DETERMINATION OF WATER POTENTIAL – MOISTURE CHARACTERISTICS OF SOIL POROUS MEDIA

Barbara Witkowska-Walczak, Ryszard Walczak
Cezary Sławiński

EDITED BY
Barbara Witkowska-Walczak, Ryszard Walczak

Centre of Excellence for
Applied Physics in Sustainable
Agriculture AGROPHYSICS

Institute of Agrophysics
Polish Academy of Sciences

EU 5th Framework Program
QLAM-2001-00428

Barbara Witkowska-Walczak, Ryszard Walczak,
Cezary Sławiński

DETERMINATION OF WATER POTENTIAL-MOISTURE CHARACTERISTICS OF SOIL POROUS MEDIA

Lublin 2004

EC Centre of Excellence AGROPHYSICS
Centre of Excellence for Applied Physics
in Sustainable Agriculture
QLAM-2001-00428

Reviewed by: Prof. dr hab. Ryszard Dębicki

ISBN 83-87385-93-X

Edition: 180 copies

Project of cover: Grzegorz Józefaciuk
Computer edition: Irena Kulawiak

Printed by: ALF-GRAF, ul. Kościuszki 4, 20-006 Lublin, Poland

PREFACE

The amount of water contained in a soil and the energy state of water in the soil are important factors affecting the growth and yield of plants. The other soil properties depend strongly on water content, *ei* consistency, plasticity, strength, compactibility, penetrability, stickiness and trafficability. In clayey soils, swelling and shrinking associated with addition or extraction of water change the total porosity and bulk density of the soil, as well as its pore size distribution. Soil moisture also determines the air content and gas exchange in the soil, affecting the respiration of roots, the activity of microorganisms and oxidation-reduction processes.

Water content (moisture) and matric potential are functionally related to each other and the graphical representation of this relationship is called the water retention curve of soil. The relationship is not unique, however, it is affected by direction and rate of change of soil moisture and is sensitive to changes in soil volume and structure. Moisture and matric potential vary widely in space and time as the soil is wetted by rain, drained by gravity and dried by evaporation and root extraction, from saturation to air dryness.

This elaboration, based on the papers of Hillel (1971, 1998), Iwata *et al*. (1988), Kutilek and Nielsen (1994) and the own works of the authors, can be a source of basic information on static water characteristics of soil not only for soil scientists, but also for specialists in the other branches, *eg* agronomy, ecology, environmental protection, forestry, geomorphology, meteorology and water management.

The authors

I. INTRODUCTION

Water economy, the basic goal of which is rational management of natural water resources, their transformation into available resources and protection against destruction, is closely linked with the cycling of water in the biosphere.

One of the most important entities determining the cycling of water in the biosphere is the soil. The soil, *ie* the pedosphere, constitutes the surface layer of the earth's crust, affected by soil-forming processes. As an essential element of the lithosphere, it forms one of the most important components of terrestrial and aqueous ecosystems. Soils number among the non-renewable resources of the earth and perform several functions, above all they provide habitat for the growth of plants and transformation of mineral and organic components. Thanks to their peculiar properties, soils reflect the history of the landscape they are part of. A comprehensive knowledge of the physical, chemical and biological properties of soils is indispensable in planning their proper utilisation for human needs following the principles of sustainable development. This knowledge is also necessary for the rational utilisation of production space, regionalisation of plants, preparation of economic plans, determining crop rotation and farming methods (Connoly, 1998; Hillel, 1971; Kutilek and Nielsen, 1994; Vorosmarty *et al.,* 1989; Witkowska-Walczak, 2001).

The soil performs the function of a water distributor: it receives precipitation and retains some of it while some water is transferred to the deeper layers and groundwater and some returned to the atmosphere by way of evaporation and transpiration. Furthermore, the amount of water flowing down the surface of the earth is determined by the soil and the land gradient. In Poland, where the average annual precipitation amounts to approx. 600 mm, *ie* 186.2 km^3 of water, arable land which covers 47% of the country's area receives (assuming that the precipitation is even) 87.5 km^3 of water and this amount of water can be effectively used if the land is farmed properly (Gliński *et al.*, 1989; Lvovicz, 1965; Milly, 1994).

