The influence of soil structure and air content on gas diffusion in soils

J. W. Bakker and A. P. Hidding

Institute for Land and Water Management Research, Wageningen, the Netherlands

Summary

A method to determine the diffusion coefficient of oxygen in undisturbed soil samples is described. Calculations are made of the errors involved when neglecting the storage term, the temperature and pressure variations, the oxygen consumption and the resistance of the ambient air.

The diffusion coefficient (D_s) is a function of air porosity (ε_g) and of soil structure. It appears that at values of ε_g below 0.20, the effect of a puddled soil surface becomes evident. In puddled soils D_s is at equal ε_g far below the D_s in non-puddled soils. These differences increase with decreasing ε_g.

Calculated are, for a tulip crop in spring time, the gradients of oxygen concentration in the soil atmosphere needed for the supply of oxygen to the plant roots. The calculation shows that a wet soil crust can limit the gas exchange by diffusion to a very great extent.

Introduction

Mechanical harvesting of root and bulb crops grown on heavy clay soils is often extremely difficult. For this reason the use of lighter soils for these crops is preferable. Heavy clay soils, however, may be improved for mechanical harvesting by lowering the clay content of the topsoil by means of ploughing up lighter soils or by mixing the topsoil with sand. A disadvantage of some lighter soils, however, is their unstable structure, which can easily be destroyed by rain or long lasting wet circumstances. In such a case soil particles often are washed down and fill up greater pores. This 'puddling' causes a high-density layer at the surface. These puddled surface soils have a very low permeability for water and for air, which may cause insufficient aeration and severe damage to crops.

In the Netherlands, up till now the structure of a topsoil in regard to aeration has been characterized by the air porosity measured at field capacity (pF 2.0) or during a wet period (Boekel, 1963; Boekel and Pelgrum, 1966). The relation between plant growth and air porosity is, however, different for different soils.

An improvement in the evaluation of the structure with regard to aeration could be the determination of the diffusion coefficient of gases in a soil, because the diffusion coefficient mainly determines the gas transport in the soil (Buckingham, 1904; Romell, 1922). The diffusion coefficient for gas in water is about 10^{-4} times that in air. It is evident that the main gas diffusion takes place through the air-filled pores of the soil

only. Besides volume, shape and especially continuity of the air-filled pores is important for gas diffusion. One may expect that in puddled soils more blocked pores are present than in soils with a good structure.
The influence of structure and air porosity on gas diffusion in surface soils was investigated, with special attention to wet puddled soils.

Review of literature

The gas diffusion in the soil is commonly given as a ratio D_s/D_a, D_s being the coefficient of diffusion in the soil (in $cm^2 . sec^{-1}$) and D_a being the coefficient of diffusion of the same gas in air at the same temperature and pressure. The ratio D_s/D_a depends on the soil only and not on the gas used for diffusion measurement (Penman, 1940).
The relationship between D_s/D_a and air porosity ε_g (cm^3 air . cm^{-3} soil) can be described over a wide range of air porosity values by an equation of the form

$$D_s/D_a = \gamma \, \varepsilon_g^{\mu} \text{ (Currie, 1960)} \tag{1}$$

For a short range of ε_g values, however, the relationship can be described by a linear equation of the form

$$D_s/D_a = a \, (\varepsilon_g - b) \tag{2}$$

In Table 1 several published data of these relationships are given.
It is remarkable that only few measurements of diffusion at low air porosities are available, the range that is of special importance. The factor b, which is interpreted as the volume of blocked pores (van Duin, 1956; Call, 1957) is higher in wet soils than in dry soils at the same ranges of ε_g. It should, however, be kept in mind that the factor b cannot be an exact indication of the amount of blocked pores only, because in a wet soil it increases with an increase in ε_g (Penman, 1940; Currie, 1960, 1961; Gradwell, 1961, 1965; see Table 1), which is not in accordance with the mentioned interpretation of b.
In disturbed soil samples (Table 1a and b) differences in diffusion between the different soils are small. The reason may be that sieving and artificially packing the samples results in a disappearance of the existing differences in structure of the original soils.
Investigations between samples with different structure (Table 1c) show a significant influence of structure on the relation between ε_a and D_s/D_a, especially at lower values of ε_g.

