A SMALL LYSIMETER SYSTEM TO INVESTIGATE OXYGEN AND CARBON DIOXIDE PROFILES IN SOILS WITH WATER TABLES

by

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A dissertation submitted in accordance with the requirements for the degree

Magister Scientiae Agriculturae

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University of the Free State

Bloemfontein

South Africa.

January 2015

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DECLARATION

I declare that this dissertation hereby submitted by me for the in Magister Scientiae Agriculturae degree in Soil Science at the University of the Free State is my own independent work and has not previously in its entirety or part been submitted to any other University. I also agree that the University of the Free State has the sole right to the publication of this dissertation.

…… ……..

Signature Date

ACKNOWLEDGEMENTS

Taking a look back at the end of this journey, I can't help but wonder if I would be standing at the finishing line, which once seemed so far away, without the people who have been there with me along the way to give me direction, guidance, and encouragement. They have carried me through the times where it just seemed to hard to go on.

From the Department of Soil, Crop and Climate Sciences, University of the Free State, Prof. L.D. Van Rensburg my supervisor for his enduring patience, long hours during weekends and full commitment to my success, and unconditional efforts of providing the best possible academic training that have been the source of my conviction, confidence, and motivation to reach higher standards. As a good teacher and mentor, he saw my strengths and weaknesses, and pushed me to reach my full potential. Prof. C.C. du Preez my co-supervisor for his advice, patients and informative conversations. The staff members, especially, Dr. J Barnard, Dr. Z. Bello and Mrs. R. van Heerden for their assistance and help throughout the study, as well as Miss C. Britz who have been of great assistance along the way.

I am immensely grateful to Omnia Fertilizer for the bursary and assistance to complete my Magister Scientiae Agriculturae degree. Special gratitude to the team of Statistic Analysts, Dr. J. Habig, Mrs. S. Opperman and Mrs. V. Nolan for their advice and analysis of some of my data.

Lastly, my wife, Mrs. Ellie Schoonwinkel for her encouragement, motivation and loving support.

DEDICATION

I dedicate this thesis to my grandparents, Lourens en Lena van Rensburg that gave me the opportunity to study and experience student life to the full, as well as Nan Schoonwinkel who always encouraged me to be the best that I can.

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January 2015

In South Africa a huge amount of time and energy has been spent on evapotranspiration research over the past 30 years, mainly to predict the amount of plant available water needed to prevent crop stress. In the quest to conserve water losses due to transpiration, researchers tended to neglect the importance of soil-air concentrations in relation with soil water. Rising water tables caused by recharged groundwater through irrigation is one of the most important factors that change soil-air concentrations. For measurements, researchers found lysimeters more convenient due to the fact that they can simulate transient or constant water table conditions, which is otherwise very difficult to study in agricultural fields. The dissertation focuses mainly on the development of a monolith-lysimeter to measure soil water and soil-air regimes under rising water table conditions for different soils.

The research was conducted on five soils (sandy Hutton, loamy-sand Hutton, Bainsvlei, Sepane and Valsrivier) sampled in small (200 kg) lysimeters. A disturbed and undisturbed Bainsvlei soil was sampled at the experimental farm of the Department of Soil, Crop and Climate Science (University of the Free State) at Kenilworth in the Bloemfontein district while the remaining four undisturbed soils were sampled at the Orange-Riet River Irrigation Scheme. A total of 6 lysimeters was arranged in the glasshouse of the University of the Free State located on the main campus in Bloemfontein, South Africa.

The aim was firstly to develop and test a small weighing-lysimeter system for measuring soil temperature, soil water and soil air (oxygen and carbon dioxide) responses under water table conditions in a disturbed and undisturbed Bainsvlei soil monolith. These monolith lysimeters were used to characterize the influence of the lysimeter compared to *in situ* data that determined the accuracy of the method. After saturation of the soils with de-aired water from the bottom, drainage curves were determined by measuring weight-loss with both a weighing bridge and a capacitance DFM probe. Results showed that the shapes of the drainage curves for both

the disturbed and undisturbed soils were similar due to the similarity of the easily drainable pores. However, the water retention was significantly lower in the undisturbed soil compared to the disturbed soil. Furthermore, a water-table height control system was used for both raising, and to keep the water table steady at three heights while soil air measurements took place. According to the results it was found that an undisturbed soil is better to use for studying O_2 and CO_2 concentrations in soils. This conclusion is supported by the results which showed the sampling method with disturbed soil induced significantly higher O_2 and lower CO_2 concentrations, respectively compared to that of the undisturbed soil. Overall, the results indicated that the proposed small weighing-lysimeter system contribute towards the understanding of a very important subject, namely soil aeration.

Secondly, the monolith-lysimeter technique developed was used to evaluate five undisturbed soils in their O_2 and CO_2 response to a rising water-table over a period of 6 days. The water table was set at each height for six consecutive days for measurements where after it was raised to the next height. It was found that the O_2 and CO_2 concentration profiles were significantly influenced by the rise in water-table heights for the five soils under investigation. However, there were some distinct differences in the gas profiles observed between the sandy soils (sandy Hutton, loamy-sand Hutton and Bainsvlei) compared to the clay soils (Sepane and Valsrivier) due to differences in physical composition. The results further showed that time had significantly influenced O_2 and CO_2 concentrations over the 6-day period. As O_2 concentrations gradually decreased, CO_2 concentration gradually increased for all five soils. The only difference between the two soil groups was the intensity of respiration that resulted in lower O_2 and higher CO_2 concentrations for the clay soil group than for the sandy soil group over the 6 day period.

KEY WORDS: Oxygen and carbon dioxide concentrations; undisturbed soil; monolith; small lysimeter; water table heights

CHAPTER 1

MOTIVATION AND OBJECTIVES

1.1 Motivation and problem statement

There are an estimated 230 million ha of irrigated land in the world available of which 20% is seriously affected by waterlogging (Ghassemi *et al*., 1995). South Africa has an estimated 1.3 million ha of irrigated land for both commercial and subsistence agriculture (Perret, 2002; Bembridge, 2000). An estimated 20% (260 000 ha) of these irrigated soils have shallow water tables in or just below the rooting depth. Soils from most irrigation schemes become water logged before reaching their full potential due to the presence of frequently high water tables (Backeberg & Groenewald, 1995).

Reasons for waterlogged soil are poor water management, unsuitable soils, poor internal and external drainage, topography, inefficient irrigation systems, blocked drainage systems, periods of high rainfall, infrastructure deficiencies emanating from inappropriate planning and design (Perret & Touchain, 2002), poor operational and management structure and lack of technical knowledge (Bembridge, 2000).

With respect to technical knowledge, not much research has been done in South Africa on the quality and quantity of soil air during or near waterlogged situations. Waterlogging is a common problem in many soils especially irrigated soil and it has substantial adverse effects on the growth of crops (Belford *et al*., 1985; Mc Donald & Gardner, 1987). However, it is difficult to study these substantial adverse effects of waterlogging under field conditions because waterlogging events are transient and of variable duration. Waterlogging and water table height can be accurately controlled using columns of soil (lysimeter) removed from the field [Cannell *et al*., 1980a (as cited by Barrett *et al*., 1986).

As soil water and soil air are in relation to one another it is also important to measure the influence of soil water on soil air during waterlogged conditions. Changes in soil-air composition due to waterlogging can lead to a decrease in biological processes as well as the increase of gasses like methane (CH_4) and nitrogen oxide (NO) which is toxic for plants (Glinski & Stepniewski, 1985). Amundson & Davidson (1990) did a study on composition of soil atmosphere by analyzing extracted air samples and found that it is better to measure the composition of soil air than air volume alone, because it related more directly to problems that might exist. According to Payne & Gregory (1988) extracted air samples for determining soil-air

composition may not always be that accurate due to better-aerated pores in the soil or leakage into the sampling tubes. Analysis of extracted samples can comprise of many components such as O_2, CO_2, CO, C_2H_4, CH_4, N_2O, H_2, NH_3, NO and NO_2, although all components cannot always be determined in every situation (Glinski & Stepniewski, 1985). For this study, the main focus was placed on only two components measured directly from the soil with a hand apparatus (MultipleRAE IR gas monitor instrument).

A similar study where small lysimeters was used was done in the U.S.A. by Callebaut *et al.* (1982) with a sandy loam and loamy sand, repacked in Perspex cylinders with a diameter of 192 mm and a length of 900 mm. Soil columns were saturated from the bottom using distilled water until completely saturated. After soil was kept saturated for 1 week, water tables were maintained at various depths for 1 week at each depth. Several measurements took place which includes redox potential, water pressure head, O_2 and CO_2 concentrations, respectively. Changes in the soil aeration status, was found mainly at 300 mm or less above the water table. For gas sampling, two probes (diffusion chambers) and five bare micro platinum electrodes were inserted at different depths in the soil columns. For the determination of O_2 and CO_2 concentrations in gas samples, a gas chromatograph with two parallel columns both at 50°C and a thermal conductivity detector at 100°C was used (Callebaut *et al*., 1982).

Stotzky (1965) (reviewed by Glinski & Stepniewski, 1985) stated that every method of measurement is ultimate and ideal and that the research on methods never ceases. Methods used to measure soil respiration can be divided into laboratory techniques where all techniques involve the measurement of O_2 consumed by, or CO_2 evolved from, known quantities of soil incubated under controlled environmental conditions and field methods, based on measurement of CO_2 evolution and comprise chamber methods, method of CO_2 profile in the soil and micrometeorological methods (De Jong *et al*., 1979; reviewed by Glinski & Stepniewski, 1985).

Glinski & Stepniewski (1985) also reviewed techniques for the measurement of O_2 and found successfully used older techniques like the paramagnetic oxygen analyzers based on principles where only O_2 is attracted by a magnetic field, and also the polarography membrane covered sensors that require small samples that can be used in the field with the portable O_2 meter (Uhling *et al*., 1981). Some other methods to measure the composition of soil air are the gas chromatograph technique which helped to make measurements more reliable (Hillel, 1998; Glinski & Stepniewski, 1985), and the membrane covered electrodes techniques described by (Phene, 1986).

1.2 Hypothesis

High water tables will affect soil air quantity and quality negatively over time, and it will differ in different soil types.

1.3 Objectives of the study

It is well established that for most agricultural crops, water logging reduces oxygen and increases carbon dioxide concentrations that will eventually impact negatively on crop yield. The foregoing process is most closely associated as a result of the lack of oxygen for biological processes and plant use. In this study the general aim was to evaluate the effect of water table heights (water logging) on soil air composition and content in five different soil types. Specific objectives were to:

i) Develop and test a lysimeter for measuring soil temperature, soil water and soil air response under glasshouse conditions.

ii) Quantify O_2 and CO_2 profiles under water table conditions for irrigated soils.

1.4 Layout of thesis

This thesis consists of five chapters. Chapter one deals with the motivation and objectives of the study. Chapter two reviews the literature relevant to soil water-air relationships, influencing factors and measurement techniques. Chapter three contains the procedure for building a small lysimeter wherein the water-table heights can be controlled. The chapter deals also with procedures for sampling undisturbed soil monoliths and to equip them for extracting soil air at different depths in the profile. The chapter compares results from oxygen and carbon dioxide concentrations profiles in a disturbed versus undisturbed sandy loam soil. Research findings from this chapter were applied in Chapter 4 using a range of undisturbed soils to measure O_2 and CO_2 profiles under water table conditions. Thus, this chapter presents some results on oxygen and carbon dioxide profiles of the different soils. Chapter 5 was allocated to summarise the thesis and to make some recommendations for future research.

CHAPTER 2

LITERATURE REVIEW

2.1 Introduction

The main focus of this chapter is essentially to give a short review of the theoretical background relevant to soil air content and the factors affecting it, biological soil processes, soil water characteristics and the measurement of these parameters and processes. Different methods have been reviewed for soil air-water measurements to suit a glasshouse experiment. Special attention was given to the development of a lysimeter suitable for accurate soil air and water measurements, filled with undisturbed soil.

2.2 Aeration

2.2.1 Soil air content and composition

According to Drew & Stolzy (1996) the content of soil air is directly related to the bulk density of a soil in relationship with soil water, minerals and organic matter in the three-phase system (Figure 2.1 & Figure 2.2). This system is a schematic composition of a medium-textured soil at a condition considered optimal for plant growth. Soil air fills the part of the soil volume that is not occupied by water. The air content (V_a), often called soil air-filled porosity, is therefore equal to the difference between the total porosity of the soil (V_t) and its current water content (V_w) by volume (Glinski & Stepniewski, 1985). Soil water and soil air are related so that an increase in one is associated with a decrease in the other (Hillel, 1998).

$$V_a = V_t - V_w \qquad 2.1$$

Porosity is highly variable in clayey soils as the soil alternately swells, shrinks, disperses, aggregates, compacts and cracks (Hillel, 1998). In swelling soils the air content decreases in a nonlinear manner with an increase in water content both by weight and volume. In non-swelling soils, the air content decrease in a linear manner with an increase in water. The amount of air in a soil profile has an effect on both reserve oxygen and the oxygen diffusion coefficient that have a direct effect on root growth. This affects the rate of gas exchange in soil. The rate of gas exchange in soil, together with the rate of soil respiration determines the composition of soil air (Glinski & Stepniewski, 1985).

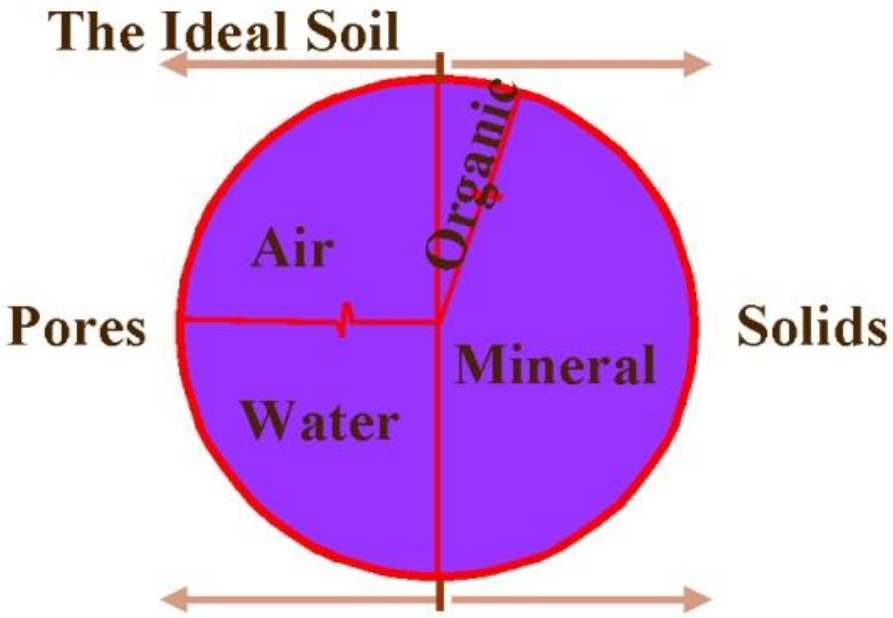


Figure 2.1 The three-phase system illustrating the relationship of air and water in composition of a fertile mineral soil (Drew & Stolzy, 1996).

When most of the air-filled pores are filled with water, anaerobic conditions will occur in the soil. Anaerobic conditions can occur during periods of prolonged rainfall, irrigation, in low-laying positions in the landscape (wetlands) and when impenetrable layers occur in the soil. During anaerobic conditions soil micro-organisms utilize other compounds as electron acceptors during respiration. Under these conditions gasses such as methane (CH_4), nitrogen oxide (NO), ethylene (CH_2=CH_2) and hydrogen sulphide (H_2S) will be produced instead of carbon dioxide (CO_2). These gasses are harmful and even toxic for most plants.

2.2.2 Soil density

The density of solids (P_s), in most mineral soils is about 2600-2700 kg m^{-3} where there is no air in the soil (Hillel, 1998). Calculations can be done based on Figure 2.2.

Mean particle density: $P_s = M_s/V_s$ 2.2

Where M_s = Mass of solids (kg) and V_s = Volume of solids (m^3)

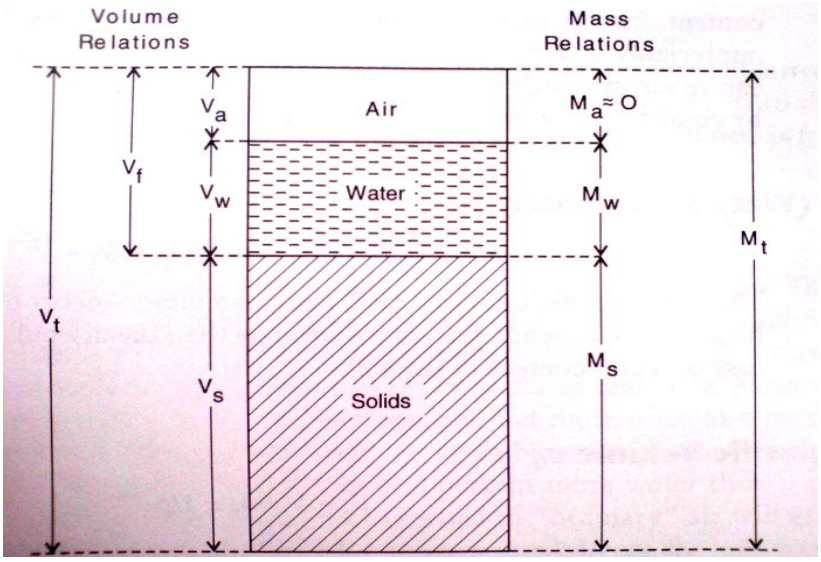


Figure 2.2 Schematic diagram of soil as a three-phase system (Hillel, 1998). V represents volume, M represents mass and the subscripts a, w, s, f and t are air, water, solids, fraction and total, respectively.

The dry bulk density expresses the ratio of the mass of solids to the total soil volume and can be estimated by:

$$P_b = M_s/V_t = M_s/(V_s+V_a+V_w) \qquad 2.3$$

Where V_s = Volume of solids (m^3) and V_a = Volume of soil air (m^3) and V_w = Volume of soil water. These solids and pores together (P_b) are always smaller than P_s and is about 1300 to 1350 kg m^3. The pore space in a soil is known as porosity (V_f):

$$V_f = (P_t - P_s)/P_s \qquad 2.4$$

The value of porosity generally ranges from 0.3 to 0.6 (30-60%). Fine textured soils tend to be more porous than coarse-textured soils, though the mean size of individual pores is greater in the former. As mentioned earlier, porosity is highly variable in clayey soils as the soil alternately swells, shrinks, disperses, aggregates, compacts and cracks. The total porosity of soils reveals nothing about the sizes or shapes of various pores in the soils (Hillel, 1998).

Soil wetness (water content) can be expressed in various ways relative to the mass of solids, total mass or volume of solids, or to the total volume or the volume of pores. Thus one of the indexes is defined as follows:

Mass wetness (θ_g) is the mass of the water relatively to the mass of the dry soil particles where the definition of dry soil refers to the mass when dried over a 24 hour period at 105°C in an oven.

$$\theta_g = M_w/M_s \qquad 2.5$$

At saturation, when all pores are filled with water, the water content will be higher in clayey than in sandy soils. Water content at saturation can range between 25 and 60% in different soils depending on the bulk density (Hillel, 1998).

2.2.3 Gas transport through a soil environment

The exchange of gasses between the soil and the atmosphere takes place under the influence of both pressure gradients (mass flow) and concentration gradients (diffusion flow). Exchange of considerable amounts of gasses such as oxygen and carbon dioxide is of most importance in the soil (Glinski & Stepniewsk, 1985; Hillel, 1998). Gasses transport by both these kinds of flow may take place in the soil and can occur by several mechanisms (Glinski & Stepniewsk, 1985; Payne & Gregory, 1988). With changes in temperature between various parts of the soils as well as in atmospheric pressure, the contraction and expansion of air within the pore space may cause some exchange between those various areas in soil profiles (Hillel, 1998).

There is a movement of air between the atmosphere and the soil. During the day, soil is warmer than the atmosphere and the soil gases expand and pass to the atmosphere rapidly. During the night the soil gets cooler than the atmosphere and the gases flow into the soil from the atmosphere. Boyle's law stated that any increase in the barometric pressure of the atmosphere should cause a compression and subsequent decrease in the volume of soil air, thereby allowing penetration of atmospheric air into soil pores. Changes in soil air pressure can also occur during tillage or compaction by machinery (Jury *et al*., 1991; Hillel, 1998).

Air blowing across a bare surface with a wind speed of 24 km h^{-1} can penetrate coarse sand to a depth of several centimetres. The infiltration of rainwater displaces soil air from the pores and enriches the soil by carrying "new" oxygen in dissolved form into the pores. Diffusion rather than convection is the more important mechanism of soil aeration although convection can significantly contribute to soil aeration, particularly at shallow depths and in soils with large pores (Jury *et al*., 1991; Hillel, 1998).

2.2.3.1 Mass flow

Pressure differences between soil air and the outer atmosphere induce convective flow into or out of the soil. The factors that cause mass flow in soil are changes in soil temperature, atmospheric pressure changes, wind and soil water (rain, irrigation, evaporation, movement of ground water table). Water that penetrates during infiltration causes displacement of soil air or can sometimes compresses soil air. The movement of shallow ground water table can push air upwards or drawing it downward and plant roots also extract soil water (Glinski & Stepniewski, 1985; Jury *et al*., 1991; Hillel, 1998; Jury & Horton, 2004).

Convection is similar to water flow in soil and the flow is proportional to the pressure gradient involved. Air is compressible and dependent on the density and viscosity. Gravity does not affect the gas flow in soil and gas is not attracted to mineral surfaces in soil. Jury & Horton (2004) stated that gas flow can be described by a formula similar to Darcy's Law:

$$J_c = -K_a\left(\frac{dP}{dz}\right) \qquad 2.6$$

Where J_c = air flux density (m s^{-1}), K_a = air conductivity (m s^{-1}), z = Distance (m) and P = air pressure in head unit. Recalling that the density of a gas depends on its pressure and temperature, the assumption can be made that soil air is an ideal gas in which the relation of mass, volume and temperature is given by the following equation (Jury *et al*., 1991; Hillel, 1998).

$$PV = nRT \qquad 2.7$$

Where P = Pressure, V = Volume, n = Number of moles of the gas, R = Universal gas constant per mole and T = Absolute temperature.

2.2.3.2 Diffusion

Concentration diffusion is the primary mechanism of gas exchange in a soil medium under normal field conditions. Transport of oxygen and carbon dioxide occurs partly in the gaseous phase and partly in the liquid phase where gas transport takes place in the direction of the decreasing concentration (Glinski & Stepniewski, 1985; Hillel, 1998). According to Jury & Horton (2004) both phases of the diffusion process can be described by Fick's law (Jury *et al*., 1991; Hillel, 1998).

$$J_g = -D\frac{\partial c}{\partial x} \qquad 2.8$$

Where J_g = Diffusive flux of a gas (mass diffusing across a unit area per unit time), D = Diffusion coefficient (area/time), c = Concentration (mass of a diffusing substance per unit volume) and $\partial c / \partial x$ = Concentration gradient.

In porous bodies, the diffusion coefficient depends on fractional volume of the continuous gas phase and it is not affected by the shape of a solid surface or distribution of pore size. The gaseous diffusivity (D) in bulk air varies with the molecular weight of the diffusing gas (normally higher for gases of lower molecular weight) and with temperature and air pressure (Hillel, 1998; Bird *et al*., 2001). Hillel (1998) stated that at atmospheric pressure and standard conditions of temperature (25°C), it varies between 0.05 and 0.28 cm^2 sec^{-1}, while Jury & Horton (2004) stated that under the same conditions it will range between 0.015 and 0.25 cm^2 sec^{-1}. Jury *et al* (1983) recommended using an average value of 0.05 cm^2 sec^{-1}. There are standard values for several gases in air and water (Table 2.1). In aggregated soils, gas diffusion takes place in interaggregated macro pores due to the fact that macro pores are readily drained (Jury *et al*., 1991; Hillel, 1998).

Table 2.1 Diffusion coefficient at standard temperature and pressure (Hillel, 1998)

	Diffusion coefficient (m^2 sec^{-1})
CO_2 in air	1.64×10^{-5}
O_2 in air	1.98×10^{-5}
H_2O vapour in air	2.56×10^{-9}
CO_2 in water	1.6×10^{-9}
O_2 in water	1.9×10^{-9}
N_2 in water	2.3×10^{-9}
NaCl in water	1.3×10^{-9}

Gas must diffuse through a longer path length to get from one point to another because the cross-section area available for flow can be reduced to a certain extent by solids and liquid barriers (Jury & Horton, 2004). Since this path of diffusion is then much smaller than the width of the pores, gaseous diffusivity is little affected by the distribution of pore size or the shape of the solid surface (Hillel, 1998). On the other hand, flux is affected by pore tortuosity but can be calculated by modifying the diffusion flux in air by a gas tortuosity factor (Jury & Horton, 2004;

Hillel, 1998). Hillel (1998) studied the diffusion of carbon dioxide through packed soil cores and recommended a linear relation where 0.66 is a tortuosity coefficient:

$$D_s/D_o = 0.66\, f_a \qquad 2.9$$

Where D_s = Diffusion coefficient in soil and D_o = Diffusion coefficient in bulk air. The smaller D_s can be expected to be some function of the air-filled porosity f_a. The suggested coefficient is about two-thirds the length of the real average path of diffusion in soil. Because tortuosity itself should depend on the fractional volume of air-filled pores it is expected that this constant coefficient will have only a limited range of variation (Hillel, 1998).

Some other values found by investigators on different soils and ranges of air and water content were those of Van Bavel (1952) that initiated $D_s/D_o = 0.61\, f_a$, and was found to give good agreement with observation in a sieved and repacked medium (Moldrup *et al*., 2000). As air-filled porosity fell to around 10%, the ratio D_s/D_o decreased to about 0.02 (Grable & Siemer, 1968).

2.3 Factors that influence aeration

2.3.1 Texture

Some factors that determine the extent of the difference between atmospheric and soil air constituents include depth in the soil profile and soil pore size distribution.

Oxygen levels generally decrease with depth in the soil profile due to slow diffusion rates of oxygen from the surface through the soil. Soils with large pores promote more rapid oxygen diffusion into and through the soil, and carbon dioxide movement out of the soil. Soils with small pores have slower oxygen diffusion into the soil and carbon dioxide diffusion out of the soil (Watson & Kelsey, 2005).

Watson & Kelsey (2005) stated that sandy soils generally have low total porosity but large individual pores and clay soils have high total porosity but small individual pores. Soils with large pores generally have good drainage (less water) and aeration, while soils with small pores generally have poor drainage and aeration. Thus, sands have good drainage, while clays have poor drainage and are more likely to become anaerobic (deprived of oxygen) as microbes use oxygen more rapidly than it is replenished through diffusion.

The composition and amount of soil air is determined by the water content of soil unless the soil is very dry. In a well-aerated soil the O_2 content will be higher than that of a poorly aerated soil.

The concentration of CO_2 in a soil profile will increase with depth as O_2 concentrations will decrease (Lal & Shukla, 2004).

2.3.2 Water table height

Water tables can be used to benefit crop growth but can also be negative when oxygen levels decrease to cause oxygen stress conditions (Garcia-Vila *et al*., 2008).

According to Bennett *et al.* (2009) waterlogging is the major factor influencing the availability of oxygen levels in the root zone. Lal & Shukla (2004) confirmed that if there is an increase in the degree of saturation in the soil, the O_2 content will be reduced. This is a very common scenario in poorly drained and undrained soils where water logging results in O_2 deficiency. This is often caused by flood irrigation and wet weather that cause water to replace air in the soil (Surya *et al.*, 2006). Due to the presence of perched water tables in the surface soil layers and the depth of the groundwater, O_2 deficiency will cause respiration and root growth to be constrained (Bennett *et al*., 2009).

Bennett *et al.* (2009) stated in a literature review on soil physical conditions and drainage that it can be assumed that 10% of volume of air-filled pores is the lowest value at which air can be exchanged in the soil". Zhang *et al.* (2004) confirmed that in a duplex loamy sand over clay soil near Kojonup in Western Australia, air-filled porosities was below 10% at 0.1 m depth when average water tables were at a depth of 0.3 m.

Waterlogging varies spatially and temporally in the landscape. It is unclear how it should be measured in terms of, changes in concentrations of soil gases like O_2 (Belford *et al*., 1980; Cannell *et al*., 1980a; Barrett-Lennard *et al*., 1986) and redox potentials (Armstrong *et al*., 1985).

2.3.3 Soil compaction

Watson & Kelsey (2005) stated that compaction reduces total air-filled (no capillary) pore space, reduces average pore size and increases mechanical resistance to root penetration. Water infiltration and gas diffusion is reduced, soil O_2 concentration is decreased and CO_2 concentration can increase, possibly to toxic levels with the loss of macro pore space.