The influence of the soil on the water balance of the particular regions has been known for a long time and research has showed that two properties of the soil exert an essential influence on the relationship between the components of water balance: the water capacity and permeability of the soil (Ahuja *et al.*, 1984; Crawford *et al.*, 1995). The knowledge of these properties, the influence of various factors on them and the methods of their regulation allows to control water circulation processes in the biosphere and, therefore, regulate the amount of water in aqueous systems, surface and underground water reservoirs as well as provide plants with proper development conditions while using water in the most efficient way (Mioduszewski, 1996).

The above is particularly important in areas used for agricultural production where adequate management of water resources leads to sustainable development, through the use of suitable land improvement and agrotechnical measures, crop rotation (taking into account plant varieties which have a lower demand for water) and the creation of protection zones around arable fields, pastures and forest land (Kaca, 1996;Vyn and Raimbault, 1993).

The amount and level of water in the soil considerably influences plant growth and development. Moreover, in agricultural areas with a balanced water economy, the water capacity and permeability of the soil may be changed through the modification of its physical properties such as aggregate structure and density by means of agrotechnical measures (Hakansson and Lipiec, 2000, Hamblin, 1982; Mbagwu and Auerswald, 1999; Oades and Waters, 1991; Quirk and Murray, 1991; Russell, 1971; Shainberg, 1997; Williams *et al.*, 1983). However, for agrotechnical measures to bring the desired results, the direct links between soil structure and density, *and* its hydrophysical properties must be determined because, despite several studies on the subject, these links have not been sufficiently explained so far.

Changes in the aggregate structure of soils can be achieved by applying appropriate agrotechnical and structure-forming measures. It is extremely important to learn about the close relationships between the size and durability of the elements of aggregate structure and the water-air properties of soils which condition plant growth and development because it will lead to the maximisation of vegetal production and, at the same time, provide guidelines for controlling the range of the components of the soil's water balance. Moreover, obtaining such information will make it possible to predict the amount of crops with the use of numerous developed models for predicting plant growth and productivity. These models include physical sub-models that primarily use data concerning the influence of the soil's aggregate structure on its water retaining capacity and the possibility of its movement in the soil profile (Haverkamp and Reggiani, 2002, Kosugi, 1994; Walczak, 1984; Witkowska-Walczak, 2000).

1. Soil properties

1.1. Soil structure

The development of the soil begins when organisms, usually the most primitive forms, penetrate the rock-mantle. The soil is not homogenous; it is a polydispersive, multi-component system that varies in time and space and consists of three phases (Kutilek and Nielsen, 1994; Turski *et al.*, 1999, Zawadzki, 1999):

- the solid phase that contains the original minerals, the products of their weathering as well as organic matter, a product of a peculiar part of the solid phase, *ie* a complex of soil organisms;

- the liquid phase, *ie* soil water or, rather, the soil solution;
- the gaseous phase, *ie* the soil air.

The ratio of the above-mentioned phases in a specific volume of soil varies in different soil types and profiles, depending on the origin and evolution of the soil. The ratio of these phases, shaped by natural factors, is relatively stable for a given soil (Fig. I.1). Under the influence of agrotechnical measures, particularly in the arable layer, this ratio may fluctuate considerably, even in the growing season. It particularly applies to changes in the volume of the space occupied by the air and the soil solution.

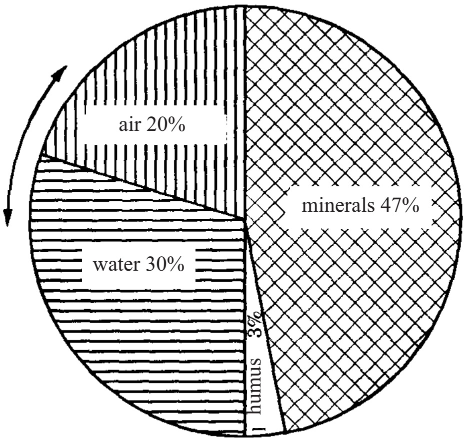

Fig. I.1. The share of the particular phases in mineral soil. The arrow indicates the possibility of changes depending on the share of various groups of soil pores (Buckman and Brady, 1971).