Determination gas diffusion coefficient

Principle
The determination of the gas diffusion coefficient of a soil sample is based on the measurement of the amount of gases which diffuse per unit time and per unit concentration difference.
The diffusion in one direction can be described as:

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Table 1   Relation between air porosity ε_g and gas diffusion in soils D_s/D_a

Author	Year	Material	ε_g	$D_s{}'D_a$
a. Disturbed dry material				
Penman	1940	sand, soil and glass spheres	0.18–0.50	0.66 ε_g
			> 0.50	1.0 $(\varepsilon_g - 0.25)$
Currie	1960	sand, glass spheres, carborundum	0.20–0.55	0.7 $(\varepsilon_g - 0.05$ à 0.1)
				0.9 $\varepsilon_g^{1.4}$
van Bavel	1952	sand	0.415	
		mixtures of sand and soil	0.315	} 0.6 ε_g
		mixtures of sand, soil and glass spheres	0.250	
b. Disturbed wet material				
Taylor	1949	sand (porosity 0.39–0.45)	0.04–0.15	0.55 ε_g
		ibid	0.15–0.45	1.0 $(\varepsilon_g - 0.09)$
Currie	1961	sand (porosity 0.38)	0.05–0.18	0.25 ε_g
		ibid	0.18–0.30	0.8 $(\varepsilon_g - 0.15)$
		ibid	> 0.16	6.0 $\varepsilon_g^{3.4}$
Taylor	1949	loam	0.1 –0.49	0.67 ε_g
Penman	1940	'Rothamsted subsoil'	0.15–0.5	0.66 ε_g
Call	1957	several soils	0.14–0.39	0.66 $(\varepsilon_g - 0.1)$
Gradwell	1965	clay loam	0.02–0.09	0.06 $(\varepsilon_g - 0.02)$
		ibid	0.09–0.16	0.3 $(\varepsilon_g - 0.07)$
Grable and Siemer	1968	silty clay loam (6.8 % organic matter)	0.2 –0.4	5.25 $\varepsilon_g^{3.36}$
c. Soils with different structure				
Blake and Page	1948	clay soils measured in situ		
		blocky structure	0.05–0.3	0.67 ε_g
		soil with more fine aggregates	0.15–0.3	1.3 $(\varepsilon_g - 0.1)$
Gradwell	1961	Kaolien aggregates	0.12–0.33	0.65 $(\varepsilon_g - 0.1)$
		ibid dense packing	0.22–0.33	0.7 $(\varepsilon_g - 0.17)$
		undistulbed samples from pasture topsoil (silt loam)	0.02–0.4	0.27 ε_g
Domby and Kohnke	1956	silt loam		
		1 with a soil crust 2 uncrusted	0.13	$D_{s1}/D_{s2} = 0.55$
		1 with a soil crust 2 uncrusted	0.25	$D_{s1}/D_{s2} = 1.0$

$$F \cdot A = \frac{dq}{dt} = - D_s \frac{dc}{dx} \cdot A \tag{3}$$

F = rate of transfer per unit area of section (g . cm^{-2} . sec^{-1})
A = area (cm^2)
q = amount of gas (g)
t = time (sec)
c = concentration (g . cm^{-3})
x = distance along the line of flow (cm)

Measurements can be made either at a constant concentration gradient, using for example saturated vapour of CS_2, acetone or ethylene dibromide (Penman, 1940 and Call, 1957), or by a non-steady state technique. In the latter case the diffusion coefficient

is calculated from the decrease in gas concentration with time in a cap placed over the soil sample, which at the other end is connected with the free atmosphere. Taylor (1949) and Currie (1960) applied such a technique using oxygen or hydrogen.

In our experiments the non-steady state method with oxygen was used.

Sampling and measuring

Undisturbed samples of the top 5 cm soil layer were taken with a large stainless steel cylinder 6 cm high and 7.5 cm radius. A thin layer of the lower surface, which might have been smeared during sampling, was cut off with a small sharp knife. Small gaps between soil sample and the cylinder were filled with a plastic glue. The sample was placed on a supporting gauge wire (0.7 mm) box with openings of 2.0 mm. A cap (height 10 cm, area 124 cm²) was placed on the sample ring and an air-tight connection was made with plastic kit (see Fig. 1). The oxygen concentration in the cap was measured with a polarographic oxygen analyser (Beckman, Model 777) of which the membrane-covered electrode was placed in the cap. The oxygen concentration was registrated continuously with a recorder for a period of one to several hours after filling the cap with oxygen. Before and after each diffusion run the oxygen concentration of the ambient air under the sample was registrated. To prevent high evaporation losses, the humidity of the air was maintained at a high level during measurement.

