1915) note that oxygen deficits are remarkably small on a warm day preceded by a frosty night. Romell (1922) found, however, that this effect was very small.

Water movement in a soil profile resembles a piston in its action as air is pumped into and out of the soil. When the water content of a layer increases, air is pressed out and conversely when the water content decreases, air is drawn into the soil. Oxygen-rich surface water infiltrating through the profile may also transport dissolved O_2 down to the lower soil layers (Russell, 1961). However, the overall effect of water movement on soil air replacement is one of inhibition. Increased degree of saturation causes expulsion of soil air which is a positive effect in the short run, but in the long run reduction of soil air volume has the negative effect of blockage of air pathways (Flühler & Läser, 1975).

The effects of wind are variable and depend on pore structure of the upper soil layers and the amount of plant cover on the soil surface. In coarse, bare soil such as a ploughed field, pressure differences between points on the irregular surface during wind gusts are relatively large. These effects must decrease rapidly with depth in the profile. Soils covered by vegetation and with a compact structure are less affected by wind because friction is greater in the narrower pores of such a soil and the plant cover reduces wind action.

Romell (1922) concluded that the contribution made by convective flow to air exchange in covered soils is only 1 to 2 % of the total. Measurements carried out in auger holes 1-2 cm in diameter (Jonsson & Lindström, 1985) has given stable equilibrium values for O_2 content, indicating that occasional contributions by convection are negligible. The relative contributions of diffusion and convection to gas transport were not measured directly in this study.

2.4.4. Theoretical comparison of gas transport mechanisms in model capillary systems

2.4.4.1 Transport through a straight, cylindrical capillary with constant diameter

From the previous sections, we know that friction forces are significant when gas transport in a pore system occurs by effusion and convection. This means that pore size distribution has a considerable influence on rate of transport by these processes. Information on the relative importance of various transport mechanisms can be obtained by considering a model of a pore system. In its simplest form, the system consists of one uniformly straight, cylindrical capillary for which different diameters can be chosen. Fick's, Knudsen's, and Hagen-Poiseuilles laws apply:

$$q_D = -D \cdot \frac{M}{RT} \cdot \pi \cdot r^2 \cdot \frac{dp}{dx} \qquad \text{(Fick's law)}$$

$$q_E = -\frac{8}{3} \cdot \sqrt{\frac{g \cdot M}{2 \cdot \pi \cdot R \cdot T}} \cdot \pi \cdot r^3 \cdot \frac{dp}{dx} \qquad \text{(Knudsen's law)}$$

$$q_S = -\frac{\pi \cdot r^4}{8 \cdot \eta} \cdot \frac{dp}{dx} \qquad \text{(Hagen-Poiseuille's law)}$$

Assume that the transported gas is water vapour at 20°C and capillary radius is chosen for a range of values from 10^{-3} - 10^{-8} m. Gradient is taken as 1 N/m^2 and m flow path in all cases. Results of these calculations are shown in Table 3.

Table 3. Flow of water vapour (g/s) through a uniform cylindrical capillary as a result of a gradient of 1 N/m^2 and m path. Theoretical calculations are made using formulae given in the text.

Capillary radius (r, m)	Convection (q_S, g/s)	Effusion (q_E, g/s)	Diffusion (q_D, g/s)	Remarks
10^{-3}	$0{,}408 \cdot 10^{-1}$		$0{,}569 \cdot 10^{-12}$	
10^{-4}	$0{,}408 \cdot 10^{-5}$		$0{,}569 \cdot 10^{-14}$	
10^{-5}	$0{,}408 \cdot 10^{-9}$		$0{,}569 \cdot 10^{-16}$	Convection and diffusion in real systems
10^{-6}	$0{,}408 \cdot 10^{-13}$		$0{,}569 \cdot 10^{-18}$	
- - - radius corresponding to wilting point - - -				
10^{-7}	$0{,}408 \cdot 10^{-17}$	$0{,}909 \cdot 10^{-20}$	$0{,}569 \cdot 10^{-20}$	- - - - - - - - -
10^{-8}		$0{,}909 \cdot 10^{-23}$		Effusion in real systems
10^{-9}		$0{,}909 \cdot 10^{-26}$		

In theory, flow by convection is the greatest at any of the capillary radius values chosen in Table 3. The model takes no account of the fact that flow becomes turbulent beyond a certain radius.

Flow by effusion is not significant in soils since it occures at pore sizes less than 0.2 μ diameter. Such pores are found below the wilting point of soils and thus remain water-filled except near the soil surface, where they are air-dried.

2.4.4.2 Transport through a straight cylindrical capillary with varying diameter

A system where pores do not have a constant diameter is nearer to that which occurs in reality. Shape of such capillaries in a model is shown in Fig. 10. Mean cross-sectional area of a capillary is A_m.

It can be shown that both diffusion and convection through such a system is slower than that through a straight, cylindrical capillary with constant diameter, both with cross-sectional area = A_m (see Krischer, 1963). Theoretically, transport by diffusion through capillaries with varying diameter sinks to a fraction of that through a straight capillary with constant diameter according to the expression:

$$q/q_{A_m} = \frac{\left(1 + \frac{l_2}{l_1}\right)^2}{\left(1 + \frac{A_1}{A_2}\frac{l_2}{l_1}\right) \cdot \left(1 + \frac{A_2}{A_1}\frac{l_2}{l_1}\right)}$$

and with mass flow

$$q/q_{A_m} = \frac{\left(1 + \frac{l_2}{l_1}\right)^3}{\left(1 + \frac{A_1^2}{A_2^2}\frac{l_2}{l_1}\right) \cdot \left(1 + \frac{A_2}{A_1}\frac{l_2}{l_1}\right)^2}$$

These functions are represented in Fig. 10 for the case of $l_1 = l_2$ and $A_1 > A_2$. It can be seen from the curves obtained that the effects of changes in diameter are relatively lower for diffusion and higher for mass flow. Flow is rapidly impeded if A_1/A_2 deviates from a value of 1.

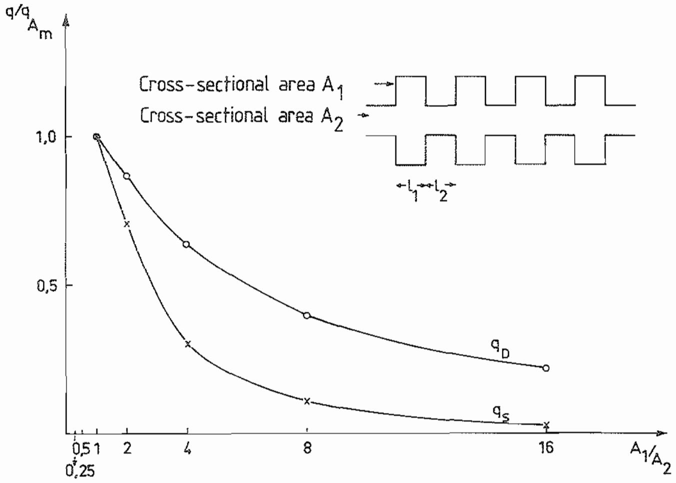


Figure 10. Theoretical flow (q) through a straight, cylindrical capillary of varying diameter, with maximum and minimum cross-sectional area A_1 and A_2 resp. and averaging A_m expressed as ratio of q to q_{Am}, flow through a straight, cylindrical capillary with cross sectional area A_m, as a function of the ratio A_1/A_2 q_D = diffusive flow, q_S = convective flow.

2.4.4.3 Conclusions

From the previous discussion it can be seen that the diffusion process is relatively effective in exploiting a particular pore system regardless of its geometry. Geometrical irregularities in the pore system have little effect on diffusive flow but convective flow depends on the presence of large, regular pores to be fully effective. This explains the findings on model profiles constructed with beads (Fig. 11 a). The finer the pores, the less opportunity there is for convection to occur.

During the course of a year, changes take place in structure of a naturally-occurring soil. Fig. 11 b shows how the structure of the upper part of a heavy clay profile in eastern Sweden can vary with the seasons. The change from coarse, blocky structure after winter ploughing to a denser structure during the summer means that the possible influence of the different air transport mechanisms can vary during the year. In winter, the wind can act directly on the bare, irregular soil surface, increasing convective flow. During the growing season, most gas exchange must take place by diffusion since convection is inhibited by the crop cover which cancels the effects of temperature and wind variations and prevents creation of total pressure gradients.

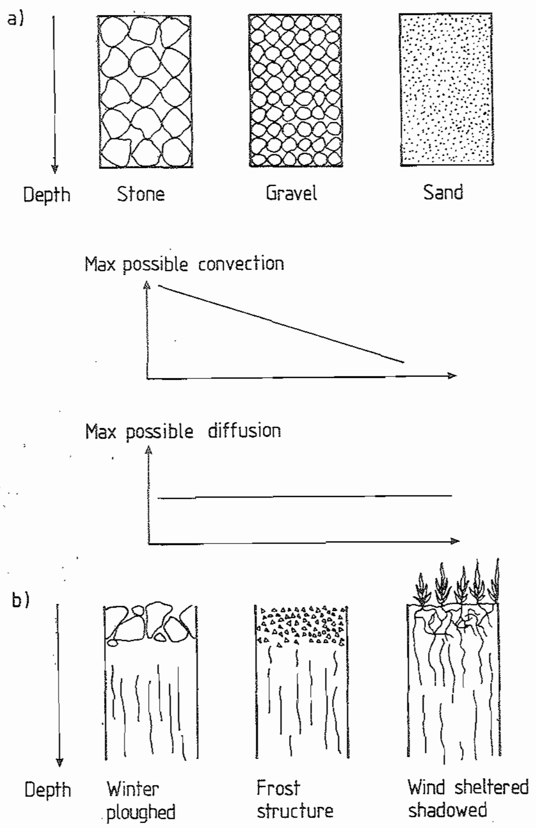


Figure 11. Principles of maximum diffusion and convection in a) some model systems and b) some topsoils of natural profiles.

2.5 Water vapour and humidity

Soil air differs from atmospheric in its high degree of humidity (100 % or just a little lower). The laws of gas movement which have been discussed earlier in this report apply also to water vapour in the soil air. Replacement of water vapour is complicated, however, by the fact that it is in vapour pressure equilibrium with the soil water phase, i.e. a water molecule may oscillate between the soil air and soil water phases.

2.5.1 Condensation

A liquid which is in equilibrium with its own vapour has a zone of saturated vapour pressure over its surface. Saturation vapour pressure is the maximum partial pressure of the vapour at a particular temperature. If partial pressure exceeds saturation vapour pressure (by for instance temperature fall), water vapour condenses (dewpoint).

Saturation vapour pressure - temperature values are available from physical or mathematical tables.

Vapour pressure over the surface of water in a capillary is lower than that over free water. Curvature of the meniscus in a capillary causes a pressure deficit in relation to atmospheric pressure. This pressure difference is counteracted by hydrostatic pressure of the liquid in the capillary after capillary rise. The greater the initial pressure difference (i.e. the higher the capillary rise), the more energy required to evaporate a molecule from the surface of the liquid inside the capillary hence the lower vapour pressure over its surface (Johansson, 1944). Lower vapour pressure over the meniscus means that saturation vapour pressure decreases and water vapour condenses in the capillary even if the air outside is not at saturation pressure. This process is responsible for hygroscopicity of materials.

The condensation phenomenon means that transport of water vapour through a porous medium cannot be considered as simply gas transport. Capillary flow of the liquid phase becomes a very important part of the transport process. Thus, in addition to diffusion and convection of gases, capillary forces must also be included.

2.5.2 Moisture movement in capillary models

The basic principles of water vapour movement can be determined by theoretical studies on moisture movement in simple capillary systems (Johansson, 1948).

The most simple of all capillary models is a straight, cylindrical tube. Consider such a tube connecting two chambers which have a water vapour pressure difference ΔP. In the more humid of these chambers, vapour pressure is above saturation level and in the less humid, it is under saturation level. Conditions in the tube are shown in fig. 12 and the following equations apply:

for the liquid phase $q = k_C \cdot \pi \cdot r^2 \cdot \frac{\Delta P_1}{l_1}$ (Darcy's law)

for the gas phase $q = k_D \cdot \pi \cdot r^2 \cdot \frac{\Delta P_2}{l_2}$ (Fick's law)

for the entire tube $q = k \cdot \pi \cdot r^2 \cdot \frac{\Delta P}{l}$ and $\Delta P = \Delta P_1 + \Delta P_2$

These equations can be summarized in a single expression:

$$q = \pi \cdot r^2 \cdot K_D \frac{\Delta P}{(l - l_1) + K_D/K_C \cdot l_1} \qquad (1)$$

Symbols as in Fig 12.

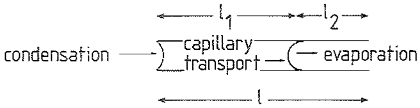


Figure 12. Vapour transport through partially water-filled capillaries (After Johansson, 1948).

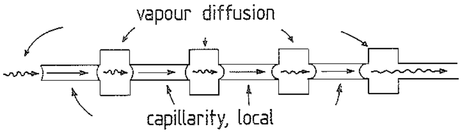


Figure 13. Local capillarity. (After Johansson, 1948).

Equation (1) shows how flow through a tube is altered by changes in its degree of saturation. It can be shown that transport through the tube is independent of whether the tube is completely saturated with water when $r = 0.05\ \mu$. When $r < 0.05\ \mu$ saturation of the tube inhibits transport and when $r > 0.05\ \mu$, saturation increases transport. This means that pore size distribution of a system determines hydraulic conductivity at a particular degree of saturation, i.e. the coefficient for unsaturated hydraulic conductivity.

When the largest pore in a system is less than 0,1 μ, hydraulic conductivity decreases with increasing degree of saturation. When the smallest pore is greater than 0,1 μ, hydraulic conductivity increases with degree of saturation. When the smallest pore is less than 0,1 μ and the largest pore is greater than 0,1 μ, hydraulic conductivity first decreases, then increases as degree of saturation goes from 0 to 100 %.

The situation in a straight, cylindrical capillary is complicated if its diameter is not constant. If wider parts of the capillary are air-filled and if vapour pressure at one end of the capillary exceeds saturation vapour pressure, water is drawn through necks in the capillary (Fig. 13) by capillary forces. Rate of flow through the capillary is determined by its degree of saturation and by diameter of the necks (minimum diameter of the tube). This type of flow is known as local capillarity. Another type of capillary force, namely moisture equalizing capillarity, can occur in a system of tightly packed beads. During wetting, initial condensation of water vapour occurs in the form of ring-like lamellae round contact points on the beads. Water in these lamellae is drawn from one side to the other by local capillarity. When lamellae are so large that they blend together, capillary forces arise causing moisture to be equally distributed throughout the system. Transmission of moisture through the system is dependent on its degree of saturation and on bead size distribution. The latter determines pore size distribution which, as mentioned previously, is one of the factors determining hydraulic conductivity at a particular degree of saturation.

In a system of beads, convective flow of gases may also become significant and may under certain circumstances cancel out the effects of both capillary forces and vapour diffusion.

3. EXPERIMENTAL INVESTIGATIONS ON GAS TRANSPORT THROUGH SOIL FRACTIONS AND NATURAL SOILS

Studies on gas transport in soil can be carried out along two main lines. The first is concerned with moisture movement, especially in the surface layers of a soil (i.e. evaporation effects). The second is concerned with 'pure' gas movement, applied also to such problems as root oxygen supply. In order to consider complicated moisture movements it is first necessary to understand the simpler process of gas movement. The experiments which were carried out in this investigation were studies on gas and vapour diffusion and convection through soil samples, without the complications of condensation or capillarity. Thus the conclusions which could be drawn dealt mainly with oxygen supply problems. These may, however, provide a basis for future studies of vapour movement.

3.1 Diffusion investigations

Kinetic theory of gases was developed during the latter part of the 19th century by research workers such as Maxwell (1868) and Stefan (1871, 1890). This theory was soon being applied to the question of air exchange in soil (Buckingham, 1904). The work of Romell (1922) provided a model for the principles of gas exchange which is still valid. Diffusion was considered to be the most important process in this model. Much research has been carried out on diffusion through soils. Methods used and results obtained from such research have varied greatly. As an introduction to experiments carried out in this investigation, a review of the more important methods and results from the literature are described below.

3.1.1 Review of methods and results from the literature

Penman (1940) published what has become a classic experiment on diffusion through dry and moist soils. Vapour from a volatile liquid (CS_2) was allowed to diffuse through a soil sample. Mass of vapour which diffused was determined by repeated weighings of the CS_2 container and the soil sample to which it is connected. Vapour pressure over the surface of the CS_2 liquid was calculated from vapour pressure/temperature tables. Since dimensions of the sample and of the surface area of the evaporating liquid were known, the coefficient for diffusion through the sample could be calculated.

Penman (1940) found the following relationship between the diffusion coefficient of bulk air, D_o, and the diffusion coefficient of the sample, D:

$$\frac{D}{D_o} = 0.66 \cdot \frac{f_a}{100} \qquad (1)$$

where D/D_o is relative diffusion (see section 2.4.1) and f_a is the air content of the soil in %.

Shortly after Penman's results were published, they were verified by Hagan (1941), whose results agreed with Penman's in all but some minor details.

Both Penman and Hagan carried out their experiments on disturbed soil samples. Blake & Page (1948) investigated gas diffusion under field conditions in soils, comparing diffusion from a CS_2 container in soil with that from a CS_2 container in free air. Their results should be considered with some caution from a quantitative point of view since their results are so widely scattered that the regression line calculated to fit the data has a high degree of error. However, the results make an important qualitative statement on properties of the soil system; that diffusion in soil varies greatly between points and between sites even if the air content is identical. This means that expression relating diffusion to air-filled porosity or pore area (e.g. Penman's equation) must be regarded as producing average values with a high error. Blake & Page (1948) were the first to indicate that diffusion through soil cannot be described by a simple first order equation relating porosity to diffusion. de Vries (1950) carried out some theoretical analyses of diffusion through soil. He calculated relative diffusion as a function of air content and particle shape of a porous material, using Burger's formula for electrical conductance of granular material. de Vries confirmed the main findings of Blake & Page (1948) for he found that shape of primary particles influences diffusion rate, i.e. that values should vary between points.

All the investigations described above deal with steady state diffusion. Taylor (1949) introduced a transient state method of diffusion analysis. Taylor's method involved registration of concentration changes in a gas mixture in a chamber from which diffusion took place via gas-transporting pores of a soil sample to the atmosphere. The chamber was initially filled with N_2 gas and changing composition of the gas with time was determined in gas analysis apparatus. Taylor's diffusion coefficient, D, was calculated using the formula:

$$D = \frac{\lambda^2}{t} \cdot \ln \frac{C_o}{C_o - C} \qquad (2)$$

where λ = material constant,

C_o = O_2 concentration in the atmosphere,

C = O_2 concentration in the chamber at time t after the starting point.

Taylor described conductivity properties of a system to diffusion with the constant $1/\lambda^2$, where λ is equivalent diffusion path in the system. $1/\lambda^2$ was considered to be of the same order of size as D/D_o in Penman's equation. Taylor found $1/\lambda^2$ to have a value of 0.668 $f_a/100$ for an Ontario loam but that this value is not valid for all soils. The constant $1/\lambda^2$ is composite in nature and derivations show it to include some apparatus constants which make it difficult to analyse. For these reasons, it was not adopted in subsequent research.

van Bavel (1952) and Flegg (1953) refined the methods first described by Penman (1940) but did not change any of the basic principles. van Bavel changed Penman's expression for relative diffusion from $0.66 \cdot f_a/100$ to $0.68 \cdot f_a/100$ although his results were somewhat scattered. Flegg's experiments verified Penman's findings.

In general, Penman's result held good till the end of the 1950's. According to them, the principle equation for relative diffusion was of the form:

$$D/D_o = a \cdot \frac{f_a}{100} + b \qquad (3)$$

where constant 'a' was 0.5 - 0.8 and constant 'b' was usually 0.

Trends in several other investigations indicated that the situation was not as simple as in equation 3. Buckingham (1904) had proposed a quadratic equation and Blake & Page (1948) and Taylor (1949) found indications that 'a' and 'b' were not constants but functions of soil properties. de Vries (1950) showed that this, in theory, should be the case.

In 1960, the English scientist J.A. Currie published results from an investigation of gas diffusion through porous material. He investigated dry and moist, single-grained and aggregated material, using a transient state method similar to Taylor's and H_2 as the diffusing gas. Currie proposed that for transient state diffusion, the following relationship applied at point x in the sample:

$$dC/dt = \frac{D}{f_a/100} \cdot \frac{d^2C}{dx^2} \qquad (4)$$

This equation has analogous solutions in thermodynamics - see, for example, Carslaw & Jaeger (1959).

Solution of equation (4) is complicated and the diffusion coefficient calculated from it is larger than that calculated by Taylor (1949). This is because equation (4) takes into account concentration changes within the soil sample.

Currie in his discussion relates his results to de Vries' (1950) theoretical model from Burger's formula and to Bruggeman's formula for electrical conductance of a granular system. For diffusion these formulae are written:

$$D/D_o = \frac{f_a/100}{1 + (k-1)\ (1-n/100)} \qquad \text{Burger}$$

$$D/D_o = (f_a/100)^m \qquad \text{Bruggeman}$$

where the constants k and m are particle shape factors. Over 20 different types of dry material were investigated.

Currie concluded that diffusion formulae analogous to Burger's and Bruggeman's equations did not provide a completely true picture of the system. However, the equations fitted well for some materials when suitable values of the constants k and m were chosen. Currie regarded k and m as parameters of pore geometry.

For moist materials, Currie found two different types of relationship between relative diffusion and porosity. Single-grained materials showed a parabolic curve whereas aggregated material showed a curve with two distinct phases, the first of which was parabolic. The obvious discontinuity between the two phases of the aggregated system occurred at the point on the air content axis when water began to drain from inside aggregates; as long as water drained from inter-aggregate spaces, the system functioned as a single grained soil. Currie concluded that diffusion in single-grained and the first phase of aggregated soils is mainly a function of air content (f_a) and shape factors k, m.

Diffusion in the second phase of aggregated soils is a function of at least five factors e.g. air content, f_a, air content of the aggregates, n_a, external form of aggregates, k, internal form of aggregates, k_a and water content of the soil, f_w. Form factors and m calculated for aggregated material have minimum values at the boundary between phases. These factors are thus an important measure of complexity of the air-filled pore system and Currie (1965) suggested that they be used as parameters of soil structure.

Currie in his investigations (1960, 1965) showed that there is no unique relationship between diffusion and air content. Diffusion through a porous material is a complicated function of airfilled porosity and nature of the pore system.

Bakker & Hidding (1970) carried out a number of experiments on diffusion in naturally occurring Netherlands soils in order to determine the effects of pan formation on O_2 transport through the soil surface. Their experimental method was similar to that used by Currie (1960) and, since naturally occurring soils were used, results obtained were widely scattered.

Rather than performing detailed theoretical analyses of their results, Bakker & Hidding calculated average values and observed general trends in these. For some soils, they also calculated the oxygen supply situation using diffusion theory as a starting point.

Millington & Shearer (1971) employed statistical principles to put forward a theory for diffusion in porous systems. Graphical representations of the function they produced relating diffusion and air content are very similar to the empirical curves produced by Currie.

Calculation using Millington & Shearer's formula requires subdivision of the pore system into interaggregate and aggregate porosity. Such a subdivision is almost impossible without detailed knowledge of the actual soil system, so the formula should perhaps be regarded as a theoretical complement to experimental data rather than a practical method of calculation.

Other theoretical analyses of diffusion equations for porous systems have been carried out by Troeh et al., (1982).

All investigations cited above except Blake & Page (1948) and Bakker & Hidding (1970) were carried out on disturbed soil samples. An essential difference between disturbed soil samples and studies on naturally occurring soils is the spread of results. Disturbed samples, which are sieved and compacted in the laboratory, show a much smaller variation of results than natural soils. The wide range of results from natural soils is a result of the form factors described by Currie (1961) which vary between points in the soil profile. Internal pore geometry is thus irregular and form factors cannot be assigned values which are exactly correct for every point in the system. All values assigned to factors for naturally occurring soil must be regarded as averages of more or less disparate data. This is also true for many other soil physical parameters.

3.1.2 Diffusion experiments by a steady state method

3.1.2.1 Materials and methods

A method similar to that of Penman (1940) was devised for experimental investigation of diffusion. The diffusing gas was carbon disulphide (CS_2), which is evolved from its volatile liquid form at room temperature. An apparatus was constructed so that all CS_2 produced from a quantity of its liquid in a chamber was forced to diffuse through a soil sample. Mass of gas which diffused was determined by weighing.

The apparatus was designed so that a cylinder containing the soil sample could be screwed on to the CS_2 container. When screwed in position, the cylinder came in contact with a sealing O-ring of fluorpolymer rubber (viton rubber). This quality of rubber was necessary to contain the aggressive CS_2 gas. The sample cylinder and CS_2 container were both made from aluminium. The apparatus is pictured in Fig. 14.