Aeration problems accrues when soils become compacted by natural causes as a consequence of their textural composition, water regime, or the manner in which they were formed in place and also when clay soils shrink upon drying (Huang *et al*., 2006). Compact soils of fine texture may suffer from poor aeration due to water logging (gaseous exchange may not be so rapid to

remove CO_2 from soil air and to supply oxygen to the roots). This also happens when there is an excessive amount of readily decomposable organic matter added to the soil (Watson & Kelsey, 2005). According to Ewa (2004) repeated wheel slipping by a tractor with a weight of only 2 000 kg, can produce soil conditions in which aeration can be limiting for crop growth. This can be overcome by the use of dual wheels that resulted in lower values of soil bulk density and associated greater soil aeration. Ploughing the soil is a common cultivation method to prevent soil compaction. Compaction has an influence on soil aeration and soil aeration has an influence on yield (Huang *et al*., 2006).

2.4 Biological processes in the soil environment

2.4.1 Soil respiration

Soil respiration is the result of biological activity in the soil of all soil organisms and is usually refers to as carbon dioxide efflux from the soil surface. The most important respiratory activities are those of soil microorganisms and plant roots (Bingrui & Guangsheng, 2008). Root respiration in soil has a wide variation and its contribution may exceed that of the microorganisms in some cases. These two components, microbial respiration and root respiration, are also interrelated to one another. Roots in soil contribute to respiration, not only by their own respiration, but also by stimulating microbial respiration due to the incorporation of root exudates and the decaying residues of dead roots. However, root respiration and microbial respiration depend essentially on the same factors although not always in the same manner. Thus, for example, the roots of higher plants are more sensitive to drought than microorganisms. Only roots are affected by soil resistance to penetration, while only microbial respiration is affected by organic matter content. The oxic respiration process where oxygen gets abundant can be summarized as follows:

$$C + O_2 \rightarrow CO_2 + \text{energy} \qquad 2.10$$

This activity can be measured by the amount of carbon dioxide efflux or oxygen taken up in unit of time per volume or mass unit in soil (Glinski & Stepniewski, 1985). De Jong & Schappert (1972) made calculations for respiration distribution with depth on the basis of the measurement of carbon dioxide content and its diffusion coefficient distribution within the soil profile. This shows that approximately 90% of soil respiration is concentrated in the humus horizon of soil.

Soil respiration creates a concentration gradient where O_2 flow in the soil through the process of diffusion and pushes the CO_2 out of the soil (Glinski & Stepniewski, 1985; Lal & Shukla, 2004). Some examples of respiratory activities of different organisms are presented in Table 2.3.

Table 2.2 Respiration rate of different organisms (Glinski & Stepniewski, 1985).

Organisms	**O_2 uptake (cm^3 hr^{-1} kg $body^{-1}$)**
Reptiles	5 – 1,100
Amphibia	14 – 130
Insects	37 – 15,000
Bacteria	Up to 1,200,000
Fungi	Up to 10,000
Algae	Up to 40,000

Production of gas and absorption processes in soil involve gasses such as O_2, CO_2, CO, C_2H_4, CH_4, N_2O, H_2, NH_3, NO and NO_2, where O_2 and CO_2 are of major importance (Figure 2.3). O_2 is mainly consumed, and CO_2 is evolved in the process of respiration (Bingrui & Guangsheng, 2008; Glinski & Stepniewski, 1985).

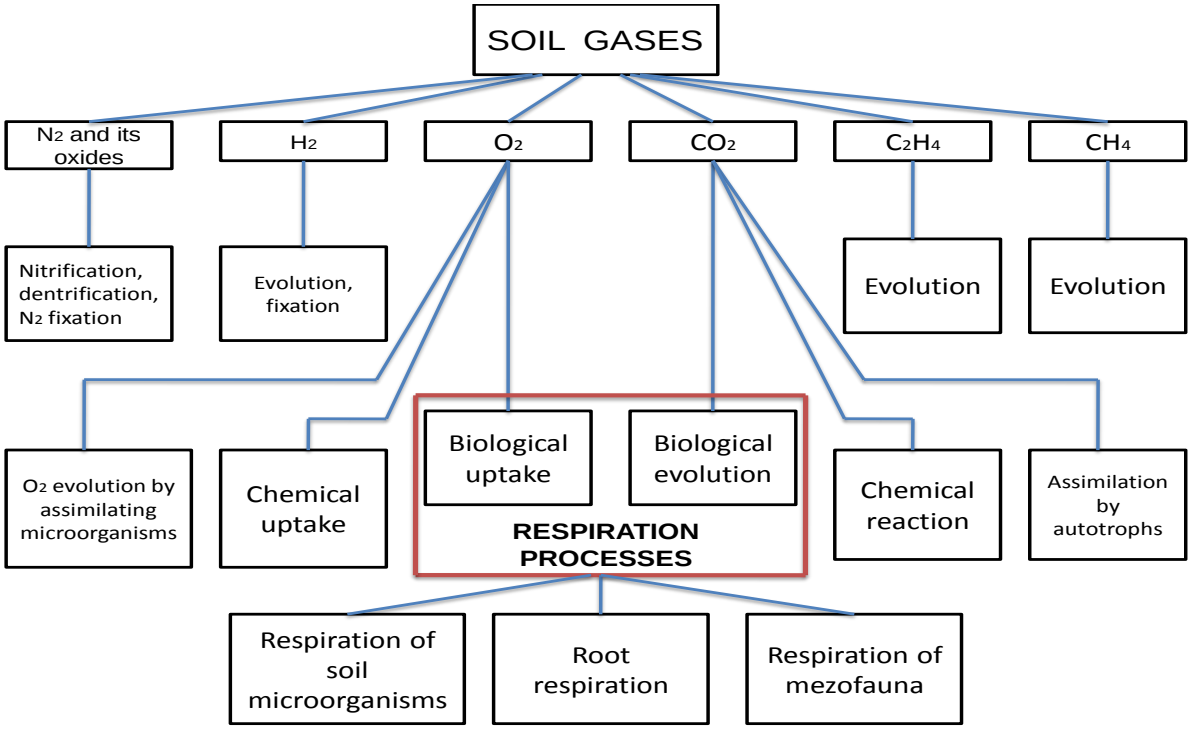


Figure 2.3 Soil gases and there processes effecting the composition of soil air (Glinski & Stepniewski, 1985).

2.4.1.1 Soil microbial respiration

Apart from root and meso-fauna respiration, respiratory activity is the result of the activity of the entire microbial population in the soil. The rate of microbial activity depends on the composition, size and metabolic activity of the population and these factors are then again influenced by numerous factors. Prominent among the latter factors are soil temperature (Lloyd & Taylor, 1994; Fang & Moncrieff, 2001), soil water (Davidson *et al*., 2000), oxygen concentration and organic matter content. Some other factors involve carbon dioxide, soil air-filled porosity, bulk density and aggregate size, soil reaction, soil minerals, mineral fertilization, heavy metals and pesticides.

Table 2.4 show the result of soil microbial activity in different soils, given by varies authors. These values are normally in the range from 0.2 to 20 $cm^3\ O_2\ kg^{-1}\ hr^{-1}$ (Glinski & Stepniewsk, 1985). According to Sarlistyaningsih *et al.* (1996) low oxygen supply in soil during water logging, which resulted mainly from activity of microorganisms, was the major factor causing the reduction in survival of lupin seed in waterlogged soil.

Table 2.3 Respiratory activity of different soil forms (Glinski & Stepniewsk, 1985).

Soils	**O_2 uptake ($cm^3\ kg^{-1}\ hr^{-1}$)**	**CO_2 evolution ($cm^3\ kg^{-1}\ hr^{-1}$)**	**References**
Clay soils	2.2 – 28.0		Croswell & Waring, 1972.
Loamy sand	0.9	0.9	Glinski & Stepniewski, 1973.
Sandy loam soil		0.14 – 1.63	Glinski & Stepniewski, 1973.
Silty loam	1.0 – 4.0		Lyda & Robinson, 1969.

2.4.2 Factors influencing microbial respiration

2.4.2.1 Temperature

Microorganisms are divided into three groups with respect to their temperature requirement for respiratory activity, with each group reacting differently to varies temperatures: Kriophiles (with an optimum temperature < 20°C), mesophiles (with an optimum temperature between 20°C and 40°C) and thermophiles (with an optimum temperature > 40°C) (Golebiowska, 1975).

The maximum point of respiration in soil usually occurs between the temperature ranges of 40 to 70°C. These temperatures can change even in the same soils with different water contents. The respiratory maximum is usually not reached at normal soil temperature and an increase is

observed in oxygen uptake or carbon dioxide production rate with temperature rise (Bingrui & Guangsheng, 2008; Salonius, 1978).

The influence of temperature on soil respiration is commonly described using Van Hoff's equation. Van Hoff's equation stated that the reaction rate increases by a factor of 2 to 3 with a temperature rise of 10°C.

$$Inq = c + T \qquad \frac{InQ_{10}}{10} \qquad 2.11$$

Where q = Respiration rate at temperature T, c = Constant and Q_{10} = the Q-ten temperature coefficient. The Q_{10} coefficient shows how many times the respiration intensity increases when the temperature increases by 10°C (Han *et al*., 2007; Zhou *et al*., 2008). Glinski & Stepniewsk (1985) stated that the Q_{10} value is in the range of 2 to 3 for chemical and 1.2 to 1.3 for physical (diffusion conductivity) processes. Abrasimova (1979) obtained values of 1.6 to 7.6 in several cultivated soils of the U.S.S.R. An increase in altitude from 100 to 5200 m above sea level, together with a decrease in mean annual temperature from 25°C to 0°C was accompanied by a drop of about 100-fold in aerobic bacteria counts, as well as by a 10-fold reduction in oxygen uptake per gram organic matter in mountain soils (Franz, 1976).

2.4.2.2 Soil water

Soil water content has significant effects on quantity and activity of microorganisms (Howard & Howard, 1993; Davidson *et al*., 2000). Microorganisms are divided into three groups with respect to their water requirement for respiratory activity, with each group reacting differently to water deficiency. The first group is called hygrophiles (majority of bacteria, yeast and some fungi), where activity disappear at soil water tension above 7100 Pa. The second group is called mesophiles (majority of fungi and some bacteria), where activity disappear at soil water potential intervals from 7100 Pa to 30 000 Pa. The third group is called xerophiles (Some *Phycomycetes*), where activity disappear at water tension above 30 000 Pa (Prusinkiewicz, 1974).

Bingrui & Guangsheng (2008) found that the relationship between soil water and microbial activity is curvy-linear with a maximum at a certain point of optimum. The soil respiration rate will decrease at both low and high water content with a maximum rate usually within soil water potential intervals of 10 to 1 000 kPa. Reduction in respiration rate at low water contents is due to a low availability of water for respiration where the soil respiration rate decreases with decreasing water availability and stop when the soil water content falls below 4.2%. This might

be the wilting point of the soils in *L. chinensis* steppe because it was found that the permanent wilting point of soils was 4% in sandy loam (Lüdeke *et al*., 1994). Reduction in the respiration rate at high water contents is due to a limitation of oxygen, caused by the pores filled with water (Bingrui & Guangsheng, 2008).

The factor of time is very important in respiratory activity and it should be emphasized that respiratory activities, at a constant water level, decrease in time. Respiration activity was lower at a certain water content, obtained by drainage of a saturated soil sample than in the case of the same water content achieved by adding water to the dry sample. This was due to different rates of development of microbes and by different degrees of substrate utilization during the two processes (Bingrui & Guangsheng, 2008).

When air-dried soil is wetted (recharged), the respiration rate will be relatively high and then decrease after several days (Croswell & Waring, 1972) or weeks to be more specific (Das, 1970). Wetted air-dry soil is also related to an increase in the number of microbes due to increased decomposability of organic matter after prolonged drying, caused by chemical processes such as oxidation (Croswell & Waring, 1972).

According to Jager & Bruins (1975), the repeated cycles of wetting and draining increase the respiratory intensity which implies an increased rate of organic matter depletion compared to constant water saturated conditions.

2.4.2.3 Oxygen

Oxygen stress will occur when the rate of oxygen supply falls below the rate of oxygen demand by respiratory processes in soil. Since the storage of oxygen in soil is relatively low compared to the quantity required for respiration, these conditions can develop quite quickly (Hillel, 1998). Ericson & van Doren, (1960), as cited by Hillel (1998) stated that plant growth depends more on the occurrence and duration of oxygen stressed periods than on average conditions. When there is a decrease in soil O_2 concentration, there will be a decrease in the aerobic microbial population. The anaerobic microbial population will then increase which will change the soil respiration (Surya *et al.*, 2006).

2.4.2.4 Organic matter content

Soil respiration is always higher in cropped than in fallow land, due to more organic matter available for root and microbial respiration in the form of live roots and the decay of dead roots. Organic materials in soil are accompanied by an increase in soil microbial activity and the rate of

mineralization (Hillel, 1998). According to Glinski & Stepniewski (1985) it can be assumed that 75-80% of organic material incorporated in soil undergoes mineralization, while the remainder is transformed into more stable specific humic substances. Microbial respiration is influenced by organic amendment due to the decomposability of that material (kind and age). For example, organic material contains a wide range of components, like cellulose, hemicelluloses, lignin, compounds soluble in water, compounds soluble in ether and alcohol, and proteins and water. Soluble compounds decrease as the plant ages, while that of components more resistant to microbial decomposition like lignin, cellulose and hemicelluloses increase. Figure 2.4 shows the influence of organic amendment on soil respiration. The results illustrate the increase in O_2 uptake with an increase in the rate of residue application. There was a sharp increase in O_2 consumption during the first two weeks after the O_2 consumption tends to decrease over time in both residue treatments, irrespective of crop type.

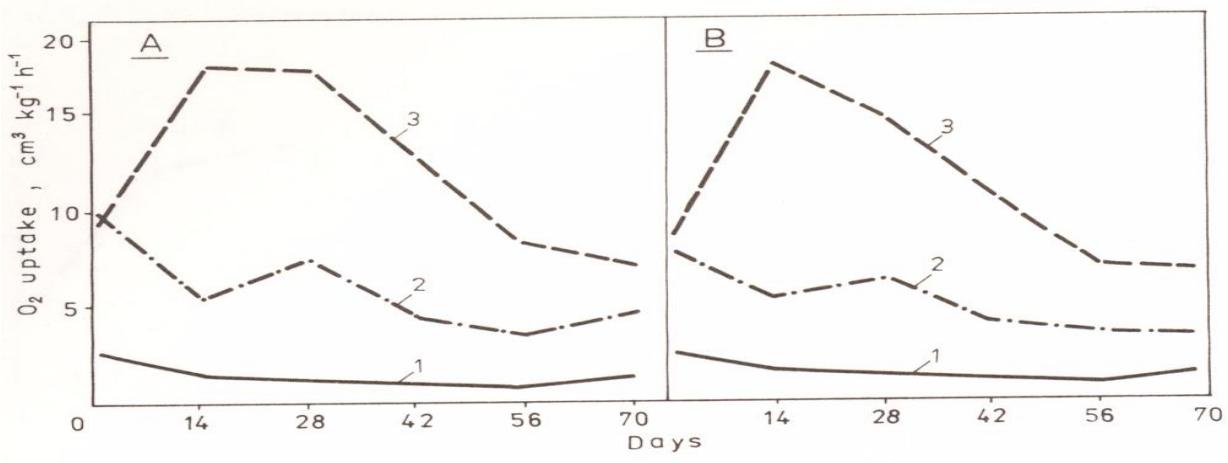


Figure 2.4 The influence of crop residues of oats (A) and cotton (B) on the total respiratory activity of a fine sandy loam soil. Line 1 is the control, Line 2=1% crop residue added and Line 3 =4% crop residue added (Lyda & Robinson, 1969 reviewed by Glinski & Stepniewski, 1985).

According to Reddy & Patrick (1974), organic matter breakdown is faster under aerobic conditions than under anaerobic conditions due to higher oxygen availability. Soil organic matter contributes to soil productivity which contributes to crop productivity when decomposed. When there is a decline in organic matter, large amounts of plant nutrients, especially nitrogen, are released in the soil. A 2% decrease in organic matter can release as much as 442 kg ha^{-1} of nitrogen and also improves soil physical properties like aggregation and water holding capacity.

Kirschbaum (1995) did a study on the influence of temperature on the decomposition of soil organic matter in Australia. He found that a 1°C increase in temperature could lead to a loss of

over 10% soil organic carbon where there is an annual mean temperature of 5°C. In regions with the same temperature increase, soil organic carbon loss will be 3% where soil temperatures reach 30°C.

2.4.3 Diurnal and seasonal effects on soil respiration

Respiration rates in soil vary from season to season, day to day and hour to hour, and are related to microbial activity and the stage of crop growth (Hillel, 1998). Soil air temperature plays an important role at both diurnal and seasonal respiration (Bingrui & Guangsheng, 2008). Diurnal variation of soil respiration is directly related to soil temperatures with O_2 uptake rates that can increase twofold from early morning to mid-afternoon (Hillel, 1998). According to Lal & Shukla (2004), respiration rates in the summer can be up to ten times higher than in the winter.

In Table 2.5 it is shown how soil-air composition varies significantly between cropping systems in the soil as plants consume some gases and microbial processes release others (Lal & Shukla, 2004).

Table 2.4 Measured O_2 and CO_2 content (% by volume) in soil air collected during summer and winter at 150 mm depth (Lal & Shukla, 2004)

Cropping systems		O_2	CO_2
Arable land manured and cropped	Summer	20.74	0.23
	Winter	20.31	0.37
Arable land unmanured and cropped	Summer	20.82	0.19
	Winter	20.42	0.21

2.5 Interaction of soil and water

2.5.1 Methods of soil water measurements

Measurements of soil water can be classified into direct and indirect methods to express it quantitatively (Hillel, 1998; Topp & Ferré, 2002; Muńoz-Carpena, 2004). Direct methods involve drying a soil sample in an oven to determine the gravimetric soil water content, expressed in g/g. The volumetric water content can be calculated and displayed in mm^3 mm^{-3} from the gravimetric soil water content, if the soil bulk density is known (Topp & Ferré, 2002). The procedure for calculating gravimetric water was further examined by Hillel (1998) and Topp & Ferré (2002).

Indirect methods involve the measurement of soil water on a volumetric basis or soil water potential. This is based on estimating some chemical and physical soil properties such as dielectric constant, heat capacity, electrical conductivity, hydrogen content or magnetic susceptibility that ultimately relates to soil water content. Thus, absolute water content is estimated by a calibrated relationship with some other measurable variables (Muńoz-Carpena, 2004).

2.5.1.1 Determination of gravimetric and volumetric water contents

The gravimetric method involves the collecting of soil samples from the field, weighing the samples before it is oven-dried at a temperature range of 100-105°C to a constant mass. The samples are weighed again to determine the mass of the water content in relation to the mass of the dry soil (Topp & Ferré, 2002). The following equation can be used to determine water content:

$$\theta_g = M_w/M_s \qquad 2.12$$

Where M_w is the mass of water and M_s is the mass of the oven dried soil. If the gravimetric metric water content is determined the volumetric water content can be determined by equation 2.13. The volumetric soil water content is expressed as the volume of water in the volume of undisturbed soil. Volumetric water contents are related to soil bulk density and estimated by:

$$\theta_v = \theta_g \times P_b \qquad 2.13$$

This technique provides advantages such as low costs, easy operation and accuracy (±10 g/kg) (Muńoz-Carpena, 2004). The disadvantages of this procedure are that it is time consuming (minimum of 2 days per measurement) and samples cannot be extracted at exactly the same location since the destructive nature of the technique.

2.5.1.2 The neutron water meter

The neutron probe technique is a popular indirect method used *in situ* by scientists in the past (e.g. McKenzie *et al*., 1990; Kamgar *et al*., 1993; Corbeels *et al*., 1999; Evett *et al* 2002; Heng *et al*., 2002 & Yao *et al*., 2004). The neutron probe as shown in Figure 2.5 consists of a probe, pulse counter, a cable that connects the probe with the pulse counter and a transport shield with display and a keyboard (Bell, 1987).

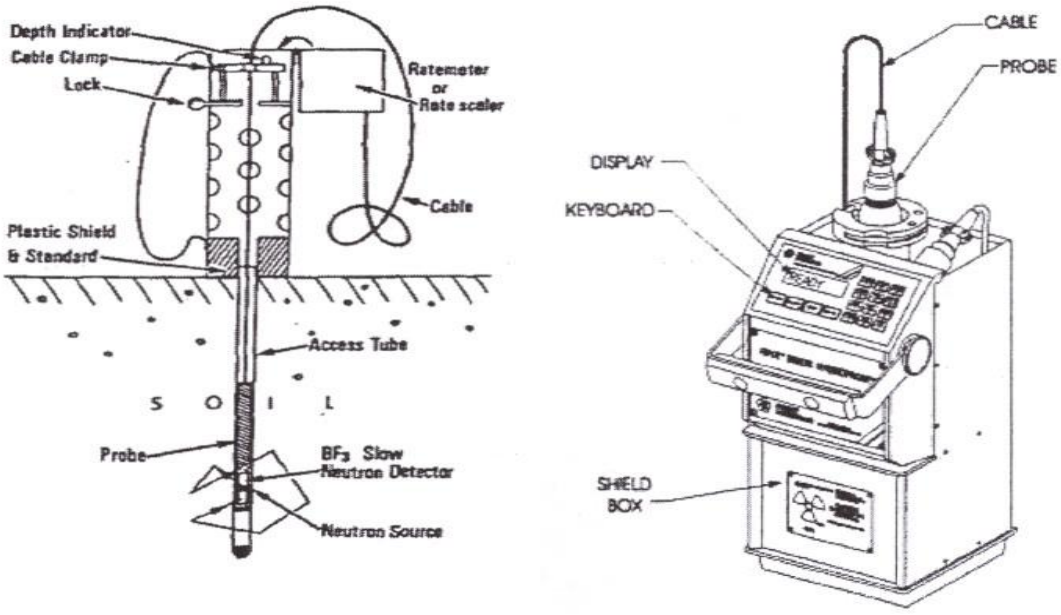


Figure 2.5 Schematic diagram of a neutron probe (Bell, 1987; Evett, 2000).

The access hole where the probe is lowered should have a liner of aluminium or polythene tube, sealed at the bottom and with no cavities between the soil and the lining material. The access hole is constructed by a power-auger where the lining tube is subsequently pushed down. The soil water profile, as well as the total water content can be determined by lowering the probe successively to greater depths (Marshall & Holmes, 1996). The neutron probe expels high-energy neutrons at a rate of 10^{27} s^{-1} from the radioactive source of the probe into the surrounding soil where collision with nuclei of atoms takes place. Hydrogen is predominantly in this situation because it is the most effective in slowing down fast neutrons due to similar masses of a proton and a neutron. Thermal neutrons will be back scattered to the detector where an electrical pulse are created and measured. The mean count rate is displayed as the amount of water in a unit volume of the soil. The volume of soil influencing the count increases with decreasing water content and has a radius of 25 cm at $\theta = 0.1$ (Marshall & Holmes, 1996; Hignett & Evett, 2002).

Soil water content is regulated by the concentration of hydrogen nuclei around the access tube as water is also the main source of hydrogen in the soil matrix. Each different element in soil has a different “scattering and capture cross section” and this cross section is constant for a unit volume of soil and proportional to bulk density (Bell, 1987). Calibration is necessary for accurate soil water measurements in different soils and at different soil depths as absolute water content cannot be measured automatically (Van Bavel *et al*., 1961). Although neutron probes come with

a factory calibration used in common soil types and for routine soil water determination, recalibration is important for accurate measurements (Dickey, 1990). It is recommended by Tyler (1988) and Klenke & Flint (1991) that different calibration equations are needed for different types of soil for better water management.

2.5.1.3 Capacitance techniques for water determination

Capacitance probes were commercially developed and tested under laboratory (Dean *et al*., 1987) and field (Bell, 1987) conditions in the 1980's, but eventually became very popular in the 1990's with the advent of microelectronics (Paltineanu & Starr, 1997) for measuring *in situ* soil water content (Fares & Alva, 2000). Ever since, capacitance probes have been used extensively for water monitoring in a wide range of soils to enhance irrigation management (Alfa & Fares, 1999).

There are different systems available with most of them consisting of probes, a data logger, oscillator equipment and computer software (Gardner *et al*., 1998). Soil water determination involves the measurement of the soil dielectric constant by measuring the capacitance between two electrodes of a probe in the soil. In essence the capacitance probe determines the velocity of an electromagnetic wave through the soil environment (Muńoz-Carpena, 2004). The theoretical background is described in more detail by Dean *et al*. (1987).

The capacitance change of the three-phase system (soil, air and water) in the soil matrix is mainly governed by the water content of the soil as the dielectric constant of water is 81, while the dielectric constant of other components i.e. soil minerals is 2-5 and soil air is 1 (Muńoz-Carpena, 2004). Therefore the measurement of a soil's dielectric constant (capacitance of the soil system) is primarily determined by the volumetric water content in the soil. Capacitance of the soil-access tube system is measured in terms of resonant frequency around the probe in the soil environment. As water content in the soil volume increase, the resonant frequency will decrease. The capacitance measurements are thus directly proportional to the water content of the measured soil (Dean *et al*., 1987).

Measurement of the dielectric constant by capacitance probes are extremely sensitive to air space around the access tube due to the wide variation of dielectric properties of soil constituents (k_{water} = 81 and K_{air}= 10), (Figure 2.6). According to Paltineanu & Starr (1997) the sensitivity within a 0.1 m radius of the centre of the probe is 99% and 92% within a 0.03 m radius. Hence, even very small variation in the soil properties close to the access tube, must be taken in consideration.

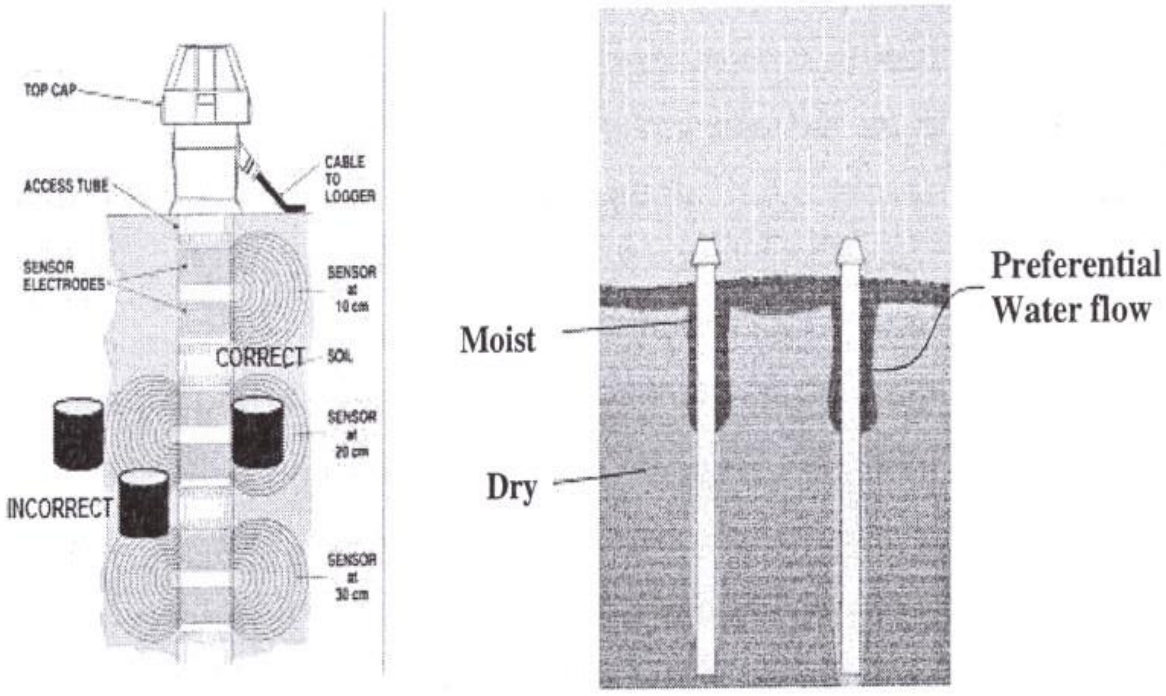


Figure 2.6 Schematic diagram of capacitance probe (Starr & Paltineanu, 2002).

Some other capacitance sensors are portable sensors that are lowered into the access tube where soil water is measured at different depths in just a few seconds (Bell, 1987; Evett *et al*., 2002; Heng *et al*., 2002; Geesing *et al*., 2004). Sensors where placed at different depths into an access tube and the output frequency is logged at regular intervals (Paltineanu & Starr, 1997; Fares *et al*., 2004). Other sensors are installed directly into the soil (Ould Mohamed *et al*., 1997; Seyfried & Murdock, 2004).