Each first run was made with the sample having its in situ air porosity; later runs were made after successively drying the sample gradually to raise the air porosity.

Calculation

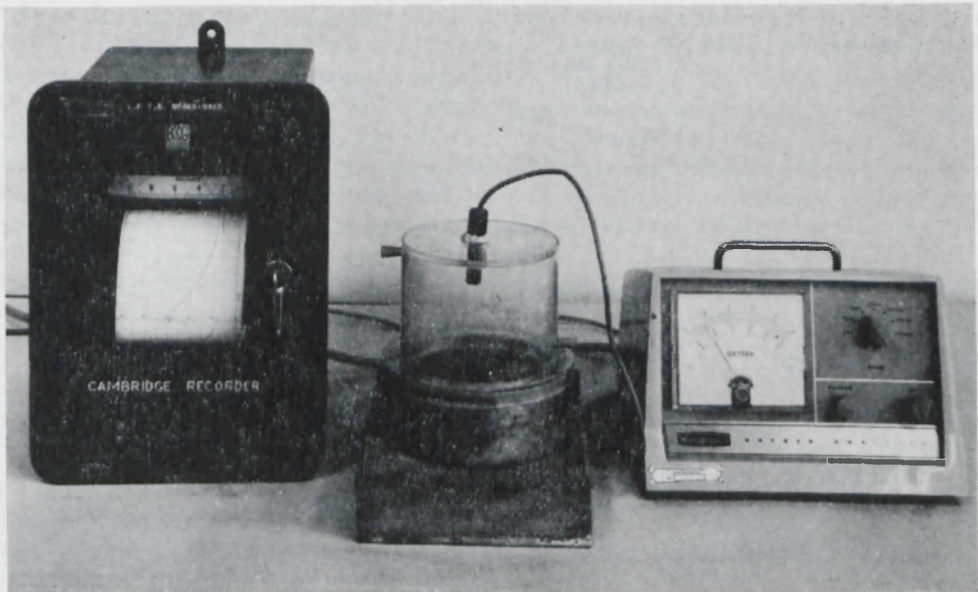Neglecting storage of gases in the sample, Eq. 3 becomes

Fig. 1   Apparatus for the determination of oxygen diffusion through soil samples

$$\frac{dq}{dt} = - D_s \frac{C_c}{h_s} . A \tag{4}$$

C_c = concentration difference between both ends of the sample (g . cm^{-3})
h_s = height of the sample (cm)
Concentration gradients of oxygen outside the soil sample are assumed to be negligible, compared with those inside the sample. The relation between concentration change in the cap and amount of gas diffusing through the sample can be given by

$$\frac{dq}{dt} = - \frac{dC_c}{dt} h_c . A \tag{5}$$

where $h_c . A$ = volume of the cap.

Combining Eq. 4 and 5 gives

$$\frac{dC_c}{dt} . h_c = \frac{D_s C_c}{h_s} \tag{6}$$

Solving Eq. 6 for the boundary conditions t = o, $C_c = C_{co}$ and t = t, $C_c = C_{ct}$ gives:

$$D_s = \frac{h_s . h_c}{t} 2.303 \log \frac{C_{co}}{C_{ct}} \tag{7}$$

According to Eq. 7 there is a linear relationship between t and log C_{ct} which is in accordance with the results of the diffusion measurements as represented in Fig. 2. Taking into account the storage of oxygen in the sample, one has to substitute in Eq. 3, according to the law of continuity, the equation

$$\frac{dF}{dx} = - \frac{1}{\varepsilon_g} \frac{dc}{dx} \tag{8}$$

which results in the general diffusion equation

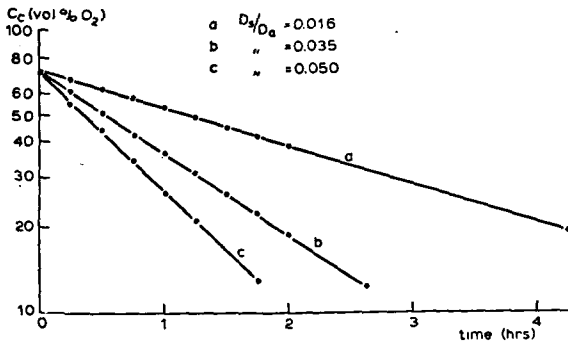$$\frac{dc}{dt} = \frac{D_s d^2 c}{\varepsilon_g dx^2} \tag{9}$$