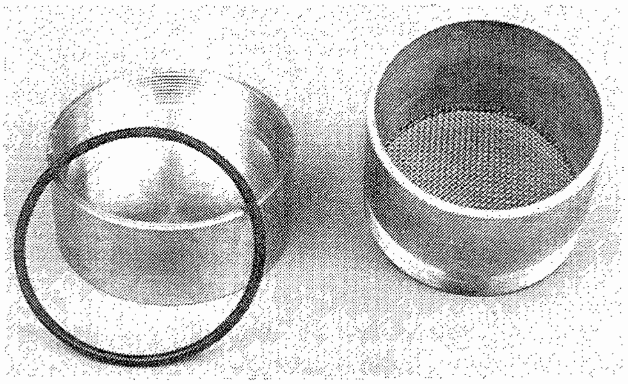


Figure 14. Sample cylinder and CS_2-cup for experiments with steady state diffusion.

The soil sample was held in place in the cylinder by a steel mesh which was fixed in a groove at the cylinder base. To contain material with smaller particles, a closemeshed nylon gauze was laid on top of the steel mesh. The assembled apparatus was placed in a specially adapted cupboard which could accommodate 10 samples simultaneously. A low powered fan inside the cupboard disposed of the poisonous and noxious CS_2 gas.

The sample to be tested was packed into the cylinder by vibration and compaction. During packing, the cylinder was fitted with an extension cylinder, held in place by a rubber ligature. A pressure was applied to the surface of the sample and the cylinder was vibrated. The pressure and extension cylinder were then removed and the upper surface of the sample was planed off with a knife.

30 ml of liquid CS_2 were added to the container; to avoid splashing during manipulation of the apparatus, the CS_2 liquid was

applied on a cotton wool wad with volume 50 mls. Finally, the sample cylinder was screwed on to the CS_2 container and the apparatus was placed in the fume cupboard.

Vapour pressure over CS_2 liquid is a function of temperature. Immediately above the liquid is a zone of saturation vapour pressure, the magnitude of which at a particular temperature is obtained from a standard curve. Since CS_2 vapour pressure at the upper surface of the sample is zero, the total pressure difference between the surface of the liquid and surface of the sample is equivalent to the saturation vapour pressure.

Since the junction between the sample cylinder and CS_2 container is sealed with an O-ring, all CS_2 gas evaporating from the liquid is transported through the soil sample. The situation is initially non-steady state (see Penman, 1940) but after some time, steady state conditions are reached by the system. In actual fact, steady state conditions are not perfect since the level of the CS_2 liquid in the container falls as evaporation proceeds.

In the present investigation, the system was allowed to stand for two hours for steady state conditions to be achieved.

After this, the apparatus was weighed once or twice an hour. The diffusion coefficient was calculated from Fick's formula (see 2.4.1)

$$q = D \cdot \frac{M}{RT} \cdot \frac{P_o}{P} \cdot \left(\frac{T}{T_o}\right)^2 \cdot \frac{A}{l} \cdot \Delta P \cdot t \qquad (1)$$

where q = mass of gas diffusing per unit time,
D = diffusion coefficient at temperature T_o and pressure P_o,
M = molecular weight of the diffusing gas,
T = absolute temperature,
P = air pressure,
T_o = normal temperature (293°K),
P_o = normal pressure (760 mm Hg),
A = cross-sectional area of the sample,
l = length of sample,
ΔP = driving force (pressure gradient),
t = time,
R = universal gas constant.

3.1.2.2 Discussion of method

The situation which arises in this seemingly simple apparatus is, in actual fact, quite complicated, due to disruption of transport from the surface of the liquid to the atmosphere. The steel mesh interferes somewhat with diffusive flow, CS_2 can be adsorbed onto soil particles and mass transport occurs in the system.

Some loss of pressure gradient occurs between the surface of the liquid and the lower surface of the soil sample and a further loss occurs during passage through the steel mesh.

It can be shown from the continuity equiation that the loss of gradient mentioned above increases with relative diffusion rate of the sample. Pressure loss in the sample has been found by calculation to be a function of sample porosity (Table 4). In this calculation, relative diffusion through the sample was obtained using the formula devised by Penman (1940), see 3.1.1.

Table 4. Relationship between pressure loss in a soil sample (% of CS_2 saturation vapour presure) and its porosity.

Porosity, %	Pressure loss in sample, %
30	93
40	91
50	89
60	85

It can be seen that, if the process is regarded as exclusively one of diffusion, percentage of the total partial pressure gradient which is lost in the sample decreases rapidly with increasing porosity. A further complication arises, namely that CS_2 concentration above the upper surface of the sample increases as intensity of diffusive flow increases and the assumption that pressure here is zero becomes increasingly inaccurate. The significance of this is hard to evaluate. In table 4 is one per cent of the total partial pressure gradient supposed to be used above the sample.

Adsorption of CS_2 onto the soil particles was investigated experimentally, by recording changes in weight of the soil sample immediately after completing diffusion test. It was concluded that adsorption onto mineral particles was very low while adsorption onto colloidal humus particles was significant. In the case of moist soils, the low solubility of CS_2 in water meant that adsorption in the soil water phase was not a problem. Thus sorption occurred rapidly in mineral soils but took considerably longer in humus-rich soils.

The relatively long initial period of stabilization (2 hours) meant that sorption had ceased before measurement commenced, even in humus-rich soils.

Vapour evaporates from the surface of the CS_2 liquid under the influence of a partial pressure gradient. Total pressure (P) at the starting point was the same everywhere in the system so a partial pressure gradient must arise in the air between the upper surface of the sample and surface of the liquid. Consequently, diffusion of air must also occur, though in the opposite direction to diffusion of the CS_2. Since the air cannot penetrate the surface of the CS_2 liquid, the diffusive flow of air must be compensated for by mass transport of the total gas mixture in the direction of CS_2 diffusion. As a result, diffusing CS_2 vapour is transported more rapidly than it would be by diffusion alone. The diffusion coefficient calculated by equation (1) is obviously too large in this case.

According to van Bavel (1952) and Krischer (1963), this error can be corrected by introducing into equation (1) the term $P/P-P_D$ (P_D = saturation vapour pressure at the surface of the CS_2 liquid). Equation (1) becomes thus:

$$q = D \cdot \frac{M}{RT} \cdot \frac{P_o}{P} \cdot \left(\frac{T}{T_o}\right)^2 \cdot \frac{A}{l} \cdot P \cdot \ln \frac{P}{P-P_D} \cdot t \qquad (2)$$

3.1.2.3 Experimental results

Experiments were carried out on the soils and soil fractions listed in Table 5. Dry soil fractions (size classed after Atterberg's system) were mixed to produce a series of dry, single-grained soils with porosity between 21.5 and 53.6 %.

The effect of moisture content on diffusion was partly determined on a slightly damp medium sand and partly on a saturated medium sand which was drained at successive tensions using a pressure chamber technique described by Andersson (1971).

The influence of aggregation on diffusion was determined by measurements on clay aggregates and on peat.

Five replicates were used in every case and results calculated from equation (2) and corrected according to Table 4 are shown in Table 6 and Fig. 15. A correlation coefficient of 0.99 was obtained for results from single grained samples.

Table 5. Specification of soil fractions and soil mixtures used in the experiment.

Material	Components	%
Coarse sand (CSa)	0.6 - 2.0	100
Medium sand (MSa)	0.2 - 0.6 mm	100
Coarse silt (CSi)	0.006 - 0.02 mm	100
Mix 1	Fyle sand 1	50
	Fyle sand 2	50
Mix 2	2 - 6 mm	50
	0.2 - 0.6 mm	50
Mix 3	2-4 mm	53.50
	0.075 - 0.125 mm	23.25
	0.006 - 0.06 mm	23.25
Heavy clay (HC)	frost aggregated	100
Peat	Hasselfors peat	100

Table 6. Results from experiments, steady state diffusion.

Material	Air content %	Diffusion coefficient D, cm^2/s	Relative diffusion D/D_0
MSa	45.1	0.0238	0.235
MSa	40.5	0.0218	0.214
MSa	40.6	0.0218	0.214
MSa	42.6	0.0218	0.214
CSa	43.4	0.0226	0.222
CSi	53.6	0.0281	0.275
Mix 1	35.3	0.0173	0.170
Mix 2	28.0	0.0126	0.124
Mix 3	21.5	0.0073	0.072
MSa, moistened	39.0	0.0147	0.144
HC	55.5	0.0253	0.248
Peat	71.7	0.0173	0.170
MSa, h_t = 10 cm w.c.	4.1	0.0	0.0
MSa, h_t = 20 cm w.c.	5.0	0.0015	0.015
MSa, h_t = 40	20.0	0.0031	0.030
MSa, h_t = 80 cm w.c.	38.1	0.0164	0.161
MSa, h_t = 10^7 cm w.c.	40.8	0.0217	0.212

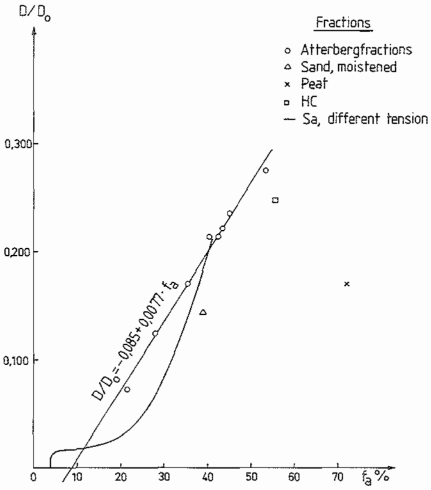


Figure 15. Steady state diffusion: Relative diffusion (D/D_0) as a function of air content of the material (f_a).

3.1.3. Diffusion experiments by a transient state method

When the experiments described above on dry Atterberg fractions were completed, the situation in moist and naturally occurring soils was investigated. Undisturbed soil samples were extracted using cylinders designed at the Department of Agric. Hydrotechnics (407 and 203.5 cm^3), see Andersson (1955). Equipment to measure diffusion rate through undisturbed samples had to be compatible with these cylinders. Problems with ensuring a CS_2 gas proof seal between standard cylinders and the CS_2 container used in the previous experiment could not be resolved by a simple but effective method. For this reason, a new method was developed for this experiment.

3.1.3.1. Materials and method

The method developed was based on the principle used by Taylor (1949), Currie (1960) and Bakker & Hidding (1970). It was possible to develop this technique so that it could be accommodated in standard laboratory equipment used in the Department for moisture regulation in soils (pressure chamber technique, see Andersson, 1971). This apparatus is designed to take 12 samples simultaneously, so it was convenient to carry out investigation of diffusion on this number of samples at a single time. This allowed for two alternative methods, either rapid analysis of one sample at a time or simultaneous analysis of up to 12 samples. The latter method was preferred as more practical since the former, rapid alternative required constant manipulation and observation.

Currie (1960) and Bakker & Hidding (1970) investigated only one sample at a time. Single sample apparatus described in the literature is connected up to gas analysis equipment or to gas electrodes which register changes in the gas chamber.

For economic reasons, it was impossible to obtain high quality gas electrodes for each of the 12 samples in the present method so a gas chromatograph was used instead. Final construction of the apparatus is shown in Figs 16 and 17.

Soil samples, which were excavated in the field or packed in the laboratory, were set on tripods. Then, with a rubber cuff acting as a seal, samples were connected to chambers. Nitrogen gas (N_2) from a bomb was circulated through the chambers. This gas was chosen because it is nontoxic and is hopefully neither produced nor consumed in the sample.

Regardless of the gas chosen to circulate through the chamber, gases from the atmosphere will enter the system and air in the chamber will, according to the laws of diffusion, have identical composition to the atmosphere after an infinite time. While gas inside the chamber diffuses out, air diffuses in.

It was considered advisable to choose one of the component gases of the atmosphere to fill chambers at the start of this experiment. This meant that mainly O_2 and N_2 diffused in opposite directions and at very similar rates, since diffusion rate is inversely proportional to the square root of molecular weight and O lies adjacent to N in the Periodic Table of the Elements.

The similarity of diffusion rates of gases in this experiment prevented mass transport which may occur because of thermal motion (self-diffusion) in two gases diffusing in opposite directions at very different rates. In narrow pore systems, this process may result in energy losses. However, Currie (1972) estimates such losses to be negligible. In any case, the process is avoided in the present experiment by choice of N_2 and O_2 as the main components in the diffusing gases. Although N_2 was chosen as the gas to circulate in the chambers in this experiment, O_2 would work just as well.

Rapid analysis of the gas mixture in the chambers was carried out by gas chromatograph (Perkin-Elmer Fraktometer T-116). Only nitrogen and oxygen concentrations were measured. The gas sample was extracted from the chamber with a gas syringe inserted through a rubber membrane similar to the injection membrane of the GC. The experiment was carried out thus: A chosen moisture tension was applied to the sample by means of a pressure chamber. When equilibrium was reached between applied tension and capillary tension in the pores (water removal ceases), the samples were weighed and placed in the diffusion apparatus. Nitrogen gas from the bomb was circulated through the samples until their N_2-content was around 99%, a process which took a few minutes in most cases. The apparatus was then allowed to stabilize for a period of at most 10 min (or less in the case with high expected coefficient of diffusion). A gas sample was then extracted from chamber 1 and analysed. Analysis time was 1.5 minutes per sample; after 2 minutes, a sample was taken from chamber 2, after 4 min. from chamber 3 etc. Thus analysis of all 12 samples took 24 minutes. After an interval of 2 1/2 hours, sampling and analysis was repeated.

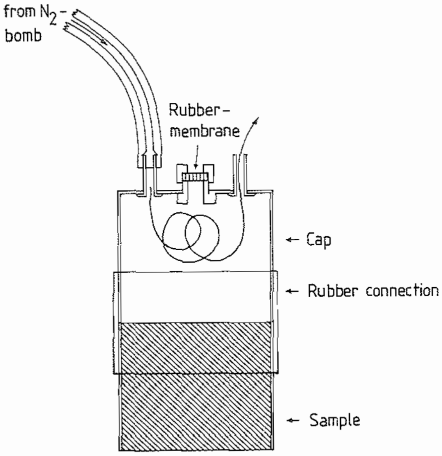


Figure 16. Schematic representation of apparatus used in transient state diffusion experiments.

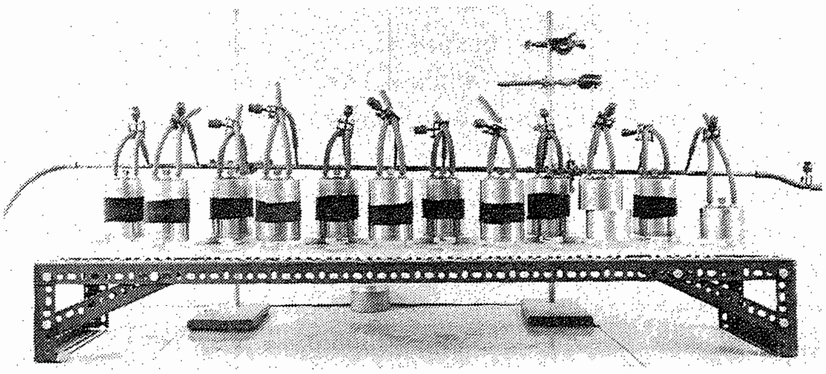


Figure 17. Multi-sample apparatus used in transient state diffusion experiments. Unit No. 1 (numbered from right to left) is shown minus sample and cuff, unit No. 3 without cuff.

Results of gas analyses were produced by the gas chromatograph in the form of a chromatogram showing peaks of nitrogen and oxygen concentration. The exact amount of each gas in the chamber could then be calculated by comparing the peaks obtained to calibration curves. It was assumed here that O_2 and N_2 were the only gases to diffuse, i.e. the effects of other components of the atmosphere such as argon, krypton and carbon dioxide were ignored.

Fig. 18 shows a typical chromatogram.

Using values obtained for composition of the gas mixture in chambers at times 1 and 2, the diffusion coefficient of samples can be calculated.

From Fick's first law:

$$dq = -D \cdot A_s \cdot \frac{C}{l_s} \cdot dt$$

where dq = mass of gas transported
D = diffusion coefficient
A_s = cross-sectional area of sample,
C = concentration difference between top and bottom of the sample,
l_s = length of sample,
t = time.

Furthermore,

$$dq = -A_c \cdot l_c \cdot \frac{dC}{dt} \cdot dt$$

where A_c = cross-sectional area of chamber,
l_c = length of chamber,
dC = change in concentration of gases in the chamber at any instant

In the apparatus, $A_s = A_c$, therefore:

$$D \cdot \frac{C}{l_s} \cdot dt = l_c \cdot \frac{dC}{dt} \cdot dt$$

$$D \cdot dt = l_c \cdot l_s \cdot \frac{dC}{C} \cdot dt \quad ;$$

$$D = \frac{l_c \cdot l_s}{t} \cdot 2.303 \cdot \log \frac{C_1}{C_2} \qquad (1)$$

C_1 and C_2, concentration differences between top and bottom of the sample are easily calculated when N_2 concentrations in the chamber are known and concentration of N_2 in the external air is 78 %.

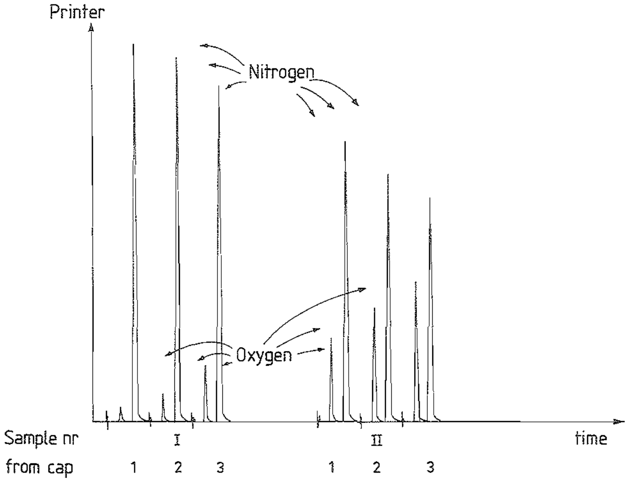


Figure 18. Diagram of a typical chromatogram.

Equation (1) is used by Taylor (1949). However, this equation gives too low a value for the diffusion coefficient, since the relationship between N_2 concentration and the length coordinate of the sample is assumed to be a straight line in derivation of the equation. This is not, in fact, the case, since N_2 concentration inside the chamber changes continually and thus mass of gas leaving the chamber at any instant and concentration gradient in the sample also change. Equation (1) must be corrected for this to give an accurate estimation of the diffusion coefficient.

Consider a layer in the sample with thickness dx. The diffusive flow into this layer at any instant and per unit area is:

$$dq_{in} = -D \cdot \frac{dC}{dx} \cdot dt$$

and the corresponding relationship for flow out of the layer is:

$$dq_{out} = -D \cdot \frac{d}{dx} \cdot (C + \frac{dC}{dx} dx) \cdot dt$$

Changes in the gas content of the layer are described by:

$$dq_m = \frac{f_a}{100} \cdot \frac{dC}{dt} \cdot dx \cdot dt$$

According to the continuity equation:

$$dq_{in} = dq_{out} + dq_m$$

Thus

$$-D \cdot \frac{dC}{dx} \cdot dt \cdot \frac{\frac{dC}{100}}{f_a} \cdot dx - D \cdot \frac{dC}{dx} \cdot dt - D \cdot \frac{d^2C}{dx} \cdot dx \cdot dt$$

$$\frac{dC}{dt} = \frac{D}{\frac{f_a}{100}} \cdot \frac{d^2C}{dx^2} \qquad (2)$$

This very important equation, often referred to as Fick's second law, describes transient state diffusion. The equation is mathematically analogous to Fourier's law of heat conduction.

Several solutions of Fourier's law have been proposed (see Carslaw & Jaeger, 1959). Currie (1960) and Bakker & Hidding (1979) chosed the solution:

$$C = \sum_{n=1}^{\infty} \frac{2\,hC_0 \cdot e^{-D \cdot \alpha_n^2 \cdot t/\frac{f_a}{100}} \qquad \sin \alpha_n \cdot x}{\sin \alpha_n \cdot l_s \cdot \left[l_s \cdot (\alpha_n^2 + h^2) + h\right]} \qquad (3)$$

where $h = \frac{f_a}{100} \cdot \frac{1}{l_c}$ and

α_n with $n = 1, 2, 3 \ldots \infty$ are the positive roots to

$$\alpha \cdot \tan \alpha \cdot l_s = h \qquad (4)$$

for $x = l_s$ and $t >> 0$, this simplifies, since the second and subsequent terms in the series are negligible, to

$$\frac{C}{C_o} = \frac{2 \cdot h \cdot e^{-D \cdot \alpha_1^2 \cdot t / \frac{f_a}{100}}}{l_s \cdot [\,(\alpha_1^2 + h^2) + h\,]} \qquad (5)$$

This approximates to:

$$D = \frac{f_a}{100 \cdot \alpha_1^2 \cdot t} \cdot 2.303 \cdot \log \frac{C_1}{C_2} \qquad (6)$$

Comparing a coefficient calculated from equation (6) with one from equation (1), it is seen that neglecting the summation term in equation (1) gives too low a value for D. The main difference between the two processes represented by these equations is demonstrated in Figs. 19 and 20, where patterns of flow are shown in an apparatus designed by Andersson (1971).

In practice, it is simplest to calculate D from equation (1) and to then correct it according to:

$$\frac{D(1)}{D(6)} = \frac{l_c \cdot l_s}{t} \cdot 2.303 \cdot \log \frac{C_1}{C_2} \cdot \frac{100 \cdot \alpha_1^2 \cdot t}{f_a \cdot 2.303 \cdot \log {}^{C_1}/C_2}$$

$$D(6) = D(1) \; \frac{f_a}{100 \cdot \alpha_1^2 \cdot l_c \cdot l_s} \qquad (7)$$

Numerical values for α_1 are obtained from a table for solution of equation (4) (see Carslaw & Jaeger, 1959). The correction factor can then be calculated for various combinations of f_a, l_c and l_s.

Some examples of the correction factor are shown in Table 7. The general equation for the correction factor can be calculated from these (in accordance to Bakker & Hidding (1970)):

$$y = (1 + 0.34 \cdot \frac{f_a}{100} \cdot \frac{l_s}{l_c})$$

and equation (7) can thus be written:

$$D(6) = D(1) \cdot (1 + 0.34 \frac{f_a}{100} \cdot \frac{l_s}{l_c})$$

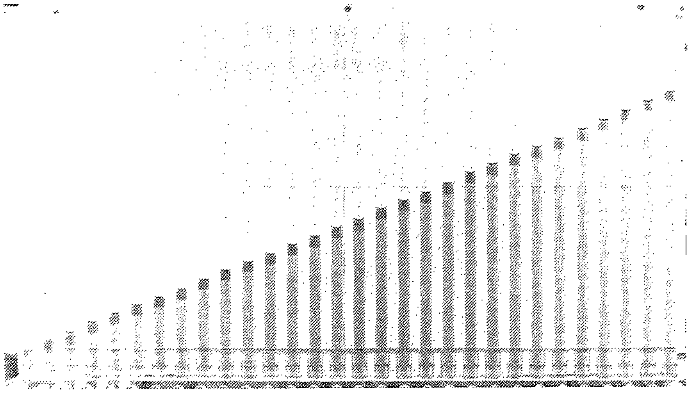


Figure 19. Representation of N_2 - concentration gradient in a sample, as described by equation (1). The apparatus was designed by Andersson (1971) to simulate groundwater flow, heat flow and diffusion. The gas chamber is assumed to be on the right of the picture and N_2 diffuses to the left.

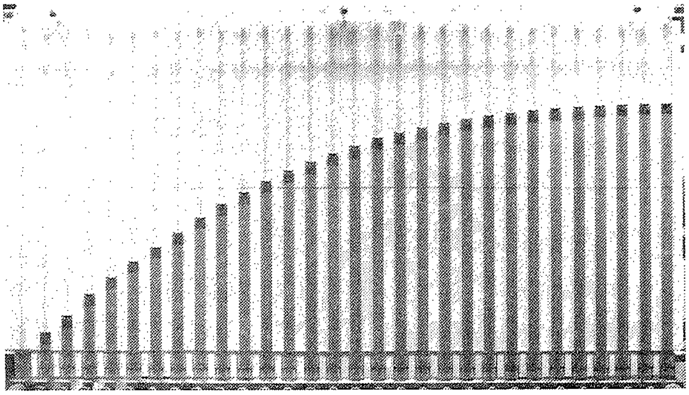


Figure 20. Representation of N_2 - concentration according to equations (2) and (6). Apparatus and assumptions as in Fig. 19.

Table 7. Correction factors for transient state diffusion at different combinations of sample air content (f_a), sample length (l_s) and chamber length (l_c). α from Carslaw & Jaeger (1959), see text.