The calibration of capacitance probes consist of estimating the scaled frequency (*SF*) which is generated by comparing the probe response in the soil to the response in air and water. A calibrating equation can then be determined by regressing *SF* against the real volumetric water content. Sentek (2004) describes a non-linear power model that is used to determine the standard regression equation for capacitance probe calibration:

$$\theta_v = a\,(SF)^b \qquad 2.14$$

θ_v is volumetric water content, a and b are fitting coefficients and *SF* is scaled frequency. *SF* is a normalized value of 0 in air and 1 in water that is calculated using the following equation described by (Sentek, 2004):

$$SF = (F_A - F_S)/(F_A - F_W) \qquad 2.15$$

where SF is the scaled frequency, F_A is the frequency reading in the PVC tube suspended in air, F_S is the frequency reading in the PVC access tube installed to a certain depth in the soil, and F_W is the frequency reading in the PVC access tube suspended in 100% water.

If soil water is unevenly distributed in the soil profile it will have an effect on the *SF* of the capacitance probe which necessitates the regression of *SF* on θ_v (Hulme, 1997). This is problematic as the equation should be created by *SF* on θ_v rather than θ_v on *SF* (Hulme, 1997; Heng *et al*., 2002). Some models used in previous studies were a non-linear power model (Enviroscan-Diviner 2000), a two parameter power model (Paltineanu & Starr 1997; Heng *et al*., 2002; Geesing *et al*., 2004), and a three-parameter power model (Baumhardt *et al*., 2000; Fares *et al*., 2004). These modules were found to give acceptable accurate measurements (Equation 2.16). In addition Bell (1987), Ould Mohamed *et al*. (1997) and Evett *et al*. (2002) also described a linear model for calibrating the capacitance probe (Equation 2.17).

$$SF = a\theta_v^{b} \qquad 2.16$$

$$SF = a + b\theta_v \qquad 2.17$$

Some capacitance probes come with a factory calibration based on different soils that can estimate relative water change. This calibration method is not always sufficient for scientific studies (Paltineanu & Starr, 1997; Alfa & Fares, 1998; Starr & Paltineanu, 1998; Alfa & Fares, 1999).

According to Paltineanu & Starr (1997) and Muńoz-Carpena (2004), calibration of a probe for a specific soil is needed due to the influence of not only soil water on dielectric constant but also physical and chemical characteristics such as texture, organic matter content, shrinking and swelling of clay soil and salinity (also supported by Baumhardt *et al*., 2000; Morgan *et al*., 2001; Fares *et al*., 2004). There are only a few scientific studies on the calibration procedure of capacitance sensors (Bell, 1987; Paltineanu & Starr, 1997; Baumhardt *et al*., 2000).

2.5.2 Soil hydraulic properties

Soil form the fundamental building block of all agricultural activities. Soil water movement is an important component to be understood and can help solve problems related to irrigation, water holding capacity, subsurface drainage contribution to groundwater and water disposal. Information on soil hydraulic properties for a certain area is useful for adequate and effective management of soil and water for example irrigation scheduling and management (Klute & Dirksen, 1986).

2.5.3 Soil water retention characteristics

Jury & Horton (1994) defines water retention characteristics (presented as θ-h) as the relationship between the amount of water in the soil and the energy potential with which the water is bound by the soil. The θ-h relationship is unique for different soils as each soil has a different capacity to retain water against an energy potential. The θ-h relationship is influenced by both the variation in soil particle size distribution and soil structure that is affecting the pore size distribution and the number of given pore size in each size class (Dexter, 2004).

The θ-h relationship will be strongly affected by soil particle size distribution at suction heads > 100 kPa and to a lesser extend at lower suctions where soil structure have more of an influence. The θ-h relationship is a very important soil property that has an influence on critical soil processes such as irrigation scheduling, drainage and hydraulic conductivity (Kern, 1995).

2.5.3.1 Methods of characterizing soil water retention characteristics

There are several methods available for measuring water retention characteristics with a complete discussion given by Dirksen (1999). Direct methods will be presented to determine θ-h relationships in the laboratory by means of hanging water columns and pressure plates. According to Dirksen (1999) and Bohne (2005) the entire range from saturated to almost completely dry soil can be represented by using the hanging water columns and pressure plate apparatus.

Jury *et al*. (1991) stated that a hanging water table is normally used to obtain equilibrium by exposing saturated, undisturbed soil samples at suctions up to 10 kPa. Limitation to hanging water columns, suction is due to the length of the hanging water column. However, the pressure plate apparatus is normally used for a suction range of 30 to 1500 kPa (Jury *et al*., 1991). According to Reeve & Carter (1991) the pressure plate method is very accurate with a coefficient of variation of 1-2% attainable. Reduction in efficiency of the pressure plates can occur where there is clogging of the plates by soil particles or alga growth (Townened *et al*., 2001).

According to Hillel (1998) the θ-h relationship in a low suction (0-100 kPa) is strongly influenced by the soil's natural pore size distribution and structure. However, Dirksen (1999) recommended that undisturbed soil samples should be used for a low suction range. When suction exceeds 100 kPa it is acceptable to use disturbed soil samples as the water retention characteristics of a soil is not influenced by soil structure anymore.

2.5.4 Capillary rise

Capillary rise can be defined as the up flow of water from a water table into an active plant root zone (Sepashah & Karimi-Goghari, 2005; Hillel, 1998). Hillel (1998) describes an equation where the equilibrium capillary height, is in relation with the radii of the pores.

$$hc = (2y \cos \alpha)/rP_w g \qquad 2.18$$

y = Surface tension, r = Capillary radius, P_w = Water density, g = Gravitational acceleration and α = Wetting angle (normally taken as zero). It was found that an increase in silt-plus-clay content in the soil will increase the height of capillary rise above the water table (Hillel, 1998; Ehlers *et al*., 2003;). This is due to the smaller pores of clay. However, all pores are not uniform or have a constant radius, hence the height of capillary rise will differ in different pores (Hillel, 1998).

Where a root accessible water table is present, the total amount and number of irrigations can be reduced. According to Gharmarnia *et al.* (2004) water table depths of 0.7 and 1.5 m can have a respectively contribution of 20% and 40% to evapotranspiration demand of different crops by capillary up flow. Grismer & Gates (1988) reported that under arid conditions, the water table can supply as much as 60–70% of a crop's water requirement. Non-saline groundwater under irrigated and non-irrigated conditions contributed 72–90% and 91–100% of plant water use, respectively (Sepashah & Karimi-Goghari, 2005).

The successful use of water tables can supplement water supply to crops but will depend on several factors such as the water table depth, soil physical properties, plant root distribution and soil salinity (Ehlers *et al*., 2003).

Three different soils with different textures were used to illustrate the relationship between water table depth and the contribution from the water table as a percentage of evapotranspiration (ET) (Figure 2.7). As the water table depth increased, there was a decrease in the contribution of the water table as a percentage of ET (Sepashah & Karimi-Goghari, 2005). Streutker *et al.* (1981), as cited by Ehlers *et al*. (2007) found that water tables can reduce the irrigation requirements of cotton and wheat by 50%.

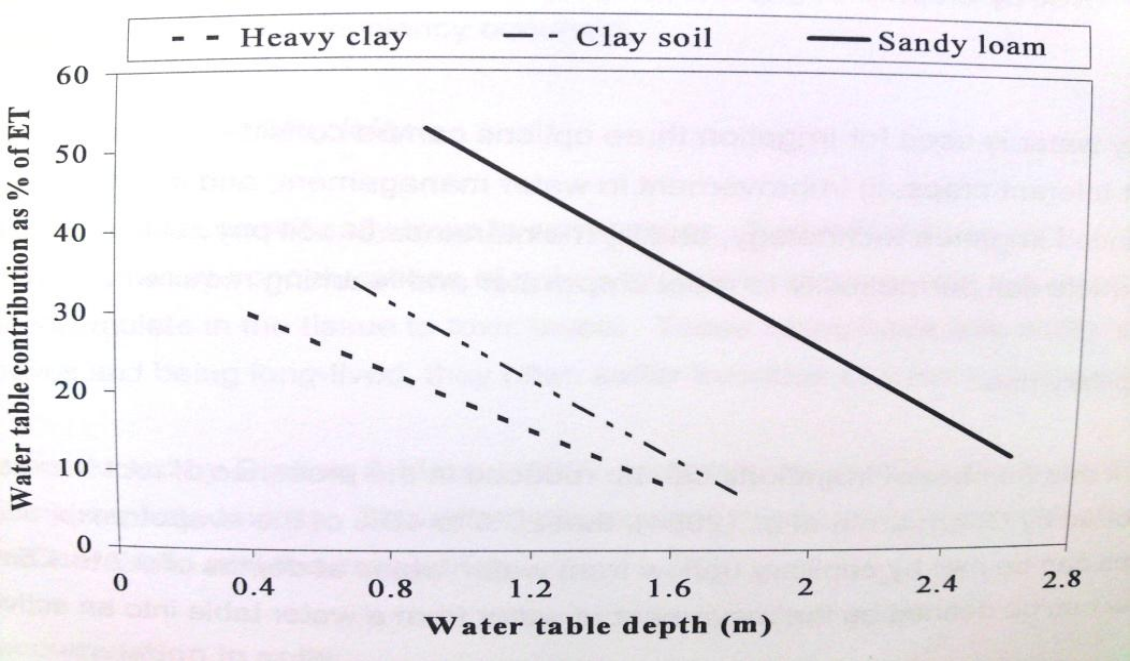


Figure 2.7 The effect of water table depth and soil texture on water uptake from water tables (Grismer & Gates, 1988).

Research conducted by Ehlers *et al.* (2003) found that the hydraulic properties of the soil layer above the water table control the upward flow rate and height which regulates the supply rate from a water table.

In this study the modelled upward cascading approach was used in SWAMP (Soil Water Management Program) as well as the finite difference water redistribution approach in SWB (Soil Water Balance). Both models showed a positive correlation between simulated and measured results and therefore these models can be used to measure capillary flow from different water table depths.

Before irrigation or drainage systems were introduced, the rate of the upward seepage from the equilibrium depth of the water table could be estimated. When the top soil is dry and the seepage rate equalled both the rate of capillarity rise from the saturated zone and the rate of evaporation, the rate of capillarity can be found from the steady state relationship between depth of water table, hydraulic properties of the soil and water content as displayed in Figure 2.8 (Oosterbaan, 1994).

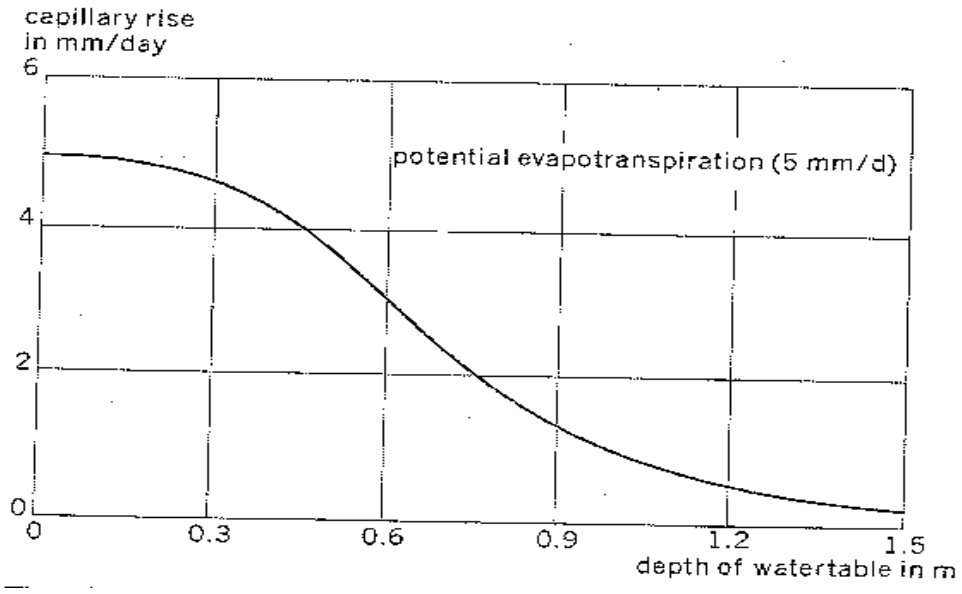


Figure 2.8 Relationship between water table depth and the rate of capillary rise in a clay soil (Oosterbaan, 1994).

2.5.5 Redox potential and pH

Redox potential can be described as an electrical measurement that shows the tendency of a soil solution to transfer electrons to or from a reference electrode. This potential can indicate whether the soil is aerobic, anaerobic, and whether chemical compounds such as Fe oxides or nitrate have been chemically reduced or are present in their oxidized forms (Vepraskas & Faulkner, 2001). According to Callebaut *et al.* (1982) redox potential is direct in relation to soil pH. Redox potential will stay in equilibrium with soil pH even with the pH changing in different soil types (Callebaut *et al.,* 1982).

Redox potential describes the electrical state of a matrix. Thus this potential is an important parameter controlling the persistence of many organic and inorganic compounds in a soil environment (Callebaut *et al.,* 1982).

Figure 2.9 shows how the redox potential of a silt loam soil in the U.S.A. responded rapidly to changes in the aeration status of the soil. The maximum redox potential was in aerobic conditions at +600 mV, while the minimum was measured at -300 mV in anaerobic conditions. The redox potential measurements were made by connecting the Pt electrode to a pH meter, using a saturated calomel half-cell as reference electrode (Reddy & Patrick, 1974).

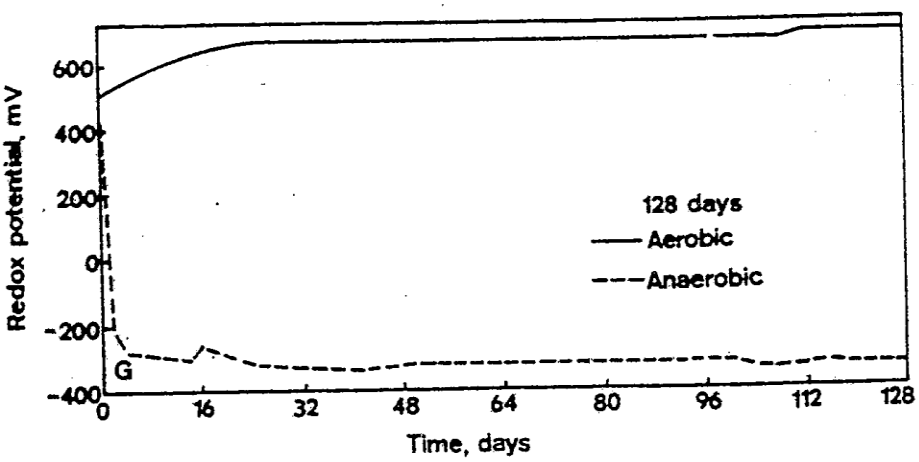


Figure 2.9 The effect of alternate aerobic and anaerobic conditions on redox potential of a silt loam soil (Reddy & Patrick, 1974).

In Figure 2.10 the redox potential is presented as a function of water table depth in two different soils. Redox potential in saturated soil never exceeded the + 200mV in both the sandy loam and the loamy sand soil. As the water table was lowered to 22 cm, a significant increase in redox potential was observed at 5 and 25 cm depths beneath the soil surface in the sandy loam and loamy sand soil profiles. At these depths the redox potential fluctuated between 0 and 350mV in both soils with the pH of the sandy loam and the loamy sand ranged respectively between 6.8 - 7.0 and 6.0 - 6.4. As redox potential measurements are pH-dependent, they are normally adjusted to pH 7 by a factor, usually -59mV. If redox potential differs between two soils, its more due to pH differences in the soils and it should not differ when values are expressed at a constant pH of 7 (Callebaut *et al*., 1982).

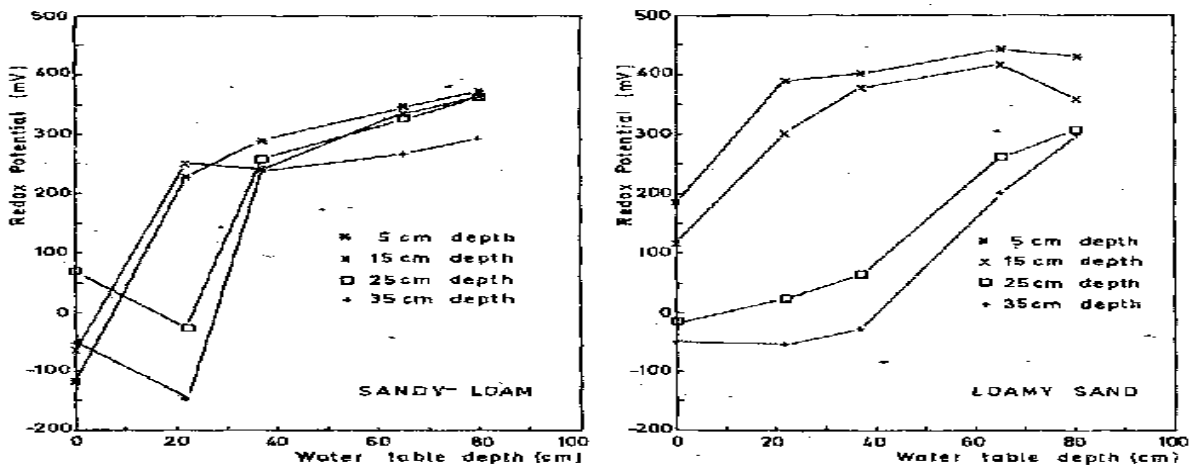


Figure 2.10 Redox potential as a function of water table depth, at different depths in a sandy loam and a loamy sand soil profile (Callebaut *et al*., 1982).

The method used by Callebaut *et al.* (1982) was to measure redox potential (*Eh*) with a platinum electrode, which internal resistance ranged from 100 to 150 ohms. Vorenhout *et al*. (2004) described a method which was also described by Cogger *et al.* (1992) and Mansfeldt (2003) where a small piece of Pt is placed on a copper wire. The copper wire with the Pt piece is placed in the soil which contains a reference electrode at a chosen distance. Fluctuations measured in redox potential can be very large and depth dependent.

Figure 2.11 illustrates the redox potential as a function of oxygen concentration in two different soils. Redox potential is fairly well correlated with oxygen concentration at low O_2 content (6 to 15%) in the loamy sand and sandy loam soils. However, a greater scattering of data appears in well-aerated conditions (>15%).

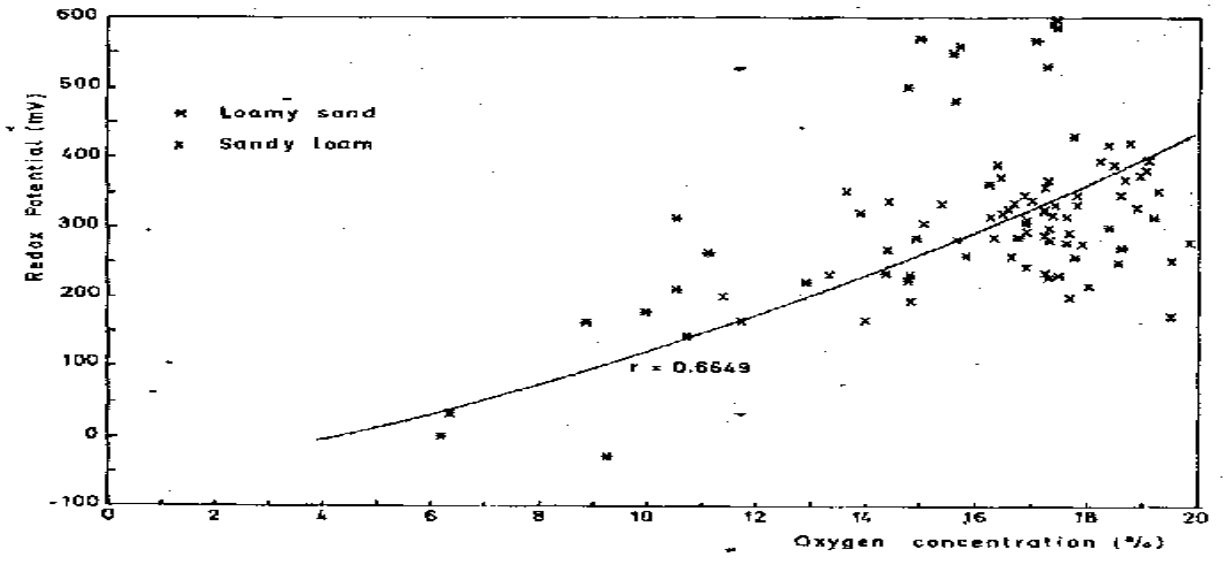


Figure 2.11 Redox potential as a function of O_2 concentrations in a sandy loam and a loamy sand soil (Callebaut *et al*., 1982).

2.6 Summary

The literature review describes how soil air is directly related to physical properties and in relation with soil water in the three-phase system (Drew & Stolzy, 1996). It was clear that there are direct and indirect factors that not only affect, but determine the soil air quantities and concentration. The strengths and limitations of different methods used for soil-air measurements in a lysimeter were also summarized. From these methods a unique design could be established to suit the purpose of this study. Furthermore, there are no detailed descriptions on how to construct a movable 200 kg lysimeter containing an undisturbed soil sample. There is

also a relative abundance of literature dealing with theory and application of soil air concentrations in small lysimeters in the presence of a fluctuating water table and in the absence of a crop. This information is needed for improving understanding of the effects of fluctuating water tables on O_2 and CO_2 concentrations in irrigated soils of South Africa. This study is intended to help fill this gap.

CHAPTER 3

DEVELOPMENT AND EVALUATION OF A SMALL LYSIMETER TO STUDY OXYGEN AND CARBON DIOXIDE CONCENTRATION IN SOILS WITH RISING WATER TABLES

3.1 Introduction

Lysimeters are well-established instruments for studying water balance and water flow processes in soils. Evidence for this can be found in Table 3.1 that summarise some of the prominent lysimeter-experiments conducted world-wide. Depending on the purpose of the study, lysimeters can be installed in the field, laboratory, greenhouse or glasshouses. Hence, the instruments vary in shape, size, depth and mass. Users have the option to either fill the lysimeters with disturbed soils or to install *in situ* sampled (undisturbed) profiles. Based on their mass, Bello & van Rensburg (2014) grouped lysimeters into large (> 5000 kg), medium (1000 - 5000 kg), small (10 - 1000 kg) and very small (< 10 kg) classes. Small and very small lysimeters can be automated easily with load cells, data recording systems or data loggers and multiplexes. Although the cost and installation of these lysimeters are favourable compared to the medium and large lysimeters, the calibration of load cells remains an effort.

A common problem is that long-term irrigation leads to increases in groundwater recharge, which in turn result in rising water tables (Belford *et al*., 1985; Mc Donald & Gardner, 1987). Rising water tables can be utilized positively to reduce the crop water requirements (Belford *et al*., 1985; Mc Donald & Gardner, 1987). Researchers found lysimeters convenient because they can simulate transient or constant water table conditions, which is otherwise very difficult to study in agricultural fields (Cannell *et al*., 1980a; Barrett-Lennard *et al*., 1986). For example, Ehlers *et al*. (2003) studied the contribution of root accessible water tables towards the irrigation requirements of field crops using large non-weighing field lysimeters in which the water tables were kept at constant heights throughout the seasons. They found that water uptake ranged from 38 to 63% for wheat, 25 to 53% for maize, 30 to 55% for peas and 21 to 45% for groundnuts.

On the other hand, poor water management under rising water tables has the potential to cause waterlogging which may result in substantial adverse effects on the growth and yield of crops (Belford *et al*., 1985; Mc Donald & Gardner, 1987). The reason for this is that the internal oxygen supply from above-ground parts to the roots is inadequate to meet the plants respiration

requirement. Adequate root respiration therefore demands that the soil itself to be aerated. In order to avoid oxygen deficiency or that excess carbon dioxide builds up in the root zone, it is of utmost importance that gaseous exchange takes place at a sufficient rate between soil air and the atmosphere (Hillel, 1998). The problem of waterlogging is a world-wide phenomenon. Ghassemi *et al*. (1995) estimated that about 20% of the 230 million ha of irrigated land in the world is seriously affected by waterlogging.

Despite the huge impact that waterlogging has on aeration, lysimeters are seldom equipped to measure oxygen and carbon dioxide concentrations in soils subjected to waterlogging (Table 3.1). The main reason for this is the difficulty experienced in sampling of soil air and the subsequent analysis of the sample (Glinski & Stepniewski, 1985). The extraction of a sample may not always be accurate due to better-aerated pores or leakage into the sampling tubes (Payne & Gregory, 1988). The most commonly primary methods to measure O_2 and CO_2 concentrations include soil air sampling at different depths for laboratory analysis (Buyanowski & Wagner, 1983) and laboratory analysis of core soil samples (Cortassa et al., 2001). Furthermore, CO_2 flux is measured where surface flux is determined from changes in gas concentrations with an enclosed chamber on the soil surface (de Jong *et al*., 1979; Cropper *et al*., 1985; Drewitt *et al*., 2002). Dugas (1993) stated that there are two commonly used chambers to measure CO_2 flux which are the Li Cor Li-6200 and the Li-6400 systems. Both these chambers have systematic errors as found by Norman *et al*. (1997), and require correction factors for accurate values. This chapter describes a method for the establishment of controlled waterlogging using a small lysimeter (about 200 kg), which allows simultaneous measurements of soil air (oxygen and carbon dioxide) within the unsaturated zone and soil water content, as well as soil temperature over the profile. These results were compared between disturbed and undisturbed soil-monoliths.

3.2 Materials and methods

The experimental system consisted of disturbed and undisturbed soils that were sampled at the experimental farm of the Department of Soil, Crop and Climate Science (University of the Free State) at Kenilworth (29°01’00’’S, 26°08’50’’E, and altitude 1417 m) in the Bloemfontein district. The monoliths were transported with a pickup truck to the glasshouse of the University of the Free State located on the main campus in Bloemfontein, South Africa. These soils were then transferred to the lysimeters.

Table 3.1 A summary of some prominent lysimeter-experiments conducted world-wide

	Reference	Country	Location	Depth (mm)	Total mass (kg)	Status of the soil	E	T	ET	∆ in soil water per layers	O_2	CO_2	Temperature	Redox potential
1	Young *et al*. (1997)	Pennsylvania State, USA	Field	4000	29391	Disturbed	X	X	X	√	X	X	X	X
2	Seyfried & Murdock (2001)	Indaho, USA	Field	1220	NA	Undisturbed	X	X	X	√	X	X	X	X
3	Martin *et al*. 2001	Phoenix, Arizona, USA	Field	610	1110	Disturbed	√	X	X	X	X	X	√	X
4	Diaz-Espejo *et al*. (2005)	Berkshire, UK	Field	250	71	Disturbed	√	X	√	X	X	X	X	X
5	Virtanen *et al.* (2010)	Helsinki, Finland	Greenhouse	1000	NA	Undisturbed	√	X	X	√	X	X	√	√
6	Johnson *et al.* (2002)	Kearney, USA	Field	1950	NA	Undisturbed	X	X	√	X	X	X	X	X
7	Lorite *et al.* (2012)	Cordoba, Spain	Field	2150	NA	Undisturbed	X	√	√	X	X	X	X	X
8	Payero & Irmak (2008)	Nebraska, USA	Field	2130	NA	Undisturbed	√	X	√	X	X	X	X	X
9	Callebaut *et al.* (1982)		Laboratory		NA	Disturbed	X	X	X	X	√	√	X	X
10	Cannell *et al*. (1980b)	Cambridge, UK	Movable Glasshouse	1350	NA	Undisturbed	√	X	X	X	X	X	√	X
11	Meyer *et al.* (1985)	Griffith, NSW	Field	1400	NA	Disturbed / Undisturbed	√	X	X	X	√	X	X	X

3.2.1 Profile description and soil analysis

Description of the soil took place *in situ* from a fresh face soil pit (Figure 3.1) using the procedure specified by the South African Agricultural Research Council-Institute for Soil Climate and Water. The soil was classification as a Bainsvlei form, according to the Soil Classification a Taxonomic System for South Africa (Soil Classification Working Group, 1991). The pedological properties are summarised in Table 3.2. The soil can be described as a deep sandy loam that drains freely towards the soft plintic horizon (Bennie *et al*., 1994). Each of the five selected diagnostic horizons was physically and chemically analysed (Table 3.3) according to procedures described by The Non-Affiliated Soil Analysis Work Committee (1990). Particle size distribution was determined with the pipette method, while the bulk density was determined with the core sampling method. The exchangeable cations and the cation exchange capacity were determined with the ammonium acetate method (1 mol dm^{-3}, pH=7). Extractable phosphorous was determined using NaH_2CO_3 (Olsen). Total N was determined with Kjeldahl and organic C with Mebius procedures. This Bainsvlei soil form is very suitable for agriculture in this semi-arid climate of the Free State province.