Fig. 2 Oxygen concentration difference between both sample ends (C_c expressed in vol. %) on logarithmic scale plotted against time, during some diffusion runs

Solutions of this equation have been published for several problems of heat transfer. Mathematically the diffusion measurement here applied is analogous to the solution of Eq. 9 for heat transfer through a sheet of metal which is kept at a constant temperature at one side and at t = 0 is brought in contact with a known amount of well stirred fluid of deviating temperature at the other side, as is given by Carslaw and Jaeger (1950, p. 128, Eq. 8) and Currie (1960):

$$\frac{C_{ct}}{C_{co}} = \sum_{n=1}^{n=\infty} \frac{2\,\varepsilon_g}{h_c} \cdot \frac{e - D_s\,a_n^2 \cdot t \cdot \varepsilon_g^{-1}}{h_s\left\{a_n + \varepsilon_g^2\,h_c^{-2}\right\} + \varepsilon_g \cdot h_c^{-1}} \tag{10}$$

$$a_n = n^{th} \text{ root of } \alpha h_s \tan \alpha h_s = \frac{\varepsilon_g h_s}{h_c}$$

With the described apparatus and a value of $\varepsilon_g = 0.27$ at $D_s = 10^{-3}$ cm^2 . sec^{-1} the second term of the series is less than 1 % of the first term within 60 seconds. Therefore, only the first term of Eq. 10 was used. The solution becomes

$$D_s = \frac{\varepsilon_g}{a_t^2}\,2.303 \log \frac{C_{co}}{C_{ct}} \tag{11}$$

If diffusion coefficient calculated from Eq. 7 (D_{s1}) is compared with that from Eq. 11 (D_{s2}), it appears that neglecting of the storage term of oxygen in the sample causes too low a calculated value (D_{s1}) of D_s. An approximation of D_s for values $\frac{\varepsilon_g h_s}{h_c} \leqslant 0.5$, below which value all the experiments were carried out, is:

$$D_s = D_{s2} = \left(1 + 0.342\,\frac{\varepsilon_g h_s}{h_c}\right) D_{s1} \tag{12}$$

A convenient way for the calculation of D_s from oxygen concentration registrations is therefore the application of Eq. 7 and a correction for storage with Eq. 12. The maximum correction necessary in our experiments was 7 % at an ε_g of 0.34.

Discussion of method

Ambient air

The assumption that concentration gradients of oxygen outside the sample are of minor importance was tested on samples with a rather high D_s of 10^{-2} cm^2 . sec^{-1}. At such a high D_s one may expect the greatest influence of diffusion resistances in the air outside the sample. Measurements were carried out with and without stirring the air below the sample and in the cap by small fans. Stirring appeared to have no measurable influence on the diffusion coefficients determined.

The oxygen diffusion coefficient in air was also determined in our apparatus. Samples of air-tight material were used with holes parallel to the walls of the sample ring. At 20° C and an air pressure of 1013 millibar the coefficient measured was 0.203 cm^2 . sec^{-1}, with values given in literature of 0.200—0.216 cm^2 . sec^{-1} at 20° C and 1013 millibar (Smithsonian Physical Tables, 1956; Willey and Tanner, 1964).

Oxygen consumption
The oxygen consumption of the soil sample causes a supplemental decrease of oxygen concentration in the cap, resulting in a too high calculated value of D_s. At high microbiological respiration activities of the sample, the oxygen consumption will rise to about 2.10^{-6} g O_2 . cm^{-3} . $hour^{-1}$ causing a decrease of about 0.1 vol. $\%$ O_2 . $hour^{-1}$ in the cap. This means that real values of D_s/D_a of 0.05, 0.01 and zero are calculated to be, respectively, 0.05022, 0.0101 and 0.00009 (measurement lasting 5000 sec). Soil samples with an extremely low D_s value show a relative high error. Usually, however, oxygen consumption of such soils, which are often almost saturated, is very low (Kzrysch, 1963). In our sample series the oxygen consumption was in the order of 0.5×10^{-6} g O_2 . cm^{-3} . $hour^{-1}$, causing an error of D_s/D_a of 0.1 to 1 $\%$ at D_s/D_a values of 0.05 and 0.002, respectively.