$\frac{f_a}{100}$	l_s	l_c	α	corr.
0.01	5	5	0.01996	1.004
0.02	5	5	0.02820	1.006
0.04	5	5	0.03974	1.013
0.06	5	5	0.04850	1.020
0.08	5	5	0.05582	1.027
0.10	5	5	0.06222	1.033
0.20	5	5	0.08656	1.068
0.30	5	5	0.10436	1.102
0.40	5	5	0.11864	1.137
0.50	5	5	0.13066	1.172
0.01	10	5	0.01410	1.006
0.02	10	5	0.01987	1.013
0.04	10	5	0.02791	1.027
0.05	10	5	0.03111	1.033
0.10	10	5	0.04328	1.068
0.20	10	5	0.05932	1.137
0.30	10	5	0.07051	1.207
0.40	10	5	0.07910	1.279
0.50	10	5	0.08603	1.351

3.1.3.2 Discussion of method

A possible source of error in these experiments were concentration gradient in the air outside the sample or between the sample surface and the point of sampling in the gas chamber.

Trials with weak circulation of the external air and with sampling at different levels in the chamber showed that results were not affected by these factors. They were thus neglected in the experiment.

When biological activity is at its highest in a soil sample, oxygen consumption can be up to 2 x 10^{-6} g/cm^3 and hour (equivalent to 10 g/day and m^2 in a 20 cm deep topsoil). This can mean that values of D obtained in experiments are too low or to high, depending on choise of diffusing gases and wether one or both of them are consumed or produced in the sample. In this case however this influence should be small. The studied gas is N_2, which hopefully is not involved in rapid processes of nitrification or denitrification in the samples. The O_2 which possibly is consumed should be replaced with CO_2, assuming aerobic conditions, and so keeping the total gas-pressure constant.

In order to estimate the magnitude of the possible error, if biological activity in the sample affects the composition of the gas in the diffusion chamber, this is studied in table 8.

Table 8 compares $\log C_1/C_2 + 0.4$ with $\log C_1/C_2$. The ratio of these logarithms is at a maximum (error is minimum) when $C_2 = 8$ %. For $13 > C_2 > 3$ %, the error is under 7 %, but for $13 < C_2 < 3$ it increases

Table 8. Effects of oxygen consumption in samples. For explanation, see text. C_1 is assumed to be 20 % in all cases.

C_2 %	$C_2+0.4$ %	$\frac{C_1}{C_2}$	$\frac{C_1}{C_2+0.4}$	$\log \frac{C_1}{C_2}$	$\log \frac{C_1}{C_2+0.4}$	$\frac{\log \frac{C_1}{C_2+0.4}}{\log \frac{C_1}{C_2}}$
19	19.4	1.053	1.031	0.0224	0.0133	0.59
17	17.4	1.176	1.149	0.0704	0.0603	0.86
15	15.4	1.333	1.299	0.1248	0.1136	0.91
13	13.4	1.538	1.493	0.1869	0.1741	0.93
11	11.4	1.818	1.754	0.2596	0.2440	0.94
9	9.4	2.222	2.128	0.3468	0.3279	0.95
7	7.4	2.857	2.703	0.4559	0.4319	0.95
5	5.4	4.000	3.704	0.6021	0.5687	0.94
3	3.4	6.667	5.882	0.8239	0.7695	0.93
1	1.4	20.00	14.28	1.3010	1.1547	0.89

rapidly. When D is small, the error is large but insignificant in absolute terms. When D is large, the error is large and significant in absolute terms. Experimentally determined values of D are always smaller than its real value in the cases mentioned above.

Error in analyses of gas samples was estimated to at most 0.5 vol. %. The significance of this is shown in Table 9, where C_1 is assumed to be 20 ± 0.5 % and C_2 is set at 1, 5, 10, 15 and 19 ± 0.5 %. The combinations examined show that large and small ratios give the greatest error.

In certain cases, when the ratio of C_1 to C_2 is very large or very small, the error can be 100 %.

Table 9. Effect of an analytical error of 0,5 % by volume in oxygen concentration. See text for explanation.

C_1/C_2				$\log \frac{C_1}{C_2}$	$\frac{\log \frac{C_1 \pm 0.5}{C_2 \pm 0.5}}{\log \frac{C_1}{C_2}}$
$C_1 =$ 19,5 %	$C_1 =$ 20,0 %	$C_1 =$ 20,5 %	C_2 %		
1.000			19.5	0.0000	0
	1.053		19.0	0.0224	-
		1.108	18.5	0.0445	1.999
1.258			15.5	0.0997	0.799
	1.333		15	0.1248	-
		1.414	14.5	0.1504	1.205
1.857			10.5	0.2690	0.894
	2.000		10	0.3010	-
		2.158	9.5	0.3340	1.110
3.545			5.5	0.5496	0.913
	4.000		5	0.6021	-
		4.560	4.5	0.6590	1.095
13.00			1.5	1.1139	0.856
	20.00		1.0	1.3010	-
		41.00	0.5	1.6128	1.240

The total uncertainty in this experiment due to error in method, apparatus and analysis was thought to be kept to 10 % by careful adjustment of the time inverval between the first and second runs of sampling.

In most cases, 9000 seconds were allowed to elapse between the first and second runs of sampling and this interval was sufficient to minimize experimental error satisfactorily. In a few cases, usually when relative diffusion was large, the ratio of C_1 to C_2 became too large. The error was then great in relative and absolute terms. Such cases required a shorter time interval between samplings to reduce the quotient C_1/C_2. Even when C_1/C_2 was approximately 1 (i.e. when relative diffusion was low), the error was large but its absolute value was small so its effects could be ignored.

Furthermore, variations in the materials investigated, especially natural soils, were considerably greater than the experimental error. Replicates often differed by more than 100 % while the order of magnitude of the error was usually less than 10 %. This is also in agreement with the discussion of Currie (1960).

The influence of temperature and barometric pressure of the atmosphere were corrected for as in the previous method with steady state diffusion. In the present experiment, however, the exponent 'n' in Kennard's equation (see 2.4.1) is n = 1.75 for gases rather than n = 2 which was valid for vapours.

3.1.3.3 Experimental results

Experiments were carried out on a number of dry Atterberg fractions (see Table 5) and on a number of naturally occurring soils which were drained stepwise at a number of matric tensions from 0.15 to 5 m water column.

Natural soils were obtained from different sites and from different cultivation backgrounds. Triplicates were tested in all cases. Mechanical composition of natural soils used in the experiment is shown in Table 10 and differences in their background are described in Table 11.

Results from experiments on transient state diffusion are shown in Tables 12 - 15 and Figs. 21 - 28.

Table 10. Mechanical composition of natural soils used in transient state experiments.

Site		Level cm	Clay d 0.002 mm %	F. silt 0.002-0.006 mm %	C. silt 0.006-0.02 mm %	VF. sand 0.02-0.06 mm %	F. sand 0.06-0.2 mm %	M. sand 0.2-0.6 mm %	C. sand 0.6-2.0 mm %	Loss on ignition
Ultuna	Site I	10-20	35	12	16	13	16	2	2	4
Ultuna	Site I	20-30	38	12	16	12	15	2	1	4
Ultuna	Site I	30-40	38	14	13	13	15	2	1	4
Ultuna	Site VI	10-20	13	14	5	5	39	30	1	3
Kasby		0-5	40	15	17	17	6	2	1	2
Uddeholm		10-20	20	29	34	7	3	1	1	5
Uddeholm		20-30	17	35	33	8	3	1	1	1
Grävsta		10-20	29	13	16	18	18	2	0	4
Grävsta		20-30	28	14	16	21	15	2	0	4

Table 11. Review of experimental sites used in diffusion and air permeability studies on natural soils.

Site	Level (cm)	Previous treatment
Ultuna, site I	10-15	
Ultuna, site I	25-30	
Ultuna, site I	35-40	
Ultuna, site VI	10-15	
Kasby, H	0-5	Smeared surface with subsequent pan formation
Uddeholm, 1a, 1b and 1c	10-15	1a: permanent grassland
	15-20	1b: cultivated for barley with spade
	20-25	1c: barley growing, very heavy compaction
	25-30	
Grävsta, 1a, 1b and 1c	10-15	see Uddeholm
	15-20	
	20-25	
	25-30	

Table 12. Results from transient state diffusion experiments. Atterberg fractions and mixtures of these (see Table 5).

Material	Air content f_a %	Relative diffusion D/D_o
MSa	41.1	0.212
"	41.1	0.209
Mix 1	35.3	0.162
"	35.4	0.167
Mix 2	20.2	0.075
"	21.2	0.085
MSa	41.1	0.201
Mix 3	29.7	0.133
Mix 1	35.4	0.157
Mix 2	20.2	0.078
"	21.1	0.070
MSa	41.1	0.199
Mix 3	29.7	0.141
Mix 1	35.4	0.168
Mix 2	20.2	0.078
"	21.2	0.078
MSa	41.1	0.212
Mix 3	29.7	0.142
Mix 1	35.4	0.178
Mix 2	20.2	0.077
"	21.2	0.083

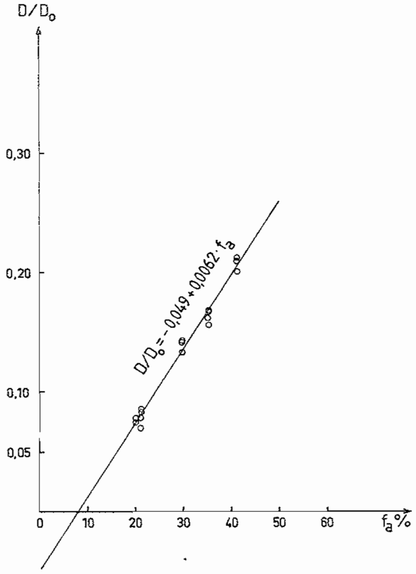


Figure 21. Relative diffusion (D/D_0) as a function of air-filled porosity (f_a) in transient state diffusion experiments. Dry Atterberg fractions.

Table 13. Results from transient state diffusion experiments. Soils from the Ultuna area (see Tables 10 & 11).
D/D_o = relative diffusion f_a = air-filled porosity, %

Sample	Matric tension applied, cm w.c.									
	15		30		50		100		150	
	D/D_o	f_a	D/D_o	f_a	D/D_o	f_a	D/D_o	f_a	D/D_o	f_a
U1, 10-15	0.0027	9.2	0.0036	9.9	0.0048	10.9	0.0124	11.7		
U1, 25-30	0.0017	5.0	0.0021	6.0	0.0026	6.5	0.0040	7.0		
U1, 35-40	0.0087	9.7	0.0109	11.5	0.0104	11.8			0.0138	13.9
U6, 15-20	0.0071	11.3	0.0112	13.3	0.0190	15.0	0.0210	16.2		
Kasby, smeared	0.0007	2.8	0.0012	5.0	0.0013	5.3			0.0063	8.2
Kasby, broken up	0.0019	3.2	0.0031	5.2	0.0034	5.2			0.0117	8.1

	200		500	
U1, 10-15	0.0222	13.7	0.0283	14.5
U1, 25-30	0.0094	8.5	0.0180	9.8
U1, 35-40			0.0192	16.1
U6, 15-20	0.0248	18.9	0.0327	21.9
Kasby, smeared			0.0201	11.2
Kasby, broken up			0.0231	10.6

Table 14. Results from transient state diffusion experiments. Soils from Uddeholm (see Tables 10 & 11).

D/D_0 = relative diffusion f_a = air-filled porosity, %

Sample		Matric tension applied, cm w.c.									
		15		30		50		150		500	
		D/D_0	f_a	D/D_0	f_a	D/D_0	f_a	D/D_0	f_a	D/D_0	f_a
1a	10-15	0.0055	4.9	0.0086	6.9	0.0115	8.5	0.0218	11.6	0.0260	13.3
	15-20	0.0032	5.2	0.0064	6.9	0.0090	8.3	0.0154	10.6	0.0233	12.6
	20-25	0.0010	1.5	0.0021	2.5	0.0024	3.0	0.0035	3.7	0.0058	4.4
	25-30	0.0023	2.6	0.0030	3.5	0.0040	4.0	0.0053	4.5	0.0085	5.4
1b	10-15	0.0037	4.3	0.0058	5.9	0.0092	7.6	0.0157	9.9	0.0220	12.5
	15-20	0.0084	6.1	0.0105	7.3	0.0143	8.9	0.0208	10.8	0.0266	13.1
	20-25	0.0006	2.3	0.0009	3.2	0.0008	3.5	0.0012	4.3	0.0045	5.2
	25-30	0.0009	1.6	0.0012	2.8	0.0016	3.1	0.0022	3.8	0.0055	4.7
1c	10-15	0.0029	1.0	0.0032	1.7	0.0045	2.4	0.0081	3.9	0.0152	5.5
	15-20	0.0012	1.1	0.0018	1.8	0.0027	2.5	0.0062	3.9	0.0111	5.5
	20-25	0.0004	0.3	0.0007	0.9	0.0013	1.4	0.0027	2.0	0.0042	2.7
	25-30	0.0004	0.7	0.0006	1.3	0.0014	1.7	0.0029	2.2	0.0044	2.7

Table 15. Results from transient state diffusion experiments. Soils from Grävsta (see Tables 10 & 11).

D/D_o = relative diffusion f_a = air-filled porosity, %

Sample	Matric tension applied, cm w.c.									
	15		30		50		150		500	
	D/D_o	f_a	D/D_o	f_a	D/D_o	f_a	D/D_o	f_a	D/D_o	f_a
1a 10-15	0.0010	0.9	0.0031	2.4	0.0032	4.3	0.0064	5.9	0.0204	8.6
15-20	0.0052	1.8	0.0063	3.5	0.0085	5.0	0.0102	6.4	0.0188	9.0
20-25	0.0073	3.9	0.0100	5.7	0.0138	7.8	0.0188	10.1	0.0223	12.8
25-30	0.0073	0	0.0099	2.2	0.0127	4.2	0.0144	6.2	0.0183	8.8
1b 10-15	0.0058	6.6	0.0079	8.4	0.0091	9.7	0.0114	10.9	0.0190	13.3
15-20	0.0069	4.0	0.0085	5.7	0.0069	7.1	0.0124	8.3	0.0200	10.7
20-25	0.0063	4.1	0.0079	5.2	0.0122	6.5	0.0167	8.2	0.0188	10.1
25-30	0.0068	3.1	0.0074	4.6	0.0090	5.8	0.0107	7.4	0.0134	9.2
1c 10-15	0.0010	0.5	0.0010	0.7	0.0016	0.7	0.0016	1.4	0.0021	3.1
15-20	0.0026	1.9	0.0031	2.6	0.0031	2.9	0.0058	3.9	0.0163	6.0
20-25	0.0026	0	0.0021	0.5	0.0031	1.0	0.0026	2.0	0.0047	3.6
25-30	0.0073	1.2	0.0084	2.9	0.0084	3.7	0.0100	5.0	0.0127	7.2

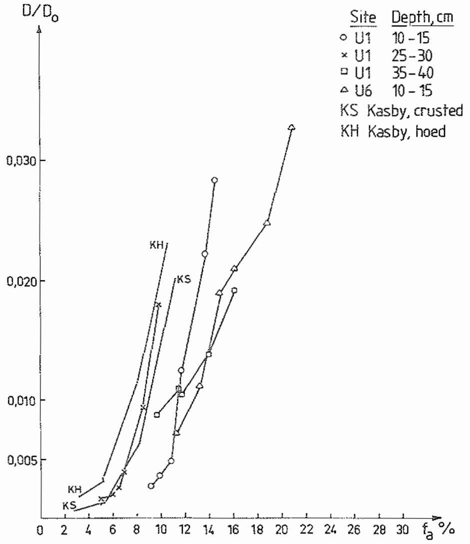


Figure 22. Relative diffusion (D/D_0) as a function of air-filled porosity (f_a) in transient state diffusion experiments. Soils from the Ultuna area.

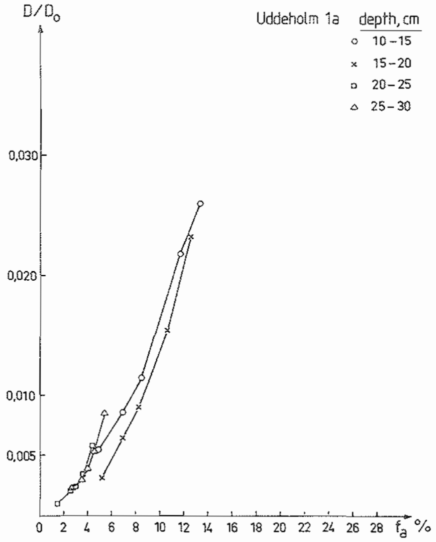


Figure 23. Relative diffusion (D/D_0) as a function of air-filled porosity (f_a) in transient state diffusion experiments. Uddeholm 1a.

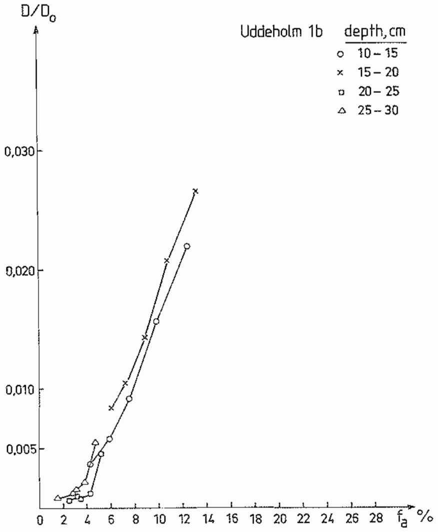


Figure 24. Relative diffusion (D/D_0) as a function of air-filled porosity (f_a) in transient state diffusion experiments. Uddeholm 1b.

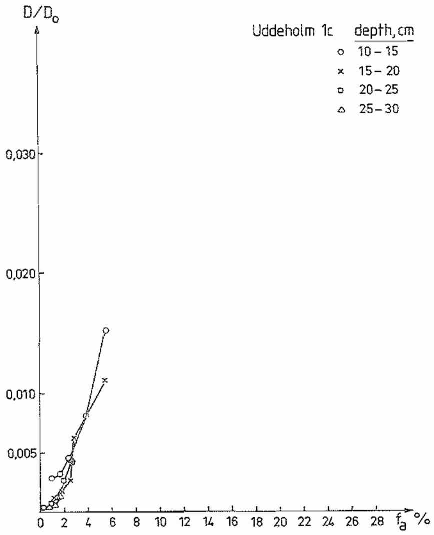


Figure 25. Relative diffusion (D/D_0) as a function of air-filled porosity (f_a) in transient state diffusion experiments. Uddeholm 1c.

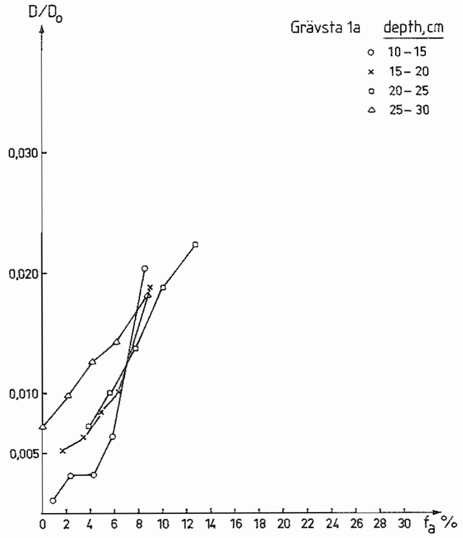


Figure 26. Relative diffusion (D/D_o) as a function of air-filled porosity (f_a) in transient state diffusion experiments. Grävsta 1a.

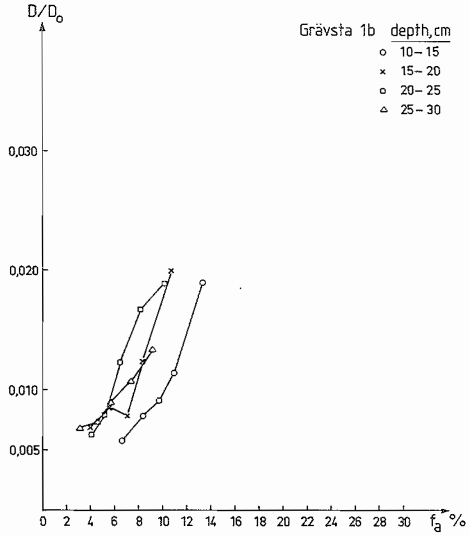


Figure 27. Relative diffusion (D/D_o) as a function of air-filled porosity (f_a) in transient state diffusion experiments. Grävsta 1b.

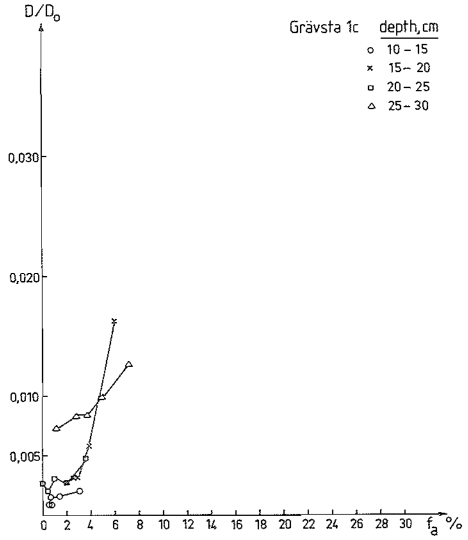


Figure 28. Relative diffusion (D/D_0) as a function of air-filled porosity (f_a) in transient state diffusion experiments. Grävsta 1c.

3.1.4. Discussion of diffusion experiments

Diffusion through dry Atterberg fractions was investigated by both the methods described. Relative diffusion as a function of air-filled porosity is shown in Fig. 15 (steady state) and Fig. 21 (transient state). It can be seen that results differ somewhat. Regression analysis of coordinates in the steady state diffusion graph gave the equation $D/D_0 = -0.085 + 0.0077 \cdot f_a$. Similar analysis of transient state figures gave the equation $D/D_0 = -0.049 + 0.0063 \cdot f_a$. In both cases, the correlation coefficient was greater than 0.99. Values obtained in experiments are thus well represented by the equations. This means also that errors due to non-systematic uncertainties are small.

Diffusion coefficients for steady state diffusion were also calculated as if the entire partial pressure difference between the surface of the carbon disulphide liquid and the external air was active in inducing diffusion through the sample. Partial pressure gradient was then equivalent to saturation vapour pressure of the CS_2

liquid. Fig. 29 shows results of these calculations compared to results from transient state investigations. Regression analysis of values produced by this steady state calculation gave an equation: D/D_0 0.051 + 0.0063 f_a. In this case there is almost perfect agreement between results obtained from the two methods.

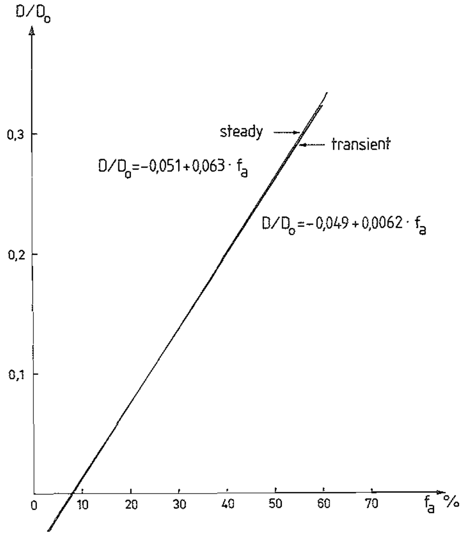


Figure 29. Corrected results from investigations of diffusion through dry Atterberg fractions and mixtures of these. Relative diffusion (D/D_0) is shown as a function of air-filled porosity (f_a) from both steady state and transient state diffusion experiments.

Considering that the apparatus used to measure steady state diffusion caused some convective flow in the system, it is likely that the partial pressure gradient between the surface of the CS_2 liquid and the sample base is cancelled out. This being the case, the corrections shown in Table 4 are inaccurate. In fact, the good agreement between uncorrected steady state results and the transient state results indicates that this is the case. It should be noted here that the lack of effect of a partial pressure gradient in the cap was also noted in transient state experiments, when level at which samples were taken from the chamber had no effect on results.

It is difficult to directly compare results from diffusion experiments carried out by two different methods. From the discussion of methods, it can be seen that there are many complicating factors involved. However, results do not need to be completely accurate quantatively in order for some general conclusions to be drawn. It is sufficient that qualitative differences between samples are correct. The following discussion will centre on qualitative considerations although quantitative aspects will be discussed in some instances (see 3.3.4.).

As mentioned, degree of error in the transient state method used was an estimated 10 % in normal cases. Results of experiments on dry Atterberg fractions can be used to check the accuracy of this figure. Table 12 shows the actual error to be less than 5 % in all cases.