Table 3.2 Profile description of the Bainsvlei soil, sampled at Kenilworth Experimental Farm

Horizon	Depth (mm)	Description	Diagnostic horizon
A	0-250	Moist state; dry colour: yellowish red (5YR5/6); moist colour: reddish brown (5YR4/4); texture: fine loamy sand; structure: apedal massive; consistence: friable; few fine normal pores; water absorption: 1 second; few roots; gradual smooth transition.	Orthic
B1	250-420	Moist state; dry colour: red (2.5YR4/8); moist colour: red (2.5YR4/6); texture: fine sandy loam; structure: apedal massive; consistence: friable; few fine normal pores; water absorption: 1 second; few roots; gradual smooth transition.	Red apedal
B2	420-700	Moist state; dry colour: yellowish red (5YR5/8); moist colour: red (2.5YR4/6); texture: fine sandy loam; few fine faint black illuvial humus mottles; structure: apedal massive; consistence: friable; common fine normal pores; water absorption: 1 second; few roots; gradual wavy transition.	Red apedal
B3	700-1200	Moist state; signs of wetness: yellowish red (5YR4/6); moist colour: reddish brown (5YR4/4); texture: fine sandy clay loam; common fine faint black illuvial humus mottles; structure: apedal massive; consistence: slightly firm; common fine normal pores; water absorption: 1 second; clear wavy transition.	Soft plintite

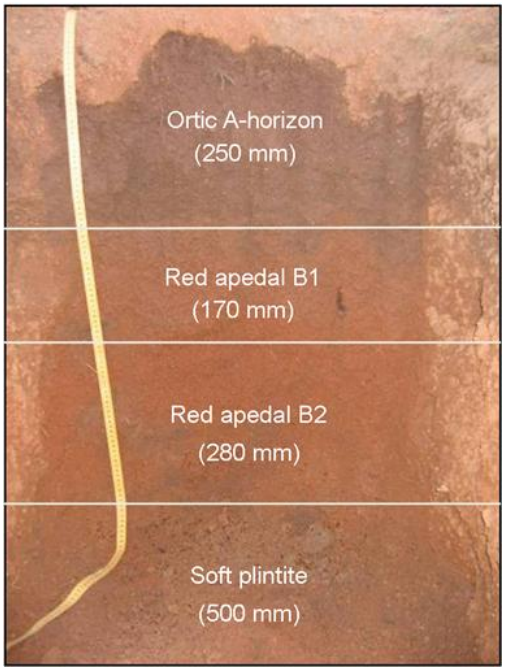


Figure 3.1 Fresh-face soil profile of the Bainsvlei form sampled at Kenilworth Experimental Farm near Bloemfontein.

Table 3.3 Soil physical and chemical properties of the Bainsvlei soil, sampled at Kenilworth Experimental Farm

Properties	**Diagnostic horizons depth (mm)**				
	A	B1	B2	B3	B4
Physical	0-250	250-420	420-700	700-1200	1200-1500
Coarse sand (2 - 0.5 mm) **%**	11.7	13.1	14.5	17.1	15.0
Medium sand (0.5 - 0.25 mm) **%**	0.8	0.6	0.3	0.5	0.9
Fine sand (0.25 - 0.106 mm) **%**	30.6	7.2	3.5	3.2	2.9
Very fine sand (0.106 - 0.05 mm) **%**	42.4	55.0	54.6	53.4	54.6
Coarse Silt (0.05 - 0.02 mm) **%**	5.8	2.6	3.3	2.4	3.9
Fine Silt (0.02 - 0.002 mm) **%**	0.6	0.8	0.8	0.6	1.2
Clay (> 0.002 mm) **%**	7.1	17.0	20.0	21.2	19.1
Bulk density (Mg m^{-3})	1.66	1.68	1.66	1.67	1.68
Texture class	loamy sand	sandy loam	sandy loam	sandy loam	sandy loam
Chemical					
Exchangeable cations:					
Ca (mg kg^{-1})	272	353	310	262	213
K (mg kg^{-1})	75	96	72	66	61
Mg (mg kg^{-1})	43	100	104	94	117
Na (mg kg^{-1})	41	39	27	20	24
CEC ($cmol_c$ kg^{-1})	3.7	4.0	4.8	4.6	3.8
pH (H_2O)	5.2	5.1	6.3	6.1	6.7
P(Olsen)	20	23	19	18	10
Extractable micronutrients: (mg kg^{-1})					
Fe	1.1	0.7	0.8	0.8	0.6
Zn	2.1	0.56	0.49	0.62	0.31
Mn	4.6	3.7	3.6	3.7	4.2
Cu	2.2	1.4	1.3	1.1	1.4
Total N (%)	0.08	0.05	0.03	0.03	0.03
Organic C (%)	0.62	0.59	0.31	0.32	0.44

3.2.2 Sampling of soil profiles

The soils were sampled using two methods; disturbed *versus* undisturbed in order to obtain a soil monolith. For the disturbed profile, the A horizon (0 - 250 mm), B1 horizon (250 - 420 mm), B2 horizon (420 - 700 mm) and the B3 horizon (700 - 1200 mm) were manually sampled and transported to the glasshouse facility.

For sampling the undisturbed monolith the soil was prepared in three successive steps. Step 1 - excavating of the undisturbed soil monolith: a cylindrical soil column with a diameter of about 400 mm was manually excavated to a depth of 1600 mm (Figure 3.2). The A-horizon (250 mm) was then manually sampled because it was already disturbed by cultivation practices employed at the farm. The rest of the column was left undisturbed, ready to install the air extraction equipment.

Figure 3.2 Series of photos demonstrating the excavating of the undisturbed soil monolith.

Step 2 - the installation of air-extraction equipment: in order to obtain air samples from the undisturbed soil monolith, three horizontal holes (8 mm in diameter with a length of 250 mm) were drilled at depths of 400 mm, 600 mm and 800 mm from the soil surface. Perforated plastic pipes (8 mm in diameter with a length of 250 mm), sealed at both ends, were installed into the drilled holes. These pipes were connected to 4 mm plastic pipes, pasted into a vertically grove along the side of the monolith as indicated in Figure 3.3. To extract an uncontaminated air sample a control valve was inserted on the end of the 4 mm plastic pipe.

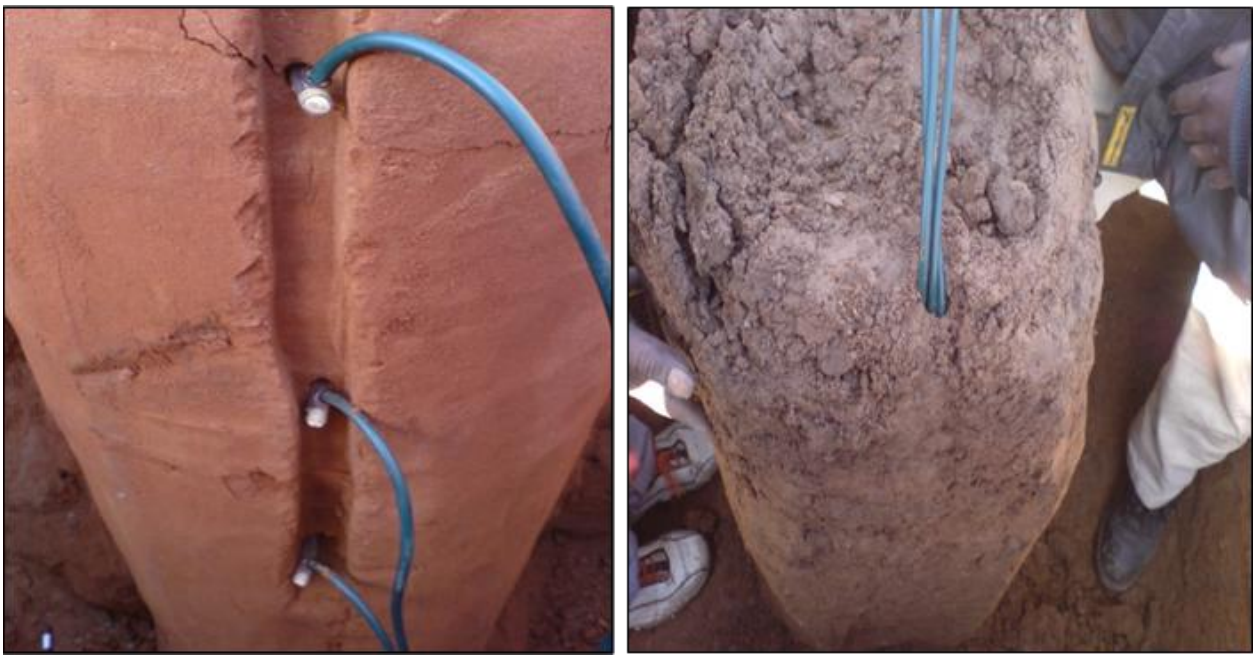


Figure 3.3 Series of photos illustrating the installation and sealing of 8 mm tubes at different depths along the monolith.

Step 3 - Sampling of the monolith: the monolith was sampled using a Class 16 polyvinyl chloride (PVC) pipe with a diameter of 300 mm, length of 1500 mm and wall thickness of 5 mm. The PVC pipe was manually pushed slowly into the soil and was excavated from around the outside of the pipe to lessen the force required. When the cylinder was inserted a metal plate was pushed underneath the monolith (Figure 3.4). For lifting and handling the monolith the metal plate was tied to the PVC pipe with ropes. The monoliths were then transported with a pickup truck to the glasshouse facility where it was further prepared for the lysimeter (Figure 3.5).

Figure 3.4 Series of photos illustrating the securing of the soil monolith.

Figure 3.5 Transporting the monolith obtained at Kenilworth experimental farm to the glasshouse facility at the University of the Free State

3.2.3 Lysimeters

3.2.3.1 Preparation of the soil monolith

The bottom of the PVC pipe, which contains the soil monolith, was filled with a 50 mm layer of gravel and covered fully with an 80% nylon shade cloth. Thereafter a PVC lid was used to seal the bottom of the PVC pipe and serve as a base of the lysimeter when revered to an upright position (Figure 3.6). The A-horizon was repacked in the lysimeter containing the previously described air extraction perforated plastic pipes at depths of 100 mm and 200 mm from the soil surface. The disturbed soil monolith was packed in similar fashion with perforated plastic pipes at depths specified for the undisturbed monolith. The depth and density of each diagnostic soil horizon was simulated during the repacking process of the monolith.

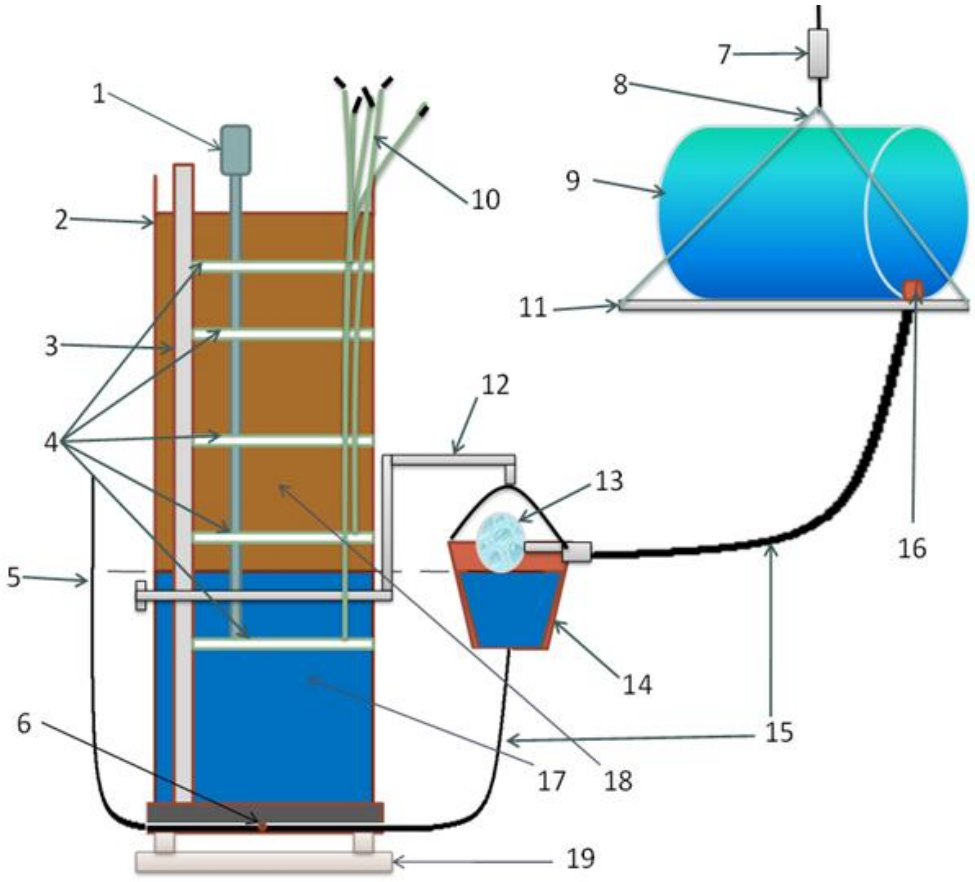
3.2.3.2 Description of the lysimeter

The lysimeter comprised of a monolith PVC column (described in the previous section), a weighing bridge and water-table height control system (Figure 3.6). The weighing bridge consisted of a load cell, Model 1260 Tedea-Huntleigh aluminium off-centre loading single point, with a capacity range of 50 - 660 kg. The load cell was connected to a Campbell Scientific CR1000 data logger. The water-table height control system consisted of three components (i) a 20 litre container fitted on a hanging platform from the roof of the glasshouse, (ii) a height-adjustable bucket attached to the PVC pipe and (iii) a manometer. Water flow from the hanging container to the monolith was controlled by a ball valve fitted in the bucket.

3.2.3.3 Management of the lysimeter

An Affrox CO_2-cylinder was connected with a plastic pipe to the opening at the bottom of the lysimeter. This was done to reduce oxygen levels in the soil to as low as possible before saturating the soil with water. Oxygen often makes air bubbles in soil that prevent soil from complete saturation with water. Carbon dioxide on the other hand is much more soluble than oxygen in water and can thereby overcome the above mentioned obstacle (Hillel, 1998).

De-aired water was used to reduce the influence of dissolved oxygen and carbon dioxide. Water was de-aired by boiling municipal water for 15 minutes using two 50 litre boilers. The 20 L container was then manually filled with the de-aired water.



1. DFM probe
2. PVC constructed lysimeter
3. Piezometer
4. 8 mm perforated plastic pipes, installed at 100 mm, 200 mm, 400 mm, 600 mm and 1200 mm depths.
5. Manometer
6. Opening, constructed with a T-plastic
7. Hanging load cell
8. Cables
9. 20 litre plastic container
10. 4 mm irrigation pipes
11. Iron plate
12. Adjustable handle bar
13. Ball valve
14. Bucket
15. 12 mm plastic pipe
16. Tap
17. Water table
18. Undisturbed soil
19. Sitting load cell

Figure 3.6 The lysimeter as constructed with the water-table height control system.

3.2.4 Temperature experiment

A two-factorial design was adopted for the temperature experiment using the monolith-lysimeter method as described in Section 3.2.3. The main treatments were two soil sampling methods (disturbed versus undisturbed profiles) and four measuring soil depths (200 mm, 400 mm, 600 mm and 800 mm from the soil surface) replicated three times over a seven-day period. The experiment lasted from 2 September 2009 to 30 November 2009 and measurements were recorded hourly (mV) using DFM capacitance probes. Temperature data was transferred from the logger with a portable device and transported to a computer via Bluetooth.

3.2.5 Internal-drainage experiment

The internal drainage process of a disturbed monolith was compared with an undisturbed monolith using the described lysimeter method. Drainage was measured between 2 September 2009 and 26 September 2009. During this period, the drainage treatment which lasted for 48 hours, was replicated three times with five days between replications. Before each drainage treatment, the lysimeters were saturated with de-aired water from the bottom of the monolith via the water supply system (Figure 3.6). Weight-loss was measured on an hourly basis during the drainage process. Internal drainage was measured with both the weighing bridge and the DFM capacitance probes. This was done to compare the two instruments and as part of the calibration process for the DFM probes.

3.2.6 Changes in O_2 and CO_2 concentrations in soil profile experiment

A two-factorial design was adopted for the soil respiration experiment using the monolith-lysimeter method as described in Section 3.2.3. The main treatments were two soil sampling methods (disturbed versus undisturbed soils) and three water-table heights (300 mm, 500 mm and 800 mm from the soil surface) replicated three times over a nine-day period. For establishing the water table treatments, the water table was set at 800 mm for three consecutive days where after it was raised to the next height. This procedure was repeated until the top layer (300 mm height) was reached. After the last measurements were taken the monolith was allowed to drain freely for five consecutive days. This drainage period was to ensure that the soil can be aerated before the start of the next replications. The soil-air experiment period lasted from 03 October 2009 to 5 November 2009.

Oxygen and carbon dioxide was measured with a MultipleRAE IR gas monitor instrument, starting at 10am each day (Figure 3.7). The instrument consisted of an oxygen and carbon dioxide sensor, internal pump, lithium-ion rechargeable battery and charger, alkaline adapter and ProRAE software for Windows. In order to extract uncontaminated air samples from the monolith, the tip of the 4 mm plastic pipe was inserted into the suction hole of the MultipleRAE IR gas monitor instrument before opening the control valve for air to be extracted. The pump extracted air at a maximum rate of 250 cm^3 min^{-1} and minimum of 150 cm^3 min^{-1} from the monolith for about thirty seconds until O_2 and CO_2 readings stabilised on the display of the instrument.

Figure 3.7 MultipleRAE IR gas monitor instrument, measuring O_2 and CO_2 from different depths in the monolith.

3.2.7 Calibration procedures

3.2.7.1 DFM probes

The DFM capacitance probes were calibrated for both temperature and volumetric soil water content at the same time.

Temperature calibration: The 1200 mm probes were placed in a PVC cylinder with a length of 1500 mm and diameter of 300 mm, filled with distilled water and placed inside a climate controlled room at the main campus of the University of the Free State (Figures 3.8). The air temperature in the climate controlled room was set at 5°C, 15°C, 25°C and 35°C respectively for 48 hours at a time. Probes were programmed to measure water content and temperature at 15 minute intervals. The variation between DFM temperature measured in the PVC cylinders compared to actual temperature was within an acceptable 0.5°C.

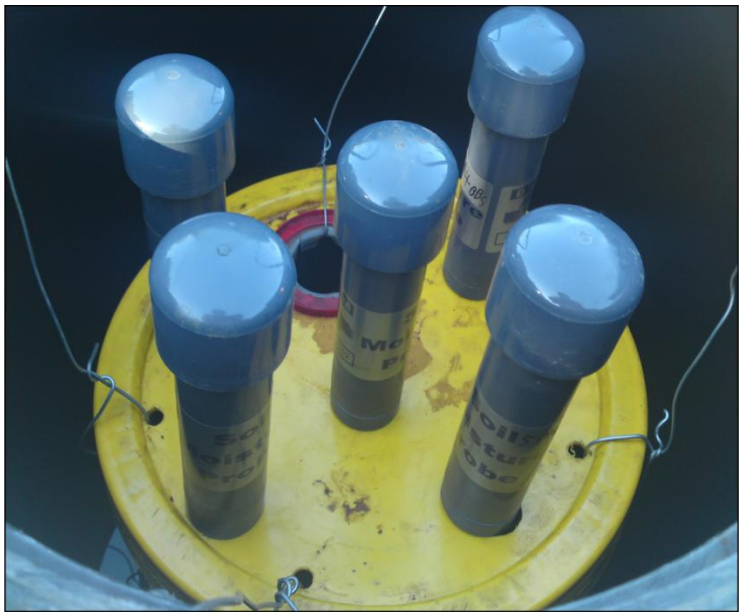


Figure 3.8 Probes placed in a PVC cylinder with distilled water inside a climate controlled room.

Temperature correction of water sensors: According to Fares *et al*. (2007) and Seyfried & Murdock (2001) the effect of temperature on capacitance readings will increase as the scaled frequency of the probe increase. There is no definite barrier as a gradual increase will be seen from ±80 MHz. The frequency used in DFM probes is between 100 and 120 MHz (Mercker, 2012). This can explain why temperature has an effect on the DFM probe readings. Figure 3.9 predicts the slope of regression for the rate of change between temperature and DFM readings. The slope was then used to develop a linear correction equation (Equation 3.1), similar to the one used by Fares *et al.* (2007):

$$DFM_{NT} = (25 - DFM_T)^* x + DFM_R \quad 3.1$$

where DFM_{NT} is the converted water content after the temperature were normalised to a ambient temperature of 25°C, DFM_T is the measured DFM temperature, x is the slopes of the different probes (water % to temperature °C), and DFM_R is the measured DFM water content in the water columns. This equation was used to convert DFM water readings at different temperatures, to a commonly-used standard ambient temperature of 25°C (Wikipedia Contributor, 2012). The temperature correction equation was tested with an independent data set and the results confirmed the statistical significance of the equation (Appendix 3.1).

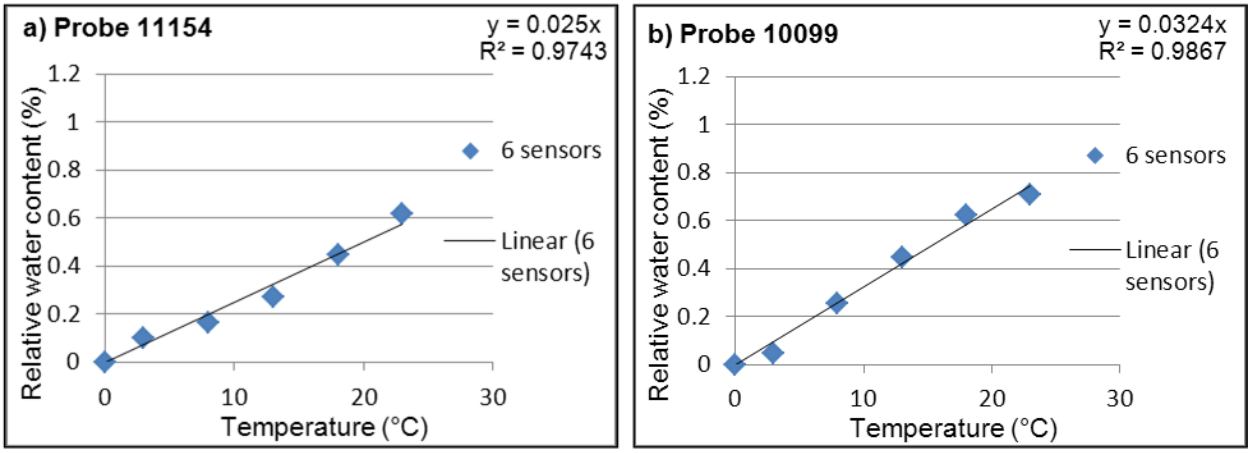


Figure 3.9 The effect of temperature variation on DFM water content (%) for probe a) 11154 and b) 10099.

Soil water calibration: The calibration procedure used was based on the method described by Zerizghy *et al*. (2013). Accordingly, three plastic drums (210 litres) with a length of 920 mm and diameter of 270 mm were packed with the A-horizon and another three drums with the B1 horizon of the Bainsvlei soil (see Table 3.3 for bulk densities used). The drums were equilibrated at three volumetric water contents, *viz.* air dried (2%), near drained upper limit (16%) and saturation (36%). Probes were then inserted into the drums and left for 24 hours to stabilise, where after measurements were recorded for another 24 hours at 15 minute intervals. Unfortunately, only the bottom three sensors of each probe were able to fit in the drums because of the probe lengths.

The data from the two horizons were combined to calculate an average equation for each probe (Table 3.4).

Table 3.4 Calibration equation and R^2 values of each probe, calibrated in different soils

Probe nr.	Calibration equation	R^2
11154	y = 0.5184x – 3.275	0.995
10099	y = 0.5224x – 2.952	0.998

3.2.7.2 Weighing bridge

There was a temperature effect on the final mass readings from the load cells that was corrected by deducting temperature correction factors for both lysimeters. Temperature and load cell readings were plotted against one another from which the slope of regression was taken as the rate of change for the readings of each load cell. The average temperature of 25 °C was taken as the point of conversion for all readings. Equation 3.1 was applied to derive a linear correction for both lysimeters. The calibration procedures involved loading known weight that ranges between 5 kg to 150 kg onto the load cells until mass measurements was satisfactory.

The volumetric water content during the drainage process measured with the DFM probes was compared with the measured values of the weighing bridge for the disturbed and undisturbed soils (Table 3.5). The results in Figure 3.10 reveal that the volumetric water content was significantly similar (P value of 0.3979) amongst the measuring devices, irrespectively of the soils.

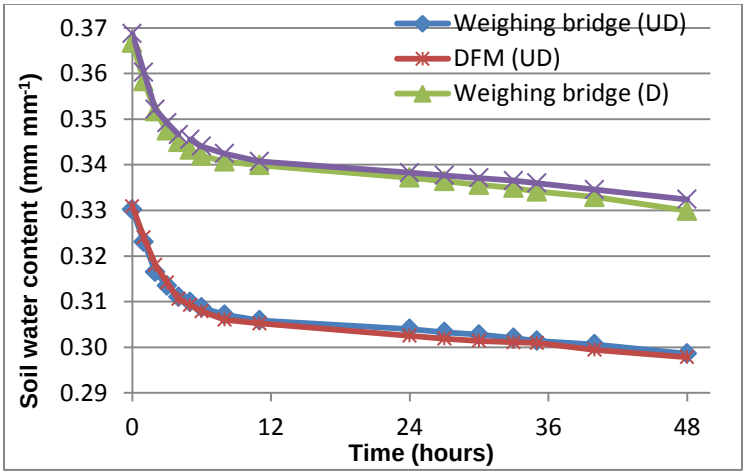


Figure 3.10 Changes in water content during the drainage test over a 120 hour period for both the disturbed and undisturbed monolith-lysimeters.

Table 3.5 Water content (mm mm^{-1}) representing the overall profile at each water-table height.

Water-table depth (mm)	300	500	800	Mean
Disturbed	0.120	0.126	0.117	0.121
Undisturbed	0.154	0.146	0.148	0.149
Mean	0.137	0.136	0.133	0.135

3.2.7.3 MultipleRAE IR gas monitor instrument

The sensors of the MultipleRAE IR gas monitor instrument were calibrated in a two-step process using fresh air and span gas. Span gas is a commercial product which contains a known concentration of given gas. For this experiment a four-gas mixture in a 34 ℓ cylinder fitted with a gas regulator and tubing was used, containing 50% LEL methane, CO_2, 20.9% oxygen and CO. The MultipleRAE IR gas monitor instrument was calibrated before each replication.

3.2.8 Statistical analysis

Temperature data was analysed using the software system STATISTICA, version 12 (Statsoft, Inc., 2013). The t-test was used at the 5% level of significance to test for significant differences. For the internal drainage experiment, the two-tailed parallel t-test was used to determine significant differences (SAS 2006). In the case of the aeration experiment, oxygen and carbon dioxide data was also analysed with STATISTICA, version 12 (Statsoft, Inc., 2013). Tukey's honestly significant difference test at the 5% level of significance was used to determine statistically significant differences between treatment means, where the analysis of variance (ANOVA) indicated significant effects of the tested variables.

3.3 Results

3.3.1 Soil temperature

Table 3.6 summarises the statistical results of mean soil temperatures over a seven-day period between the disturbed and undisturbed soils at four soil depths. There was no significant interaction between the soil-sampling methods and soil depths over the profiles. The three replications also showed no significant differences. However, there were highly significant differences for the soil-sampling method, as well as for the soil-depth treatments. The temperature of the disturbed soil was 1.7°C lower than the undisturbed soil.

Considering the soil-depth treatment, the results indicate that the temperature gradually increases from the surface to the bottom of the profiles with about 1°C. The results showed

further that temperatures at the 200 mm depth differed significantly from the 400 mm, 600 mm and 800 mm depth. Temperature at 400 mm were not significantly different from the 600 mm depth, but were significant different from the 800 mm depth. The two bottom layers were not significantly different from each other with temperatures of 23.2°C and 23.3°C, respectively. These temperatures were about 1°C higher than that of the experimental mean of 22.5°C.