Oxygen concentration measurement
During a run of 5000 sec, the error of the oxygen concentration measurement is about $\pm$ 0.5 vol. $\%$ O_2. This means an error in D_s of $\pm$ 5.10^{-5} cm^2 . sec^{-1}. For the determination of extreme low D_s values, a longer test run is necessary. The accuracy of the oxygen measurement can be held within $\pm$ 1 vol. $\%$ O_2 for 48 hours, which means an error in D_s of $\pm$ 0.4×10^{-5} cm^2 . sec^{-1}. For saturated samples, therefore, the applied method is unsuitable since D_s is in the order of 0.2×10^{-6} cm^2 . sec^{-1}. It is possible, however, to increase the sensitivity of the diffusion measurement by using thin samples and caps with a smaller volume.

Temperature and pressure
The effect of temperature and pressure on the diffusion coefficient can be represented by the equation (Smithonian Physical Tables)

$$D_1 = D_2 \left(\frac{T_1}{T_2}\right)^{1.75} \frac{p_1}{p_2} \tag{13}$$

where T = absolute temperature (°K) and p = pressure.

Temperature and pressure during the test runs were 20 $\pm$ 3°C and 1013 $\pm$ 25 millibar. This will result in a deviation of D_s of 2 $\%$ due to temperature, and of 2.5 $\%$ due to barometric pressure variation.
Apart from the direct effect on the diffusion, changes in temperature and barometric pressure during the measurement cause also a mass gas transport through the sample. The error in D_s ($\triangle D_s$) with decreasing temperature or increasing barometric pressure, when gas storage in the sample is neglected, may be given as

$$\triangle D_s = \frac{+ h_s h_c}{t} 2.303 \log (1 + x)$$

x = being the relative change of absolute temperature or pressure.
Normally the changes of barometric pressure do not exceed 2 millibar during a measurement of 5000 sec, 20 millibar in 24 hours and 25 millibar in 48 hours. The maximum change of temperature was 1.5°C in 5000 sec up to 3°C in 48 hours.
When, for example, x is 0.005 (a decrease of temperature of 1.5°C or an increase of

pressure of 5 millibar) during a measurement of 5000 sec, the error of D_s is 3.6×10^{-5} cm^2 . sec^{-1}.

Increasing temperature or decreasing barometric pressure will have an opposite but smaller effect.

Sample height

The determination of sample height (h_s) influences the accuracy of D_s as well as that of ε_g. With a careful measurement the error of h_s can be kept smaller than $\pm$ 0.5 mm, resulting in an accuracy of D_s of $\pm$ 1 % and of ε_g of $\pm$ 0.01.

Total error

The total error of the D_s values, as determined from the individual mentioned errors in the experiments, will be about $\pm$ 4, $\pm$ 7, $\pm$ 15 and $\pm$ 100 % for values of respectively 10^{-2}, 10^{-3}, 10^{-4} and 10^{-5} cm^2 . sec^{-1}.

A series of test measurements on dry gypsum samples (with D_s of about 2.10^{-3} cm^2 sec^{-1}) showed a deviation of $\pm$ 5 %.

Investigated soils

The top layer of several marine soils of different soil structure was investigated, especially in regard to the degree of puddling at the moment of sampling.

Some additional remarks on the soils mentioned in Table 2 may be given here:

a. The soils in North Friesland are often puddled very badly during the winter period. Samples 1 and 2 were taken in a field with spring wheat. The topsoil had been cultivated after the winter period. Samples 3, 4 and 5 were taken in winter wheat,

Table 2 Data of investigated soil samples

Location	Sample No	< 2 µm (%)	M [1] (µm)	Org. matter (%)	Porosity	Structure [2]
a. North Friesland	1, 2	7	42	3	0.50	—
	3, 4, 5	17	38	3	0.46	++
b. East Flevoland						
deep-ploughed (R46)	6	17	30	8	0.67	—
covered with sand (R47)	7	16	90	1.8	0.43	+
sand used		2.7	110			
original topsoil		30	30	3		
c. Schouwen, Haamstede	8	8–10	140	1.9	0.40	++
	9, 10	8–10	140	1.9	0.42	++
	11	8–10	140	1.9	0.45	—
d. Oudkarspel						
deep-ploughed [3]	12	12	50	2.3	0.52	++
deep-ploughed	13	12	50	2.3	0.54	+
original topsoil	14	25	37	1.6	0.51	+

[1] M = medium size of particles > 2 µm.
[2] ++ = surface puddled; + = surface slightly puddled; — = surface not puddled.
[3] Surface of sample artificially puddled.