Steady state investigations on dry Atterberg fractions and mixtures of these produced a positive relationship between diffusive flow and air content of the sample. In dry materials, air-filled porosity is equivalent to the total porosity. Results from all materials lie very close to a straight line with equation $D/D_0 = -0.051 + 0.0063\ f_a$. Diffusion through both tightly packed sand (Mix 3) with an air-filled porosity of 21.5 % and loose silt with an air-filled porosity of 53.6 % can be described by the same equation. The reason for this is probably that particle size is rather similar in all materials and the tortuosity is thus also similar. It is surprising that even the most approximate regression line for these results does not pass through the origin. This is probably because the conductivity to diffusive flow of pore systems in the materials investigated decreased at a particular rate within the air content range studied. It should also be noted that the proportionality factor for air content in the relative diffusion equation, 0.0063, is approximately equal to the ratio of diameter of a circle to half its circumference multiplied by 1/100, i.e. 0.0064. The diffusion coefficient per percentage unit air-filled porosity is reduced to just this value if diffusive flow is diverted by a distance of half the circumference of a sphere which has diameter = length of the system and axis = diffusion direction.

By studying diffusion through air-dried clay aggregates, peat and moist sand, some information was obtained on the influence of changes compared to a dry, single-grained system (Fig. 15). All such changes in conformation of the pore system cause diffusion to decrease. This finding agrees well with the theoretical results of de Vries (1950).

Equation 6 (page 15), which defines specific diffusion (S_D) can be used to calculate S_D for each of the samples. Table 16 shows the values obtained from such calculations for dry Atterberg fractions, moist medium sand, air-dried heavy clay, air-dried peat and drained medium sand. These results are shown graphically in Fig. 30.

Table 16. Calculated values of relative diffusion (D/D_0) and specific diffusion (S_D) for different soil materials.

Material	Air-filled porosity, %	Relative diff. D/D_0	Specific diff. S_D
Atterberg fractions, dry	20.2-53.6	$-0.05+0.006 \frac{f_a}{100}$	$0.6-\frac{5}{f_a}$
MSa, moistened	39.0	0.144	0.369
HC, air dry	55.5	0.248	0.447
Peat, air dry	71.7	0.170	0.219
MSa, moist	4.1	0.0	0.0
MSa, moist	5.0	0.015	0.306
MSa, moist	20.8	0.030	0.144
MSa, moist	38.1	0.167	0.438
MSa, dried	40.8	0.212	0.521

In dry Atterberg fractions, specific diffusion increased with air content. This means that path for diffusion became simpler as air content increased. Moistening, on the other hand, had the effect of complicating the canal system or increasing tortuosity of the diffusion path. Specific diffusion decreased by 13 % when as little as 3.6 vol. % of water was added to a dry medium sand with air-filled porosity of 42.6 %.

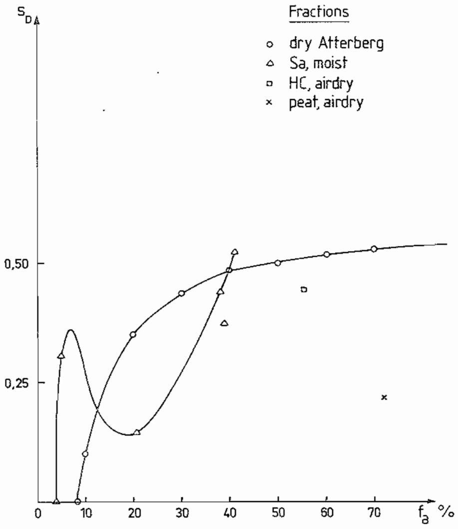


Figure 30. Calculated values of specific diffusion (S_D) as a function of air-filled porosity (f_a) for different soil materials.

A secondary pore system such as the micropore system of air-dried clay aggregates and air-dried peat seems also to increase complexity of the pore system. Specific diffusion of the more complex soil types clay and peat can be compared to that of Atterberg fractions with corresponding porosity. The pore system of peat produced an extreme effect on specific diffusion, giving a value of 0.22 as compared to an estimated 0.53 for a corresponding pure fraction (deduced from experimental values).

To illustrate the influence of degree of saturation by water on diffusion, investigations were carried out on a medium sand which was drained stepwise with 10, 20, 40 and 80 cm w.c., then air dried. Specific diffusion as a function of air-filled porosity gave a curve with two distinct points of inflexion (Fig. 30).

It should be noted, however, that the minimum point on the curve in Fig. 30 (at around 20 % air-filled porosity) occurs in the centre of the soil moisture characteristic plateau. This may be of some real significance. The maximum value on the curve probably represents the case when applied tension exceeds capillarity in a limited number of pores, i.e. that only a few large pores are drained and open for diffusion. When smaller pores open, they are at first not very continuously open, thus not very effective.

The regression equation for transient state diffusion in dry Atterberg fractions, as shown in Table 12 and Fig. 30, is $D/D_0 = -0.049 + 0.0062\ f_a$. As has been discussed previously, this equation agreed well with the steady state equation $D/D_0 = -0.051 + 0.0063\ f_a$. If both these equations are rounded off the equation becomes:

$$\frac{D}{D_0} = -0.05 + 0.006 \cdot f_a$$

Results from natural soils investigated are shown in Tables 13-15 and in Figs. 22-28. These results confirm the fact that the relationship between diffusion and air content is a complicated one.

Investigations were carried out on soils from the Ultuna area (fig. 22) and on soil samples taken from two experimental sites, Uddeholm and Grävsta, see Figs. 23-28.

Results show a clear trend, that of a positive relationship between diffusion and air content of the sample. However, results were widely scattered and diffusion at a particular air content of the sample varies from point to point within a soil and from one soil to the next. This is clearly illustrated in Fig. 22, which shows results obtained from soils of the Ultuna region. Three sites, Ultuna 1, Ultuna 6 and Kasby, are represented in this diagram. On Ultuna 1 site, three soil depths were investigated. On the Kasby site, samples were taken from an area with a hard, intact surface pan and from an area where this pan had been broken up by shallow hoeing.

The results obtained form two main groups with all the Ultuna samples except the 25-30 cm level from Ultuna 1 in one group and all the Kasby samples plus the 25-30 cm level from Ultuna 1 in the other.

Surprisingly, soil samples with the worst structure, i.e. low-humus heavy clay from Kasby and the plough pan from Ultuna 1, showed the highest relative diffusion at a particular air content, highest specific diffusion. Consider, e.g. soil from the plough pan of the Ultuna site, which has a massive structure caused by extreme mechanical compaction. At 10 % air-filled porosity, diffusion in this soil was twice that in the granular, well structured upper part of the subsoil. On the other hand, breaking up of the surface pan on the weak-structured Kasby soil had also a positive effect on specific diffusion. This agrees with the findings of Bakker & Hidding (1970).

The relationship between soil structure and relative diffusion at a particular air content is obviously complicated. This conclusion is confirmed by results from Uddeholm and Grävsta soils. From Uddeholm samples (Figs. 23-25), it was found that relative diffusion in soils under grass (1a) does not differ from that in soils under open cultivation (1b), i.e. specific diffusion was the same in both cases. Relative diffusion was, however, badly affected in intensely compacted soils (1c), but to the effect that specific diffusion was higher in this case. Compare this with the findings for the plough pan of Ultuna 1, where specific diffusion was also higher. This trend was also exhibited, though to a less extent, by the Grävsta soils (Figs. 26-28). Evidently is high specific diffusion at lower air contents a sign of bad structure in the soil.

Grävsta soils showed a slight tendency for specific diffusion to be higher in soils under grass (1a) and compaction (1c) than under open cultivation (1b). The conclusion to be drawn was that both grass and intense compaction of soil affected specific diffusion, in the same way. This would seem to imply that a number of conflicting processes are at work in natural soils.

From the results obtained, it is clear that relative diffusion in a soil cannot be defined as a function of only one variable, for example air-filled porosity. All equations given in the literature which have form $D/D_0 = a \cdot f_a$, $D/D_0 = a(f_a - b)$, $D/D_0 = f_a^c$ etc. and which express specific diffusion as a constant, can only be true for a certain point in a certain soil and at a certain time. Such formulae are only valid for use as simple models or for approximate calculations. The constants given in these formulae must be regarded as functions of many variables.

A more correct equation describing relative diffusion as a function of specific characteristics of a porous medium and external factors affecting this is:

$$D/D_o = (f, f_a, \mathcal{S}_v \cdot \mathcal{S}_p, h_t, w, \frac{dw}{dt}, \frac{dw}{dx}, c_f, c_s, m_h, B, z, y, x, X, Y)$$

where f = porosity,
f_a = air-filled porosity,
$\mathcal{S}_v$ = pore size distribution,
$\mathcal{S}_p$ = particle size distribution,
h_t = matric tension,
w = water content,
dw/dt = drying or wetting conditions
dw/dx = moisture gradient in the profile,
c_f = particle form,
c_s = particle surface,
m_h = humus content,
B = cultivation history,
z, x, y = position coordinates in profile,
X, Y = unknown factors.

With regard to this, it may be worthwhile reconsidering the ideas on gas diffusion in soils first proposed by Blake & Page (1948). The results of the present investigation agree in principle with their findings: "Diffusion is the dependent variable in a function which has many independent variables which may be functions of each other."

In discussions on soil structure, relative diffusion and associated entities such as specific diffusion and various constants have been regarded as possible parameters of soil structure, see Currie (1965). From the foregoing description of results it is obvious that relative diffusion as a function of air content is a complex parameter.

Finally, Fig. 31 shows results of regression analyses on the sum of all values obtained from Uddeholm samples and on the sum of all values obtained from Grävsta samples. Relative diffusion is expressed as a function of air-filled porosity and the curves obtained show the difference in $D/D_o = (f_a)$ between two different sites. However, no further conclusions can be drawn from this diagram.

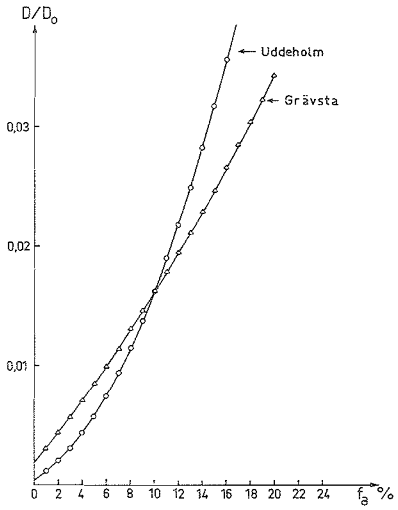


Figure 31. Results of regression analysis on 36 points from Uddeholm and 36 from Grävsta, these being the total results from triplicate samples on 12 plots per site.

3.2. Air permeability investigations

At the time of this investigation, a large number of measurements of soil permeability to air had been carried out at the Department of Agricultural Hydrotechnics by Dr. Paul Wiklert.

Since this technique was well tested and much experience of the apparatus and the method existed at the Department, it was decided to complement diffusion studies with investigations into the air permeability of samples.

3.2.1 Background in the literature

Renk (1879), Ammon (1880) and Wollny (1893) showed that permeability of a soil to air decreases with increasing sample length, water content and degree of dispersion. Buckingham (1904) discovered that what he termed a soil's permeability constant is proportional to free pore volume to the 6th or 7th power. Buehrer (1932) showed that permeability to air of a material is proportional to the square of the mean particle diameter. In other words, air permeability of a soil increases rapidly as frequency of larger pores increases. Compare also the theoretical model of flow in a straight, cylindrical capillary described in 2.4.4.1.

In more recent years, apparatus has been developed for field and laboratory investigations of air permeability of natural soils (Evans & Kirkham, 1957; Kmoch & Hanus, 1965; Andersson, 1969). As a rule, air permeability has been regarded as a parameter of soil structure.

At present, the exact soil physical significance of soil permeability to air has not been fully investigated. Evaluation of experimental results has been used to construct complicated models of, for example, relative surface area of the particle system although there is a lack of empirical data.

However, the basic effects of pore size distribution and degree of saturation on air permeability have been presented in a capillary model by Andersson (1969) and this model can be applied in further studies on air permeability of soils.

Air permeability of soils has not generally been considered with respect to soil ventilation problems, the diffusion process is regarded as the deciding process in soil air exchange and research has been concentrated on diffusion (Romell, 1922; Penman, 1940).

3.2.2 Air permeability experiments

Air permeability of soil was measured in an apparatus which, based on the continuity principle, was developed by Andersson (1969). Construction of this apparatus is shown in Fig. 32.

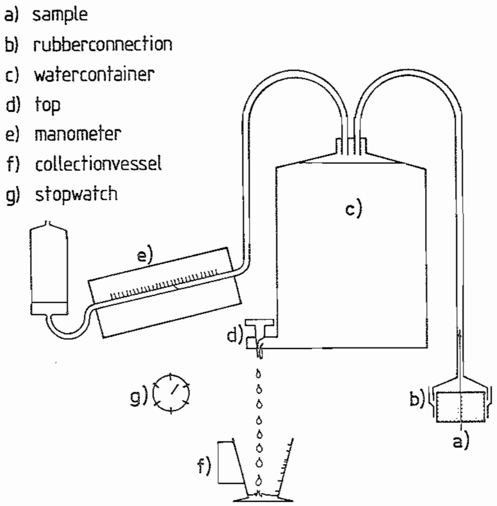


Figure 32. Diagram showing construction of apparatus used to measure air permeability of soil.

3.2.2.1 Materials and method

Air permeability measurements were carried out immediately after samples were removed from the pressure chambers (where matric suction in the soil was regulated) and before they were transferred to diffusion apparatus.

Air permeability of a sample was measured thus:
The sample, (a) in Fig. 32, enclosed in its steel cylinder, was connected to the body of the apparatus by a rubber cuff (b). The sample is thus in contact with the volume of air over the water in tank (c) via the rubber tube connected to the cuff. When tap (d) is opened, water runs out and a pressure deficit occurs in the air in tank (c). This pressure drop causes air to be drawn in through the soil sample to the tank. When the system reaches equilibrium i.e. when steady state conditions are fulfilled, the volume of water leaving via the tap per unit time is equal to the volume of air drawn in through the sample. Magnitude of the pressure gradient is shown by a manometer (e) which is water-filled and lying at an angle to the horizontal and therefore sensitive.

The pressure gradient can be regulated for each individual sample by opening tap (d) to varying degrees. Once steady state conditions are reached, the volume of water which runs from the system in a certain time interval is collected in a vessel and measured in a graduated cylinder. Time is measured with a stopwatch. Using measured values of sample dimensions, pressure gradient, volume of water which left the system and ambient temperature, the air permeability coefficient k_L can be calculated from Darcy's law.

$$q = k_L \cdot A \cdot \frac{\Delta P}{l} \cdot t \qquad (1)$$

where q = volume of water collected in time t,
k_L = permeability coefficient,
ΔP = pressure gradient,
l = sample length,
A = cross-sectional area of sample.

3.2.2.2 Discussion of method

Values of the air permeability coefficient obtained by the method above contain some uncertainties and systematic errors.

In calculation, no account was taken of air expansion during flow through the sample but, according to the universal gas law, $P \cdot V = n \cdot R \cdot T$ = constant, air must expand when it is drawn through the soil by a pressure deficit. The pressure gradient through the sample in this experiment was never more than 5 cm w.c. According to the gas law, a decrease in pressure of 5 cm w.c. means an increase in volume of 0.49%. This is therefore the greatest error which can be brought about by expansion of the gas mass in this investigation.

Air viscosity is a function of temperature. For a clean, dry air viscosity η is given by $\eta = 181.8 + 0.495(T - 293)$ micropois. Ambient temperature on the laboratory is maintained at 21°C, 294 °K. A change in temperature of 1 °K means an error of 0.27 %. In other words, the error due to temperature variations is insignificant.

An error which is difficult to determine is the effect of turbulence in the flowing air. Such turbulence can be expected to occur quite frequently, especially in larger pores or in the rubber tubing of the apparatus. The theoretically calculated equations in e.g. Andersson (1969) or in Section 2 of this work, and from which equation 1 is derived, presuppose that flow is laminar. When turbulent flow occurs, loss of flow increases and the calculated value of the permeability coefficient is too low. This error can be considerable, especially in systems with coarse pores where air permeability is great and pressure gradient held great. However, this combination was not allowed.

Magnitude of the error is therefore insignificant as long as the turbulence phenomenon is not significant and adjustment of calculations was not deemed necessary.

3.2.2.3 Experimental results

Experiments were carried out on several different Atterberg fractions and mixtures of these, on two moist medium sand fractions and on naturally occurring soils. These samples were those used in diffusion investigations. For explanation of abbreviations and soil types see Tables 5, 10 and 11 in section 3.1. As in diffusion experiments, three replicates were tested for each sample. Results of experiments are recorded in Tables 17-20 and in Figs. 33-40.

Table 17. Results of air permeability experiments. Atterberg fractions and mixtures of these, see Table 5.

Material	Air content, f_a %	Air permeability, k_L cm/min
MSa	40.5	$0.7629 \cdot 10^2$
CSi	53.6	$0.1807 \cdot 10$
Mix 2	28.0	$0.4970 \cdot 10^2$
Mix 3	21.5	0.8187
MSa I, h_t = 10 cm w.c.	4.1	0
MSa, h_t = 20 cm w.c.	5.0	$0.4767 \cdot 10$
MSa, h_t = 40 cm w.c.	20.0	$0.2505 \cdot 10^2$
MSa, h_t = 80 cm w.c.	38.1	$0.6821 \cdot 10^2$
MSa, h_t = 10^7 cm w.c.	40.8	$0.8455 \cdot 10^2$
MSa II, h_t = 15 cm w.c.	5.7	$0.11 \cdot 10$
MSa, h_t = 25 cm w.c.	15.0	$0.289 \cdot 10$
MSa, h_t = 30 cm w.c.	26.5	$0.536 \cdot 10$
MSa, h_t = 50 cm w.c.	34.1	$1.16 \cdot 10^2$
MSa, h_t = 200 cm w.c.	36.1	$1.18 \cdot 10^2$
MSa, h_t = 500 cm w.c.	36.8	$1.60 \cdot 10^2$

Table 18. Results of air permeability experiments. Soils from the Ultuna area, see Tables 10 and 11.

k_L = air permeability, cm/min f_a = air content of sample, %.

Site, level	Matric tension applied, cm w.c.									
	15		30		50		100		150	
	k_L	f_a	k_L	f_a	k_L	f_a	k_L	f_a	k_L	f_a
U1, 10-15	$5.00 \cdot 10$	9.2	$6.62 \cdot 10$	9.9	$1.15 \cdot 10^2$	10.9	$2.30 \cdot 10^2$	11.7		
U1, 25-30	$2.00 \cdot 10$	5.0	$3.41 \cdot 10$	6.0	$3.87 \cdot 10$	6.5	$8.43 \cdot 10^1$	7.0		
U1, 35-40	$2.05 \cdot 10^2$	9.7	$2.20 \cdot 10^2$	11.5	$4.08 \cdot 10^2$	11.8			$4.16 \cdot 10^2$	13.9
U6, 10-15	$1.01 \cdot 10^2$	11.3	$1.79 \cdot 10^2$	13.3	$2.92 \cdot 10^2$	15.0	$5.82 \cdot 10^2$	16.2		
KS	0.133	2.8	2.05	5.0	2.34	5.3			$9.71 \cdot 10$	8.2
KH	$1.21 \cdot 10$	3.2	$2.03 \cdot 10$	5.2	$2.63 \cdot 10$	5.2			$1.48 \cdot 10^2$	8.1

	200		500	
U1, 10-15	$1.59 \cdot 10^3$	13.7	$2.36 \cdot 10^3$	14.5
U1, 25-30	$1.17 \cdot 10^2$	8.5	$5.55 \cdot 10^2$	9.8
U1, 35-40			$5.79 \cdot 10^2$	16.1
U6, 10-15	$7.53 \cdot 10^2$	18.9	$1.08 \cdot 10^3$	21.9
KS			$7.76 \cdot 10^2$	11.2
KH			$9.06 \cdot 10^2$	10.6

Table 19. Results of air permeability experiments. Soils from Uddeholm, see Tables 10 and 11.
k_L = air permeability, cm/min f_a = air content of sample, %.

Site, level		Matric tension applied, cm w.c.									
		15		30		50		150		500	
		k_L	f_a	k_L	f_a	k_L	f_a	k_L	f_a	k_L	f_a
1a	10-15	$2.60 \cdot 10^2$	4.9	$3.08 \cdot 10^2$	6.9	$3.26 \cdot 10^2$	8.5	$4.29 \cdot 10^2$	11.6	$6.63 \cdot 10^2$	13.3
	15-20	$4.79 \cdot 10$	5.2	$6.91 \cdot 10$	6.9	$7.91 \cdot 10$	8.3	$1.21 \cdot 10^2$	10.6	$1.93 \cdot 10^2$	12.6
	20-25	9.67	1.5	$2.52 \cdot 10$	2.5	$2.96 \cdot 10$	3.0	$2.31 \cdot 10$	3.7	$3.38 \cdot 10$	4.4
	25-30	$5.51 \cdot 10^2$	2.6	$1.01 \cdot 10^3$	3.5	$1.06 \cdot 10^3$	4.0	$8.94 \cdot 10^2$	4.5	$9.12 \cdot 10^2$	5.4
1b	10-15	$2.23 \cdot 10^2$	4.3	$1.85 \cdot 10^2$	5.9	$2.90 \cdot 10^2$	7.6	$5.76 \cdot 10^2$	9.9	$4.43 \cdot 10^2$	12.5
	15-20	$7.32 \cdot 10^2$	6.1	$7.29 \cdot 10^2$	7.3	$6.54 \cdot 10^2$	8.9	$7.31 \cdot 10^2$	10.8	$1.29 \cdot 10^3$	13.1
	20-25	0	2.3	1.0	3.2	1.0	3.5	1.10	4.3	4.06	5.2
	25-30	5.85	1.6	5.78	2.8	6.49	3.1	8.88	3.8	$1.54 \cdot 10$	4.7
1c	10-15	$8.61 \cdot 10^2$	1.0	$8.75 \cdot 10^2$	1.7	$9.26 \cdot 10^2$	2.4	$9.20 \cdot 10^2$	3.9	$9.41 \cdot 10^2$	5.5
	15-20	7.12	1.1	$3.81 \cdot 10$	1.8	$3.91 \cdot 10$	2.5	$9.11 \cdot 10$	3.9	$9.32 \cdot 10$	5.5
	20-25	0	0.3	0.19	0.9	0.38	1.4	1.79	2.0	9.45	2.7
	25-30	0	0.7	0	1.3	0.24	1.7	2.50	2.2	7.97	2.7

Table 20. Results of air permeability experiments. Soils from Grävsta, see Tables 10 and 11.
k_L = air permeability, cm/min f_a = air content of sample, %.

Site, level		Matric tension applied, cm w.c.									
		15		30		50		150		500	
		k_L	f_a	k_L	f_a	k_L	f_a	k_L	f_a	k_L	f_a
1a	10-15	3.22	0.9	7.42	2.4	$1.28 \cdot 10$	4.3	$2.14 \cdot 10$	5.9	$6.83 \cdot 10^2$	8.6
	15-20	$1.58 \cdot 10^2$	1.8	$1.48 \cdot 10^2$	3.5	$1.73 \cdot 10^2$	5.0	$1.94 \cdot 10^2$	6.4	$3.12 \cdot 10^2$	9.0
	20-25	$7.34 \cdot 10^2$	3.9	$7.60 \cdot 10^2$	5.7	$8.12 \cdot 10^2$	7.8	$5.11 \cdot 10^2$	10.1	$1.06 \cdot 10^3$	12.8
	25-30	$2.26 \cdot 10^2$	0	$3.89 \cdot 10^2$	2.2	$4.39 \cdot 10^2$	4.2	$6.54 \cdot 10^2$	6.2	$7.58 \cdot 10^2$	8.8
1b	10-15	$9.88 \cdot 10$	6.6	$1.24 \cdot 10^2$	8.4	$1.60 \cdot 10^2$	9.7	$1.62 \cdot 10^2$	10.9	$3.86 \cdot 10^2$	13.3
	15-20	$1.22 \cdot 10^2$	4.0	$1.75 \cdot 10^2$	5.7	$2.18 \cdot 10^2$	7.1	$2.47 \cdot 10^2$	8.3	$5.75 \cdot 10^2$	10.7
	20-25	$1.76 \cdot 10^2$	4.1	$2.06 \cdot 10^2$	5.2	$2.14 \cdot 10^2$	6.5	$4.79 \cdot 10^2$	8.2	$9.09 \cdot 10^2$	10.1
	25-30	$1.14 \cdot 10^2$	3.1	$1.06 \cdot 10^2$	4.6	$1.51 \cdot 10^2$	5.8	$2.15 \cdot 10^2$	7.4	$2.57 \cdot 10^2$	9.2
1c	10-15	0	0.5	0.31	0.7	0.49	0.7	0.64	1.4	1.55	3.1
	15-20	$2.74 \cdot 10$	1.9	$2.63 \cdot 10$	2.6	$3.27 \cdot 10$	2.9	$5.18 \cdot 10$	3.9	$2.85 \cdot 10^2$	6.0
	20-25	$3.36 \cdot 10$	0	$2.69 \cdot 10$	0.5	$3.42 \cdot 10$	1.0	$3.58 \cdot 10$	2.0	$3.59 \cdot 10$	3.6
	25-30	$2.66 \cdot 10^2$	1.2	$3.06 \cdot 10^2$	2.9	$3.55 \cdot 10^2$	3.7	$3.83 \cdot 10^2$	5.0	$4.76 \cdot 10^2$	7.2

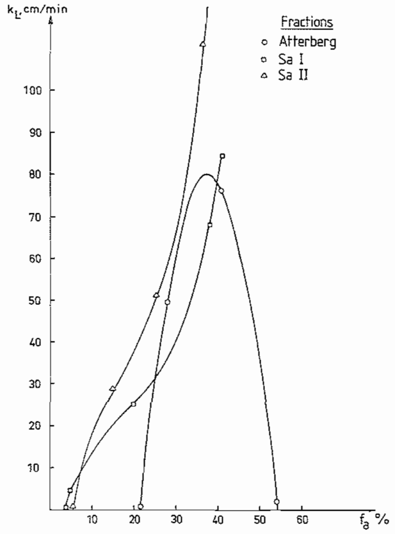


Figure 33. Air permeability (k_L) as a function of air-filled porosity (f_a) in dry Atterberg fractions and two moist medium sand fractions.