Table 3.6 Statistical results of the mean temperatures (°C) for the soil-sampling methods and soil-depth treatments

Sensor Depth (mm)	**200**	**400**	**600**	**800**	**Mean**
Disturbed	21.7	22.1	22.3	22.3	22.1[b]
Undisturbed	23.3	23.7	24.0	24.3	23.8[a]
Mean	22.5[c]	22.9[b]	23.2[ab]	23.3[a]	22.5
	P > F				
Sampling methods	<.0001				
Depth	0.0004				
Replications	0.2979				
Interaction	0.4136				

Means followed by the same letter in either rows or columns do not differ from each other at the 5% level of significance.

3.3.2 Internal drainage

The results in Figure 3.11 compare drainage curves for the disturbed and undisturbed soils, expressed in volumetric soil-water content (mm mm^{-1}) over time (hourly). The fitted polynomial functions presented in Figure 3.11 had coefficient of determinations that were close to unity with errors of 0.0015 and 0.0007 for the disturbed and undisturbed soils, respectively.

The t-test revealed that the internal-drainage process between the disturbed and undisturbed soils was highly significantly different with a P value of 0.0001. The disturbed soil was constantly wetter over the total drainage period than the undisturbed soil; mean soil-water content was 0.342 mm mm^{-1} for the disturbed soil and 0.308 mm mm^{-1} for the undisturbed soil. At the point of field saturation, the mean soil water content of the disturbed soil (0.366 mm mm^{-1}) was 0.037 mm mm^{-1} higher than the undisturbed soil, *i.e*. 44 mm per 1200 mm soil. Both these saturation points correspond well with the calculated porosity of 0.37 mm mm^{-1} due to the injection of CO_2 before recharging of the profiles.

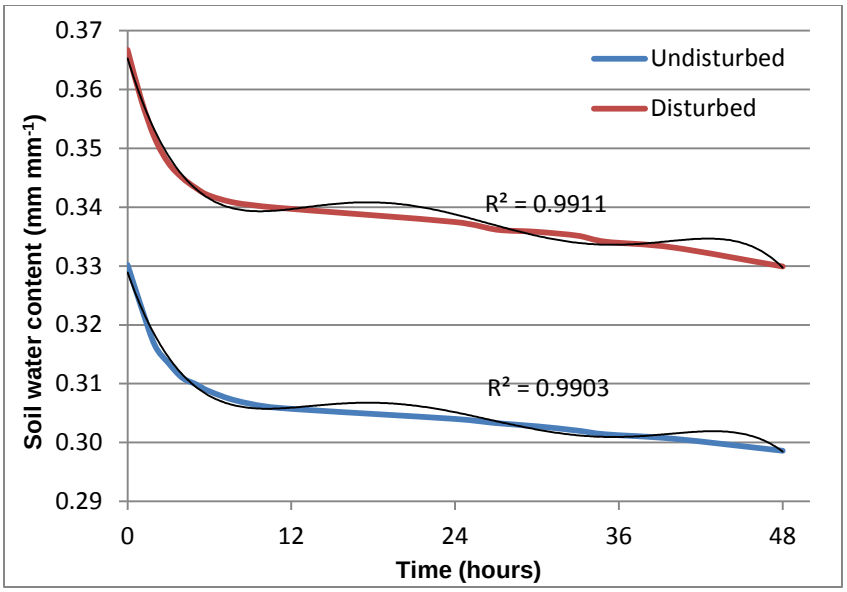


Figure 3.11 Changes in soil-water content during the drainage period measured by the weighing bridge for the disturbed and undisturbed profiles.

3.3.3 Changes in oxygen and carbon dioxide concentrations in soil profiles

The statistical results (ANOVA) on the oxygen concentrations, as affected by the soil-sampling methods and the water-table height treatments, are summarised in Table 3.7. There was a highly significant difference for the soil-sampling methods, but not for the water-table height treatments. Of particular interest, there was a highly-significant interaction between the soil-sampling methods and water-table heights.

O_2: Considering the interaction between the soil-sampling methods and water-table heights, three main findings were made. Firstly, the O_2 concentration over the disturbed soil was consistently higher than over the undisturbed soil. The difference in O_2 concentration between the disturbed and undisturbed soils gradually increased from 0.16% at the 800 mm water-table height to 0.46% at the 300 mm height. Secondly, the O_2 concentration at the 300 mm water-table height differed significantly from the 500 mm and 800 mm water-table heights for both the disturbed and undisturbed soils. Thus, it seems that the O_2 concentrations were only affected where the soil was close to saturation. Thirdly, an inverse relationship between the disturbed and undisturbed soils with regard to the O_2 concentrations was observed; Oxygen concentration gradually decreased with 0.18% for the undisturbed profile as the water-table height was raised from 800 mm to 300 mm, while the disturbed profile increased with 0.12%.

CO_2: The statistical results of the CO_2 concentrations (Table 3.7) revealed that there were highly significant differences between soil sampling methods as well as between water-table height treatments. However, there was a highly-significant interaction between the soil sampling methods and water-table heights. The interaction between the soil-sampling methods and water-table heights revealed two main findings. Firstly, the CO_2 concentration over the disturbed profile was consistently lower than over the undisturbed profile. The difference in CO_2 concentration between the disturbed and undisturbed profiles increased from 152 mg ℓ^{-1} at the 800 mm water-table height to 4795 mg ℓ^{-1} at the 300 mm. Secondly, the CO_2 concentration at the 300 mm water-table height differed significantly from the 500 mm water-table height for both the disturbed and undisturbed profiles. The increase in CO_2 concentrations for the undisturbed soil was more prominent than the disturbed soil. It is interesting to note that the CO_2 concentrations for the disturbed soil were between 2.5 and 3 times lower than the undisturbed soil.

Table 3.7 Statistical results on the mean O_2 and CO_2 concentrations of the unsaturated zone for the soil-sampling methods and water-table height treatments.

		Water-Table Height (WT)			Mean
Soil gas	**Soil Treatment (ST)**	**300 mm**	**500 mm**	**800 mm**	
O_2 (%)	**Undisturbed**	20.33[a]	20.46[b]	20.51[b]	20.44[a]
	Disturbed	20.79[d]	20.69[c]	20.67[c]	20.72[b]
	Mean	20.56[a]	20.57[a]	20.59[a]	
CO_2 (mg l^{-1})	**Undisturbed**	7100[d]	2981[c]	1552[a]	3877[a]
	Disturbed	2305[b]	1239[a]	1400[a]	1648[b]
	Mean	4702[a]	2110[b]	1476[c]	
P-values	**O_2**	**CO_2**			
Soil sampling treatment (SST)	0.001	0.001			
Water-table height (WT)	0.260	0.001			
Interaction SST x WT	0.001	0.001			

Means followed by the same letter in either rows or columns do not differ from each other at the 5% level of significance.

3.4 Discussion

The main purpose of this study was to develop a small lysimeter system to study the effect of rising water tables on O_2 and CO_2 concentration in the unsaturated zone. Another objective was to investigate the suitability of using a disturbed or undisturbed soil sample in the proposed lysimeter. With respect to the main purpose, sufficient evidence was provided that demonstrates that it was possible to construct the small lysimeter system using a PVC pipe, electronic weighing bridge and the water-table control system. Thus, the remainder of the discussion will focus on the evaluation of the small lysimeter system using three experiments concerning soil temperature, internal drainage and the change in soil oxygen and carbon dioxide concentrations.

Regarding the soil temperature experiment, the lysimeters were placed on the floor in the temperature-controlled glasshouse which implies that the monoliths were exposed to the temperature fluctuation of the surrounding environment. During the soil-temperature experiment in September the mean temperature increased from 22.5°C at the soil surface to 23.5°C at the bottom of the profiles. The 1°C decrease in temperature over the profiles compares well with values measured in the field near Bloemfontein by Nhlabatsi (2010). He predicted with a 50% probability that the topsoil (150 mm from the soil surface) and bottom (750 mm) will be 16°C and 17°C, respectively. The difference over the profiles was similar, irrespectively of whether it was measured in a glasshouse-lysimeter or in the field. On the other hand, mean profile temperature was 5°C warmer in the glasshouse-lysimeter than in the field. This was expected because of the fixed temperature at 20°C in the glasshouse. This temperature difference may have an impact on both soil water evaporation and transpiration studies, because more energy is available as latent heat in the glasshouse compared to the field (Hillel, 1998). This problem can be overcome by insulation of the wall of the lysimeter column as suggested by Virtanen *et al*. (2010). Considering the sampling-method treatments, the results showed that temperature was significantly higher (1.7°C) in the undisturbed (23.8°C) profile compared to the disturbed (22.1°C) profile. According to Tuli *et al*. (2005) this can be ascribed to soil structure changes induced by the disturbance. When soils are disturbed, heat flow processes are also disturbed due to the changes in pore-geometrical characteristics such as tortuosity, connectivity and constriction (Vogel, 1997; Wildenschild *et al*., 2005). Thus, it is obvious that the thermal conductivity of the disturbed profile was reduced probably due to the lost in contacts between soil particles.

On the subject of the internal-drainage experiment, two main findings can be made from the 48-hour drainage experiment. Firstly, the drainage experiment succeeded in showing that the drainage process, as indicated by the shape of the drainage curves, was similar for the disturbed and undisturbed soils. The drainage process of both these treatments compared also well with the *in situ* drainage results of Chimungu (2009), determined at the same site where the disturbed and undisturbed soils were sampled. The pores that represent the drainage over 48 hours were regarded as easily drainable pores. The easily drainable pores for the disturbed, undisturbed and *in situ* profiles were 10.1%, 9.6% and 10.3%, respectively. The second main finding relates to the water retention caused by disturbance of the soils during preparation of the monoliths. This can be derived from the actual water contents after 48 hours of drainage which amounted to 0.329 mm mm^{-1}, 0.298 mm mm^{-1} and 0.276 mm mm^{-1} for the disturbed, undisturbed and *in situ* profiles, respectively. Using the *in situ* profile as reference, the disturbed soil showed 19% higher water content and the undisturbed soil 8%. The slightly higher water retention for the undisturbed soil can be ascribed to some disturbance of the A-horizon compared to the undisturbed A-horizon of the *in situ* profile. This also explains the slight difference between the disturbed and undisturbed soil as well as the higher difference between the disturbed and *in situ* profiles. The difference in the water retention of the soils can be attributed to an increase in the meso and micro-pore fractions caused by mechanical disturbance during the preparation phase of the monoliths. This can be proved by subtracting the water content (0.097 mm mm^{-1}) determined at -1500 kPa for the *in situ* profile, from the water contents of the disturbed and undisturbed soils measured at 48 hours of drainage. These differences in water content represent the remaining meso and micro-pore fractions, which were 0.232 mm mm^{-1}, 0.201 mm mm^{-1} and 0.179 mm mm^{-1} for the disturbed, undisturbed and *in situ* profiles, respectively. In a previous study done by Tuli *et al*. (2005), differences were reported between disturbed and undisturbed soils regarding water permeability. Differences in soil characteristic and permeability functions were solely due to soil structure. The elimination of soil structure in the loamy disturbed sample changed the soil water retention parameters with the change of pores.

Concerning the soil O_2 and CO_2 concentration experiment, sufficient evidence was provided that is was possible to extract soil air samples at different depths in the unsaturated zone. For example, the mean O_2 concentrations (20.44%) as well as the CO_2 concentrations (3877 mg L^{-1}) for the unsaturated soil sample compares well to those reported for a similar soil type studied by Lal & Shukla (2004). Furthermore, the results indicated that it is better to use an undisturbed soil sample for studying O_2 and CO_2 concentrations in soils. This conclusion is supported by the

results which showed that the disturbed soil sampling method induced a significantly higher O_2 and lower CO_2 concentration compared to the undisturbed sampling method. For example, CO_2 concentrations for the disturbed soil were between 2.5 and 3 times lower than the undisturbed soil. Tuli *et al*. (2005) also recommended undisturbed soils as they argueed that the disturbance of the soil had caused an increase in the macro-pore fraction of the soil. Soils with large pores promote more rapid oxygen diffusion into and through the soil, and carbon dioxide movement out of the soil (Watson & Kelsey, 2005). In a similar study Meyer *et al*. (1985) found that CO_2 concentrations of a disturbed soil was almost three times that of the undisturbed soil due to greater pore space and better drainage. In the internal-drainage experiment it was found that mechanical disturbance of the soil did not caused a difference in the easily drainage porosity. Thus, the differences between the O_2 and CO_2 concentrations in the disturbed and undisturbed soils can only be ascribed to the connectivity of the pores. In this case the disturbance probably induced a higher connectively within the so-called easily drainable pores.

3.5 Conclusion

This chapter has provided a method for constructing a small weighing lysimeter which comprises of a PVC pipe, electronic weighing bridge and the water-table control system. This system caters for studies that focus on water tables and their effect on aeration of the unsaturated zone. As shown, additional information on soil water and temperature distribution can also be obtained by inserting capacitance water sensors (for example DFM probes) in the soils.

Three experiments concerning soil temperature, internal-drainage and the change in soil oxygen and carbon dioxide concentrations were conducted to evaluate the small weighing lysimeter system. From these experiments two main conclusions were made:

Firstly, the similar shape of the drainage curves proved that the drainage processes was similar for both the disturbed and undisturbed soils. This was based on the fact that the easily drainable pores were similar for both the disturbed and undisturbed soils. However, the water retention was significantly higher in the disturbed soil.

Secondly, it was found that a undisturbed soil is better to use for studying O_2 and CO_2 concentrations in soils. This conclusion is supported by the results which showed the disturbed soil sampling method induced a significantly higher O_2 and lower CO_2 concentration compared to the undisturbed sampling method.

Overall, the results of this chapter indicate that the proposed small weighing-lysimeter system contributed towards the understanding of a very important subject, namely soil aeration.

CHAPTER 4

OXYGEN AND CARBON DIOXIDE PROFILES IN UNDISTURBED SOILS WITH WATER TABLES

4.1 Introduction

Globally there is a growing need to understand how the unsaturated zone reacts to anthropogenic impacts, not only because it is either the largest sink or source of carbon, but also because it supports the growth of natural plants and crops. For example, over-irrigation results in the accumulation of water in the sub-soil which physically reduces the soil volume of the unsaturated zone. Uncontrolled reduction of the unsaturated zone will unfortunately lead to water logging. This is a huge problem as there is an estimated 230 million ha of irrigated land in the world available of which 20% is seriously affected by waterlogging (Ghassemi *et al*., 1995). South Africa, with an estimated 1.3 million ha of irrigated land, is not excluded from this phenomenon (Water Research Commission, 1996). An estimated 20% (260 000 ha) of these irrigated soils have shallow water tables in or just below the rooting depth. Soils from most irrigation schemes become waterlogged before reaching their full potential due to the presence of frequently high water tables (Backeberg & Groenewald, 1995). The reasons for waterlogged soil are poor water management which relates to unsuitable soils, topography, inefficient irrigation systems, blocked drainage systems and other infrastructure deficiencies (Perret & Touchain, 2002).

Most of the recent research on soil aeration focussed on the direct influences of O_2 deficiencies on plant growth (Belford *et al.,* 1985). This is understandable because a lack of O_2 will impair root respiration, water uptake and nutrient uptake (Meyer *et al*., 1985). These processes are essential for optimum germination and to sustain growth and yield of crops. Oxygen shortages for crops are mainly determined by the oxygen diffusion rate (ODR) technologies such as described by Glinski & Stepniewski (1985). These measurements are crop specific as indicated by Lal & Shukla (2004). They showed, for example, that the limiting ODR value for rye and barley are 50 µg m^{-2} s^{-1} and 25 µg m^{-2} s^{-1}, respectively. However, these types of measurements do not provide information on the distribution of O_2 and CO_2 concentrations of the unsaturated zone.

There is a distinct lack on information regarding the distribution of O_2 and CO_2 concentrations in the unsaturated zone of water table soils. Most measurements are conducted at or near the soil surface as described by various authors (de Jong *et al*., 1979; Cropper *et al*., 1985; Drewitt *et al*., 2002). However, the above mentioned measurements are still limited to surface CO_2 fluxes lacking details regarding subsurface CO_2 dynamics (Parkin & Kaspar, 2004). The composition of soil air is highly variable and depends on numerous factors such as texture, structure, bulk density, water content, porosity and land-use conditions (Lal & Shukla, 2004). Despite this complexity, only a few references were found in the literature on O_2 and CO_2 profiles of the unsaturated zone with regards to water table conditions (Lal & Taylor., 1969; Turcu *et al*., 2005; Siqinbatu *et al*., 2013). Therefore, the objective of this chapter was to quantify the effects of a rising water table over time on oxygen and carbon dioxide concentrations in the unsaturated zone of five soils in the absence of crops.

4.2 Materials and methods

4.2.1 Experimental design

The five soils were: a sandy Hutton, loamy-sand Hutton, Bainsvlei, Sepane and Valsrivier. A two-factorial experimental design was adopted for each of the soils. The main treatments were four water-table heights, *viz*. 800 mm, 500 mm, 300 mm and 150 mm from the soil surface kept constant over a 6-day period. For establishing the water table treatments, the water table was set at each height for six consecutive days where after it was raised to the next height. This procedure was replicated three times during this study. Treatments were replicated three times over a period of 92 days with a ten day recovery period between replications. During this recovery period the soils were allowed to drain freely. This drainage period was to ensure that the soil can be aerated before the start of the next replications. The ten days was needed for the clay soils to drain completely as all five soil forms were treated simultaneously. This experiment was conducted from 17 January to 17 April 2010.

4.2.2 Profile sampling, description and soil analysis

The five undisturbed soil profiles were sampled using the three-step method described in Chapter 3 (see section 3.2.2). The first profile was sample at the experimental farm of the Department of Soil, Crop and Climate Sciences (University of the Free State) at Kenilworth (29°01’00’’S, 26°08’50’’E, and altitude 1417 m) in the Bloemfontein district. The remaining four profiles were sampled at the Orange-Riet River Irrigation Scheme on the property of Mr. Krause, Mr. Mulke and Mr. Galama as demarcated with red dots in Figure 4.1. The monoliths were then

transported with a pickup truck to the glasshouse of the University of the Free State located on the main campus in Bloemfontein, South Africa.

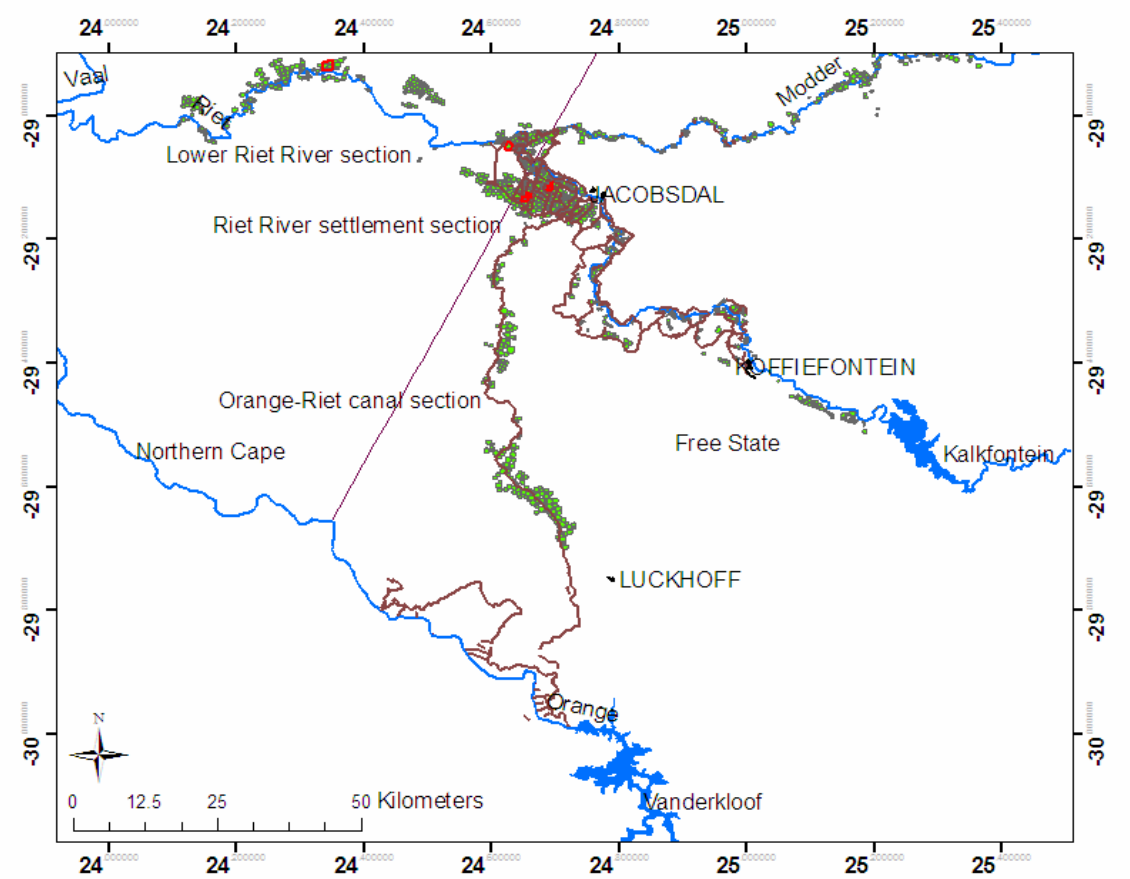


Figure 4.1 Lay-out of Orange-Riet River Irrigation Scheme where four of the soil profiles were sampled.

Description of the soil took place *in situ* from fresh face profiles (Figure 4.2 a-d) using the procedure specified by the South African Agricultural Research Council-Institute for Soil Climate and Water. The soils were classified as a) Two Hutton forms (sandy Hutton and loamy-sand Hutton), b) Bainsvlei form, c) Sepane form and d) Valsrivier form according to the Taxonomic Soil Classification System for South Africa (Soil Classification Working Group, 1991). The physical and chemical soil properties are summarised in Table 4.1 to Table 4.5 for the respective soil forms. The physical analysis included particle size distribution (pipette method) and bulk density (core sampling method), while the chemical analysis consisted of exchangeable cations and the cation exchange capacity (ammonium acetate method; 1 mol dm^{-3}, pH=7), pH (water) and P (Olsen). Extractable phosphorous was determined using NaH_2CO_3

(Olsen). Total N was determined with Kjeldahl and organic C with Mebius procedures. All above methods are available in The Non-Affiliated Soil Analysis Work Committee (1990).

Both the Hutton soils (Figures 4.2 a and 4.2 b) consisted of a orthic topsoil on a red Apedal subsoil on unspecified material with only the clay content being different between the two soils. These deep soils are generally perceived as highly suitable for irrigation (Class 1) due to their excellent internal and external drainage characteristics (Soil Classification Working Group, 1991). The Bainsvlei soil can be described as a deep sandy-loam that drains freely towards the soft plintic horizon. Soft plintic horizons are renowned for their low permeability and therefore poor external drainage. Due to the drainage restrictions these soils are regarded as a Class 3 soil type, thus less suitable than the Hutton soils for irrigation (Van Rensburg *et al*., 2012). However, they are seen as prime soils for the cultivation of field crops under dry land conditions. The two clayey soils have a similar sequence of horizons, namely an orthic A, followed by a pedocutanic B and then unconsolidated material. Due to the strong structure and high clay content of both these soils, internal drainage is seen as highly restricted and hence they are generally grouped under class 4 (Valsrivier) and Class 5 (Sepane) irrigable soils. The clay contents were lower in the Sepane soil compared to the Valsrivier soil, which is surprising as the Sepane soil has signs of wetness which is assosiated with poor drainage. This can be explained by the fact that the Sepane soil was located in a depression, recieving runoff water as well drainage water from nearby artificial drains. The water table was very shallow as can be seen in Figure 4.2 c, causing reducing conditions in the unconsolidated material. The Valsrivier soil was sampled from a well-managed centre pivot with no water table present, with no signs of wetness present (Figure 4.2 d).

Figure 4.2 Fresh-faced profiles of the a) Hutton soil, sampled at the farm of Mr. Mulke near Jacobsdal, b) Bainsvlei soil, sampled at Kenilworth Experimental Farm near Bloemfontein, c) Sepane, sampled at the farm of Mr. Galama near Jacobsdal and d) Valsrivier soil, sampled at the farm of Mr. Mulke.

Table 4.1 Soil physical and chemical properties of the sandy Hutton soil

Properties	Diagnostic horizons depth (mm)				
	A	B1	B2	B3	B4
Physical	0-250	250-420	420-700	700-1200	1200-1500
Coarse sand (2 - 0.5 mm) %	2.4	2.7	3.6	3.0	4.4
Medium sand (0.5 - 0.25 mm) %	38.5	37.0	40.0	42.4	43.2
Fine sand (0.25 - 0.106 mm) %	41.3	41.2	38.0	37.1	34.1
Very fine sand (0.106 - 0.05 mm) %	11.4	11.4	10.3	10.1	9.0
Coarse Silt (0.05 - 0.02 mm) %	0.6	0.3	0.3	0.8	0.3
Fine Silt (0.02 - 0.002 mm) %	4.0	2.0	3.0	2.0	2.0
Clay (> 0.002 mm) %	3.0	6.0	6.0	6.0	6.0
Bulk density (Mg m^{-3})	1.617	1.607	1.607	1.607	1.613
Texture class	sand	sand	sand	sand	sand
Saturated hydraulic conductivity (mm h^{-1})	52	33	33	33	26
Chemical					
Exchangeable cations:					
Ca (mg kg^{-1})	554	438	408	428	480
K (mg kg^{-1})	90	112	156	172	174
Mg (mg kg^{-1})	180	226	244	264	292
Na (mg kg^{-1})	54	78	76	46	30
CEC ($cmol_c$ kg^{-1})	5.3	5.5	6	7.8	8
pH (H_2O)	7.58	7.68	7.57	7.88	7.64
P(Olsen)	44	34	18	15	17
Extractable micronutrients: (mg kg^{-1})					
Fe	4.58	4.62	3.46	3.42	3.72
Zn	0.38	0.16	0.16	0.32	0.28
Mn	5.04	4.08	3.9	2.78	2.16
Cu	0.16	0.18	0.16	0.16	0.18
Total N (%)	0.04	0.03	0.03	0.03	0.02
Organic C (%)	0.91	0.71	0.46	0.38	0.3

Table 4.2 Soil physical and chemical properties of the loamy sand Hutton

Properties	**Diagnostic horizons depth (mm)**				
	A	B1	B2	B3	B4
Physical	0-250	250-420	420-700	700-1200	1200-1500
Coarse sand (2 - 0.5 mm) %	2.8	3.3	3.7	3.8	3.6
Medium sand (0.5 - 0.25 mm) %	35.8	32.6	35.0	35.8	32.2
Fine sand (0.25 - 0.106 mm) %	44.3	43.8	40.8	40.4	43.5
Very fine sand (0.106 - 0.05 mm) %	9.7	9.4	9.7	8.6	9.7
Coarse Silt (0.05 - 0.02 mm) %	2.4	2.3	0.8	0.6	0.8
Fine Silt (0.02 - 0.002 mm) %	2.0	2.0	2.0	2.0	2.0
Clay (> 0.002 mm) %	4.0	8.0	10.0	10.0	10.0
Bulk density (Mg m^{-3})	1.594	1.629	1.629	1.629	1.671
Texture class	sand	loamy sand	loamy sand	loamy sand	loamy sand
Saturated hydraulic conductivity (mm h^{-1})	50	31	31	31	23
Chemical					
Exchangeable cations:					
Ca (mg kg^{-1})	700	730	898	542	548
K (mg kg^{-1})	98	114	92	138	176
Mg (mg kg^{-1})	98	120	164	174	306
Na (mg kg^{-1})	50	56	66	46	46
CEC ($cmol_c$ kg^{-1})	5	6.8	6.8	2	2.6
pH (H_2O)	6.55	6.86	7.32	7.84	8.27
P(Olsen)	38	34	48	29	12
Extractable micronutrients: (mg kg^{-1})					
Fe	3.74	0.4	0.38	1.34	1.44
Zn	2.2	0.7	0.26	0.28	0.2
Mn	9.92	3.18	2.22	1.64	0.66
Cu	0.48	0.34	0.32	0.38	0.42
Total N (%)	0.04	0.03	0.03	0.3	0.2
Organic C (%)	1.51	0.95	0.78	0.66	0.58