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without tillage after winter, and showed severe symptoms of surface puddling.
b. In East Flevoland methods are investigated for improving the topsoil, in relation to possibilities of tillage. Sample 6 was taken from parcel R46 (deep-ploughed), with an excellent structure and a high content of organic matter in the topsoil. Sample 7 was obtained from parcel R47, which was covered with a sand layer of 8 cm and mixed to a depth of 18 cm. Some signs of puddling existed, probably because of the low vertical permeability of the layer under the topsoil, which had been excessively compressed by machinery during land levelling. Due to the low permeability of this layer, the topsoil remained wet for a long period.
c. In the region behind the dunes near Haamstede, dune sand and clay soils have been mixed, showing a very dense structure (Kuipers, 1960). Samples 8, 9 and 10 were taken from plots with a dense and puddled topsoil with signs of anaerobic processes and bad crop growth. Sample 11 was obtained from a plot with a better structure and a good crop growth.
d. In deep-ploughed fields near Oudkarspel heavy loamy topsoil was exchanged with a light loamy subsoil. The lighter soil had a better structure whereas the original heavy soil did show a lower air content and a lower water permeability in winter and spring.

Results and discussion

The results of the diffusion measurements as given on logarithmic scale in Fig. 3 clearly demonstrate the influence of soil structure on the relationship between air porosity and the diffusion coefficient. The diffusion coefficient in the samples of puddled soils are, with the exception of Sample 6 from East Flevoland, considerably lower than the diffusion coefficient in non-puddled soils at equal air porosity values. For the non-puddled topsoils the relation between ε_g and D_s apparently can be described by the equation $D_s/D_a = 0.85\ \varepsilon_g{}^{2.0}$ and for the puddled topsoils by $D_s/D_a = 2.0 \cdot \varepsilon_g{}^{3.0}$. At an ε_g of 0.07, D_s in puddled soils is about one tenth of D_s in non-puddled soils. At an ε_g of 0.14, D_s in puddled soils is one fourth of D_s of non-puddled soils only. At $\varepsilon_g > 0.20$, however, differences become negligible. These results are in agreement with those of Domby and Kohnke (1956) on soil samples with a soil crust.
At low air porosities all measured values of D_s are lower than the determinations published elsewhere (see the values in Table 1, taken from Taylor, 1949; Currie, 1961; Blake and Page, 1948; Gradwell, 1961). It is possible that the structure of the topsoil in our samples differs from that used by these authors. At values of ε_g above 0.20, the values of D_s of the non-puddled soils fit the relationship given by Gradwell for dense-packed Kaolien aggregates (see Table 1).
The oxygen concentration in the soil atmosphere, necessary for good crop growth, depends on the respiration activity of the roots, the diameter of the roots and the thickness of the water film surrounding them (Grable, 1968; Lemon, 1962).
The percentage ranges from < 2 vol. $^0/_0$ in dry soils and thin roots up to > 20 vol. $^0/_0$ under very wet conditions. Under wet conditions where a puddled surface inhibits the gas transport the drop of oxygen concentration in the water film will be considerable and the necessary oxygen concentration in the soil atmosphere therefore should be high, probably as high as the values of about 10 vol. $^0/_0$ given by Cannon (1925), Harris and van Bavel (1957), Geisler (1969).
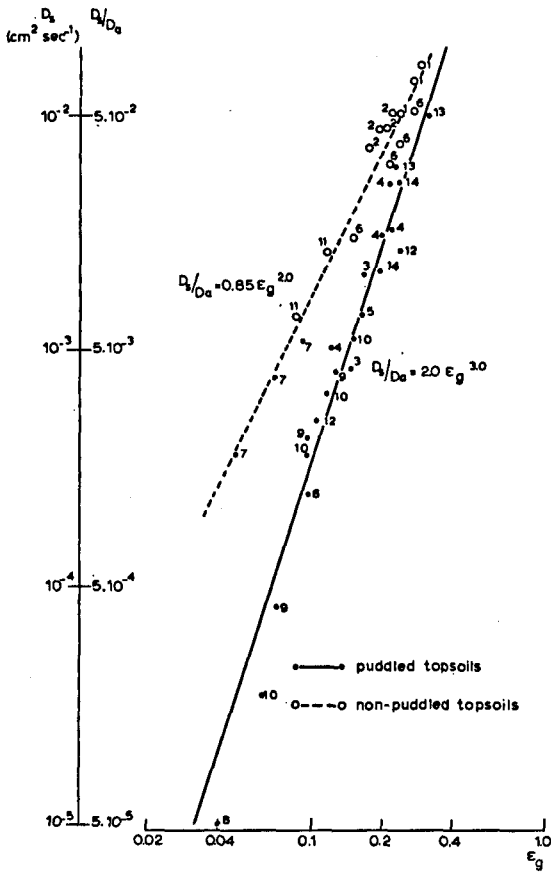
Calculations of the gas transport in the soil profile have been made with decreasing D_s and decreasing respiration activity at increasing depth (Wesseling, 1957; van Duin, 1956). Data on respiration activity in winter and early spring, however, are scanty, whereas for tulips, for example, this period seems to be of special importance. A simple model illustrates this problem on a soil with a surface layer of 5 cm thick without respiration activity, overlying a homogeneous active layer from 5 to 30 cm deep (Fig. 4).
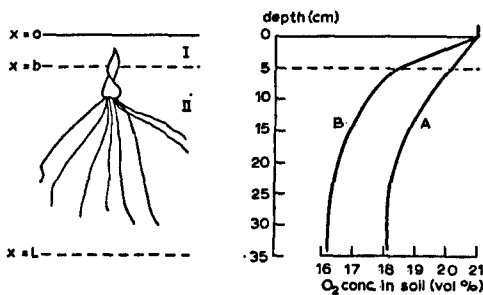