The degree of fineness is the basic parameter which characterises the solid phase of the soil as it consists of particles of different diameters, single or bound in soil aggregates of different degrees of stability. The share of soil particles of specific dimensions in the total amount of soil is called the granulometric distribution.

The texture of the soil, *ie* its fineness in its solid phase conditions the magnitude of the specific surface area and sorption properties of the soil. The finer the soil, the greater its specific surface and the larger its sorption complex which directly depends on the contents of the colloidal fraction in the soil.

The organic matter (organic compounds) is formed by dead soil components that contain carbon. It can undergo mineralization or humification. In average soil conditions the mineralization dominates over humification. With excessive moisture and lack of oxygen, part of the organic matter may undergo peat formation.

Soil humus is the part of soil organic matter which underwent the humification process. Humus can be formed from all compounds that reach the soil in organic residue, and the decisive role in their decomposition is played by enzymes, primarily enzymes of soil microorganisms (Turski *et al.*, 1999).

1.2. Soil compaction

Each soil formation, depending on its texture, type of structure and the influence of external factors, constitutes a system with a distinct degree of compaction (Turski *et al.*, 1999). The measure of soil compaction is soil density that is directly related to the density of the solid phase and the volume of free soil spaces. *Solid phase density*, expressed in g cm^{-3}, is the weight of volume units of solid soil particles. Its values strictly depend on the density of the components (quartz: 2.65, orthoclase: 2.56, plagioclase: 2.60-2.76, calcite: 2.71, micas: 2.76-3.0, silty minerals: 2.0-2.7, heavy minerals: 2.9, organic matter: approx. 1.2 g cm^{-3}). The solid phase density of mineral soils ranges between 2.55 and 2.70 g cm^{-3}. Most frequently it stands at 2.60-2.65 which is similar to the density of the main component, quartz. Soils rich in organic matter exhibit a considerably lower solid phase density, approx. 1.4-1.8 g cm^{-3}, depending on the presence of mineral admixtures.

In contrast to solid phase density, whose value is essentially constant for a given soil (it can be modified only by the introduction of large amounts of organic matter), soil density is variable and fluctuates within broad limits. Soil density (bulk density, compaction) is the weight of a soil volume unit with a natural dry structure at the temperature of 105°C. It is also expressed in g cm^{-3}. The density of humus levels of mineral soils most frequently ranges between 1.0 and 1.6 g cm^{-3}. The density of sandy soils stands at 1.3-1.7 g cm^{-3}; in the case of clay and silty soils it ranges between 1.1 and 1.6 g cm^{-3}. Soils with aggregate structure exhibit a lower density (between 0.9 and 1.4 g cm^{-3}) than cohesive soils (1.5-1.7 g cm^{-3}). Organic matter reduces soil density by reducing solid phase density on the one hand and the participation in the formation of stable aggregates, whose presence increases the volume of free soil spaces, on the other.

Soil density (compaction) varies over time. During the growing season the soil exhibits various density levels that can considerably deviate from the optimum level for a given plant. Both considerable soil loosening and its compaction are detrimental to most plants. Considerable soil loosening is accompanied by the increase of the volume of large pores, which results in the increase of water permeability and, in consequence, the runoff of water as a result of gravity. At the same time the volume of medium and small pores that enable the replenishment of water shortages by means of capillary rise decreases. If compaction increases the volume of large pores decreases and the volume of medium and small pores increases. Then the air capacity and permeability of the soil decreases, whereas the content of water inaccessible to plants as well as the soil's resistance against the plants' roots and working elements of farming equipment decreases. Plants growing on a highly-compacted soil suffer from air shortage and cannot take advantage of considerable amounts of water bound in small pores in the soil with a force too strong for the suction force of the plants' roots. Soil compaction can change under the influence of natural factors and factors related to human activity.