Table 4.3 Soil physical and chemical properties of the Bainsvlei soil

Properties	**Diagnostic horizons depth (mm)**				
	A	B1	B2	B3	B4
Physical	0-250	250-420	420-700	700-1200	1200-1500
Coarse sand (2 - 0.5 mm) **%**	11.7	13.1	14.5	17.1	15.0
Medium sand (0.5 - 0.25 mm) **%**	0.8	0.6	0.3	0.5	0.9
Fine sand (0.25 - 0.106 mm) **%**	30.6	7.2	3.5	3.2	2.9
Very fine sand (0.106 - 0.05 mm) **%**	42.4	55.0	54.6	53.4	54.6
Coarse Silt (0.05 - 0.02 mm) **%**	5.8	2.6	3.3	2.4	3.9
Fine Silt (0.02 - 0.002 mm) **%**	0.6	0.8	0.8	0.6	1.2
Clay (> 0.002 mm) **%**	7.1	17.0	20.0	21.2	19.1
Bulk density (Mg m^{-3})	1.66	1.68	1.66	1.67	1.68
Texture class	loamy sand	sandy loam	sandy loam	sandy loam	sandy loam
Saturated hydraulic conductivity (mm h^{-1})	27	17	17	17	10
Chemical					
Exchangeable cations:					
Ca (mg kg^{-1})	272	353	310	262	213
K (mg kg^{-1})	75	96	72	66	61
Mg (mg kg^{-1})	43	100	104	94	117
Na (mg kg^{-1})	41	39	27	20	24
CEC ($cmol_c$ kg^{-1})	3.7	4.0	4.8	4.6	3.8
pH (H_2O)	5.2	5.1	6.3	6.1	6.7
P(Olsen)	20	23	19	18	10
Extractable micronutrients: (mg kg^{-1})					
Fe	1.1	0.7	0.8	0.8	0.6
Zn	2.1	0.56	0.49	0.62	0.31
Mn	4.6	3.7	3.6	3.7	4.2
Cu	2.2	1.4	1.3	1.1	1.4
Total N (%)	0.08	0.05	0.03	0.03	0.03
Organic C (%)	0.62	0.59	0.31	0.32	0.44

Table 4.4 Soil physical and chemical properties of the Sepane soil

Properties	**Diagnostic horizons depth (mm)**				
	A	B1	B2	B3	B4
Physical	0-250	250-420	420-700	700-1200	1200-1500
Coarse sand (2 - 0.5 mm) %	2.4	0.9	2.3	1.6	1.4
Medium sand (0.5 - 0.25 mm) %	5.8	3.4	3.3	2.1	1.3
Fine sand (0.25 - 0.106 mm) %	22.1	19.2	15.0	10.0	13.8
Very fine sand (0.106 - 0.05 mm) %	15.9	16.7	18.1	21.6	22.7
Coarse Silt (0.05 - 0.02 mm) %	6.8	6.4	7.9	7.7	3.6
Fine Silt (0.02 - 0.002 mm) %	16.0	19.0	15.0	17.0	16.0
Clay (> 0.002 mm) %	31.0	34.0	38.0	39.0	41.0
Bulk density (Mg m^{-3})	1.455	1.531	1.531	1.531	1.614
Texture class	clay loam	clay loam	clay loam	clay	clay
Saturated hydraulic conductivity ($mm\ h^{-1}$)	7	6	6	6	4
Chemical					
Exchangeable cations:					
Ca ($mg\ kg^{-1}$)	790	522	982	856	790
K ($mg\ kg^{-1}$)	160	124	138	116	152
Mg ($mg\ kg^{-1}$)	180	202	248	228	260
Na ($mg\ kg^{-1}$)	86	31	52	54	74
CEC ($cmol_c\ kg^{-1}$)	2.3	2.8	2.8	1.6	1.9
pH (H_2O)	8.57	8.61	8.82	8.74	8.58
P(Olsen)	45	36	36	41	26
Extractable micronutrients: ($mg\ kg^{-1}$)					
Fe	2.04	1.9	3.02	2.86	9.64
Zn	3	0.88	0.6	0.46	0.68
Mn	15.7	6.52	3.46	2.72	2.82
Cu	2.3	2.4	1.56	096	0.66
Total N (%)	0.08	0.05	0.04	0.03	0.03
Organic C (%)	3.4	3.2	3.9	3.95	2.2

Table 4.5 Soil physical and chemical properties of the Valsrivier soil

Properties	**Diagnostic horizons depth (mm)**				
	A	B1	B2	B3	B4
Physical	0-250	250-420	420-700	700-1200	1200-1500
Coarse sand (2 - 0.5 mm) %	0.3	0.1	0.2	0.5	1.1
Medium sand (0.5 - 0.25 mm) %	2.1	1.4	1.2	1.1	1.9
Fine sand (0.25 - 0.106 mm) %	19.9	14.8	13.6	14.5	13.3
Very fine sand (0.106 - 0.05 mm) %	17.0	14.6	15.7	18.0	17.0
Coarse Silt (0.05 - 0.02 mm) %	8.1	7.2	6.4	5.2	6.7
Fine Silt (0.02 - 0.002 mm) %	16.0	16.0	15.0	12.0	11.0
Clay (> 0.002 mm) %	37.0	45.0	47.0	48.0	48.0
Bulk density (Mg m^{-3})	1.485	1.678	1.678	1.678	1.567
Texture class	clay	clay	clay	clay	clay
Saturated hydraulic conductivity (mm h^{-1})	10	8	8	8	6
Chemical					
Exchangeable cations:					
Ca (mg kg^{-1})	226	306	604	960	818
K (mg kg^{-1})	400	362	272	270	236
Mg (mg kg^{-1})	180	180	320	360	106
Na (mg kg^{-1})	60	62	32	72	36
CEC ($cmol_c$ kg^{-1})	2.5	2	2.6	1.5	2.3
pH (H_2O)	7.7	7.9	8.84	8.91	8.72
P(Olsen)	48	39	16	26	17
Extractable micronutrients: (mg kg^{-1})					
Fe	3.46	4.32	1.06	2.08	1.12
Zn	2.26	0.42	0.26	0.28	0.64
Mn	14.64	9.2	3.3	2.04	1.5
Cu	1.86	1.96	1.34	1.04	0.8
Total N (%)	0.09	0.06	0.04	0.03	0.04
Organic C (%)	3.6	1.0	1.1	2.92	3.06

4.2.3 Lysimeters

The details of the lysimeters used for this study were dealt with in Chapter 3 and will not be repeated here. For convenience, however, concise descriptions are given below. The set-up of the lysimeters in the glasshouse is displayed in Figure 4.3

The same preparation method as described in Section 3.2.3.1 was used where the bottom of the PVC pipes, which contained the five soil monoliths, was filled with a 50 mm layer of gravel and covered fully with an 80% nylon shade cloth. Thereafter a PVC lid was used to seal the bottom of each PVC pipe and serve as the base of the lysimeter. The A-horizon was repacked in the lysimeter containing 8 mm air extraction perforated plastic pipes at depths of 100 mm and 200 mm from the soil surface.

Each of the lysimeters used comprised of a monolith PVC column, a calibrated weighing bridge (Section 3.2.3.2) and a system to control the height of the water-table (Figure 4.3). The construction and use of both the weighing bridge and the water-table-height control system is described in Section 3.2.3.2.

The use of CO_2 from a CO_2-cylinder to reduce oxygen levels in the soil was used for all the lysimeters as described Section 3.2.3.3. De-aired water was also used to reduce the influence of dissolved oxygen and carbon dioxide.

Oxygen and carbon dioxide were measured with a MultipleRAE IR gas monitor (Section 3.2.6.). These measurements started at 10am each day after the gas monitor was calibrated (Section 3.2.7.3.) These measurements enabled the calculations of air-filled porosity, O_2 content and CO_2 content (Table 4.7):

The volumetric air-filled porosity (f_a) was calculated with equations 4.1 and 4.2.

$$f_a = f - \theta_v \quad 4.1$$

$$f = 1-(\rho_b/\rho_d) \quad 4.2$$

Where f represents the total porosity (%), θ_v the volumetric soil water content measured with the DFM capacitance device (%), ρ_b is the bulk density of a soil layer (Mg m^{-3}) and ρ_d the particle density assumed to be 2.65 Mg m^{-3}.

To simplify the calculation of O_2 and CO_2 content in the different soil layers (Table 4.7), standard conditions where assumed; 1 mol of a gas occupies 22.4 L, *i.e*. 32 g of O_2 in 22.4 L and 44 g CO_2 in 22.4 L. The gas contents would be expressed in g m^{-2} (Hillel 1998).

Figure 4.3 Some of the monolith lysimeters installed in the glasshouse: a) displays the water-table-height control system and b) shows the 4 mm plastic pipes for air extraction as well as the logger of the DFM probe visible above the soil surface.

4.2.4 Temperature analysis

A two-factorial design was adopted for the temperature experiment using the monolith-lysimeter method as described in Section 3.2.3. The main treatments were five soils (sandy Hutton, loamy-sand Hutton, Bainsvlei, Sepane and Valsrivier) and four measuring soil depths (200 mm, 400 mm, 600 mm and 800 mm from the soil surface) replicated three times over a seven-day period. The experiment was conducted during the summer and measurements were recorded hourly (mV) using calibrated DFM capacitance probes (Section 3.2.7.1). Temperature data was transfered from the logger with a portable device and transported to a computer via Bluetooth.

4.2.5 Statistical analysis

The oxygen and carbon dioxide data of all five soil monoliths were analysed using the data analysis software system STATISTICA, version 12 (Statsoft, Inc., 2013). Tukey's honestly significant difference test at the 95% level of significance was used to determine statistically significant differences between treatment means, where the analysis of variance (ANOVA) indicated significant effects of the tested variables. The mean temperature at each depth was statistically analysed, using the same system and procedure as described for oxygen and carbon dioxide.

4.3 Results

4.3.1 Soil temperature

Table 4.6 shows that mean temperature differences between the five soils ranged from 30.0°C to 30.7°C at all four measuring depths. Although there was slight fluctuation in soil temperatures, there were no interaction between the soils and the depth of measurement with a p-value of 0.06. There were also no significant differences between the soils or between measurement depths with p-values of 0.095 and 0.222, respectively. The small variation in temperatures could be ascribed to the fact that temperatures in the glasshouse were kept stable during this study.

Soil temperature contributes greatly to microorganism activity. It is therefore important to include soil temperature when estimating respiration in the soil (Lloyd & Taylor, 1994; Reichstein *et al*., 2005; Kyaw Tha Paw U *et al.,* 2006). The influence of temperature on soil respiration is commonly described using Van Hoff's equation. Van Hoff's equation stated that the reaction rate increases by a factor of 2 to 3 with a temperature rise of 10°C (Bingrui & Guangsheng, 2008). The greatest difference between the mean soil temperatures at all four depths for each soil form was 0.5°C between the sandy Hutton and the Valsrivier soils. Based on the temperature results between soils, it was evident that there would be no significant effects on oxygen and carbon dioxide concentrations between the soils for the purpose of this study.

Table 4.6 Mean temperatures (°C) in the subsoil of the five soils, measured at 10am each day with DFM probes

Soil forms	Sensor Depth (mm)				
	200	**400**	**600**	**800**	**Mean**
Sandy Hutton	30.2	30.1	30.3	30.3	30.2
Loamy-sand Hutton	30.4	30.2	30.1	30.4	30.3
Bainsvlei	30.4	30.3	30.1	30.3	30.3
Sepane	30.2	30.3	30.6	30.7	30.5
Valsrivier	30.6	30.7	30.7	30.6	30.7
Mean	30.4	30.3	30.4	30.5	30.4
	P > F				
Soils	0.099				
Depth	0.222				
Interaction (Soils x Depth)	0.067				

Means followed by the same letter in either rows or columns do not differ from each other at the 5% level of significance

4.3.2 Changes in oxygen and carbon dioxide concentrations in soil profiles

4.3.2.1 Sandy Hutton soil

Measurements of O_2 and CO_2 concentrations are indicators of soil microbial respiration. It was assumed under aerobic condition, that a higher activity of soil microbes goes inside with a decrease in O_2 and increase in CO_2 concentrations. According to the statistical analysis, there were no interactions between the water-table heights and time for both O_2 and CO_2 with P values of 0.999 and 0.111, respectively (Appendix 4.1 and 4.3). However, Figure 4.4 (a-b) showed that there were highly significant differences between the mean values of the water-table heights for both O_2 and CO_2 with P values of 0.0001 for both parameters. There was a consistent decrease in O_2 concentrations as the water-table height raised from 800 mm (20.44%) to 500 mm (20.34%) and then to 300 mm (20.26%). From the 300 mm to the 150 mm water-table height, O_2 increased to 20.31%. Furthermore, Figure 4.4 (a-b) depicts an inverse relationship between O_2 and CO_2 over the water-table heights. There was consistent increase in CO_2 concentrations as the water-table height raised from 800 mm (2913 mg L^{-1}) to 500 mm (3509 mg L^{-1}) and then to 300 mm (4248 mg L^{-1}), respectively. With a further increase in the water-table height, CO_2 decreased significantly to 3657 mg L^{-1}.

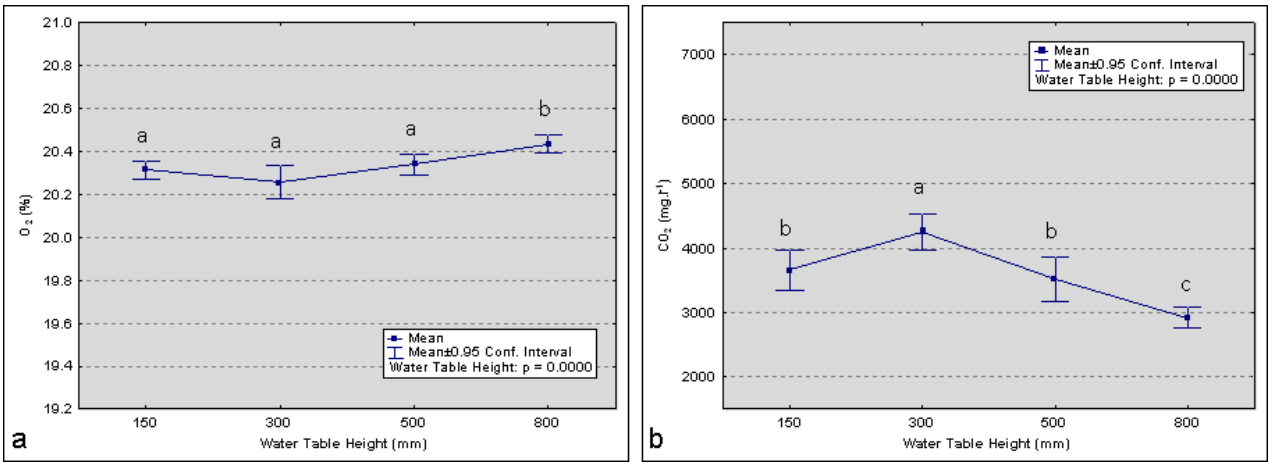


Figure 4.4 (a-b) The change in mean O_2 and CO_2 concentrations at each water-table height (mm) for the sandy Hutton soil.

Based on Figure 4.5 (a-b), there was an inverse relationship between O_2 and CO_2 over time (6 days). The statistical analysis showed that there were highly significant differences between the mean values of both O_2 and CO_2 over time (6 days) with P values of 0.0001 for both parameters. Concentration of O_2 decreased gradually from 20.42% on day 1 to 20.25% on day 6, while CO_2 concentration increased from 2980 mg L^{-1} on day 1 to 4145 mg L^{-1} on day 6. It is interesting to note that the significant differences started one day earlier for CO_2 (after 2 days) compared to O_2 (after 3 days).

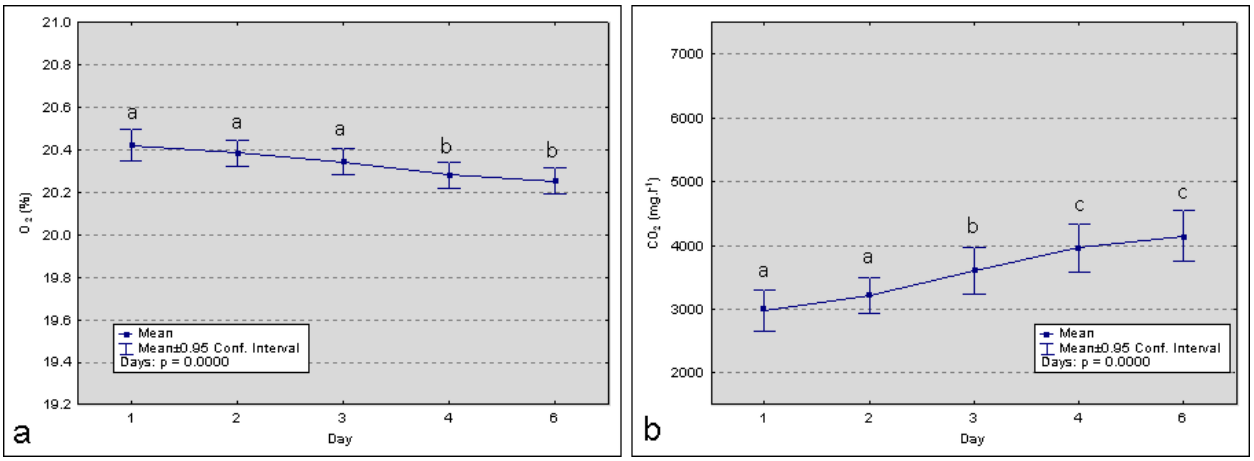


Figure 4.5 (a-b) The decrease of O_2 and increase of CO_2 concentrations over time (6 days) for the sandy Hutton soil.

4.3.2.2 Loamy-sand Hutton soil

From Figure 4.6 (a-b) and Figure 4.7 (a-b), it is clear that the effect of the treatments was similar to those of the sandy Hutton. Firstly, there were no interactions between the water-table heights and time for both O_2 and CO_2 with P values of 0.814 and 0.392, respectively (Appendix 4.5 and 4.7). Secondly, there were highly significant differences for O_2 and CO_2 concentrations between the mean values of the water-table heights with similar P values of 0.0001. The results showed that there was a decrease in O_2 concentration as the water-table height was raised from 800 mm (20.50%) to 500 mm (20.45%) and then to 300 mm (20.36%). Thereafter, there was a slight increase in O_2 concentrations as the water-table height was further raised from 300 mm to 150 mm (20.37%). There was a consistent increase in CO_2 concentrations as the water-table height raised from 800 mm (3385 mg L^{-1}) to 500 mm (3506 mg L^{-1}) with a significant increase to 300 mm (4415 mg L^{-1}), respectively. As the water table was further raised, CO_2 decreased to 4197 mg L^{-1}. Hence, as in the case of the sandy loam, the results indicate an inverse relationship between O_2 and CO_2 over the water-table heights. Thirdly, there was a constant depletion of oxygen and accumulation of CO_2 over time; oxygen concentration decreased gradually starting at 20.52% on day 1 to 20.32% on day 6, while CO_2 concentration increased from 2966 mg L^{-1} on day 1 to 4145 mg L^{-1} on day 6. The results indicate that there were significant differences for all six days except for day 3 and day 4. Figure 4.7 (a-b) depicts an inverse relationship between O_2 and CO_2 over time (6 days). These results indicate that the significant difference of O_2 and CO_2 concentrations appeared one day earlier than those of the sandy Hutton soil.

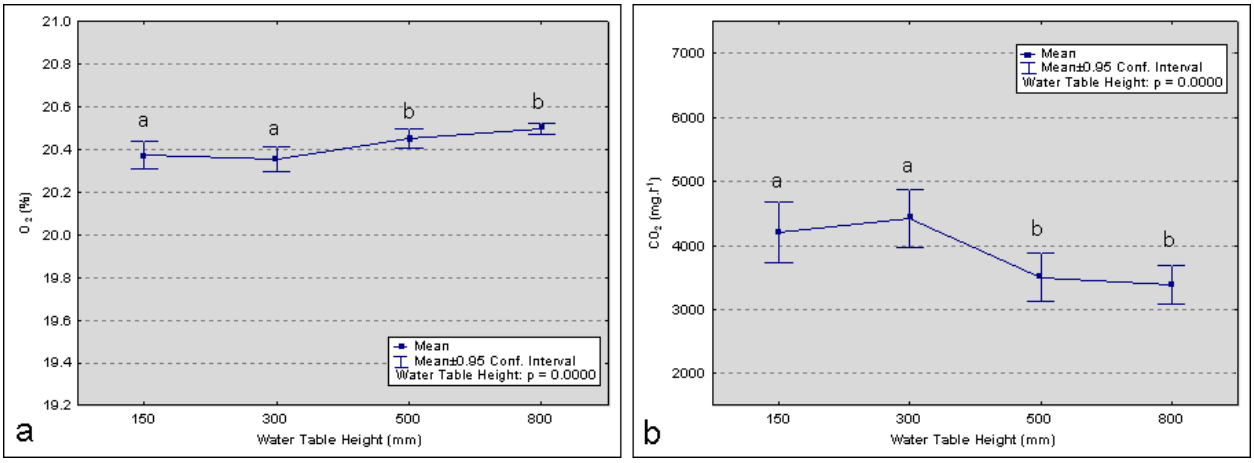


Figure 4.6 (a-b) The change in mean O_2 and CO_2 concentrations at each water-table height (mm) for the loamy sand Hutton soil.

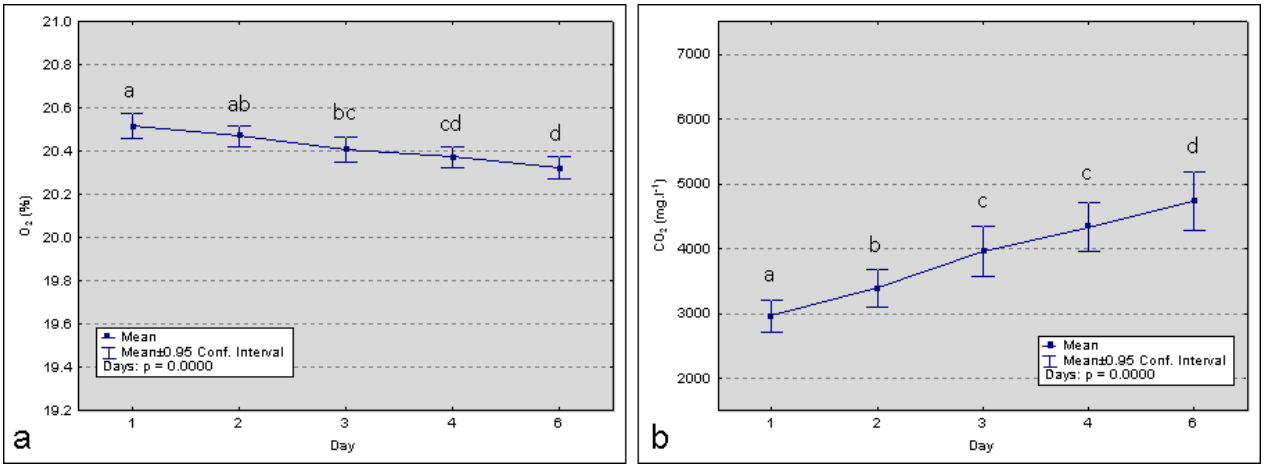


Figure 4.7 (a-b) The decrease of O_2 and increase of CO_2 concentrations over time (6 days) for the loamy sand Hutton soil.

4.3.2.3 Bainsvlei soil

Despite the fact that the Bainsvlei soil had on average between 5% and 10% more clay than the two Hutton soils (Tables 4.1 - 4.3), analysis of variance showed that the water-table height and time treatments had similar effects on both O_2 and CO_2 concentrations (Figure 4.8 a-b and Figure 4.9 a-b). Firstly, there were no interactions between the water-table heights and time for both O_2 and CO_2 concentrations (Appendix 4.9 and 4.11). Secondly, both O_2 and CO_2 concentrations differed significantly between water-table height treatments. There was a decrease in O_2 and an increase in CO_2 concentrations as the water-table height raised from 800 mm (20.49% and 2466 mg L^{-1}, respectively) to 500 mm (20.46% and 2981 mg L^{-1}, respectively) and then to 300 mm (20.33% and 3550 mg L^{-1}, respectively). This inverse relationship between O_2 and CO_2 continued with concentrations of 20.39% and 3107 mg L^{-1}, respectively, as the water table raised from the 300 mm to the 150 mm. Thirdly, there were significant differences between the mean concentrations values of both O_2 and CO_2 over time (6 days). Oxygen concentrations decreased as CO_2 concentrations increased with values of 20.51% and 2508 mg L^{-1} on day 1 to 20.32% and 4145 mg L^{-1} on day 6, respectively.

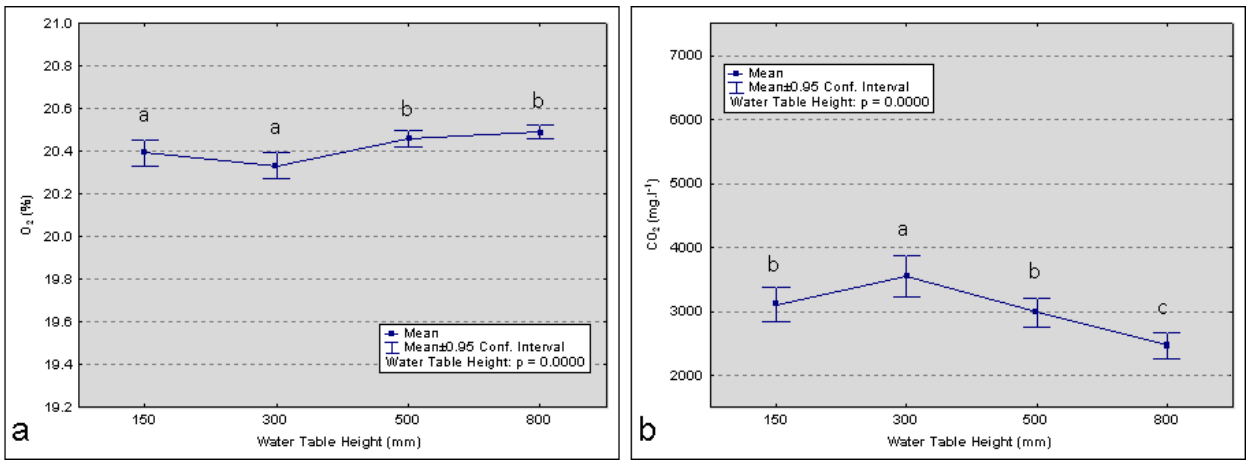


Figure 4.8 (a-b) The change in mean O_2 and CO_2 concentrations at each water-table height (mm) for the Bainsvlei soil.

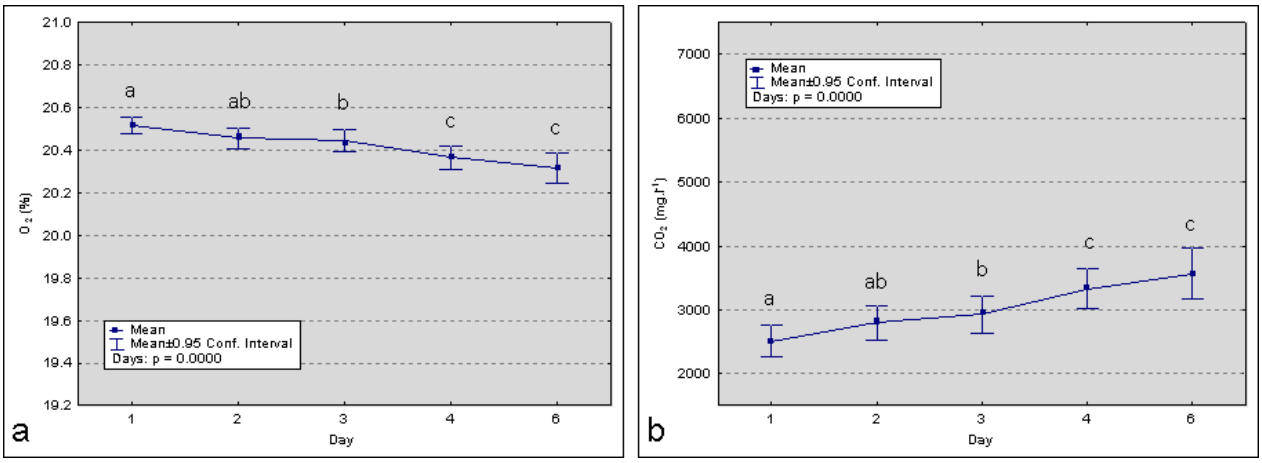


Figure 4.9 (a-b) The decrease of O_2 and increase of CO_2 concentrations over time (6 days) for the Bainsvlei soil.

4.3.2.4 Sepane soil

The results of analysis of variance for the loamy-clay Sepane soil are displayed in Figure 4.10 (a-b) and Figure 4.11 (a-b). Generally, the outcome was similar to that of the sandy Hutton, loamy-sand Hutton and the sandy-loam Bainsvlei soils reported earlier, *viz*. no interaction between main treatments, and highly significant differences for both water-table height and time treatments with respect to O_2 and CO_2 concentrations (Appendix 4.13 and 4.15). However, there were some distinct differences observed for the Sepane soil compared to the other three soils that were sandier.