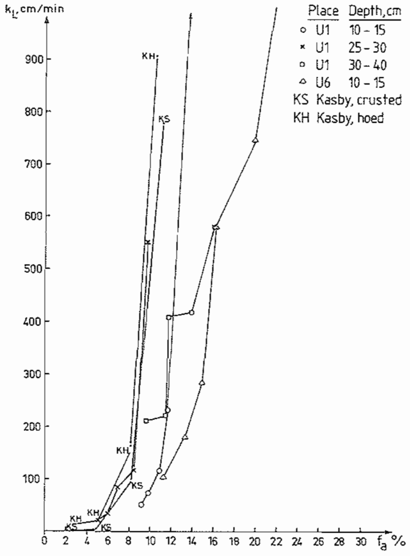


Figure 34. Air permeability (k_L) as a function of air-filled porosity (f_a) in soils from the Ultuna area.

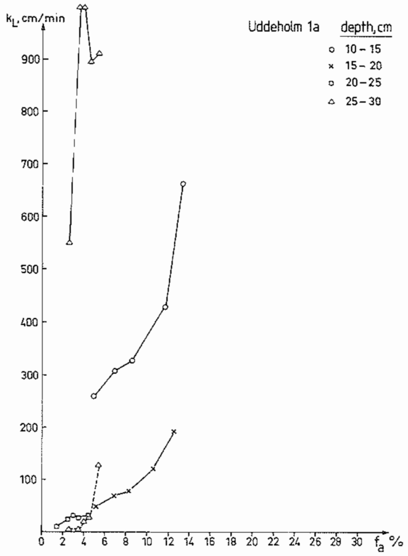


Figure 35. Air permeability (k_L) as a function of air-filled porosity (f_a), Uddeholm 1a. For explanation of dashed line, see text.

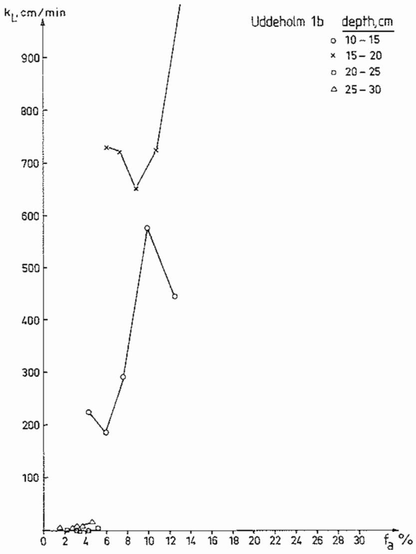


Figure 36. Air permeability (k_L) as a function of air-filled porosity (f_a), Uddeholm 1b.

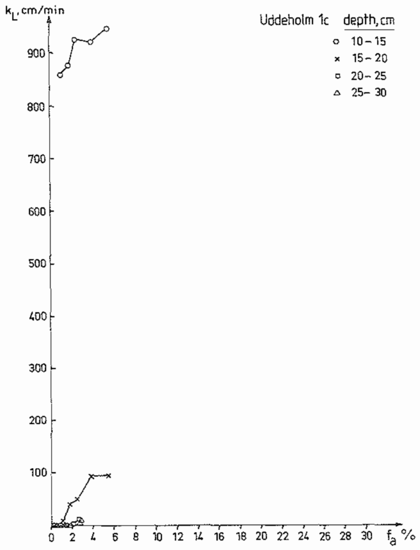


Figure 37. Air permeability (k_L) as a function of air-filled porosity (f_a), Uddeholm 1c.

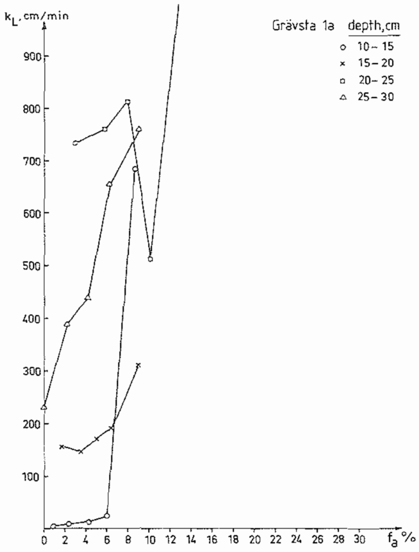


Figure 38. Air permeability (k_L) as a function of air-filled porosity (f_a), Grävsta 1a.

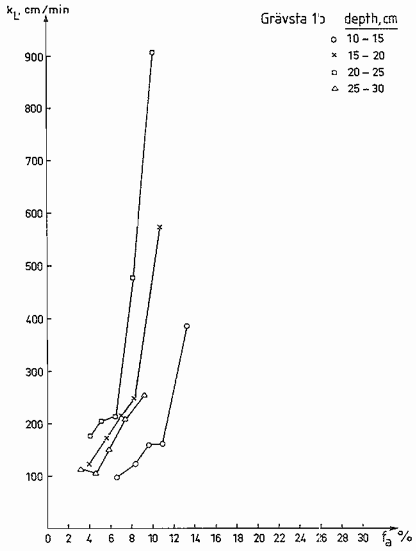


Figure 39. Air permeability (k_L) as a function of air-filled porosity (f_a), Grävsta 1b.

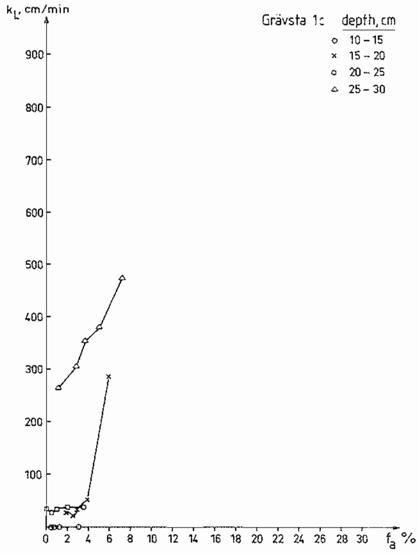


Figure 40. Air permeability (k_L) as a function of air-filled porosity (f_a), Grävsta 1c.

3.2.3 Discussion of air permeability experiments

Fig. 33 shows results of air permeability measurements as a function of air-filled porosity on dry Atterberg fractions and two moist medium sand fractions. The curve joining points for dry materials forms a maximum, for reasons which may seem difficult to explain. However, as discussed in Chapter 2, air permeability is not only proportional to total cross-sectional area of canals but also to the square of their radius. Curves such as that in Fig. 33 may be a result of this double proportionality, and illustrates the danger of regarding air permeability as only proportional to air content.

Different dry soil samples can show a range of totally different relationships between air content and air permeability. There is thus no general equation relating air content and air permeability such as that for air content and diffusion.

On the other hand, if a system is first saturated, then drained and aerated, a positive relationship between air content and air permeability is obtained for any soil. Pore dimensions determine the maximum possible level of permeability while air content and pore tortuosity determine the fraction of this which actually occurs. An example of this is shown in Fig. 33 where air permeability through two drained medium sand fractions is represented as a function of air-filled porosity. One sand fraction is slightly coarser than the other and the curves obtained are similar in shape even if actual values of permeability are somewhat higher in the coarser sand.

It is obvious that for a particular system in moist condition, the relationship between air permeability and air-filled porosity is always positive. However, no safe predictions can be made from a known value of air-filled porosity of the absolute amount of air permeating through soil materials of different textures. Results showed a great difference even between two fractions in the same textural class, medium sand in the Atterberg classification. Air permeability measurements would seem to provide a very sensitive means of determining size and continuity of a pore system.

Results from natural soils from the Ultuna area which were drained at a range of applied moisture tensions are shown as a function of air-filled porosity in Fig. 34. As was also the case in diffusion experiments, results obtained formed two groups, one with soils or soil layers having poor structure, the other with well structured soils and soil layers. The samples with poor structure, Kasby soil and the plough pan (25-30 cm level) from Ultuna 1, produced very similar curves. As in diffusion experiments, results were closely grouped. Kasby soil on which the surface pan had been broken had, however, a somewhat higher air permeability per unit volume than soil with the surface pan intact. Soils with good structure produced more scattered results and could be said to form a group only in that they differed from poorly structured soils.

Air permeability measurements on soils from Uddeholm and Grävsta, which were a part of an experiment investigating the effect of cultivation pattern on soil structure, showed how closely air permeability depends on structure of the pore system (see Figs 35-40).

Results from Uddeholm soils provided information on the parameters which can be measured in air permeability investigations. Uddeholm soil is of special interest in that the profile is divided into two layers with very different characteristics. The topsoil (10-20 cm level) consists of humus-improved silt, the subsoil of pure silt. This difference was reflected in results, where the subsoil is represented by the curves lying close to the horizontal axis in Figs. 35-37. There was, however, a significant exception in the case of soil from Uddeholm 1a. All results from the 25-30 cm level (the lower subsoil sample) were more than double those from other levels for a particular air content. This was due to the presence of a few coarse pores in one of the three replicates used. The enormous air permeability value obtained for this one sample was sufficient to influence the average result for that site. When this extreme value was discarded, the curve shown as a dashed line in Fig. 35 was obtained. Another example of this occurred in the 10-15 cm level from Uddeholm 1c (intensely compacted) see Fig. 37. Air permeability in this layer was ten times greater than that in other layers because of a few large pores in one sample.

Air permeability differences between soils used for grass and those which were carefully cultivated (1a and 1b plots respectively) were small. A slight improvement of air permeability in the grass subsoil could be noted; this was probably due to improved rooting in the upper parts of the subsoil.

In contrast to results obtained in diffusion experiments, intense compaction had a strong influence on air permeability, reducing it as shown in 1c results.

Similar conclusions can be drawn for Grävsta soils, Figs. 38-40. The influence of grass was relatively insignificant. The phenomenon observed in diffusion experiments, where the curve for grassed soils (1a) is shifted to the left in relation to that for open cultivated soils (1b) could also be observed here. Intense compaction (1c) had a strong negative influence on air permeability in Grävsta soils.

A consistent trend in the natural soils investigated was that dk_L/df_a increased within the air content range. With regard to what has been said about the importance of large pores in determining air permeability and to the fact that the corresponding effect is not as obvious for diffusion, this trend seems surprising. The finer pores which are aerated as air-filled porosity increases are obviously very important for the continuity of the pore system.

Different samples gave rise to a series of different relationships between air-filled porosity and air permeability. Samples having identical air contents can, as shown in the figures, have air permeability values which differ by 1000 times. In some cases air permeability was already significantly large at air contents of a few percent while in other cases it was still zero at this air content. The reason for this is the double proportionality, to total canal area and to the square of canal radius.

Most of the trends observed in diffusion experiments could be observed, magnified or extended with respect to the axes, in air permeability investigations. Relative diffusion could be expressed as a function of a large number of known and unknown variables. Air permeability of a material can be expressed in the same way:

$$k_L = f(f,\ f_a, \mathscr{P}_v, \mathscr{P}_p,\ h_t,\ w,\ \frac{dw}{dt},\ \frac{dw}{dx},\ c_f,\ c_s,\ m_h,\ B,\ z,\ x,\ y,\ \ldots\ X,\ Y)$$

where symbols are as given on p. 69.

3.3 Concluding discussion of experimental work on gas transport in soils

3.3.1. Summary of results

Results of experiments show that there is a positive relationship between diffusion and air content of a soil. In natural soils, however, the potential diffusion at a particular air content varies widely. Variation is great even between replicates and even greater between different samples (soil types, levels, pre-treatments). Very exact relationships can sometimes be obtained for certain well defined groups of materials, e.g. the dry Atterberg fractions investigated (see Table 12).

Results obtained for different samples were often unexpected. Relative diffusion was occasionally seen to be increased both by intense compaction and by grass.

In conclusion, results show that relative diffusion as a function of air-filled porosity was not affected in a clear-cut fashion by external factors such as compaction, permanent grassland and pan formation. Furthermore, a large variation between soil types was shown. Use of relative diffusion or specific diffusion as a structure parameter is thus not without complications. Classification of materials according to weak or strong structure on empirical basis gives a different result to classification on the basis of increasing or decreasing diffusion.

Air permeability investigations would seem to provide a better method of classification. The disparity of results between replicates, soil types and levels in the profile was even greater than in diffusion results but the effects of the different cultivation patterns were shown more clearly. In most cases, a certain effect of grass on soil structure and air permeability could also be seen, e.g. on dk_L/df_a. Unfortunately, there were cases where air permeability was great in samples of soil with poor structure, see for example results from Uddeholm, Fig. 35. Such results could lead to inaccurate classification on the basis of air permeability.

3.3.2 Choice of independent variables in graphic representation of results

Following the examples given in the literature, air-filled porosity was taken as the independent variable in these experiments but it should be noted that matric tension is a possible alternative.

Tensions were applied to the soil sample in a pressure chamber and gave rise to a certain moisture content and thus to a certain air content. Tension has the advantage that it forms the base of the function of the soil as a moisture storing and moisture transporting system. The system's function in the field is determined to a great extent by how the system responds to applied matric tension. Tension is a particularly suitable variable if movement of water vapour is to be taken into account because in such a case the gradient is not primarily a concentration difference in the air phase but a tension gradient in the water phase. Furthermore, since classification of soils was not particularly accurate when air filled porosity was chosen as the independent variable, it is necessary to find a better means of classification, possibly on the basis of matric tension.

Figs. 42-47 show relative diffusion as a function of matric tension for natural soils. Figs. 48-54 show air permeability of natural soils as a similar function.

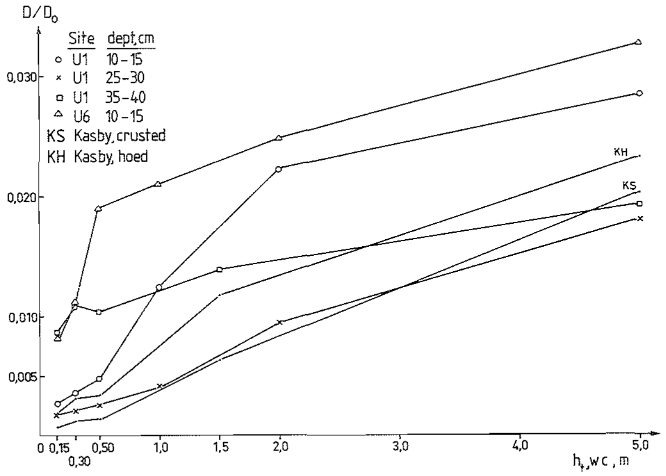


Figure 41. Relative diffusion (D/D$_0$) as a function of matric tension (h_t). Ultuna soils.

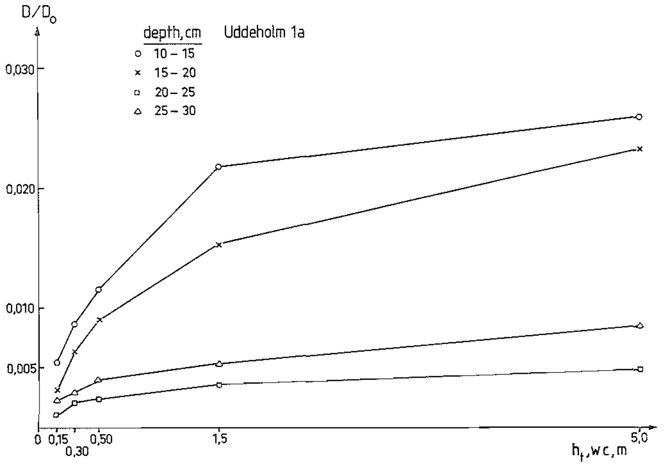


Figure 42. Relative diffusion (D/D$_0$) as a function of matric tension (h_t). Uddeholm 1a.

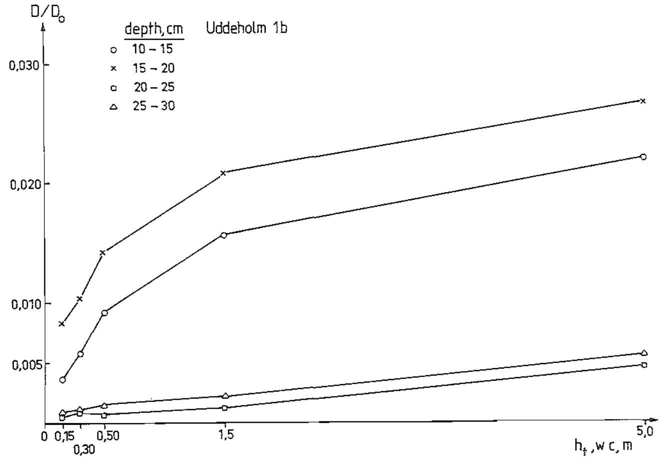


Figure 43. Relative diffusion (D/D_0) as a function of matric tension (h_t). Uddeholm 1b.

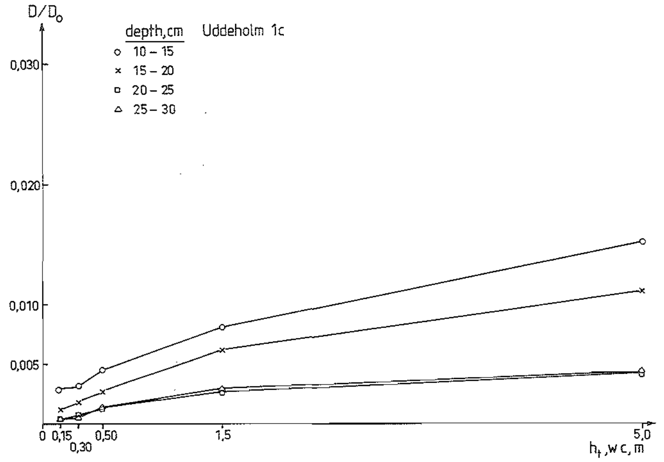


Figure 44. Relative diffusion (D/D_0) as a function of matric tension (h_t). Uddeholm 1c.

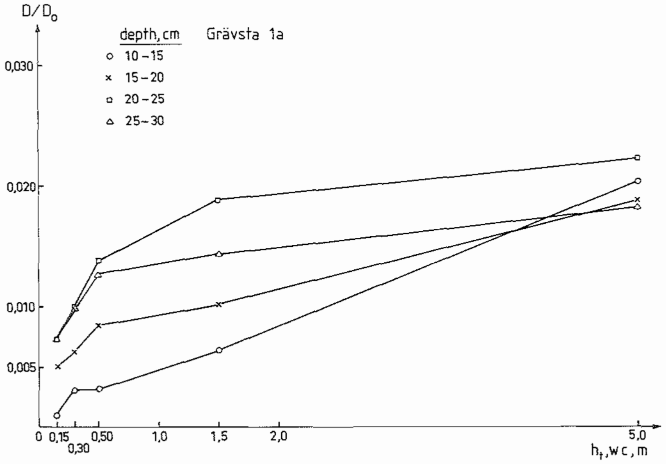


Figure 45. Relative diffusion (D/D$_0$) as a function of matric tension (h_t). Grävsta 1a.

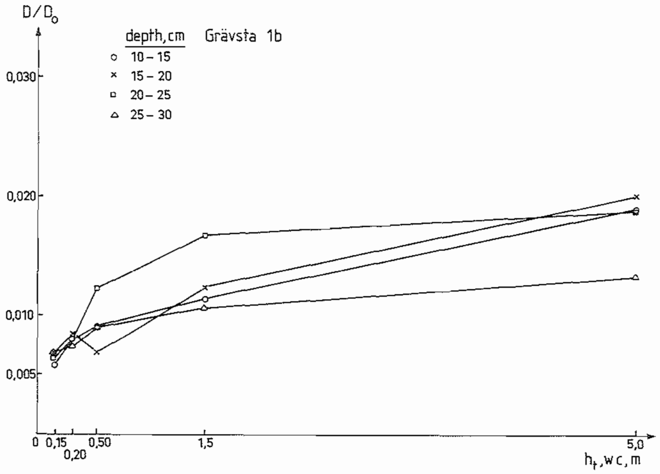


Figure 46. Relative diffusion (D/D$_0$) as a function of matric tension (h_t). Grävsta 1b.

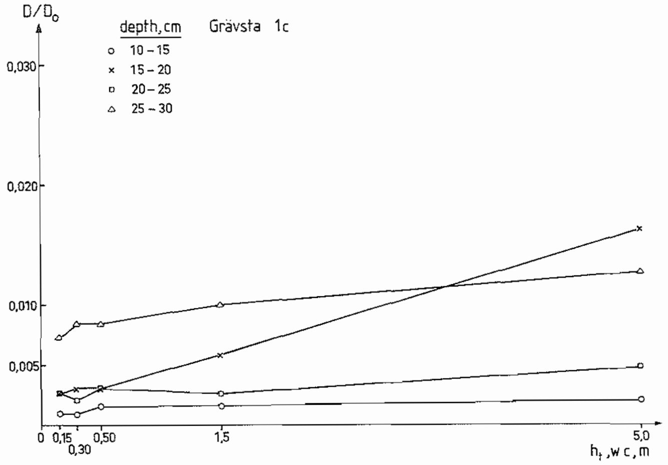


Figure 47. Relative diffusion (D/D_0) as a function of matric tension (h_t). Grävsta 1c.

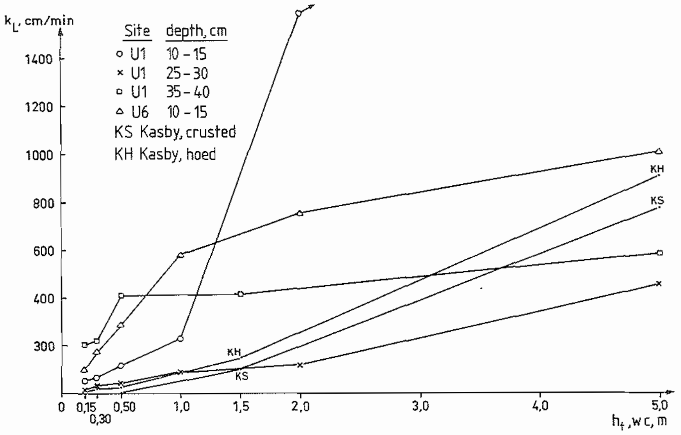


Figure 48. Air permeability (k_L) as a function of matric tension (h_t). Ultuna soils.

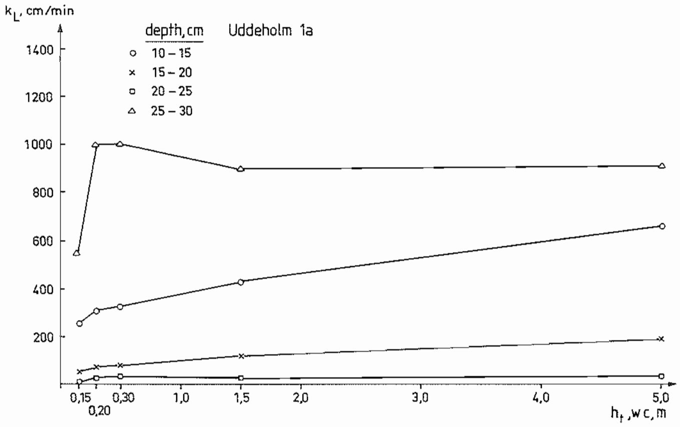


Figure 49. Air permeability (k_L) as a function of matric tension (h_t). Uddeholm 1a.