Fig. 4 Computed oxygen concentration in soil atmosphere of soils with a non-puddled (Curve A) and puddled topsoil (Curve B). ε_g of Layers 1 and 2 = 0.10; α of Layer 1 = 0; α of Layer 2 = 0.132 × 10-9 g O_2. cm^{-3}. sec^{-1}; $L = 35$ cm and $b = 5$ cm

The concentration gradient of oxygen in the inactive Layer 1 is given by the equation

$$dc = -\frac{\alpha}{D_s} L_1 dx \qquad (14)$$

and for the active Layer 2:

$$dc = -\frac{\alpha}{D_s} d (L_1 x_1 - \tfrac{1}{2} x_1^2) \text{ (van Bavel, 1951)} \qquad (15)$$

α = oxygen consumption (g O_2 . cm^{-3} . sec^{-1})
L_1 = L — b
x_1 = x — b
D_s = diffusion coefficient in the respective layer.

The value of α in Layer 2 is taken to be 0.132×10^{-9} g . cm^{-3} . sec^{-1}. With an ε_g in both layers of 0.10, D_s of non-puddled soil is in the order of 1.5×10^{-3} cm^2 . sec^{-1} (Fig. 3). The drop in concentration calculated for these conditions in Layer 1 will be 0.8 vol. % O_2 (Eq. 14) and in Layer 2 2.1 vol. % O_2 (Eq. 15), so that the total drop in concentration over 30 cm is 2.9 vol. %. With a puddled surface layer D_s of Layer 1 will be 0.45×10^{-3} cm^2 . sec^{-1} (Fig. 3). The drop in concentration in Layer 1 will be 2.8 vol. % causing a total decrease of 4.9 vol. % O_2.

At decreasing values of ε_g, D_s decreases quickly, especially in puddled soils. When ε_g is 0.07, for instance, D_s of a puddled and a non-puddled soil is 0.08×10^{-3} and 0.75×10^{-3} cm^2 . sec^{-1}, respectively. Assuming an unaltered α of Layer 2, the resulting total decrease in concentration will be 5.8 and 19.9 vol. % O_2 with a non-puddled and a puddled surface layer, respectively. These figures show that under wet conditions the structure of the surface layer may become an extremely important factor for the aeration of a soil.