At the 800 mm water-table height, O_2 was more depleted in the Sepane soil (20.10%) compared to the three more sandy soils (average 20.48%). In the Sepane soil the O_2 concentration started with a steep increase as the water-table height was raised from 800 mm (20.10%) to 500 mm (20.33%), where after there was a consistent decrease with the rise of the water-table height to 300 mm (20.30%) and 150 mm (20.20%), respectively. Considering the CO_2 concentration trends above the water-table height treatments, it was clear that the CO_2 profile responded inversely to the O_2 profile. Accordingly, there was a decrease in CO_2 as the water-table height was raised from 800 mm (4348 mg L^{-1}) to 500 mm (3198 mg L^{-1}) where after CO_2 increased as the water table was further raised to 300 mm (3523 mg L^{-1}) and to 150 mm (3884 mg L^{-1}), respectively.

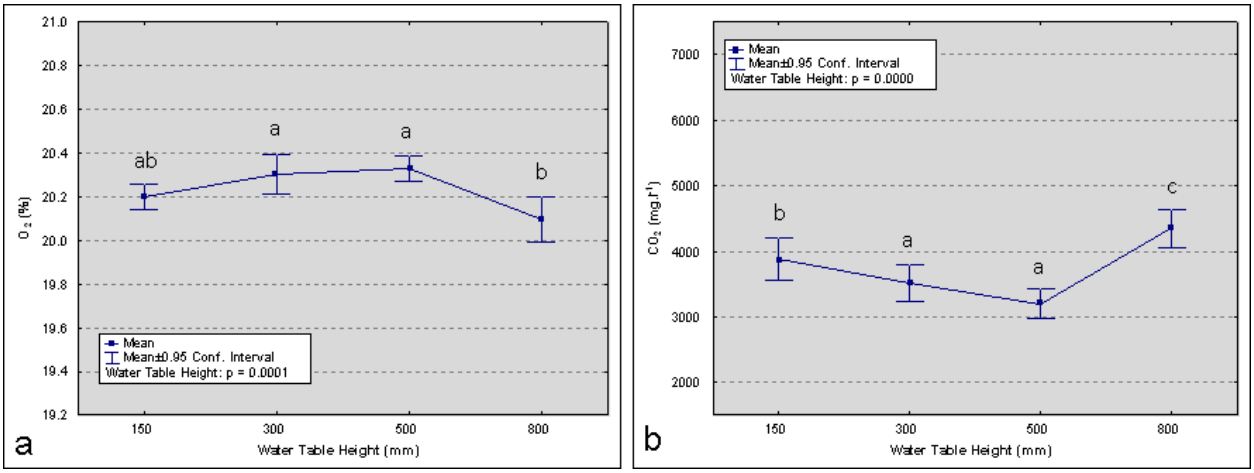


Figure 4.10 (a-b) The change in mean O_2 and CO_2 concentrations at each water-table heights (mm) for the Sepane soil.

The results of the time treatments showed a similar trend as those of the three sandier soils, except that the O_2 concentrations were consistently lower and the CO_2 concentrations were consistently higher for the Sepane soil form. Oxygen concentration decreased gradually from 20.32% on day 1 to 20.13% on day 6, while CO_2 constantly increased from 3197 mg L^{-1} on day 1 up to 4282 mg L^{-1} on day 6.

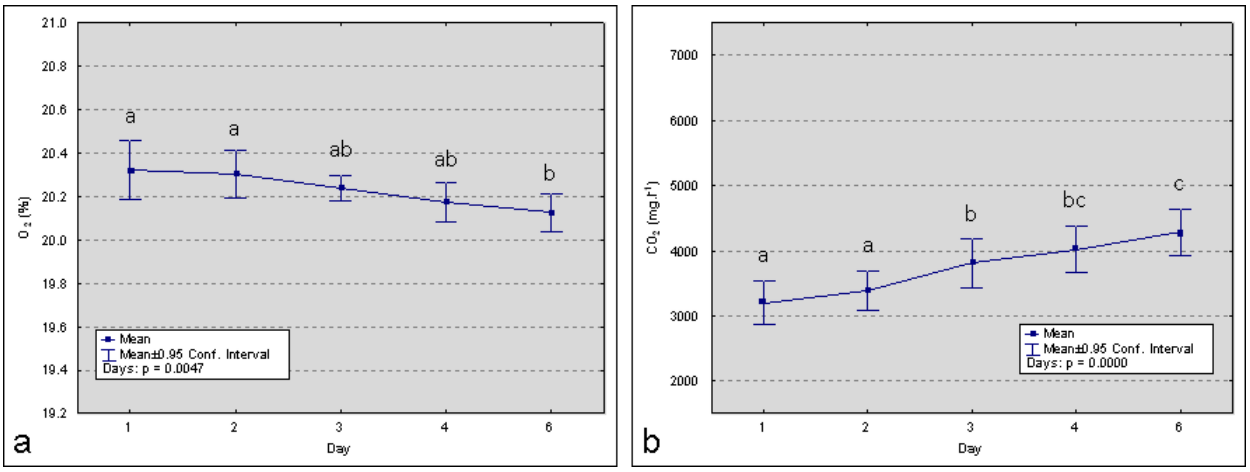


Figure 4.11 (a-b) The decrease of O_2 and increase of CO_2 concentrations over time (6 days) for the Sepane soil.

4.3.2.5 Valsrivier soil

For the Valsrivier soil, as for the Sepane soil, there was no significant interaction between main treatments, and highly significant differences for both water-table heights and time treatments with respect to O_2 and CO_2 concentrations (Appendix 4.17 and 4.19). The oxygen concentration increased significantly as the water-table height raised from 800 mm (19.94%) to 500 mm (20.23%), where after O_2 concentration decreased consistently as the water-table height was further raised to 300 mm (20.17%) and 150 mm (20.08%). Carbon dioxide significantly decreased as the water-table height raised from 800 mm (5643 mg L^{-1}) to 500 mm (3918 mg L^{-1}) where after CO_2 increased as the water table was further raised to 300 mm (4433 mg L^{-1}) and to 150 mm (5053 mg L^{-1}). There was a decrease in O_2 concentrations from 20.21% on day 1 to 19.94% on day 6, while CO_2 consistently increased over time from 4019 mg L^{-1} on day 1 to 5478 mg L^{-1} on day 6 (Figure 4.12 a-b and Figure 4.13 a-b).

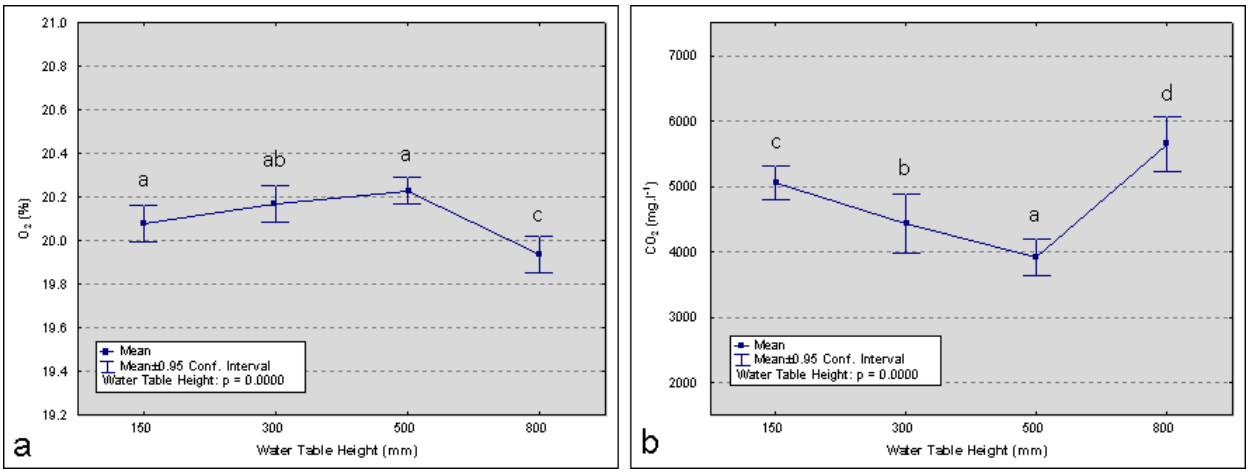
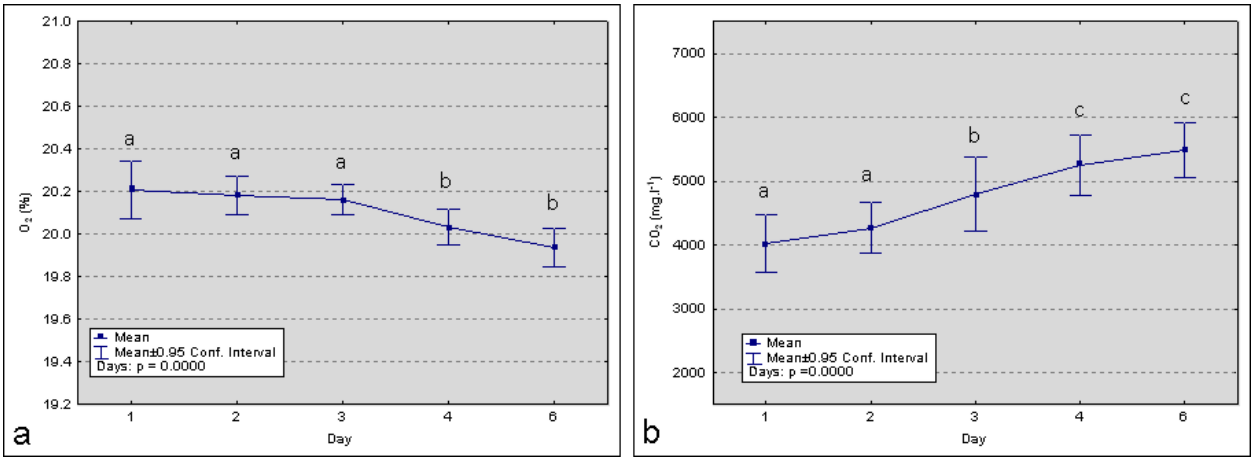


Figure 4.12 (a-b) The mean change in O_2 and CO_2 concentrations at each water-table heights (mm) for the Valsrivier soil.



Figures 4.13 (a-b) The decrease of O_2 and increase of CO_2 concentrations over time (6 days) for the Valsrivier soil.

4.3.3 Unsaturated zone

The calculated values of air-filled porosity, O_2 content and CO_2 content for the unsaturated zone of the five soils are summarised in Table 4.7. As expected, the air-filled porosity decreased as the water table was raised from 800 mm to 150 mm in all soils. Air-filled porosity showed the widest range for the sandy Hutton soil that varied between 12.1% at the 800 mm water-table height to 1.7 % at the 150 mm height. The narrowest range was observed for the Valsrivier soil with values that decreased from 8.1% to 2.8% from the 800 mm to the 150 mm water-table height, respectively.

The results showed clearly that the O_2 content, expressed as g m^{-2}, decreased as the water table raised from 800 mm to 150 mm in all five soils. Oxygen content had the widest range for the sandy Hutton soil that varied between 43.4 g m^{-2} at the 800 mm water-table height to 6.2 g m^{-2} at the 150 mm water-table height. The narrowest range was observed at the Valsrivier soil with values that decreased from 28.5 g m^{-2} to 9.9 g m^{-2} from the 800 mm to the 150 mm water-table height, respectively. As in the case of O_2 content, the CO_2 content (g m^{-2}) also decreased as the water table height raised from 800 mm to 150 mm in the five soils. Again, the Valsrivier soil showed the highest CO_2 value (565 g m^{-2}) at the 800 mm water-table height and the sandy Hutton soil the lowest value (78 g m^{-2}) at the 150 mm water-table height.

Table 4.7 Calculated values of air-filled porosity, oxygen content and carbon dioxide content at different water-table heights in the five soils

Soil forms	Water-table heights	Volumetric soil-water content	Volumetric air-filled porosity	Air-filled porosity (L)	$[O_2]$ (%)	$[O_2]$ (g)	$[O_2]$ (g m^{-2})	$[CO_2]$ (mg L^{-1})	$[CO_2]$ (g)	$[CO_2]$ (g m^{-2})
Sandy Hutton	**150**	0.345	0.017	1.5	20.31	0.431	6.2	3657	5.4	78
	300	0.325	0.038	3.2	20.26	0.938	13.4	4248	13.8	197
	500	0.288	0.074	6.4	20.34	1.852	26.6	3509	22.4	321
	800	0.242	0.121	10.4	20.44	3.027	43.4	2913	30.2	433
Loamy-sand Hutton	**150**	0.341	0.018	1.5	20.37	0.437	6.3	4197	6.3	91
	300	0.324	0.035	3.0	20.36	0.868	12.5	4415	13.2	189
	500	0.291	0.068	5.8	20.45	1.698	24.4	3506	20.4	292
	800	0.244	0.114	9.8	20.50	2.865	41.1	3385	33.1	475
Bainsvlei	**150**	0.325	0.017	1.4	20.39	0.412	7.9	3107	4.4	83
	300	0.313	0.029	2.4	20.33	0.710	10.2	3550	8.7	124
	500	0.287	0.054	4.7	20.46	1.365	19.6	2981	13.9	200
	800	0.250	0.092	7.9	20.49	2.300	33.0	2966	29.4	414
Sepane	**150**	0.376	0.024	2.1	20.20	0.592	8.5	3884	8.0	114
	300	0.366	0.033	2.9	20.30	0.830	11.9	3523	10.1	145
	500	0.339	0.060	5.2	20.33	1.503	21.6	3198	16.6	237
	800	0.305	0.094	8.1	20.10	2.316	33.2	4348	35.1	503
Valsrivier	**150**	0.327	0.028	2.4	20.08	0.691	9.9	5053	12.2	175
	300	0.317	0.038	3.2	20.17	0.936	13.4	4433	14.4	207
	500	0.289	0.066	5.6	20.23	1.631	23.4	3918	22.1	317
	800	0.274	0.081	7.0	19.94	1.988	28.5	5643	39.4	565

4.4 Discussion

4.4.1 Water-table height

The main finding of this study was that a rise in the water-table towards the surface of the five soils significantly influenced concentration profiles of O_2 and CO_2. However, there were some distinct differences observed between the five soils with regards to either their O_2 or CO_2 profiles. The responses can generally be divided into two groups, *viz*. the sandy (sandy Hutton, loamy-sand Hutton and Bainsvlei) and clay (Sepane and Valsrivier) soils, respectively.

In the case of the sandy soils, O_2 showed a decrease as the water-table height was raised from 800 mm to 300 mm, where after it increased as the water-table was further raised to 150 mm. On the other hand, CO_2 concentrations responded with a gradual increase as the water-table height was raised from 800 mm to 300 mm. Thereafter CO_2 concentrations significantly decreased as the water table was further raised to 150 mm.

For the clay soils, O_2 concentrations showed a steep increase as the water-table height was raised from 800 mm to 500 mm, where after there was a consistent decrease when the water-table height was further raised to 300 mm and 150 mm, respectively. Compared to O_2, there was a decrease in CO_2 concentrations as the water-table height was raised from 800 mm to 500 mm, where after CO_2 increased as the water table was further raised from 500 mm to 150 mm.

The design of this study did not allow the determination of the reasons for O_2 and CO_2 responses in the two soil groups. However, several authors proposed some credible explanations (Callebaut *et al.,* 1982; Deyo *et al*., 1993; Hillel, 1998; Lal & Shukla, 2004; Tuli *et al*., 2005; Watson & Kelsey, 2005). They described the O_2 and CO_2 responses to textural and structural differences in soils. Watson & Kelsey (2005) found an increase in porosity with a decrease in silt-plus-clay content. In this study, as depicted in Figure 4.14 a, the porosity decreased as the water table advanced to the surface for both soil groups. However, it was interesting to note that the porosity for the sandy soils were greater than those of the clay soils for water table heights up to 300 mm. Above the 300 mm height, the porosity of the clay soils was greater than those of the sandy soils. This phenomenon could be ascribed to macro pores that developed in the clay soils due to cracks formed during the drying proses as proved in Figure 4.15 (a-b). The exchange of O_2 and CO_2 at the soil surface and associated diffusion into the soil is more rapid in sandy soils compared to clay soils (Watson & Kelsey, 2005).

One aspect that the two soil groups had in common was an inverse relationship between O_2 and CO_2 concentrations over the water-table heights as depicted in Figure 4.14 b. However there were some distinct differences in the concentration range of O_2 and CO_2 for the two soil groups; 20.26 - 20.49% oxygen for the sandy soil group compared to 19.9% - 20.33% for the clay soil group and 2466 - 4415 mg L^{-1} carbon dioxide for the sandy soil group compared to 3198 - 5643 mg L^{-1} for the clay soil group. These results confirmed the general finding that the clay soils were more anaerobic than the sandy soils under water table conditions (Callebaut *et al.,* 1982; Deyo *et al*., 1993; Lal & Shukla, 2004; Tuli *et al*., 2005; Watson & Kelsey, 2005).

Other than the O_2 and CO_2 concentrations that described the intensity of the two gasses, the O_2 and CO_2 content highlighted the capacity of the soil groups to store these gasses (Figure 4.14 c-d). The figures illustrated that the capacity for O_2 and CO_2 storage decreased as the water table advanced to the soil surface. However, the two soil groups differ in their response to O_2 and CO_2 content. The difference could be seen in the slopes of the linear functions given in Figure 4.14 (c-d). These slopes indicated that the capacity for the two gasses decreased at a rate of 0.051 g O_2 m^{-2} mm^{-1} rise of the water table for the sandy soils compared to 0.035 g O_2 m^{-2} mm^{-1} for the clay soils. The CO_2, on the other hand, decreased from 0.51 g m^{-2} mm^{-1} for the sandy soils to 0.61 g m^{-2} mm^{-1} for the clay soils.

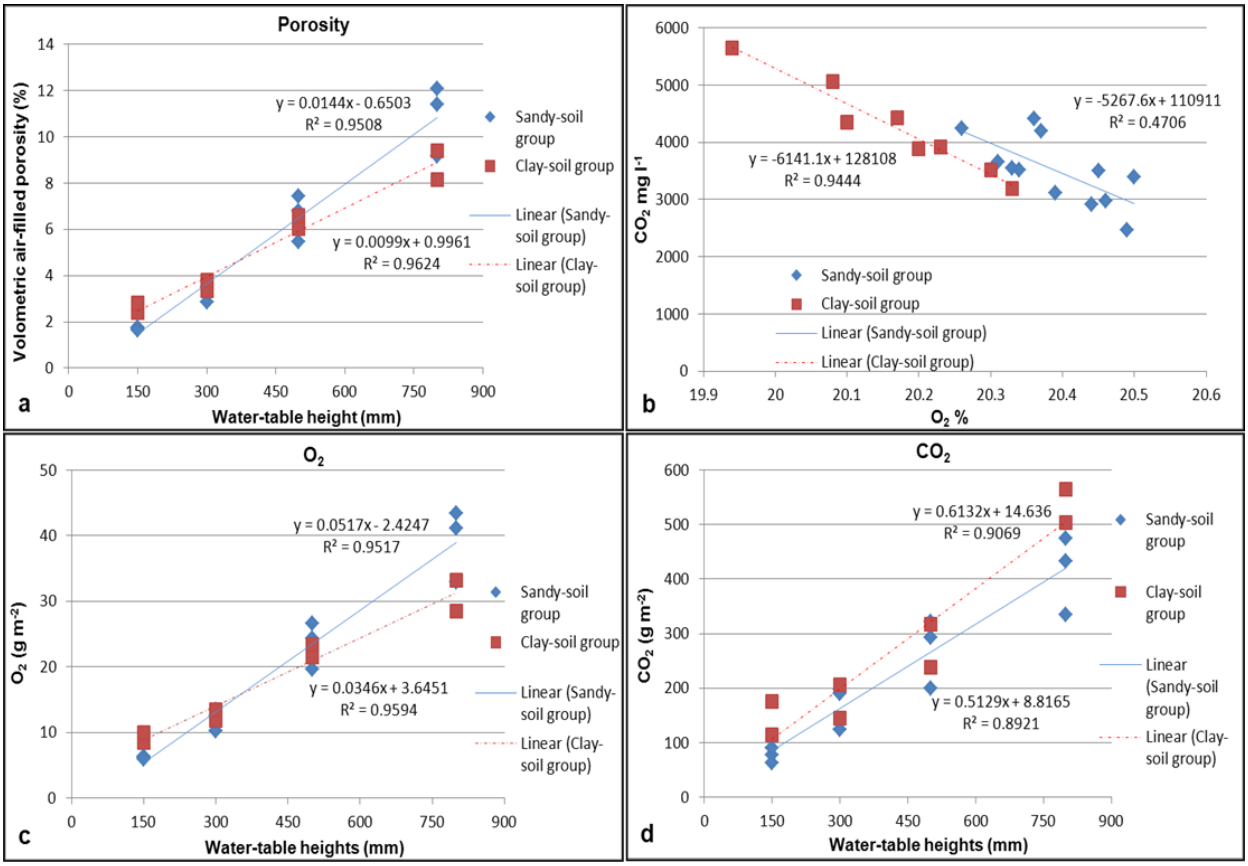


Figure 4.14 Relationships between water-table heights, and a) volumetric air-filled porosity, b) O_2 and CO_2 relationship c) O_2 content and d) CO_2 content for the sandy soil group (sandy Hutton, loamy-sand Hutton and Bainsvlei) and the clay soil group (Sepane and Valsrivier).

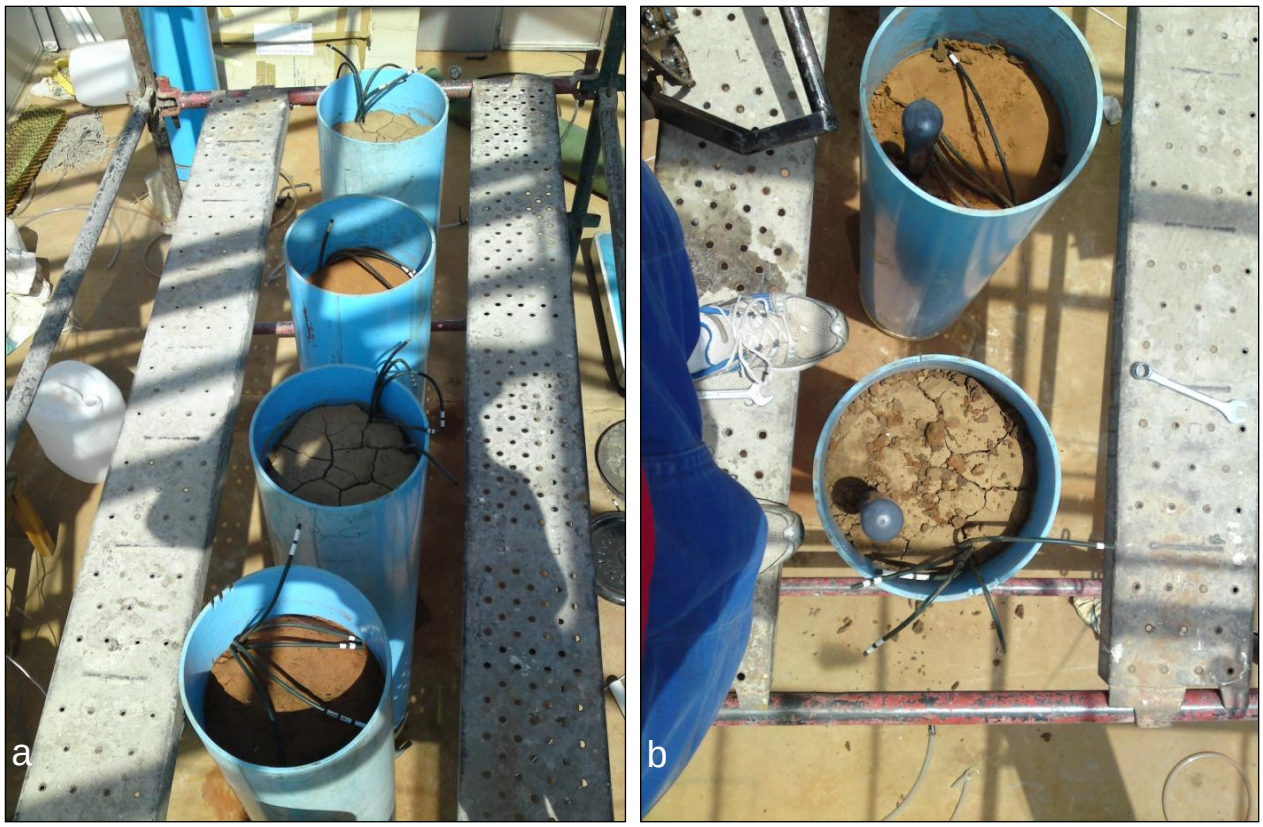


Figure 4.15 Cracks shown in the a) Sepane and b) Valsrivier soil forms.

4.4.2 Time

Based on the reported results, the main finding was that time had significantly influenced O_2 and CO_2 concentrations over the 6-day period. There was an inverse relationship between O_2 and CO_2 concentrations for all five soils. As O_2 concentrations gradually decreased, CO_2 concentration gradually increased over the 6 day period irrespective of soil. The only difference between the two soil groups was the intensity of respiration that resulted in lower O_2 and higher CO_2 concentrations for the clay soil group over the 6 day period. For this period the mean oxygen concentration was 20.4% for the sandy soil group compared to 20.17% for the clay soil group. The mean O_2 values for the sandy soils compare well to those reported by Lal & Shukla, (2004), with a value of 20.34%. The clay soils had a slightly higher mean O_2 values compared to the average of 18.7% found by Hanslin *et al*. (2005). In this study the average CO_2 concentrations were 3494 mg L^{-1} for the clay soils and 4251 mg L^{-1} for the sandy soils. This compare well with the average of 4075 mg L^{-1} found by Lal & Shukla (2004).

Watson & Kelsey (2005) found that lower O_2 concentrations, as measured for the clay soil group, can be ascribed to the high silt-plus-clay that have smaller pores and will slow down the O_2 diffusion rates into and through the soil. Clay soils with small pores generally have poor drainage and aeration. Due to respiration processes over time, clay soils are more likely to become anaerobic as microbes use oxygen more rapidly than it is replenished through diffusion than in sandy soils (Watson & Kelsey, 2005; Kyaw Tha Paw U *et al.,* 2006). According to Bingrui & Guangsheng (2008), respiratory activities can also be influenced by water content. They found that microbial activity increased with an increase in water content to the drained upper limit, where after activities decreased with a further increase in soil water content. Water availability for microbial activity above the water table is mainly through capillary rise from the water table surface (Sepashah & Karimi-Goghari, 2005).

4.5 Conclusion

The monolith-lysimeter technique developed in Chapter 3 was used to evaluate five soils (sandy Hutton, loamy-sand Hutton, Bainsvlei, Sepane and Valsrivier) in their O_2 and CO_2 response to a rising water-table over a period of 6 days.

The main findings of this study were: firstly, that the O_2 and CO_2 concentration profiles were significantly influenced by the rise in water-table heights for the selected soils. However, there were some distinct differences in either the O_2 or CO_2 profiles observed between the sandy soils (sandy Hutton, loamy-sand Hutton and Bainsvlei) compared to the clay soils (Sepane and Valsrivier). In the case of the sandy soils, O_2 showed a decrease as the water-table height was raised from 800 mm in the subsoil to just below the 300 mm topsoil, where after it increased as the water-table was further raised to near the surface (150 mm). For the clay soils, O_2 concentrations increased steeply as the water-table height was raised in the subsoil from 800 mm to 500 mm, where after there was a consistent decrease when the water-table height was further raised to the topsoil at 300 mm and 150 mm. The CO_2 profiles were inversely related to the O_2 profiles for the sandy and clay soils.

Secondly, time had significantly influenced O_2 and CO_2 concentrations over the 6 day period. As O_2 concentrations gradually decreased, CO_2 concentration gradually increased over the 6 day period for all five soils. The only difference between the two soil groups was that the intensity of respiration resulted in lower O_2 and higher CO_2 concentrations for the clay soil group over the 6 day period.

CHAPTER 5

SUMMARY AND RECOMMENDATIONS

5.1 Introduction

Despite the fact that a significant area of the irrigated land in South Africa is affected by waterlogging, not much research has been conducted on soil air during waterlogging conditions in local soils. Waterlogging is defined as a condition where the soil profile is saturated with water and in most cases the water table is stagnant, which imply that soil organisms and crops are prone to oxygen and carbon dioxide stress. Both these stresses are related to air exchange processes in soils. Theories on air movement in soils and the exchange thereof are well developed, but there is insufficient evidence on how oxygen and carbon dioxide concentrations in the unsaturated zone respond to varying water table heights. One of the reasons that restrict oxygen and carbon dioxide profiling in soils is the lack of methods to extract and measure soil air.