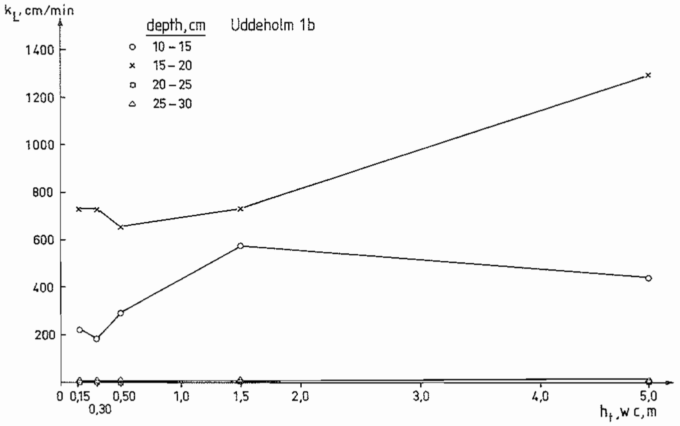


Figure 50. Air permeability (k_L) as a function of matric tension (h_t). Uddeholm 1b.

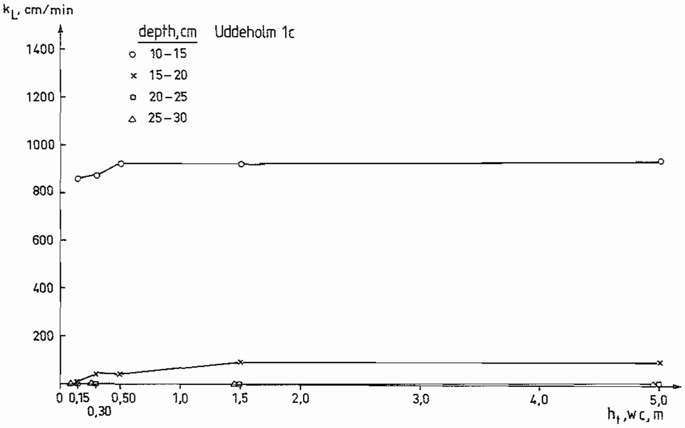


Figure 51. Air permeability (k_L) as a function of matric tension (h_t).
Uddeholm 1c.

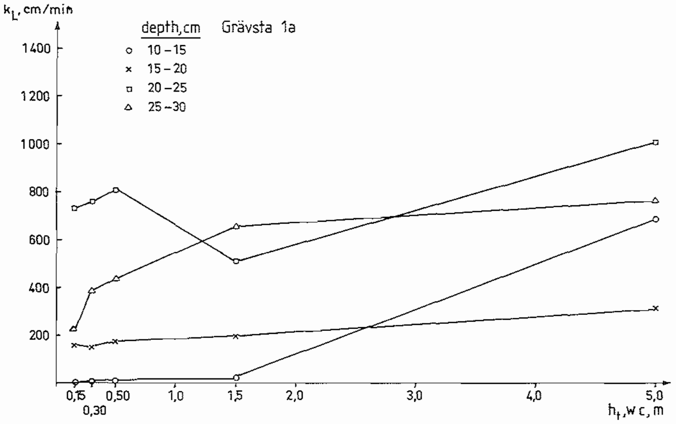


Figure 52. Air permeability (k_L) as a function of matric tension (h_t).
Grävsta 1a.

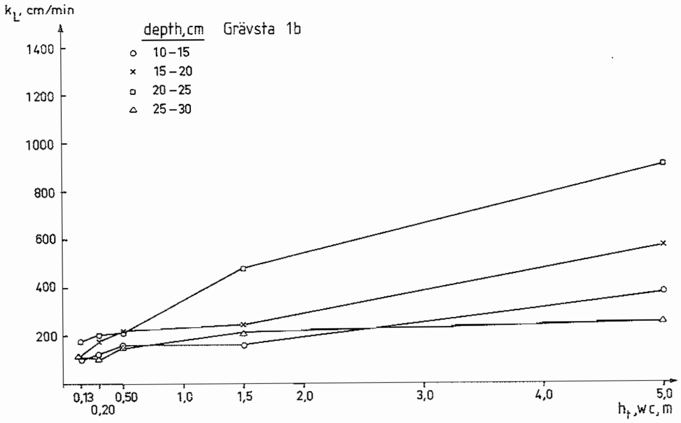


Figure 53. Air permeability (k_L) as a function of matric tension (h_t). Grävsta 1b.

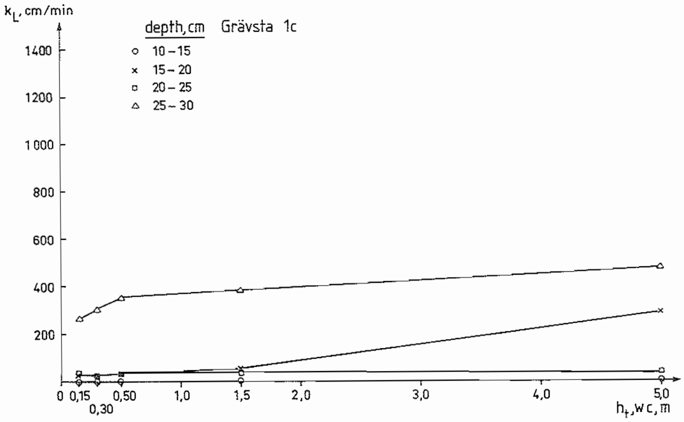


Figure 54. Air permeability (k_L) as a function of matric tension (h_t). Grävsta 1c.

This means of presentation improves classification of materials considerably. A general assessment of results shows systematic differences between the effects of various externally applied factors such as compaction, pan formation and grass production. Furthermore, differences can be seen in the reaction pattern of the different soil types.

Results from the heavy clay from Ultuna (U1) illustrate the function of some characteristic layers of a Central Swedish clay. In the topsoil, which has been compacted by heavy machinery, matric tension must be as high as 1 metre in order to open an effective transport system for air. This tension opens up a well developed network of air-filled canals throughout the sample.

In the very compact plough pan (25-30 cm), the structure of which has been destroyed by repeated compaction and smearing, the pore system opens very slowly with increasing tension. The plough pan seems to lack both a well developed macro- and micro-pore system.

The curve representing the upper part of the subsoil differs from both the other curves for this soil in that it begins at a relatively high value of relative diffusion which increases only slightly with increasing tension. This layer is characterized by a well developed macropore system enclosing rather compact aggregates.

Of the other soils from the Ultuna area, the clay layer U6, 10-15 cm appears as a positive extreme. Relative diffusion from this sample was greater than for all other samples from soil layers and different plots at every tension stage.

The low humus heavy clay from Kasby was the negative extreme, showing the lowest values for all soils in the Ultuna group. In ponded condition, this clay has diffusion properties indentical to those of the U1 plough pan. This indicates how poor the function of the Kasby soil is for aeration purposes. When the surface pan is broken up, values increase to around half those of the U1 topsoil.

In general, sorting of the above mentioned samples on the basis of increasing relative diffusion at a particular tension gave similar results to a visual assessment on the basis of structure.

This proved to be the case for Uddeholm soils also, in fact classification of these soils was extremely accurate. For all tension steps, soils occurred, according to increasing relative diffusion, in the order:- subsoil from open cultivation (1b, 1c), subsoil from grass (1a), topsoil from intensely compacted soil (1c), topsoil from grass and open cultivation (1a, 1b). This is the same order obtained when soils were classified on the basis of their visible structural features (Wiklert, 1972).

The situation in Grävsta soils is more complex, since results from certain levels distort the picture. The simple pattern found in Ultuna and Uddeholm soils is not apparent here, although similar trends can be observed. Certain irregularities can be explained by the fact that the soil consists of a relatively complex system, a medium clay with very good structure. Thus the effects of external cultivation factors such as grass are less obvious than in the uncomplicated Uddeholm silt. Thus relative diffusion between tensions 0.15 and 1.5 m w.c. was markedly lower in the surface layer of the grass (1a) than in all other levels of grass and cultivated soils alike (1a and 1b). At 5 m w.c. however, relative diffusion was highest in all layers of the soil under grass.

An irregularity also occurred in the compacted plot (1c) which otherwise reacted as expected and showed very low relative diffusion. The 15-20 cm level differed from this pattern by showing continually increasing relative diffusion within the observed tension range.

Deviations in the case of grass soils can be explained by the consolidating effect of grass on topsoil structure compared to a system of loosely layered, well structured aggregates. The irregular result from the compacted soil may have been due to the presence of a straw layer or similar discontinuity in the profile.

A characteristic common to all samples which reacted positively was that $dD/D_0/dh_t$ is great when h_t is small and that diffusion at h_t = 0.15 m w.c. is relatively large.

The reasonably good classification of diffusion material obtained when tension was chosen as the independent variable is probably due to the fact that characteristics of a pore system are clearly revealed by the system's reaction at a particular tension. At a tension of h_t, all pores with equivalent diameter $d = 0.3/h_t$ are waterfilled. All such pores are thus inactive as transport paths for diffusive or convective flow. All air passages opening into such pores are also inactive as transport paths. This means that size and shape distribution of the pore system is reflected in the transport coefficients if they are expressed as functions of matric tension. It may seem that air-filled porosity should provide an equally good basis for classification. This was obviously not the case in practice, probably because there is no general relationship between air - filled porosity values and diffusion and air permeability such as exists between tension and function. Different soil types and different cultivation histories result in different pore size distributions. These give rise to different water and air conductivities, water storage abilities etc. There is, however, no clear relationship between these characteristics and air content of the material. Function of a system at a particular air-filled porosity value cannot be predicted since nothing is known of the size and shape of the air-filled pores. At a particular tension, however, pore size distribution determines the water and air content of the soil.

Air permeability expressed as a function of matric tension gave a similar picture to that described above for diffusion. With some exceptions which do not affect the general trend, the order in which samples from different cultivation treatments on a particular soil can be placed was the same in both diffusion and air permeability investigations.

3.3.3. Relationship between diffusion and air permeability in a capillary model

Experiments indicated that there are certain relationships between diffusion and air permeability, regardless of whether the independent variable is air content or matric tension. In both cases air permeability can be said to extend or stretch the curves representing diffusion results. This means that it should be possible to use the relatively simple and cheap air permeability measurements to assess the effects of different cultivation factors, e.g. compaction, on the ecologically important process of diffusion.

Experimental results confirmed this hypothesis. It can also be shown theoretically that a relationship exists between diffusion and air permeability in a porous system. This has been discussed by Grable (1971), Edling (1973), and Eriksson (1982).

The starting point for calculations was Andersson's (1969) theoretical analysis of the k-value of a capillary system. Mass flow per unit time through a straight, cylindrical capillary with radius x is given by:

$$q = \frac{\pi x^4}{8 \eta} \cdot \frac{dp}{dz}$$

η = gas viscosity
$\frac{dp}{dz}$ = gradient

Assume that the frequency of pores in class x is $\varphi(x)$ (frequency function). Then the number of pores in class x is

$$\frac{f_a}{100} \cdot \frac{\varphi(x)}{100} \cdot A \frac{1}{\pi x^2}$$

where n = porosity of the system,
A = cross-sectional area of the system.

The mass of gas flowing through capillaries of class x is thus:

$$\frac{dq}{dx} = \frac{f_a \cdot A}{100} \cdot \frac{\varphi(x)}{100} \cdot \frac{x^2}{8 \eta} \cdot \frac{dp}{dz} = \frac{f_a \cdot A}{10^4 \cdot 8 \eta} \cdot \varphi(x) \cdot x^2 \frac{dp}{dz}$$

If R is the greatest and r the smallest air filled pore, the mass of gas flowing through the system per unit time is:

$$q_s = - \frac{f_a \cdot a}{8 \cdot 10^4 \cdot \eta} \cdot \frac{dp}{dz} \int_r^R \varphi(x)\, x^2\, dx \qquad (1)$$

In the same way, the mass of gas which diffuses through the system can be calculated.

Mass of gas diffusing per unit time through a straight, cylindrical capillary with radius x is given by:

$$q = -D_o \cdot \frac{M}{RT} \pi x^2 \cdot \frac{dp}{dz}$$

where D_o = diffusion coefficient in free air

M/RT = unit conversion term

dp/dz = partial pressure gradient.

According to the above, the mass of gas diffusing through a system is:

$$q_d = -\frac{f_a \cdot A}{10^4} \cdot D_o \cdot \frac{M}{RT} \cdot \frac{dp}{dz} \cdot \int_r^R \mathscr{S}(x)dx$$

The relationship between the mass flowing to diffusing through the system under the influence of partial or total pressure gradient can be written:

$$\frac{q_s}{q_d} = \frac{\frac{-f_a \cdot A}{8 \cdot 10^4 \cdot \eta} \cdot \frac{dp}{dz} \cdot \int_r^R x^2 \mathscr{S}(x)\, dx}{\frac{-f_a \cdot A}{10^4} \cdot D_o \cdot \frac{M}{RT} \frac{dp}{dz} \int_r^R \mathscr{S}x \, dx} \tag{2}$$

This simplifies to:

$$\frac{q_s}{q_d} = \frac{1}{D_o} \cdot \frac{RT}{8\eta \cdot M} \cdot \frac{\int_r^R x^2 \mathscr{S}(x)\, dx}{\int_r^R \mathscr{S}(x)\, dx} \tag{3}$$

This formula shows that in a defined model system, there is a specific relationship between air permeability and diffusion through the system. Equation 3 can be solved for various values of $\mathscr{S}(x)$. The simplest solution is for a rectangular system where $\mathscr{S}(x) = a$ for $r < x < R$. The relationship between convective and diffusive flow through a straight capillary for this value of $\mathscr{S}(x)$ and for a (partial) pressure gradient of dp/dz is obtained from equation 3.

$$\frac{q_s}{q_d} = \frac{C}{3} (R^2 + Rr + r^2)$$

where $C = \dfrac{RT}{D_o 8 \eta M}$

Two conclusions can be drawn from this simple example. Firstly, q_s/q_d increases parabolically from $C/3\ R^2$ to CR^2 as r, the smallest pore radius, increases from r to R for $\wp(x) = a$. Secondly, q_s/q_d increases parabolically according to $C/3 \cdot R^2$ as R increases from 0 to ∞ and when $r = 0$.

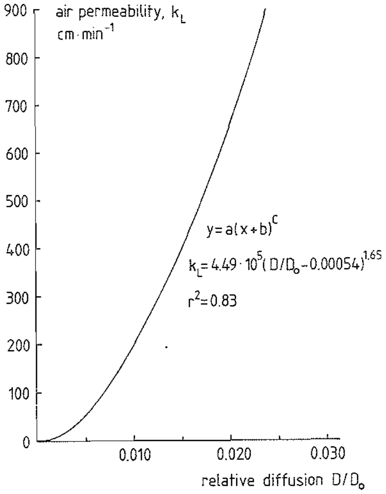
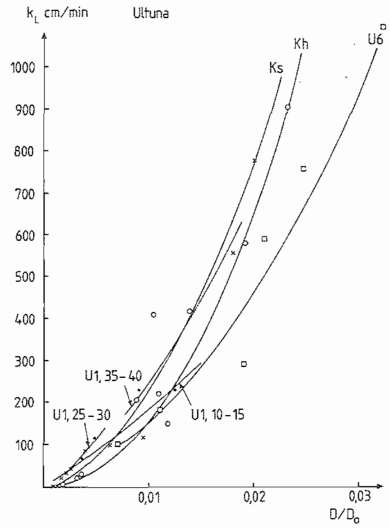
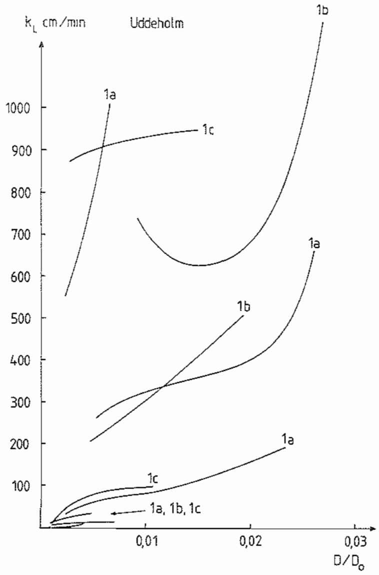


Figure 55. Relationship between relative diffusion, D/D_o, and air permeability, k_L, cm/min. Three soils, Ultuna, Uddeholm, Grävsta and totally 136 pairs of values. After Eriksson (1982).

In real systems, the situation is considerably more complicated. Eriksson (1982) has tested different equations on the results contained in this report (136 pairs of coordinates). An exponential function, $k_L = a(D/D_o + b)^c$, gave the best fit curve for data (see Fig. 55). If, however, all pairs of coordinates are plotted on a graph, a somewhat confused picture without obvious trends is obtained. On the other hand, if sampling sites and different treatments are considered separately, more regular groupings of points are obtained for some cases, see Figs. 56, 57, 58.





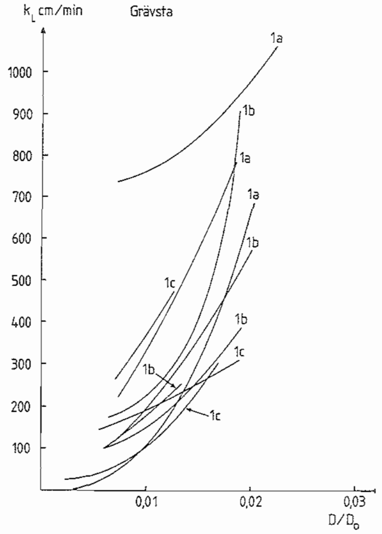


Figure 56, 57, 58.
Relationship between relative diffusion, D/D_0, and air permeability, k_L, cm/min. Ultuna, Uddeholm and Grävsta.

Air permeability measurement is thus a suitable soil physical method, unrelated to quantitative values and some methods, but related to qualitative values and other methods on aggregated soil.

On the single grain soil studied, air permeability investigations produced no significant or obvious trends from this point of view.

3.3.4 A theoretical model for soil air composition with depth in a soil profile

A theoretical model of soil air composition (with respect to O_2 and CO_2) can be developed if certain assumptions are made concerning transport coefficients and production and consumption of gases in the soil profile. The model was constructed on steady state conditions, i.e. derivatives of transport coefficients and activity with time = 0. In principle, diffusion was regarded as the sole transport process acting in the soil. In instances where mass transport became significant, it was considered as an addition to diffusion coefficients.

Discussions of this nature have previously been published by Romell (1922), van Bavel (1951), van Duin (1957), Wesseling (1960), Currie (1962), Edling (1973) and van Wijk (1980). The mathematical presentation below differs in some details from the research cited above, but is similar in principle.

3.3.4.1 Presentation of the model

Biological activity in the soil leads to the consumption of oxygen and production of carbon dioxide. In the following discussion, only O_2 consumption will be considered although the process of CO_2 production is identical in principle. The concentration of O_2 gas in air above the soil surface, C_0, can be measured in for instance percent, mm Hg or g/cm^3. P_0 was assumed to be independent of time, t, and to be constant in magnitude. Activity in the soil, a, was measured as O_2 consumption. a is a function of time, t, and depth below the surface, z, such that $a = a(z, t)$. The diffusion coefficient is also a function of depth and time, $D = d(z,t)$. All activity was assumed to cease at depth l where a watertable, compacted layer or inactive layer began.

The layer dz lies at depth z under the surface, see Fig. 59. O_2 content of soil air at this depth = p. Diffusion into the layer dz with time occurs according to Fick's law:

$$dQ_{in} = -D\ (z,t)\ \frac{dp}{dz}\ dt \qquad (1)$$

In the same layer, consumption of gas with time occurs according to:

$$dQ_{cons} = a\ (z,t)\ dz\ dt \qquad (2)$$

The mass of gas diffusing through layer dz is equivalent to the mass of gas consumed in the entire body of soil underneath:

$$dQ_{out} = \int_z^l a\ (z,\ t)\ dz\ dt \qquad (3)$$

According to the continuity equation

$$dQ_{in} = dQ_{cons} + dQ_{out} \qquad (4)$$

Therefore

$$-D\ (z,t)\ \frac{dp}{dz}\ dt = a\ (z,t)\ dz\ dt + \int_z^l a\ (z,t)\ dz\ dt \qquad (5)$$

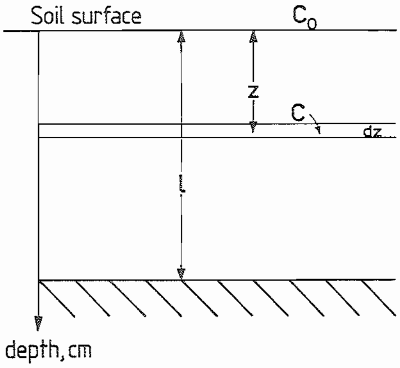


Figure 59. The soil profile, see text.

If conditions are assumed to be steady state (independent of t), equation 5 becomes

$$-\frac{dp}{dz} = \frac{1}{D(z)} \quad (a\ (z)\ dz + \int_z^1 a\ (z)\ dz) \qquad \text{(b)}$$

Oxygen deficit at depth z is then:

$$\underline{p} = -\int_0^z \frac{a\ (z)\ dz + \int_z^1 a\ (z)\ dz}{D\ (z)}\ dz \qquad \text{(c)}$$

This equation can be rewritten:

$$\underline{p} = -\int_0^z \frac{a\ (z)\ dz}{D\ (z)}\ dz - \int_0^z \frac{\int_z^1 a(z)\ dz}{D\ (z)}\ dz \qquad \text{(d)}$$

The first term becomes negligible in the second order and can be ignored. Thus:

$$\underline{p} = -\int_0^z \frac{\int_z^1 a\ (z)\ dz}{D\ (z)}\ dz \qquad \text{(e)}$$

This means that if O_2 content of free air is P_o, then the O_2 concentration of soil air at depth z is:

$$p = p_o - \int_0^z \frac{\int_z^1 a\ (z)\ dz}{D\ (z)}\ dz \qquad \text{(6)}$$

This equation has been produced in different forms by a number of authors, e.g. Romell (1922), van Bavel (1951), Currie (1962), Edling (1973) and van Wijk (1980). In some of these sources, the mathematics used are so advanced that the equation is difficult to apply in any particular case. This is one reason for seeking simple analysis of the terms involved and for investigating possible applications of the equation.

If a (z) and D (z) are constants, independent of depth z, solution of the equation is extremely simple and need not be discussed further here.

If a (z) is an easily integrated function while D(z) remains constant or vice versa, solution is also simple.

If a(z) and D(z) are different functions of depth, it must be possible to solve the term $\int_z^1 a(z)\ dz/D(z)$. In complex cases, however, terms which cannot be integrated may be obtained.

In many practical cases, it may be simpler to work within sets of intervals for which a(z) and D(z) are regarded constant.

The term $\frac{\int_z^l a(z)dz}{D(z)}$ gives the slope of the O_2 gas concentration curve, i.e. the gradient dp/dz at depth z. The term $\int_z^l a(z) dz$ gives the total activity below depth z in the soil profile per unit area of cross-section. This is an important factor. The significance of the term $\frac{\int_z^l a(z)dz}{D(z)}$ is, that the gradient arising in level z is the ratio of total activity under depth z to the actual diffusion coefficient in level z.

Consider a soil profile which can be divided into a number of homogenous layers 1, 2, 3 ....n_l, which end at depth l_1, l_2, l_3 ... l, see Fig. 60. Activity and actual diffusion coefficient in these layers are A_1, A_2, A_3 ... A_l and D_1, D_2, D_3 ... D_l.

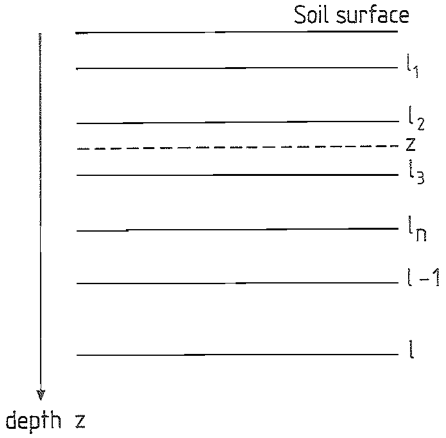


Fig. 60. The soil profile with different layers.