References

Bavel, C. H. M. van, 1951. A soil aeration theory based on diffusion. *Soil Sci.* 72: 33–46.
Bavel, C. H. M. van, 1952. Gaseous diffusion and porosity of porous media. *Soil Sci.* 73: 91–104.
Bergman, H. F., 1959. Oxygen deficiency as a cause of disease in plants. *Bot. Rev.* 25: 418–485.
Blake, G. R. & Page, J. B., 1948. Direct measurement of gaseous diffusion in soils. *Proc. Soil Sci. Soc. Am.* 13: 37–42.
Boekel, P., 1963. Soil structure and plant growth. *Neth. J. agric. Sci.* 11: 120–127.
Boekel, P. & Pelgrum, A., 1966. Bodemfysisch onderzoek op enkele percelen tulpen in Noord-Holland. Groningen, I.B. Rapport No 4, pp. 8.
Buckingham, E., 1904. Contribution to our knowledge of the aeration of soils. *Bull. Div. Soils U.S. Dep. Agric.* 25, pp. 51.
Call, F., 1957. Soil fumigation. V. Diffusion of ethylene dibromide through soils. *J. Sci. Food Agric.* 8: 143–150.
Cannon, W. A., 1925. Physical features of roots with especial reference to the relation of roots to aeration of the soil. *Publs Carnegie Instn* 368, pp. 168.
Currie, J. A., 1960. Gaseous diffusion in porous media. I. A non-steady state method. *Br. J. appl. Phys.* 11: 314–324.
Currie, J. A., 1961. Gaseous diffusion in porous media. III. Wet granular material. *Br. J. appl. Phys.* 12: 275–281.
Duin, R. H. A. van, 1956. On the influence of tillage on the conduction of heat, diffusion of air and infiltration of water in soil (Dutch). *Versl. landbouwk. Onderz.* 62.7.

Domby, C. W. & Kohnke, H., 1956. The influence of soil crust on gaseous diffusion. *Proc. Soil Sci. Soc. Am.* 20: 1–5.

Geisler, G., 1969. Die Wirkung der Sauerstoff und Kohlensäurekonzentration in der Bodenluft auf Sprosz- und Wurzelwachstum von Wintergerste, Mais und Erbsen. *Bayer. landw. Jb.* 46: 259–278.

Grable, A. R. & Siemer, E. G., 1968. Effects of bulk density, aggregate size and soil water suction on oxygen diffusion, redox potentials and elongation of corn roots. *Proc. Soil Sci. Soc. Am.* 32: 180–186.

Gradwell, M. W., 1961. A laboratory study of the diffusion of oxygen through pasture topsoils. *N.Z. Jl Sci.* 4: 250–270.

Gradwell, M. W., 1965. Soil physical conditions of winter and the growth of rye grass plants. I. The effects of compaction and puddling. *N.Z. Jl agric. Res.* 8: 238–269.

Harris, D. G. & Bavel, C. H. M. van, 1957. Growth, yield and water absorption of tobacco plants as affected by the composition of the root atmosphere. *Agron. J.* 49: 11–14; 176–184.

Kuipers, S. F., 1960. A contribution to the knowledge of soils of Schouwen-Duiveland and Tholen accorling to the conditions prior to 1953 (Dutch). *Versl. landbouwk. Onderz.* 65.7.

Lemon, E. R., 1962. Soil aeration and plant-root relations. *Agron. J.* 54: 167–170.

Penman, H. L., 1940. Gas and vapour movements in the soil (I and II). *J. agric. Sci. Camb.* 30: 447–462; 570–581.

Romell, L. G., 1922. Der Bodenventilation als ökologischer Faktor. *Meddn St. Skogsförs Anst.* 19: 125–359.

Smithonian Physical Tables, 9th ed. New York, Smithsonian Institution, 1956.

Taylor, S. A., 1949. Oxygen diffusion as a measure of soil aeration. *Proc. Soil Sci. Soc. Am.* 14: 55–60.

Wesseling, J., 1957. Some aspects of the water government in agricultural soils (Dutch). *Versl. landbouwk. Onderz.* 63.5.

Willey, C. R. & Tanner, C. B., 1964. A steady-state method for laboratory measurement of the oxygen diffusivity of porous materials. *Proc. Soil Sci. Soc. Am.* 28: 733–737.