5.2 Development and evaluation of a small lysimeter

Consequently an attempt was made to develop a lysimeter that can reflect on oxygen and carbon dioxide changes in the unsaturated zone above the water table. In order to achieve the above objective the research was divided into four tasks. Task one was to design and assemble a small lysimeter with a water supply system that can control the water-table height in the lysimeter. The basic design of the lysimeter system is depicted in Figure 3.6. Accordingly, the design comprises of three components, *viz.* a soil container, a weighing bridge and a pressure head control system. The 1.5 m long soil container was sourced from a Class 16 PVC pipe used in high pressure water supply systems. The diameter was 0.3 m with a wall thickness of 5 mm. One end of the pipe was sealed with a 5 mm thick plate cut from a commercially available PVC sheet. Thus, this container made it possible to house a soil monolith sampled from irrigable soil The weighing bridge (scale) consisted of a load cell, Model 1260 Tedea-Huntleigh aluminum off centre loading single point, with the capacity range of 50 - 660 kg. The load cell was connected to a Campbell Scientific CR1000 data logger which allowed the scale to measure the change in soil water content on a continuous basis. The manually controlled water-table height control system consisted of three parts, namely a 20 litre plastic container fitted on a hanging platform, a height adjustable 10 litre plastic bucket attached to the soil container and a manometer. Water

flow from the hanging container to the monolith was controlled by a ball valve fitted in the bucket. The lysimeter was filled with A and B horizons of a Bainsvlei soil and saturated with water and then allowed to drain. The weighing bridges were also successfully calibrated by loading known weights onto the loadcells. Based on the results of the drainage experiment, it was concluded that the lysimeter was sufficiently accurate in measuring changes in water content of the soil.

Task two concentrated on the sampling of a soil monolith by using a 3-step procedure. Step 1 refers to the careful excavation of a soil pillar with dimensions slightly larger than the PVC pipe used for the lysimeter (Figure 3.2). Step 2 focused on the installation of the perforated tubes and valves for air-extraction at desired soil depths as can be seen in Figure 3.3. In this case an 8 mm diameter tube with a length of 250 mm was inserted horizontally at 400 mm, 600 mm and 800 mm, connected to 4 mm plastic tube pasted into a vertically grove and fitted with a control valve to extract an uncontaminated air sample with the MultipleRAE IR gas monitor instrument. Step 3 implied the final sampling of the soil monolith by lowering the PVC pipe over the soil column while smoothing the soil edges. A metal plate was pushed underneath the soil column and tied (Figure 3.4), ready to be lifted and then transported to the glasshouse. The last part of this step was to place the monolith (soil column) on the weighing bridge in the glasshouse and to couple it to the water-table height control system (Figure 3.5).

Task three was designed to populate the lysimeter with instruments to measure soil air and to measure mass losses due to water losses by drainage. Standard procedures were followed to install DFM probes in the monolith to measure water content and temperature at six fixed soil depths. The capacitance DFM probes were successfully calibrated for both temperature and volumetric soil water content at the same time. Both instruments were used to measure internal drainage as part of the calibration process for the DFM probes. Furthermore, an equation was used on drainage data to eliminate temperature sensitivity (Wikipedia Contributors, 2012). The MultipleRAE IR gas monitor instrument was calibrated according to a two-step process using fresh air and span gas that is a commercial product which contains a known concentration of given gasses.

Task 4 implied on the testing of the lysimeter system to measure oxygen and carbon dioxide concentrations in the unsaturated zone of the soil. This study was intended to improve the understanding of how characteristics of disturbed and undisturbed soil influence soil temperature, internal drainage, O_2 and CO_2 concentrations under water-table conditions. Two

soil treatments were employed, disturbed versus an undisturbed soil monolith as described in previous paragraph. Measurements included internal drainage, soil temperature and soil air composition for the unsaturated zone in the presence of a fluctuating water table. Considering the soil treatments, the results showed that temperature was significantly higher in the undisturbed (23.8°C) soil compared to the disturbed (22.1°C) soil. According to Tuli *et al*. (2005) this can be ascribed to soil structure changes induced by the disturbance. When soils are disturbed, heat flow processes are also disturbed due to the changes in pore-geometrical characteristics such as tortuosity, connectivity and constriction (Vogel, 1997; Wildenschild *et al*., 2005). Thus, it was obvious that the thermal conductivity of the disturbed soil was reduced due to the lost in contacts between soil particles.

The results of the internal drainage curves confirmed that the manually sampling method caused structural changes in the soil. Drainage processes were more or less similar for both sample methods but differ in respect with water retained by the soils. The disturbed soil retained significantly more water (0.347 mm mm^{-1}) at the point of field saturation compared to the undisturbed soil (0.311 mm mm^{-1}), i.e. 44 mm per 1200 mm soil. Both saturation points correspond well with the calculated porosity of 0.37 mm mm^{-1} due to the injection of CO_2 before recharging of the soils. This large difference between disturbed and undisturbed soil was ascribed to the disappearance of macro pores that confirms the enormous impact of soil structure and pore-space characteristics on flow (Tuli *et al*., 2005).

The first main finding for soil air was that O_2 concentration had significantly differ between the disturbed and undisturbed soils with O_2 that were 0.3% lower in the undisturbed soil. This can be ascribed to large pores that formed as the soil was disrupted and promote more rapid oxygen diffusion into and through the soil (Watson & Kelsey, 2005). Furthermore, O_2 concentrations decreased in the undisturbed soil as the water table raised from 800 mm to 300 mm while O_2 incresed at the same time in the disturbed soil. This phenomenon was also found by Tuli *et al*. (2005) and they ascribed their results to the size difference in macropores between disturbed and undisturbed soils. Secondly, it was found that the one common aspect from the two soil samples was an inverse relationship between O_2 and CO_2 concentrations. Lastly, the CO_2 concentration also had a distinct difference between the soil samples with the undisturbed soil (3877 mg l^{-1}) that was significantly higher than the disturbed soil (1648 mg l^{-1}).

In this study it was much more difficult to obtain an undisturbed soil sample compared to a disturbed sample especially for sandy soils. However, the undisturbed sample was more accurate for reliable measurements as the disturbed sample had to much disruption of the pore-geometrical characteristics such as tortuosity, connectivity and constriction that all had a tremendous influence on both water movement and soil air concentrations. As far as measurement was concerned, the DFM probes were useful for soil water measurements in the presence of a raising water table as well as over time. Due to the ease of use and rapidly large data capture ability compared to other soil moisture sensors, it has the potential for irrigation planning and water use efficiency measurements. The MultipleRAE IR gas monitor instrument also proved to be useful in this study.

5.3 Oxygen and carbon dioxide profiles in soils with water tables

An understanding of the air composition (O_2 and CO_2 concentrations) in the unsaturated zone above the water tables is essential to manage these soils by reducing the risk for water logged conditions. There is a scarcity of scientific information that deals with O_2 and CO_2 concentrations in the unsaturated zone of irrigated soils in the absence of a crop. In this regard information is needed for improving understanding of the effects of soil and irrigation management. This study was intended to help fill this gap by measuring and comparing O_2 and CO_2 profiles in five soils that differ in irrigation suitability.

The horizons of the five soils were physically and chemically analysed that provided valuable pedological data for the interpretation of O_2 and CO_2 concentrations. The soils were classified as a sandy Hutton, loamy-sand Hutton, Bainsvlei, Sepane and Valsrivier. Furthermore, pictures were provided for each of the five soils, mainly to distinguish between their colour and structure in the various horizons. After the soils were classified, they were sampled and transported to the glasshouse where lysimeters was prepared. Thereafter a water-table height control system was applied and DFM probes inserted from the soil surface. Thereafter, O_2 and CO_2 concentrations were successfully measured with a MultipleRAE IR gas monitor instrument at different depths from the unsaturated zone.

The first main finding of this study was that the variations in O_2 and CO_2 concentrations were because of textural differences between the soil horizons. Therefore, the five soils were divided into two groups that consisted of the sandy-soil group (sandy Hutton, loamy-sand Hutton and Bainsvlei) and the clay-soil group (Sepane and Valsrivier) for further discussion. The soils

allocated to each group had the exact same soil-gas pattern over water table height and time with little variation within each soil.

The second main finding of this study was that O_2 and CO_2 concentrations was significantly influenced as the water-table was raised for both the sandy and clay soil group. However, there were some distinct differences as O_2 and CO_2 varied, respectively, between 20.26 - 20.49% and 2466 - 4415 mg l^{-1} for the sandy soil group compared to 19.9% - 20.33% and 3198 - 5643 mg l^{-1} for the clay soil group. This variation was ascribed to the exchange of O_2 and CO_2 at the soil surface and associated diffusion into the soil that is more rapid in sandy soils than clay soils (Watson & Kelsey, 2005). In the case of the sandy soil group, O_2 decreased as the water-table rose from 800 mm to 300 mm, where after O_2 increased as the water-table was further raised to 150 mm. For the clay-soil group, O_2 concentrations increased steeply as the water-table height was raised from 800 mm to 500 mm, where after there was a consistent decrease when the water-table height was further raised to 300 mm and 150 mm, respectively. As expected, CO_2 responded oppositely for both soil groups as the water-table was raised. These results can be ascribed to a decrease in porosity as the water table advance to the surface of both soil groups. However, it was interesting to note that the porosity for the sandy soil group was greater than that of the clay group for water table heights up to 300 mm. Above the 300 mm height, the porosity of the clay soils was greater than those of the sandy soils as a result of macro pores that develop in the clay soils due to cracks formed during the drying process.

Furthermore, it was found that as O_2 and CO_2 concentration describes the intensity of gasses, the O_2 and CO_2 content highlighted the capacity of the soil groups to store these gasses. The capacity for O_2 and CO_2 storage decreased as the water table advanced to the soil surface. However, the two soil groups differed in their response to both O_2 and CO_2 content. The calculations showed that the capacity of gasses decreased at a rate of 0.0517 g O_2 m^{-2} mm^{-1} in the sandy-soil group compared to 0.0346 g O_2 m^{-2} mm^{-1} in the clay-soil group while CO_2 decreased from 0.513 g m^{-2} mm^{-1} in the sandy-soil group compared to 0.613 g m^{-2} mm^{-1} in the clay-soil group. These results confirm the general finding that the clay soils are more anaerobic than the sandy soils under water table conditions (Lal & Shukla, 2004; Tuli *et al*., 2005; Watson & Kelsey, 2005).

Finally this study revealed that there were significant differences between the mean concentrations for both O_2 and CO_2 over time (6 days). Again an inverse relationship was found between O_2 and CO_2 concentrations for all five soils with O_2 that gradually decreased while CO_2 gradually increased. The only difference between the two soil groups was the intensity of

respiration that resulted in lower O_2 and higher CO_2 concentrations for the clay soil group compared to the sandy soil group. Oxygen was 20.17% for the clay soil group compared to 20.4% for the sandy soil group while CO_2 concentrations differed more significantly from 4251 mg l^{-1} for the clay soil group to 3494 mg l^{-1} for the sandy soil group. The more anaerobic conditions in clay soils can be ascribed to the higher silt-plus-clay content that cause small pores that will slow down O_2 diffusion rates into and through the soil and promote poor aeration (Watson & Kelsey, 2005).

The results of this study showed that O_2 and CO_2 concentrations can accurately be measured in sandy and clay soils to give comparable, reliable results. Furthermore, O_2 and CO_2 profiles are now better understood in the unsaturated zone of a bare soil under water table conditions. This basic information can be useful when decisions have to be made on the suitability of soils for crop production under irrigation in terms of O_2 requirements.

5.4 Recommendations

The theory of air movement in soils is in an advance stage, but air composition and air movement during or near waterlogging is not well researched. This study provides basic methodology to study aeration under specific soil-plant-atmosphere conditions. For example, farmers from large irrigation schemes in South Africa are forced to deal with shallow water table conditions. Shallow water tables can either harm the crop when aeration is restricted or can be utilized positively to reduce irrigation application amounts. Farmers require scientific information to help them managing the water tables to ensure sustainable yields. As demonstrated in the study, it is possible to create a variety of water-table heights in the small lysimeters and hence easily to create controlled conditions for studying the effect of waterlogging on crops.

Another aeration problem extensionists and advisors are confronted is the effect of compaction induced by tillage under irrigation. Sandy soils are prone to compaction and mouldboard ploughs and discs tend to compact the top soil at depths between 15 and 30 cm (Figure 5.1a). Under these conditions the roots are restricted to the top soil as shown in Figure 5.1b. The compaction leads to poor drainage and in extreme cases a thin anaerobic layer forms just above the compacted layer. This is a clear indication that aeration is a problem, but despite these visual signs of waterlogging the farmers still harvest high yields and they will continue with this hydroponic way of farming until research can convince them to change their practice. Thus the small lysimeters can be used to mimic the mentioned soil and irrigation conditions and first-hand information on aeration can be obtained.

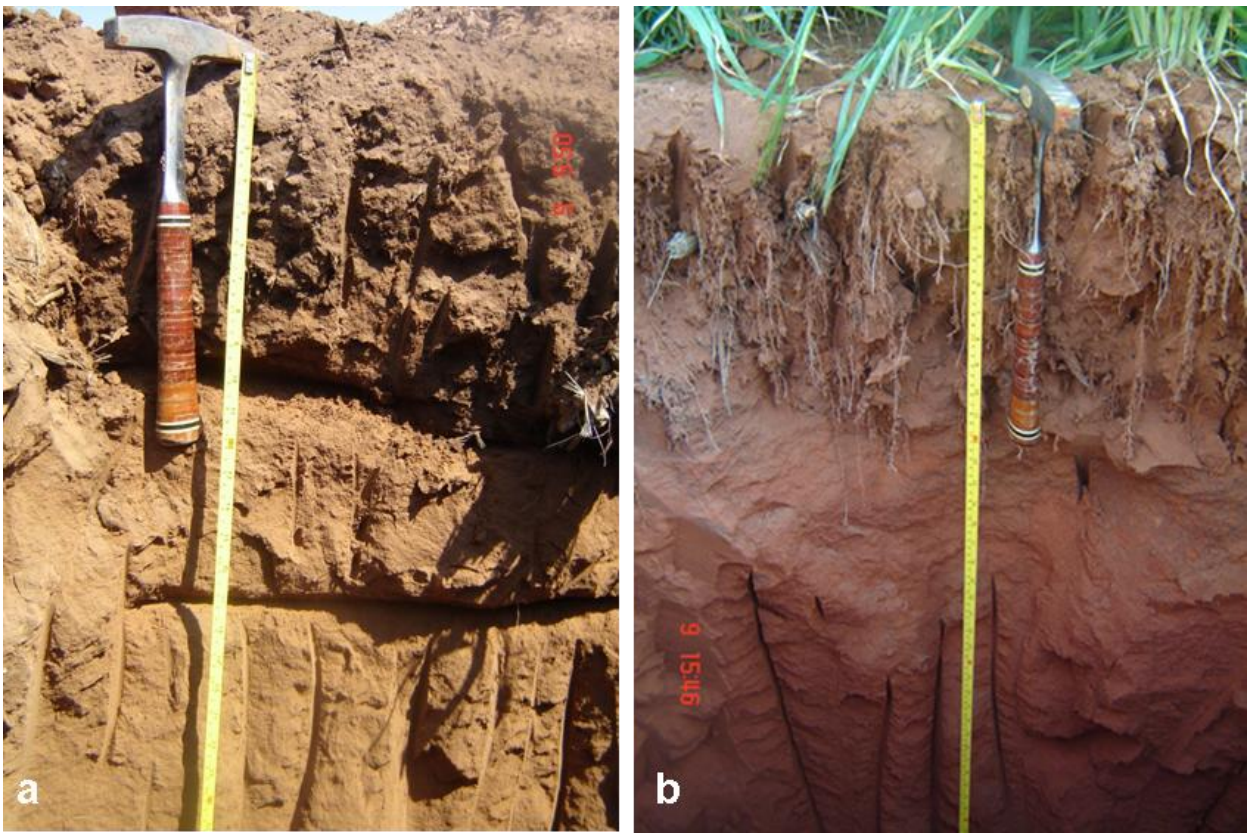


Figure 5.1 Photos illustrating the importance to study soil aeration under compaction: a) presence of a compacted layer that restricted drainage in the top soil and b) poor root development in compacted sandy soils due to poor tillage practices.

The lysimeter can also be used to study water balance components of crops under controlled conditions. For example, the barley-research team studied the effect of water stress in the major growth stages of the crop successfully. Other disciplines that can utilise the technology are wetland and environmental studies, especially studies concerning gas volatilization following point source pollution.

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APPENDICES

Appendix 3.1 Proof that sensors within probes are similar in their response to water content in water column

Probe nr.	$DFM_{NT} = (25 - DFM_T)^* x + DFM_R$							
	DFM_T	x	DFM_{NT} (each sensor)					
			S 1	S 2	S 3	S 4	S 5	S 6
11154	10	0.0259	88.86	89.45	89.70	89.90	90.17	90.99
	15	0.0259	88.94	89.52	89.74	90.01	90.42	91.01
	20	0.0259	88.97	89.62	89.84	90.05	90.51	90.99
	25	0.0259	89.07	89.64	89.93	90.13	90.58	91.18
	30	0.0259	89.14	89.66	89.97	90.17	90.62	91.27
	33	0.0259	89.15	89.67	89.99	90.21	90.61	91.27
10099	10	0.0324	88.96	89.55	89.80	90.00	90.27	91.09
	15	0.0324	89.00	89.58	89.80	90.07	90.48	91.07
	20	0.0324	89.00	89.65	89.87	90.08	90.54	91.02
	25	0.0324	89.07	89.64	89.93	90.13	90.58	91.18
	30	0.0324	89.11	89.63	89.94	90.14	90.59	91.24
	33	0.0324	89.10	89.62	89.94	90.16	90.56	91.22

P-values	
Treatment (Probes)	0.263
Sensors (S)	0.062
Interaction Probes x S	0.071

***S = Sensors**

Appendix 4.1 The ANOVA of water table height (mm) and time (Days) on O_2 of the sandy Hutton soil

Source of variation	Degrees of	Sum of	Mean sum	*F* ratio	P
Water Table Height (WT)	3	0.249	0.083	11.46	0.000015
Days (D)	4	0.239	0.060	8.254	0.000060
Interaction WT x D	12	0.011	0.001	0.126	0.999773
Error	40	0.290	0.007		
Total	59	0.790			

Appendix 4.2 The effect of water table height (mm) and time (Days) on O_2 of the sandy Hutton soil

Days (D)	Water Table Height (WT)				Mean
	150 mm	300 mm	500 mm	800 mm	
1	20.40	20.35	20.43	20.51	20.42[a]
2	20.37	20.33	20.38	20.46	20.39[a]
3	20.30	20.25	20.36	20.46	20.34[a]
4	20.27	20.18	20.29	20.38	20.28[b]
6	20.23	20.17	20.24	20.37	20.25[b]
Mean	20.31[a]	20.26[a]	20.34[a]	20.44[b]	

Means followed by the same letter in either rows or columns do not differ from each other at the 5% level of significance.

Appendix 4.3 The ANOVA of water table height (mm) and time (Days) on CO_2 of the sandy Hutton soil

Source of variation	Degrees of	Sum of	Mean sum	*F* ratio	P
Water Table Height (WT)	3	13528471	4509490	82.02	0.000000
Days (D)	4	11427739	2856935	51.96	0.000000
Interaction WT x D	12	1100842	91737	1.669	0.111445
Error	40	2199130	54978		
Total	59	28256182			

Appendix 4.4 The effect of water table height (mm) and time (Days) on CO_2 of the sandy Hutton soil

Days (D)	Water Table Height (WT)				Mean
	150 mm	300 mm	500 mm	800 mm	
1	2993	3700	2733	2493	2980[a]
2	3093	3822	3176	2789	3220[a]
3	3873	4263	3422	2868	3607[b]
4	4077	4688	3892	3175	3958[c]
6	4250	4767	4322	3240	4145[c]
Mean	3657[b]	4248[a]	3509[b]	2913[c]	

Means followed by the same letter in either rows or columns do not differ from each other at the 5% level of significance.

Appendix 4.5 The ANOVA of water table height (mm) and time (Days) on O_2 of the loamy sand Hutton soil

Source of variation	Degrees of	Sum of	Mean sum	*F* ratio	P
Water Table Height (WT)	3	0.194	0.065	16.37	0.000000
Days (D)	4	0.291	0.073	18.38	0.000000
Interaction WT x D	12	0.029	0.002	0.618	0.813967
Error	40	0.158	0.004		
Total	59	0.673			

Appendix 4.6 The effect of water table height (mm) and time (Days) on O_2 of the loamy sand Hutton soil

	Water Table Height (WT)				Mean
Days (D)	**150 mm**	**300 mm**	**500 mm**	**800 mm**	
1	20.50	20.47	20.55	20.56	20.52[a]
2	20.43	20.42	20.52	20.51	20.47[ab]
3	20.37	20.32	20.45	20.50	20.41[bc]
4	20.33	20.30	20.38	20.48	20.37[cd]
6	20.23	20.28	20.35	20.42	20.32[d]
Mean	20.37[a]	20.36[a]	20.45[b]	20.50[b]	

Means followed by the same letter in either rows or columns do not differ from each other at the 5% level of significance.

Appendix 4.7 The ANOVA of water table height (mm) and time (Days) on CO_2 of the loamy sand Hutton soil

Source of variation	**Degrees of**	**Sum of**	**Mean sum**	***F* ratio**	**P**
Water Table Height (WT)	3	11574604	3858201	33.75	0.000000
Days (D)	4	24092965	6023241	52.68	0.000000
Interaction WT x D	12	1499703	124975	1.093	0.391775
Error	40	4573240	114331		
Total	59	41740512			

Appendix 4.8 The effect of water table height (mm) and time (Days) on CO_2 of the loamy sand Hutton soil

	Water Table Height (WT)				Mean
Days (D)	**150 mm**	**300 mm**	**500 mm**	**800 mm**	
1	3017	3400	2718	2730	2966[a]
2	3620	3838	2971	3146	3394[b]
3	4353	4578	3479	3410	3955[c]
4	4633	4962	4095	3633	4331[c]
6	5360	5298	4267	4007	4733[d]
Mean	4197[a]	4415[a]	3506[b]	3385[b]	

Means followed by the same letter in either rows or columns do not differ from each other at the 5% level of significance.

Appendix 4.9 The ANOVA of water table height (mm) and time (Days) on O_2 of the Bainsvlei soil

Source of variation	**Degrees of**	**Sum of**	**Mean sum**	***F* ratio**	**P**
Water Table Height (WT)	3	0.222	0.074	19.18	0.000000
Days (D)	4	0.288	0.072	18.67	0.000000
Interaction WT x D	12	0.023	0.002	0.501	0.901697
Error	40	0.154	0.004		
Total	59	0.687			

Appendix 4.10 The effect of water table height (mm) and time (Days) on O_2 of the Bainsvlei soil

Days (D)	Water Table Height (WT)				Mean
	150 mm	300 mm	500 mm	800 mm	
1	20.50	20.47	20.54	20.55	20.51[a]
2	20.43	20.38	20.49	20.51	20.46[ab]
3	20.40	20.37	20.48	20.52	20.44[b]
4	20.33	20.27	20.42	20.45	20.37[c]
6	20.30	20.18	20.37	20.42	20.32[c]
Mean	20.39[a]	20.33[a]	20.46[b]	20.49[b]	

Means followed by the same letter in either rows or columns do not differ from each other at the 5% level of significance.

Appendix 4.11 The ANOVA of water table height (mm) and time (Days) on CO_2 of the Bainsvlei soil

Source of variation	Degrees of	Sum of	Mean sum	*F* ratio	P
Water Table Height (WT)	3	8953331	2984444	36.01	0.000000
Days (D)	4	8600651	2150163	25.94	0.000000
Interaction WT x D	12	637985	53165	0.642	0.794077
Error	40	3315060	82877		
Total	59	21507028			

Appendix 4.12 The effect of water table height (mm) and time (Days) on CO_2 of the Bainsvlei soil

Days (D)	Water Table Height (WT)				Mean
	150 mm	300 mm	500 mm	800 mm	
1	2503	2855	2625	2051	2508[a]
2	2927	3238	2712	2300	2794[ab]
3	3073	3413	2809	2418	2928[b]
4	3473	3918	3210	2725	3332[c]
6	3557	4325	3549	2835	3567[c]
Mean	3107[b]	3550[a]	2981[b]	2466[c]	

Means followed by the same letter in either rows or columns do not differ from each other at the 5% level of significance.

Appendix 4.13 The ANOVA of water table height (mm) and time (Days) on O_2 of the Sepane soil

Source of variation	Degrees of	Sum of	Mean sum	*F* ratio	P
Water Table Height (WT)	3	0.504	0.168	8.890	0.000123
Days (D)	4	0.334	0.083	4.414	0.004753
Interaction WT x D	12	0.085	0.007	0.375	0.964912
Error	40	0.756	0.019		
Total	59	1.679			

Appendix 4.14 The effect of water table height (mm) and time (Days) on O_2 of the Sepane soil

Days (D)	Water Table Height (WT)				Mean
	150 mm	300 mm	500 mm	800 mm	
1	20.30	20.45	20.45	20.09	20.32[a]
2	20.23	20.38	20.43	20.15	20.30[a]
3	20.20	20.27	20.29	20.19	20.24[ab]
4	20.17	20.23	20.25	20.04	20.17[ab]
6	20.10	20.18	20.21	20.01	20.13[b]
Mean	20.20[ab]	20.30[a]	20.33[a]	20.10[b]	

Means followed by the same letter in either rows or columns do not differ from each other at the 5% level of significance.

Appendix 4.15 The ANOVA of water table height (mm) and time (Days) on CO_2 of the Sepane soil

Source of variation	Degrees of	Sum of	Mean sum	*F* ratio	P
Water Table Height (WT)	3	10972558	3657519	30.08	0.000000
Days (D)	4	9588500	2397125	19.71	0.000000
Interaction WT x D	12	307622	25635	0.211	0.997066
Error	40	4863971	121599		
Total	59	25732651			

Appendix 4.16 The effect of water table height (mm) and time (Days) on CO_2 of the Sepane soil

Days (D)	Water Table Height (WT)				Mean
	150 mm	300 mm	500 mm	800 mm	
1	3193	2933	2745	3917	3197[a]
2	3623	3135	2888	3889	3384[a]
3	3933	3605	3191	4501	3808[b]
4	4147	3845	3465	4631	4022[bc]
6	4523	4098	3701	4804	4282[c]
Mean	3884[b]	3523[a]	3198[a]	4348[c]	

Means followed by the same letter in either rows or columns do not differ from each other at the 5% level of significance.

Appendix 4.17 The ANOVA of water table height (mm) and time (Days) on O_2 of the Valsrivier soil

Source of variation	Degrees of	Sum of	Mean sum	*F* ratio	P
Water Table Height (WT)	3	0.726	0.242	23.74	0.000000
Days (D)	4	0.631	0.158	15.46	0.000000
Interaction WT x D	12	0.129	0.011	1.054	0.421898
Error	40	0.408	0.010		
Total	59	1.894			

Appendix 4.18 The effect of water table height (mm) and time (Days) on O_2 of the Valsrivier soil

Days (D)	Water Table Height (WT)				Mean
	150 mm	300 mm	500 mm	800 mm	
1	20.23	20.33	20.35	19.91	20.21[a]
2	20.17	20.28	20.29	19.99	20.18[a]
3	20.13	20.22	20.26	20.04	20.16[a]
4	20.00	20.07	20.15	19.92	20.03[b]
6	19.87	19.95	20.10	19.83	19.94[b]
Mean	20.08[b]	20.17[ab]	20.23[a]	19.94[c]	

Means followed by the same letter in either rows or columns do not differ from each other at the 5% level of significance.

Appendix 4.19 The ANOVA of water table height (mm) and time (Days) on CO_2 of the Valsrivier soil

Source of variation	Degrees of	Sum of	Mean sum	*F* ratio	P
Water Table Height (WT)	3	25196221	8398740	96.47	0.000000
Days (D)	4	18597524	4649381	53.41	0.000000
Interaction WT x D	12	1909532	159128	1.828	0.076365
Error	40	3482349	87059		
Total	59	49185626			

Appendix 4.20 The effect of water table height (mm) and time (Days) on CO_2 of the Valsrivier soil

Days (D)	Water Table Height (WT)				Mean
	150 mm	300 mm	500 mm	800 mm	
1	4453	3455	3342	4827	4019[a]
2	4760	3878	3567	4852	4264[a]
3	4937	4320	3846	6091	4799[b]
4	5433	5112	4286	6163	5248[c]
6	5680	5402	4550	6280	5478[c]
Mean	5053[c]	4433[b]	3918[a]	5643[d]	

Means followed by the same letter in either rows or columns do not differ from each other at the 5% level of significance.