In the uppermost layer, the gradient at depth z is, according to the previous discussion:

$$\frac{dp}{dz} = \frac{\int_z^{z_1} A_1 dz + \int_{z_1}^{z_2} A_2 dz + \int_{z_2}^{z_3} + \ldots + \int_{z_{l-1}}^{l} A_l dz}{D_1}$$

For a depth z within the homogenous layer n, this can be written in general terms thus:

$$\frac{dp}{dz} = \frac{\int_{z}^{z_n} A_n dz + \sum_{n+1}^{p} A_{n+1} \, (z_{n+1} - z_n)}{D_n}$$

Oxygen concentration as a function of depth in the uppermost layer 1 is thus:

$$p = p_o - \int_0^z \frac{\int_z^{l_1} A_1 dz + \sum_{n=2}^{n=1} A_n \, (z_n - z_{n-1})}{D_1} \quad dz$$

or:

$$p = p_o - \frac{A_1(l_1 z - \frac{z^2}{2})}{D_1} - \frac{z}{D} \, (\sum_{n=2}^{n=1} A_n \, (z_n - z_{n-1}) \qquad (7)$$

For the case when the entire active layer is homogenous, i.e. when there is only one layer:

$$p = p_o - \frac{A_1}{D_1} \, (l_1 z - \frac{z^2}{2}) \;=\; P_o - \frac{z}{D} \, (l_1 A_1 - \frac{zA_1}{2}) \qquad (8)$$

For the depth l_1 we obtain in the former case (from eqn. 7)

$$p = p_o - \frac{l_1}{D_1} \left\{ \frac{A_1 l_1}{2} + \sum_{n=2}^{n=1} A_n \, (z_n - z_{n-1}) \right\}$$

and within the second layer we obtain:

$$p = p_o - \frac{l_1}{D_1} \left\{ \frac{A_1 l_1}{2} + \sum_{n=2}^{n=1} A_n(z_n - z_{n-1}) \right\} - \int_l^z \frac{\int_z^{l_2} A_2 dz + \sum_{n=3}^{n=1} A_n(z_n - z_{n-1})}{D_2} \, dz$$

and for the layer l_2, we obtain:

$$p_{l_2} = p_o - \frac{l_1}{D_1}\left\{\frac{A_1 l_1}{2} + \sum_{n=2}^{n=1} A_n(z_n - z_{n-1})\right\} - \frac{(l_2 - l_1)}{D_2}\left\{\frac{A_2(l_2 - l_1)}{2} + \sum_{n=3}^{n=1} A_n(z_n - z_{n-1})\right\}$$

For any layer n = 1, the equation is:

$$p_n = p_o - \sum_{n=1}^{n} \frac{l_n - l_{n-1}}{D_n}\left\{\frac{A_n(l_n - l_{n-1})}{2} + \sum_{n+1}^{1} A_{n+1}(l_{n+1} - l_n)\right\} \qquad (9)$$

This can be expressed verbally thus: The oxygen deficit arising in soil air within a layer at depth l can be described as "the sum of half the total activity in the layer in question and total activity in the soil underneath, multiplied by the ratio of thickness of the layer to actual diffusion coefficient".

This assumes that the soil layer above depth l can be considered homogenous and that soil air replenishment can be seen as a diffusion process. The relationship can be identified in equations 8 and 9.

This also means that from a known gradient the activity can be calculated if D is known.

3.3.4.2 Application of the model under various conditions

Using the equations derived, it is possible to study the consequences of various combinations of activity (A) and relative diffusion (D/D_o) values on the O_2 content of soil air.

Values for A were selected from several sources reported in the literature over the years giving results of similar order of magnitude (Romell, 1922; Currie, 1962; Greenwood, 1969).

As a basis for relative diffusion values, the result of D/D_o was taken from Ultuna 1, 10-15 cm at a tension of 2 m wc. This result was also doubled and used in the model.

It may be considered inappropriate for general calculations to choose one of the higher results for D/D_o obtained and, furthermore, to double this value. It will be shown, however, that it is meaningless to choose D/D_o values arising from lower matric tensions.

These do not, in according to this theory, contribute enough to soil ventilation by diffusion to supply developed crop needs.

The soil was investigated in the model to 1 m depth and it was assumed that no activity occurred below this depth. Three cases were examined:

1. constant activity and constant relative diffusion,
2. variable activity and constant relative diffusion,
3. constant activity and variable relative diffusion.

The assumed activity values in case 1 represented a total O_2 consumption of 6 and 12 l/m^3 and day. In case 2, the lowest and highest values of total activity were 6 and 12 l/m^3 and day respectively and all other values lay between these. The normal O_2 concentration of the air was assumed to be 2.88×10^{-4} g/cm^3.

Table 21. Case 1. Constant activity, constant relative diffusion.

Combin. number	A g/cm^3, s	D/D_o	D cm^2/s	z cm	p_o-p g/cm^3	p %
1	$1 \cdot 10^{-10}$	0.056	$1 \cdot 10^{-2}$	50	$0.375 \cdot 10^{-4}$	18.2
	$1 \cdot 10^{-10}$	0.056	$1 \cdot 10^{-2}$	100	$0.50 \cdot 10^{-4}$	17.3
2	$2 \cdot 10^{-10}$	0.056	$1 \cdot 10^{-2}$	50	$0.75 \cdot 10^{-4}$	15.5
	$2 \cdot 10^{-10}$	0.056	$1 \cdot 10^{-2}$	100	$1.00 \cdot 10^{-4}$	13.6
3	$1 \cdot 10^{-10}$	0.028	$0.5 \cdot 10^{-2}$	50	$0.75 \cdot 10^{-4}$	15.5
	$1 \cdot 10^{-10}$	0.028	$0.5 \cdot 10^{-2}$	100	$1.00 \cdot 10^{-4}$	13.6
4	$2 \cdot 10^{-10}$	0,028	$0.5 \cdot 10^{-2}$	50	$1.50 \cdot 10^{-4}$	10.0
	$2 \cdot 10^{-10}$	0.028	$0.5 \cdot 10^{-2}$	100	$2.00 \cdot 10^{-4}$	6.4

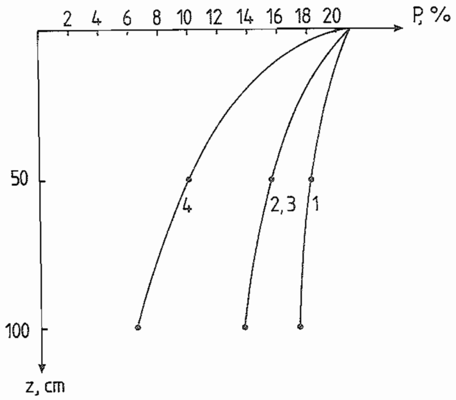


Figure 61. Case 1. Constant activity, constant relative diffusion.

Table 22. Case 2. Variable activity, constant relative diffusion. A_2 below 20 cm.

Combin. number	A_1 g/cm^3, s	A_2 g/cm^3, s	D/D_0	D cm^2/s	z cm	p_0-p g/cm^3	p %
5	3.10^{-10}		0.028	0.5 10^{-2}	20	$0.28 \cdot 10^{-4}$	18.9
		$0.5 \cdot 10^{-10}$			50	$0.48 \cdot 10^{-4}$	17.4
					100	$0.60 \cdot 10^{-4}$	16.5
6	6.10^{-10}		0.028	0.5 10^{-2}	20	$0.40 \cdot 10^{-4}$	18.0
		$0.5 \cdot 10^{-10}$			50	$0.60 \cdot 10^{-4}$	16.5
					100	$0.72 \cdot 10^{-4}$	15.7
7	3.10^{-10}		0.028	0.5 10^{-2}	20	$0.44 \cdot 20^{-4}$	17.7
		$1.0 \cdot 10^{-10}$			50	$0.84 \cdot 10^{-4}$	14.9
					100	$1.08 \cdot 10^{-4}$	13.1
8	6.10^{-10}		0.028	0.5 10^{-2}	20	$0.56 \cdot 10^{-4}$	16.8
		$1.0 \cdot 10^{-10}$			50	$0.95 \cdot 10^{-4}$	14.0
					100	$1.20 \cdot 10^{-4}$	12.2

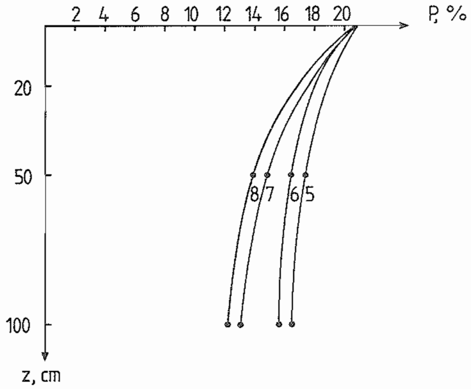


Figure 62. Case 2. Variable activity, constant relative diffusion.

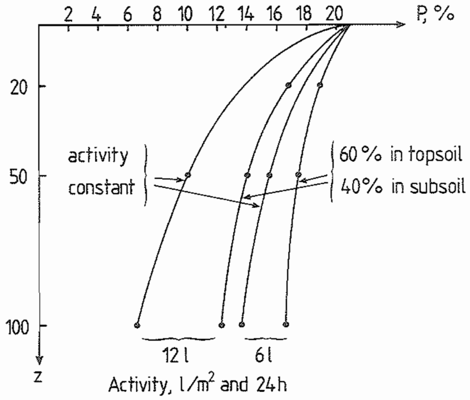


Figure 63. Total activity in the profile divided so that 60 % occurs in the topsoil and 40 % in the subsoil, compared to constant activity of the same total magnitude, constant relative diffusion.

Table 23. Case 3. Constant activity, variable relative diffusion. D_2 below 20 cm.

Combin. number	A g/cm^3, s	D_1 cm^2/s	D_2 cm^2/s	z cm	p_o-p g/cm^3	p %
9	1.10^{-10}	$0.5 \cdot 10^{-2}$	$0.25 \cdot 10^{-2}$	20	$0.36 \cdot 10^{-4}$	18.3
				50	$1.14 \cdot 10^{-4}$	12.6
				100	$1.64 \cdot 10^{-4}$	9.0
10	1.10^{-10}	$0.25 \cdot 10^{-2}$	$0.5 \cdot 10^{-2}$	20	$0.72 \cdot 10^{-4}$	15.7
				50	$1.11 \cdot 10^{-4}$	12.8
				100	$1.36 \cdot 10^{-3}$	11.0
11	1.10^{-10}	$0.25 \cdot 10^{-10}$	$0.25 \cdot 10^{-2}$	20	$0.72 \cdot 10^{-4}$	15.7
				50	$1.5 \cdot 10^{-4}$	10.0
				100	$2.0 \cdot 10^{-4}$	6.4

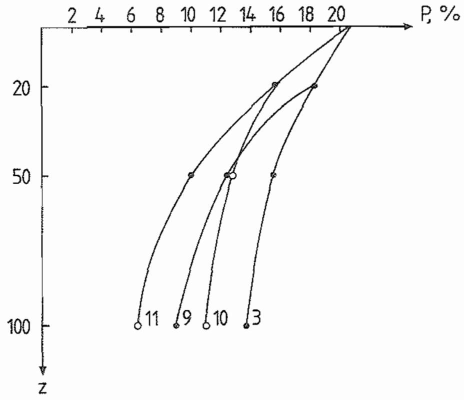


Figure 64. Case 3. Constant activity, variable relative diffusion.

The conclusion which can be drawn from these calculations is that the factors which contribute most to the development of a growth-inhibiting oxygen deficit are activity at depth in the profile and relative diffusion in the surface layers. The greater the activity deep in the profile, the greater the amount of gas which must be transported through the layers nearer the surface.

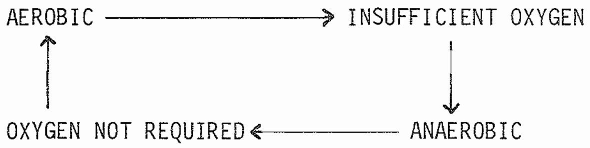
One can conclude that typical conditions causing serious O_2 deficiency in the soil profile occur when a well developed crop stand by rainfall or irrigation is supplied with a greater mass of water than can be drained away in one day (Stolzy & Flühler, 1978; Jackson, 1979). Some types of forest vegetation which form a water-holding layer of detritus which has poor drainage properties is supposed to produce the same effects (Nyholm, 1983).

Having considered the situation using experimentally obtained values of D/D_0, the O_2 content of soil air was then calculated for different soil types in particular drainage situations.

In all cases, activity was assumed to be 12 l/m^3 and day, 60 % of which occurred in the topsoil (0-20 cm) and 40 % in the subsoil (20-100 cm). Three layers were investigated in every soil - l_1, 0-20 cm = topsoil, l_2, 20-30 cm = plough pan and l_3, 30-100 cm = subsoil. Matric tension in the l_1 and l_2 layers was assumed to be 100 cm wc and that in the l_3 layer 50 cm wc. Thus we obtain for Ultuna 1, l = 100:

l_1	A_1	D_1	Deficit, g/cm^3
20	6.10^{-10}	0.012x0.189	$1.23 \cdot 10^{-4}$
l_2	A_2	D_2	
30	1.10^{-10}	0.004x0.189	$0.99 \cdot 10^{-4}$
l_3	A_2	D_2	
100	1.10^{-10}	0.01x0.189	$1.29 \cdot 10^{-4}$
		Total	$3.51 \cdot 10^{-4}$

As the O_2 content of free air is $2{,}88 \cdot 10^{-4}$ g/cm^3, the O_2 content is sufficiently low to inhibit plant growth already in the plough pan. If activity were half that given above, the deficit at 100 cm depth would be 1.75 g O_2/cm^3. This means that O_2 content of soil air stabilizes at 6.7 %. It is obvious that activity in this soil cannot increase under the conditions given and that anaerobic conditions occur locally or widely at various depths in the profile. Equilibrium for possible activity in the soil can be described according to the following model:



Aerobic and anaerobic processes occur simultaneously, a factor which is important from the economic and environmental point of view in the case of nitrogen losses through denitrification. This has also been pointed out by a number of authors, see for example the review by Stolzy & Flühler (1978).

It is important to note that the model was applied to what is regarded as a good arable soil which was in this instance at field capacity. If a silt soil, Uddeholm,is chosen for study, it can be deduced from the high capillarity of this soil that there will be no significant activity in the subsoil due to low O_2 supply or replenishment.

Some interesting observations can be made in the case of Grävsta soil. If the same assumptions are made as were used previously for Ultuna soil, the following results are obtained for Grävsta lb (careful cultivation):

l_1	A_1	D_1	Deficit, g/cm^3
20	6.10^{-10}	0.01x0.189	$1.48 \cdot 10^{-4}$
l_2	A_2	D_2	
30	1.10^{-10}	0.015x0.189	$0.33 \cdot 10^{-4}$
l_3	A_2	D_2	
100	1.10^{-10}	0.01x0.189	$1.30 \cdot 10^{-4}$
		Total	$3.11 \cdot 10^{-4}$

Oxygen content at 100 cm depth was in this case -1.7% which is a meaningless value. At half the activity above, the oxygen content becomes 9.6 %, which is almost an acceptable value.

When similar calculations were carried out for Grävsta 1c (intensely compacted) the following results were obtained:

l	A	D_1	Deficit, g/cm^3
20	6.10^{-10}	0.003x0.189	$4.94 \cdot 10^{-4}$
30	1.10^{-10}	0.003x0.189	$1.32 \cdot 10^{-4}$
100	1.10^{-10}	0.008x0.189	$1.62 \cdot 10^{-4}$
		Total	$7.88 \cdot 10^{-4}$

Here, totally impossible values are obtained because of the excessive compaction which has been applied and which has almost destroyed the soil structure. In practice, this can be likened to situations where wet autumns are followed by too early spring cultivation of the soil producing very bad soil structure. This must be counteracted by increasingly intensive tillage, which means increased traffic and compaction of the soil and so on, in the vicious circle of soil compaction (Edling & Fergedal, 1972).

Finally, the effect of a surface pan was analysed in the model. By 'pan' is meant a dense, compacted surface layer. Pan formation in this aspect usually occurs shortly after spring cultivation in cold, wet weather conditions. In this model, water was assumed to remain in the soil, i.e. there was no drainage of water from the profile. Only the upper 10 cm were investigated. To evaluate whether O_2 supply is limiting in the surface pan, values from Kasby S (surface pan) were used:

l_1	A_1	D_1	Deficit, g/cm^3
10	6.10^{-10}	0.007x0.189	$2.27 \cdot 10^{-4}$

An O_2 content of 4.4 % is achieved in the lower layer of the surface pan, given the assumptions described previously. It should be noted that the activity value used here was very high. Using values calculated from Dasberg et al. (1966), an O_2 consumption rate of only

4 % of the value used above was obtained for germinating crop (wheat, sowing rate 200 kg/ha). If this is the only activity taking place in the soil, then no significant O_2 deficit will arise.

In a situation where a thick surface pan is combined with high activity in the surface layer, O_2 supply to the upper soil layers may be influenced. Bearing in mind that pan formation as seen by the farmer occurs most often when cold, wet conditions change rapidly to warm, dry conditions, the case described above is possible. The deciding factor is, however, the amount of activity occurring in the soil. It is uncertain how closely combined activity of the germinating crop and of soil microorganisms can approach the activity value assumed in the model.

In the literature (see Ch. 2), convection is not considered to have a direct influence on gas exchange in deeper layers of the soil profile. It is considered significant only in the upper layers (Romell, 1922; Penman, 1940). However, small additions to the gas exchange process near the soil surface have indirect effects throughout the profile, since the gradient required to cause diffusion can be reduced. This means that any mass flow through the surface layer becomes very significant for the composition of soil air at depth. This conclusion is in agreement with Curries (1972).

Three factors causing mass flow are normally active - air pressure variations, diurnal temperature variations and plant transpiration. The effect of wind and movements of warmed air vary greatly with stage of crop development, soil conditions and time and these effects are difficult to distinguish.

It is possible to make a simple summary of the three factors mentioned above. Consider a surface area of 1 m^2, a watertable at 1 m depth, a total air-filled porosity in the profile of 10 vol. % and an O_2 consumption rate of 10 l/day. If air pressure varies by 10 mm Hg per day, temperature by 10^oC per day and evapotranspiration is 3 mm/day, the effects of these variations are as follows if the changes are taking place once and in one direction:

Effects of P and T changes are determined by the universal gas law. Observe an air column of 100 cm in the soil and changes are expressed in the form of compression or contraction of the air column. Then:

$$V_{P_o} \cdot P_o = \text{constant} = V_{P_1} \cdot P_1$$

$$\therefore \quad V_{P_1} = \frac{V_{P_o} \cdot P_o}{P_1} = C \frac{P_o}{P_1}$$

$$V_{P_1} = C \frac{760}{770} = 0.987 \text{ C}$$

Thus the volume of the air column decreases by 1.3 % or in this case is reduced by 1.3 cm.
For temperature variations:

$$\frac{V_{T_o}}{T_o} = \text{constant} = \frac{V_{T_1}}{T_1}$$

$$\therefore \quad V_{T_1} = \frac{V_{T_o} \cdot T_1}{T_1} = C \frac{T_1}{T_o}$$

$$V_{T_1} = C \frac{283}{293} = 0.966 \text{ C}$$

Thus the volume of the air column decreases by 3.4 % or in this case is reduced by 3.4 cm.

Evapotranspiration of 1 mm water per m^2 soil surface means removal of 1 l water and its replacement by 1 l air, i.e. 1 % of the soil air is replenished. The evapotranspiration rate of 3 mm/day assumed in this example means therefore that 3 % of the air column is drawn into the pore space. In total, 7.7 % of soil air is replaced per day in this example. This represents 7.7 l air which contain 1.6 l pure oxygen. In other words, with the assumptions given above, 1/6 of soil oxygen demand is supplied by mass flow, apart from the effects of wind and warmed air movements.

It should also be noted that as air content and depth of the aerated soil profile increase, the amount of air brought into the pores by air pressure and temperature variations also increases. The contribution made by evapotranspiration on the other hand is not directly influenced by the air content of the soil profile. This means that in damp weather, when air content of the profile falls and evapotranspiration is low, mass transport is least significant.

Mass transport is thus a rather erratic means of soil air replacement, even if it is at times very active. On the other hand, conditions for soil air replenishment by mass transport are almost ideal when pan formation occurs at the same time as crop establishment. The wind sweeps over the bare soil surface, temperature variations at the soil surface are not modified by vegetation and the desired biological activity occurs entirely under the soil surface as seed germination and growth of primary roots and cotyledons. In other words, it is not realistic to explain the effects of pan formation on crop establishment in terms of reduced oxygen supply to the seed.

It is important to know the minimum relative diffusion value necessary so that the O_2 content of a soil does not fall below the 10 %, generally regarded as critical. This is difficult to ascertain in natural soils because of variations with depth. To obtain an indication of the order of magnitude required, the simple case of the soil profile as a single entity can be considered. Soil air replenishment is then exclusively a diffusion process. From equation 8 applied to a 100 cm deep soil profile with total activity of 6 l/day:

$$1.37 \cdot 10^{-4} = \frac{100}{x} \left\{ \frac{100 \times 1.0 \cdot 10^{-10}}{2} \right\}$$

$$x = D = 0.36 \cdot 10^{-2}\ \mathrm{cm^3/s}, \qquad D_0 = 0.189\ \mathrm{cm^3/s\ at\ 10°C}$$

$$\therefore \quad D/D_0 = \frac{0.36 \cdot 10^{-2}}{0.189} = 0.019$$

In the case of activity of 12 l/day:

$$1.37 \cdot 10^{-4} = \frac{100}{x}\left\{\frac{100 \times 2.0 \quad 10^{-10}}{2}\right\}$$

$$x = D = 0.73 \cdot 10^{-2}\ \mathrm{cm}^3/\mathrm{s}, \qquad D/D_0 = 0.039.$$

From these calculations, the critical relative diffusion value for the example given is between 0.02 and 0.04. Since the activity distribution chosen in this example is unrealistic, these values of D/D_0 are too high (see Fig. 63). If activity is distributed so that 60 % occurs in the topsoil (0-20 cm) and 40 % in the subsoil (20-100 cm), the deficit at 100 cm depth will be about 60 % of that obtained with the homogenous activity distribution (see above). This means that the critical relative diffusion values would be decreased to 0.012 and 0.024. For general calculations, 0.02 can probably be accepted as a reasonable value for critical relative diffusion.

3.3.5 A theoretical model for O_2 transport from soil air into the root

Gas transport between soil air and the root occurs by diffusion through the water film which covers the active root. Simple models of conditions inside and immediately surrounding the root have previously been presented by Lemon (1962), Lemon & Wiegand (1962) and Kristenson & Lemon (1964).

The root and its surrounding water film are likened to two concentric cylinders (see fig. 65).

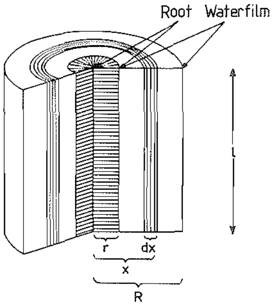


Figure 65.
Model of the root with radius = r and its surrounding water film with radius = R-r.

O_2 consumption inside the root = a, g/cm^3 and second. O_2 concentration in water at the water/air interface = C_w, obtained from Henry's law $P = K \cdot C_w$ where P = partial pressure of oxygen at the surface of the liquid and K = solubility of oxygen. The diffusion coefficient of O_2 in water = D_w. The following expression (after Fick's law) describes the mass diffusing through a cylindrical section of the water film:

$$-a \cdot \pi \cdot r^2 \cdot 1 = -D_w \cdot 2\pi \cdot x \cdot \frac{dC}{dx} \cdot 1$$

Integration, expansion and solution for x = r gives:

$$C_r = C_w - \frac{a\ r^2}{2\ D_w} \cdot \ln \frac{R}{r} \qquad (1)$$

In the root, the situation is as follows:
Consider a layer of thickness dx in the root. D_r is the diffusion coefficient of O_2 in the root. Diffusion into the layer dx is -

$$dQ_{in} = -D_r \cdot 2\pi \cdot (x + \frac{dx}{2}) \cdot \frac{dC}{dx} \cdot dt$$

Diffusion out of the layer dx is -

$$dQ_{out} = -D_r \cdot 2\pi \cdot (x - \frac{dx}{2}) \cdot \frac{d}{dx} (C + \frac{dC}{dx} dx) \cdot dt$$

Consumption of O_2 within the layer dx is -

$$dQ_{cons} = a \cdot 2\pi \cdot x \cdot dx \cdot dt$$

and $dQ_{in} = dQ_{out} + dQ_{cons}$.

Expansion, deletion of second order infinitesimal terms, integration and determination of integration constants gives:

$$C = C_r - \frac{a}{4D_r} \cdot (r^2 - x^2) \qquad (2)$$

However, C_r is defined by equation 1. O_2 concentration in a root layer at radial distance x from the root centre is:

$$C = C_w - \frac{ar^2}{2D_w} \cdot \ln \frac{R}{r} - \frac{a}{4D_r} (r^2 - x^2) \qquad (3)$$

This equation can be solved for different variables and for different intervals. This provides information on the effects of varying O_2 concentration at the water film/soil air interface, water film thickness and root radius.

The example below shows solutions of R with respect to C_r for the condition $C_0 = 0$, i.e. maximal gradient is created between the water film and the centre of the root. This corresponds to an O_2 consumption rate within the root which must not be exceeded if all parts of the root are to be supplied with oxygen.

In calculations, the following numerical values of D_W, D_r, r and a were chosen:

$D_W = 2.10^{-5}$ cm^2s $D_r = 10^{-2}$ cm^2/s (Greenwood, 1970)

r = 0.1, 0.025, and 0.01 cm a = $2.7 \cdot 10^{-7}$ g/cm^3, s.

Results of calculations are shown in Table 24.

Table 24.

Root radius cm	Thickness of film, normal O_2 content, normal activity, cm	Thickness of film, half O_2 content, double activity, cm	Thickness of film quarter O_2 content, quadruple activity, cm
0.1	0.07	0.004	0.001
0.025	0.30	0.023	0.004
0.010	$8.87 \cdot 10^4$	0.54	0.017

The values obtained agree in order of size with those discussed of Armstrong (1979) and Wesseling (1974). However, size of numerical values chosen determines absolute values of results. Choice of values of D_r is disputed in the litterature, but nevertheless important for the absolute result.

Most actively growing roots have a radius of between 0.01 and 0.025 cm. It can be seen from Table 24 that a single minute root can grow deep into a body of water, since the diffusion coefficient of O_2 in water is large enough to supply the low O_2 needs of such a root. However, the root is covered by root hairs and surrounded by microorganisms and it is also probable that an actively growing root has a greater O_2 demand than the average calculated value. Furthermore one root is never alone, but surrounded by other roots.

As stated previously, O_2 concentration in soil air is lower than that in normal free air. The case of 'normal O_2 content, normal activity' described above is thus irrelevant. The case of 'quarter O_2 content, quadruple activity' is probably that which describes reality best. The results obtained in this case are also in line with those in the literature (Letey & Stolzy, 1967), maximal water film thickness is only a few tenths of a mm.

In view of these conclusions, saturated and unsaturated hydraulic conductivity of the soil would appear to be of great significance for aeration of the rootzone. Even a few days saturation of the rootzone by water causes damage to growth (Gur et al., 1979; Jackson, 1979; Purvis & Williams, 1972 and Stepniewski, 1977). It is therefore essential that the mass of water applied by rain or irrigation is quickly drained away and only water necessary to supply crop needs is retained in the rootzone. This means that root canals are quickly available for gas transport again.

With these theories in mind, important conclusions about soil aeration and root environment should be apparent in the wh_t diagram for a particular soil.

An "average" root has diameter 0.025 cm (Esau, 1953). That part of the pore system which is accessible to the root system has therefore diameter greater than 0.025 cm. The wh_t diagrams on page 6 show that for Grävsta 1, such pores occupy a volume of 6 vol. % or

13 % of the total pore volume. For Uddeholm 1, however, they occupy hardly 1 vol. % or 2,5 % of the total pore volume. In fact, the volume of pores available to roots is increased by root hairs, which have a diameter of approx. 0.001 cm. Since root hairs cannot exist without roots, this will not be considered further.

A very much simplified picture of the root environment in relation to soil aeration can be obtained by theoretically subdividing the pore volume mentioned above into two equal parts: 1/2 = volume of pores with Ø = 0.025 cm, 1/2 = volume of pores with Ø = 0.1 cm. The former are assumed to be root/air canals and the latter drainage/ air canals. The soils mentioned above have the following number of each type of pore per cm^2 of soil surface:

Fine pores/cm^2

Grävsta $$\frac{1 \cdot \frac{6}{100} \cdot \frac{1}{2}}{\pi \cdot \frac{(0.025)^2}{4}} = 60$$

Uddeholm $$\frac{1 \cdot \frac{1}{100} \cdot \frac{1}{2}}{\pi \cdot \frac{(0.025)^2}{4}} = 10$$

Coarse pores/cm^2

Grävsta $$\frac{1 \cdot \frac{6}{100} \cdot \frac{1}{2}}{\pi \cdot \frac{0.1^2}{4}} = 4$$

Uddeholm $$\frac{1 \cdot \frac{1}{100} \cdot \frac{1}{2}}{\pi \cdot \frac{0.1^2}{4}} = 2/3$$

This means that for the square centimetre observed in this simple model, the Grävsta soil had 60 root/air canals and 4 drainage/air canals. Uddeholm soil had only 10 root/air canals and hardly 1 drainage/root canals.

Although no quantitative conclusions can be drawn from this, this qualitative trend is clear: Grävsta soil has relatively smaller anaerobic areas and surplus water drains away relatively more quickly so that the root environment is favourable. In Uddeholm soil, anaerobic conditions will occur more frequently while surplus water can only be drained away with difficulty. Roots have very much better possibilities to exploit the Grävsta soil, than they have with the Uddeholm.

Visser (1977) uses the wh_t diagram or the pF curve for a similar analysis and developed the concept of Soil Aeration Capacity (SAC). This is defined as the total internal surface (calculated by a certain formula) of all pores which are airfilled at pF 2 or $h_t = 100$ cm wc.

According to the calculations and deductions presented in this section, roots move along the surface of soil aggregates and particles at the interface of the water and air phases. Visser's line of reasoning is thus logical and ought to be investigated further. From the foregoing discussion, the following conclusions can be drawn:

When O_2 consumption is low, due to either narrow root radius (r) or to low respiration rate (a), and O_2 concentration at the surface of the water film is high, maximal thickness of the water film on the roots is larger.

When respiration is rapid and there are a large number of thicker roots, the water film is extremely thin. In natural systems, according to this, thickness of the film seldom exceeds a fraction of 1 mm. Thus roots are obliged to live at the interface between air and water. In a partially or totally saturated soil profile this means that a greater or lesser area of the profile cannot be exploited by roots.

For a soil to have a complete network of roots, it must have a well developed network of air-filled pores so that roots can come in intimate contact with soil air in all parts of the profile. This should, in fact, be the case in the upper part of the subsoil of east-Swedish clays, which have a granular structure and in which roots grow on the aggregate surface in contact with soil air and with the aggregates water and nutrient reserves.

4. BASIC DATA ON SOIL AIR

Soil air must always be considered in view of the fact that its volume, composition and the rate of gas transport through it vary. It is, in other words, a dynamic system in every respect. Various combinations of the parameters mentioned above mean that it is difficult to set precise values for these parameters at which growth of higher plants is inhibited, i.e. boundary values.

In order that the concept of soil air as an ecological factor is not totally abstract, some rules of thumb must be derived from all available experimental results. The terms given below refer to a healthy, full crop stand on arable soil, with no extremes of temperature, precipitation or fertilization.

a) Volume of soil air should exceed 10 %,

b) 0_2 content of soil air should exceed 10 %,

c) relative diffusion should be greater than 0,02,

d) air permeability of the soil should be higher than 400 cm/min.,

e) mass of 0_2 consumed should be between 0 and 20 l/m^2 and day (equivalent to 0 to 1 mole/m^2 and day or 0 to 30 g $0_2/m^2$ and day). Average values of this are 10 l, 0.5 mole or 15 g, which should be considered a good general estimate.

f) Using the values above, the 0_2 content of soil air is sufficient for 24 hours after the soil surface has been sealed by, e.g. heavy rain (assuming 1 m deep profile, 10 % air, 10 % 0_2 and 10 l/m^2, day). For each mm of water transpired by the crop, one litre of air is taken in per m^2 of soil. This is equivalent in this case to 0.21 l 0_2. This means that under the conditions given, every mm of transpiration can provide 2 % of the daily 0_2 demand.

h) A combination of high activity at depth and poor ventilation in the surface layers results in low oxygen content in the lower layers of the profile.

5. CONCLUSIONS

The most important conclusions from the literature reviewed and the experimental and theoretical analyses described in this report can be summarized thus:

a) Soil air is an extremely dynamic system; its volume, composition and intrinsic transport processes vary. Soil air must be discussed on the basis of production and consumption of its components and its replenishment. This principle can be summarized in diagram form (Fig. 66). Good or poor oxygenation situations can occur through different O_2 consumption and ventilation conditions.

b) The most important factors in O_2 concentration as a function of depth are: distribution of activity in the profile and ventilation conditions in the upper soil layers. A combination of high activity at depth and poor ventilation at the surface leads to an O_2 deficit.

c) For the soil to be completely penetrated and interwoven with roots, it must have a well developed system of cracks and canals so that roots can extend everywhere while remaining in close contact with the soil air and with soil reserves of water and nutrients.

d) Convection, or mass transport, can contribute to decreased O_2 concentration gradient in the surface layers and thus to higher O_2 concentration at depth in the soil. Mass transport can therefore contribute actively to improve oxygenation of the soil, even though it does not contribute to a great extent to the gas exchange process in the soil.

e) A complex pore system inhibits convection more than diffusion. However, there is a definite relationship between these two gas transport processes. This relationship can be determined exactly in a model. It can also be determined experimentally for aggregated soils but not for single grain soils.

f) Future research on the physics of soil air as an ecological factor should be aimed at obtaining experimental evidence for the theories discussed in this report. More research is also needed to complete basic soil physical knowledge of soil air. The wh_t relationships should be developed for use as an international soil physical reference method to describe soil aeration conditions.

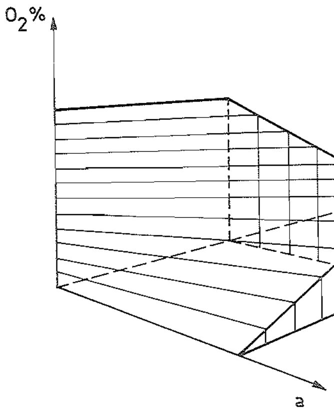


Figure 66. Schematic presentation showing soil air content as a function of O_2 consumption in soil (a) and replenishment conditions (k).

The central theory behind this report is that of Lundegårdh (1927), which stated that soil aeration status must be regarded on the basis of exchange possibilities and changes in this through consumption or production of gases in soil.

This view of soil air is summarized in Fig. 66, where a theoretical view of the relationship between O_2 content, O_2 consumption and ventilation conditions is presented diagrammatically.

6. REFERENCES

Ammon, G. 1880. Untersuchungen über die Permeabilität des Bodens für Luft. Forsch. Gebiete Agr. -Phys. 3, 209-241.

Amoore, J.G. 1961. Dependence of mitosis and respiration in roots upon oxygen tension. Proc. R. Soc. London. 154, 109.

Andersson, S. 1953. Markfysikaliska undersökningar i odlad jord. III. Om markens permeabilitet. Grundförbättring 6, 74-89.

Andersson, S. 1955 Markfysikaliska undersökningar i odlad jord. VIII. En experimentell metod. Grundförbättring 8, Spec. nr 2.

Andersson, S. 1961. Markens temperatur och värmehushållning. Stenciltryck nr 22. Inst. för Lantbrukets Hydroteknik. Uppsala.

Andersson, S. 1962. Markfysikaliska undersökningar i odlad jord. XIII. Några teoretiska synpunkter på vattenhaltskurvor, dräneringsjämvikter och porstorleksfördelningar. Grundförbättring 15, 51-108.

Andersson, S. 1969. Markfysikaliska undersökningar i odlad jord. XIX. Teoretiska modellstudier av kapillära systems k-värden som funktioner av porstorleksfördelning, bindningstryck och vattenhalt. Grundförbättring 22, 143-154.

Andersson, S. 1971. Markfysikaliska undersökningar i odlad jord. XXI. Om en apparat för modellstudier och demonstrationer av grundvattenströmningar, värmeströmningar och diffusion. Grundförbättring 24, 3-17.

Andersson, S. 1971. Markfysikaliska undersökningar i odlad jord. XXII. Härledning av och några kommentarer kring en differentialekvation för vattnets endimensionella strömning i omättad jord. Grundförbättring 24, 51-64.

Andersson, S. 1972. Markfysikaliska undersökningar i odlad jord. XXIV. Vattnets strömning till roten behandlad som ett hydrodynamiskt problem. Grundförbättring 25, 145-156.

Andersson, S. & Wiklert, P. 1970. Markfysikaliska undersökningar i odlad jord. XX. Studier av några markprofiler i Norrland. Grundförbättring 23, 3-75.

Andersson, W.B. & Kemper, W.D. 1964. Corn growth as affected by aggregate stability, soil temperature and soil moisture. Agron. J. 56, 453-456.

Armstrong, W. 1979. Aeration in higher plants. Adv. in Bot. Research 7, 226-331.

Bakker, J.W. & Hidding, A.P. 1970. The influence of soil structure and air content on gas diffusion in soils. Neth. J. Agric. Sci. 18, 37-48.

van Bavel, C.H.M. 1951. A soil aeration theory based on diffusion. Soil Sci. 72, 33-45.

van Bavel, C.H.M. 1952. Gaseous diffusion and porosity in porous media. Soil Sci. 73, 91-104.

Berglund, O. 1957. Ärtår - ett ekologiskt begrepp. Grundförbättring 10, 221-226.

Blake, G.R. & Page, J.B. 1948. Direct Measurement of Gaseous Diffusion in Soils. Soil Sci. Soc. Amer. proc. 13, 37-42.

Boynton, D. & Reuther, W. 1939. Seasonal Variation of Oxygen and Carbon Dioxide in Three Different Orchard Soils During 1938 and its Possible Significance. Proc. Am. Soc. Hort. 36, 1-6.

Brown, N.J., Fountaine, E.R. & Halden, M.R. 1965. The oxygen requirement of crop roots and soils under near field conditions. J. Agric. Sci. 64, 195-203.

Buehrer, T.F. 1932. The movement of gases through the soil as a criterion of soil structure. Arizona Agr. Exp. Sta. Tech. Bull. 39.

Buckingham, E. 1904. Contributions to our knowledge of the aeration of soils. U.S. Bur. Soils Bull. 25.

Bygg. 1972. Handbok för hus-, väg- och vattenbyggnad. Huvuddel 1 A.

Cannon, W.A. 1925. Physical features of roots with special reference to the relation of roots to aeration of the soils. Publs. Carnegie Instn. 368.

Carslaw, H.S. & Jaeger, J.C. 1959. Conduction of heat in solids. Oxford, Clarendon press.

Currie, J.A. 1960a. Gaseous diffusion in porous media. Part 1 - A non steady state method. Brit. Journ. Appl. Phys. 11, 314-317.

Currie, J.A. 1960b. Gaseous diffusion in porous media. Part 2 - Dry granular materials. Brit. Journ. Appl. Phys. 11, 318 -324.

Currie, J.A. 1961. Gaseous diffusion in porous media. Part 3 - Wet granular materials. Brit. Journ. Appl. Phys. 12, 275 - 281.

Currie, J. A. 1962. The importance of aeration in providing the right conditions for plant growth. J. Sci. Food Agric., vol 13, 380-385.

Currie, J.A. 1965. Diffusion within soil microstructure a structural parameter for soils. Journ. of Soil Sciences 16, 279-289.

Currie, J.A. 1970. Sorption and transport processes in soils. SCI monograph no. 37.

Currie, J.A. 1972. Personal communication.

Dasberg, S., Enoch, H. & Hillel, D. 1966. Effect of Oxygen and Carbon Dioxide Concentration on the Germination of Range Grasses. Agronomy Journ 58, 206.

van Duin, R.H.A. 1956. On the influence of tillage on the conduction of heat, diffusion of air and infiltration of water in soil. Versl. landbouwk. Onderz. 62. 7.

Edling, A.P.G. 1973. Några aspekter på jordluften - en undersökning av dess volym och omsättningsmöjligheter. Lic. avh. Avd f. lantbrukets hydroteknik, Lantbrukshögskolan.

Edling, A.P.G. & Fergedal, L. 1972. Modellförsök med jordpackning 1968-69. Rapporter från Jordbearbetningsavdelningen nr 31.

ELFYMA-tabellen. Tabeller-Formler-Nomogram inom områdena Matematik-Fysik-Elektroteknik. 4:e uppl. Sjöbergs förlag, Stockholm 1962.

Eriksson, J. 1982. Markpackning och rotmiljö. Inst. f. markvetenskap, avd. f. lantbrukets hydroteknik. Rapport 126.

Esau, K. 1953. Plant Anatomy. John Wiley & Sons Inc. New York & London.

Evans, D.D. & Kirkham, R.W. 1957. Measurements of air permeability of soil in situ. Soil Sci. Soc. Amer. Proc. 14, 65-72.

Fergedal, L. 1971. Jordpackning med traktor vid olika tider för vårsådd. Rapporter från Jordbearbetningsavdelningen. nr 26.

Flegg, P.B. 1953. The effect of aggregation on diffusion of gases and vapours through soils. J. Sci. Food. Agric. 4, 104-108.

Flühler, H. & Läser, H.P. 1975. A hydrophobic membrane probe for total pressure and partial pressure measurements in the soil atmosphere. Soil. Sci. 120, 85-91.

Gardner, W. & Chatelain, J. 1946. Thermodynamic potential and soil moisture. Soil Sci. Soc. Amer. Proc. 11, 100-102.

Gliński, J., Dobrański, B. Labuda, S. & Stępniewski, W. 1978. Soil oxygen conditions for the emergence of tomato seedlings in the soil. Pol. J. Soil Sci., 11, 81.

Grable, A. R. 1971. Effects of compaction on content and transmission of air in soils. Compaction of agricultural soils. 154-164. Am. Soc. Agric. Eng., S:t Joseph, Mich.

Greenwood, D, J.,1969. Effect of oxygen distribution in the soil on plant growth. Proc. of the 15th Easter School in Agr. Sci. Univ. of Nottingham. Butterworths, London, 202-223.

Greenwood, D, J. 1970. Distribution of carbon dioxide in the aqueous phase of aerobic soils. Journ. Soil. Sci. 21. 314-329.

Gliński, J., Stęphniewski, W. 1983. Soil Aeration and Its Role for Plants. CRC Press, Inc. Boca Raton, Florida.

Gur, A., Dasberg, S., Schbolnik, J., Sapir, E. and Peted, M. 1979. The influence of method and frequency of irrigation on soil aeration and some biochemical responses of apple trees. Irr. Sci. 1. 125-132.

Hagan, E.M. 1941. Movement of carbon disulfide vapour in soils. Hilgardia 14, 83-118.

Handbook of Chemistry and Physics 1969-1970. 50:th ed. The Chemical Rubber Co., Cleveland, Ohio.

Jackson, M.B. 1979. Rapid injury to peas by soil waterlogging. Journ. Sci. Food and Agriculture 30, 143-152.

Jensen, G. 1960. Effects of Temperature and Shifts in Temperature on the Respiration of Intact Root Systems. Physiologia Plantarum 13, 822-830.

Johansson, C.H. 1944. Fuktighetens adsorption och vandring i byggnadsmaterial. Teknisk Tidskrift 74, 1205-1216.

Johansson, C.H. 1948. Theoretical Investigation of the Effect of Capillary suction on Transfer of Moisture in Hygroscopic Materials Kungl. Tekniska Högskolans Handlingar nr 20.

Jonsson, B. & Lindström, J. 1985. Pers. comm.

Kennard, E.H. 1938. Kinetic Theory of Gases. Mc Graw-Hill Book Co, New York.

Kimball, B.A. & Lemon, E.R. 1971. Air Turbulence Effects Upon Soil Gas Exchange. Soil Sci. Soc. Amer. Proc. 35, 16-21.

Kmoch, M.G. & Hanus, M. 1965. Vereinfachte Methodik und Auswertung der Permeabilitätsmessung des Bodens für Luft. Z. Pflanzenernähr. Düng. Bodenk. 1, 1-10.

Kowalik, P. & Stepniewski, W. 1979. The significance of soil aeration for plants. Zescyty Problemowe Postepow Nauk Rolniczych, 220 s 61-79.

Krischer, O. 1963. Die Wissenschaftliche Grundlagen der Trocknungstechnik. 2:e Aufl. Springer Verlag. Berlin-Göttingen-Heidelberg.

Kristenssson, K.J. & Lemon, E.R. 1964. Soil Aeration and Plant Root Relations. III. Physical Aspects of Oxygen Diffusion in the Liquid Phase of the Soil. Agronomy J. 56, 295-301.

Landolt-Börnstein, 1969. Zahlenwerte und Funktionen. Zweites Band. 5 Teil. Bandteil a. Transportphänomene 1. Viskosität und Diffusion. Springer Verlag. Berlin-Heidelberg-New York.

Lemon, E.R. 1962. Soil aeration and Plant Root Relations. I. Theory. Agronomy J. 54, 167-170.

Lemon, E.R. & Wiegand, C.L. 1962. Soil Aeration and Plant Root Relations. II. Root Respiration. Agronomy J. 54, 171-175.

Letey, J. & Stolzy, L.H. 1967. Limiting distances between root and gas phase for adequate oxygen supply. Soil Sci. 103, 404-409.

Lundegårdh, H. 1927. Carbon dioxide evolution of soil and crop growth. Soil Sci. 23, 417-453.

Marshall, T.J. 1959. The diffusion of gases through porous media. Journ. Soil Sci. 10, 79-82.

Maxwell, J.C. 1868. On the dynamical Theory of Gases. Diffusion of Gases. Philosophical Magazine XXXV, ser. 4, 199-203.

Millington, R.J. & Shearer, R.C. 1971. Diffusion in aggregated porous media. Soil Sci. 111, 372-378.

Nyholm, N. 1983. Pers. comm.

Penman, H.L. 1940. Gas and vapour movements in the soil. I. The diffusion of vapours through porous solids. Journ. Agric. Sci 30, 437-462.

Purvis, A.C. & Williamson, R.E. 1972. Effects of flooding and gaseous composition of the root environment and growth of maize. Agronomy J. 64, 174-678.

Renk, F. 1879. Über die Permeabilität des Bodens für Luft. Forsch. Gebiete Agr.-Phys 2, 339-347.

Romell, L.G. 1922. Luftväxlingen i marken som ekologisk faktor. Medd. Stat. Skogsförsöksanst. 19, 131-359.

Rosenhane, S. 1663. Oeconomia. På uppdrag av Lärdomshistoriska samfundet utgiven av Torsten Lagerstedt. Almqvist & Wiksell, Uppsala 1944.

Russell, E.W. 1961. Soil Conditions and Plant Growth. Longmans, Green and Co, London.

Russell, E. J. & Appleyard, A. 1915. The atmosphere of the soil, its composition and the causes of variation. Journ. Agric. Sci. 7, 1-48.

Stefan, J. 1871. Über das Gleichgewicht u.d. Diffusion von Gasgemischungen. Sitzungsber. d. Math. Nat. Klasse der Kaiserlichen Akad. der Wissenschaften 63, 63-124.

Stefan, J. 1890. Verdampfung und Auflösung. I. Uber die Theorie der Diffusion der Gase. Sitzungsber. d. Math. Nat. Klasse der Kaiserlichen Akad. der Wissenschaften 98, 1424-1429.

Stępniewski, W.S. 1977. The effect of irrigation and increased mineral fertilization on the aeration status of a cultivated soil. Zeszyty problemome Postepow Navk Rolnizych, 197, 203-220. In english.

Stolzy, L. & Flühler, H. 1978. Measurement and prediction of anaerobisis in soils. Nitrogen in the environment. 363-426. Academic Press. New York, San Francisco, London.

Tacket, J.L. & Pearson, R.W. 1964. Effect of Carbon Dioxide on Cotton Seedling Root penetration of Compacted Soil Cores. Soil Soc. Amer. Proc. 28 , 741-743.

Taylor, A. 1949. Oxygen Diffusion in Porous Media as a Measure of Soil Aeration Soil. Sci. Soc. of Amer. Proc. 14, 55.

Troeh, F. Jabro, J. & Kirkham, D. 1982. Gaseous Diffusion Equations for porous materials. Geoderma 27, 239-253.

Wesseling, J. 1962. Some solutions of the steady state diffusion of carbon dioxide through soils. Neth. J. agric. Sci. 10, 109-117.

Wesseling, J. 1974. Crop Growth and Wet Soils. Drainage for Agriculture. No 17 in the series Agronomy, Americ. Soc. of Agronomy, Madison, Wisconsin.

van Wijk, A. 1980. Playing conditions of grass sports fields. Centre for Agricultural Publishing and Documentation, Wageningen, 124 s.

Wiklert, P. 1972. Personal communication.

Visser, J. 1977. Soil Aeration Capacity an index for soil structure, tested against yield and root development of apple trees at various soil treatments and drainage conditions. Plant and Soil 46, 221-237.

Wollny, E. 1893. Untersuchungen über die Permeabilität des Bodens für Luft. Forsch. Gebiete Agr.-Phys. 16, 193-222.

de Vries, D.A. 1950. Some remarks on gaseous diffusion in soils. Trans. 4th Int. Cong. Soil Sci., Amsterdam, Netherlands, 41-43.

Denna serie rapporter utges av Avdelningen för lantbrukets hydroteknik, Sveriges Lantbruksuniversitet, Uppsala. Serien innehåller forsknings- och försöksredogörelser samt andra uppsatser som bedöms vara lämpade för mer allmän spridning. Uppsatser av huvudsakligen internt intresse publiceras i serien Avdelningsmeddelande. Tidigare nummer i rapportserien kan i mån av tillgång levereras från avdelningen.

This series of Reports is produced by the Division of Agricultural Hydrotechnics, Swedish University of Agricultural Sciences, Uppsala. The series consists of reports on research and field trials and of articles or papers considered to be of general interest. Articles of mainly internal interest are published in a series of Divisional Communications (Avdelningsmeddelande). Earlier issues in the Report series can be obtained from the Division of Agricultural Hydrotechnics (subject to availability).

DISTRIBUTION:

Sveriges Lantbruksuniversitet

Avdelningen för lantbrukets hydroteknik

750 07 UPPSALA, Sverige

Tel. 018-171165, 171181