Soil density increases primarily as a result of:

- the self-consolidation of the soil as a result of gravity; it begins immediately after soil loosening; the settlement of the soil is facilitated by its wetting as a result of which the effect of capillary forces increases whereas the friction between the particles decreases;
- the compacting effect of the wheels of heavy machines and agricultural equipment; the intensification of plant production is accompanied by the introduction of heavy equipment that, particularly in the case of an excessively wet soil, causes its strong compaction and the destruction of soil aggregates.

Soil loosening is caused by:

- the activity of soil microflora and fauna and root systems of higher plants, primarily perennial papilionaceous plants;
- agrotechnical measures such as ploughing, harrowing, cultivator tillage and disk harrowing that are aimed at cutting and crumbling the compacted soil mass.

1.3. Soil porosity

The soil's solid phase particles form a porous system characterised by the presence of pores of various dimensions and shapes (Turski *et al.*, 1999; Zawadzki, 1999). The volume of all soil spaces filled with water or air is called the total soil porosity, expressed as the percentage of the total volume of soil or as the amount of cm^3 in 100 cm^3 (% vol., %, v/v). The level of total porosity depends on the soil's grain size distribution, its structure and the activity of loosening and compacting factors. The total porosity of humus layers in sandy soils does not usually exceed 50% (v/v) whereas in the case of clayey and silty soils that have an aggregate structure it often reaches 65% vol. In highly-compacted soils its value often drops to 35% vol. The total porosity of organic soils is considerably higher than that of mineral ones and amounts to 80-90% vol. Soil pores are basically divided into primary and secondary pores. *Primary pores* can be defined as inter-granular spaces between elementary particles that are closely-packed but not linked, *eg* in compacted sandy soils. *Secondary pores* are characteristic to formations where elementary particles are loosely-packed or form aggregates. Secondary pores include spaces whose shape, size and direction are entirely random and dependant from the shape and position of solid particles, *eg* pores between loose mineral grains, loose plant residues, or fractures and hollowed-out spaces.

With regard to diameter, and the related capacity for retaining specific forms of water, pores are divided into (Zawadzki, 1999):

- macropores: >30 μm in diameter, filled with air or, in exceptional cases, with water, *eg* after copious precipitation or sprinkling irrigation;
- mezopores: capillary pores, 30-0.2 μm in diameter, filled with water available to plants or with air, depending on soil moisture;
- micropores: microcapillary pores, <0.2 μm in diameter, filled with highly-bound water that is inaccessible to plants.

The volume of the particular groups of pores, expressed as a percentage of the total volume of soil, is called the pore structure or differential porosity. It is the volume of macro-, mezo- and micropores, not the total porosity, that indicates the character of water and air relations in the soil. Formations with the optimum total porosity can sometimes exhibit a disadvantageous pore structure which does not provide plants with the sufficient amount of water or air.

Plant growth and development takes place only if the soil is sufficiently aerated (Turski *et al.*, 1999; Zawadzki, 1999). Soil air occupies the pores from which water was removed by filtering to the deeper layers or evaporation because water occupies all pores in the soil only after intensive precipitation or sprinkling irrigation. Water runs off from aerated pores quickly because of gravitation, and the greater the diameter of the pores, the faster the runoff rate. Water flows particularly quickly through pores whose diameter exceeds 50 μm.

The amount of air contained in the soil also depends on capillary rise. Under actual soil conditions, the height of capillary rise is not as high as theoretical calculations suggest because capillary soil pores are irregular spaces of varying diameters and different shapes, and not thin tubes with an ideally circular cross-section. Maximum heights of capillary rise in field conditions range from approx. 6 cm in formations made up of particles that are 1-2 mm in diameter to 2.43 m in soils composed of particles measuring 0.01-0.2 mm. The speed of capillary rise of water is directly proportional to the capillary radius. The smaller the radius the higher the friction is which hampers the movement of water.

The phenomenon of capillary movement in the soil is of great importance because it is due to this phenomenon that water from the deeper layers can be transported upwards and used by the plants. The content of soil pores, which determine the height of capillary rise, depends on the texture of the soil. The higher the number of fine silt and clayey particles, the higher the volume of the smaller capillary pores is. Fig. I.2 presents the state of field water capacity, *ie* the state when only capillary water remains after the outflow of gravitational water.

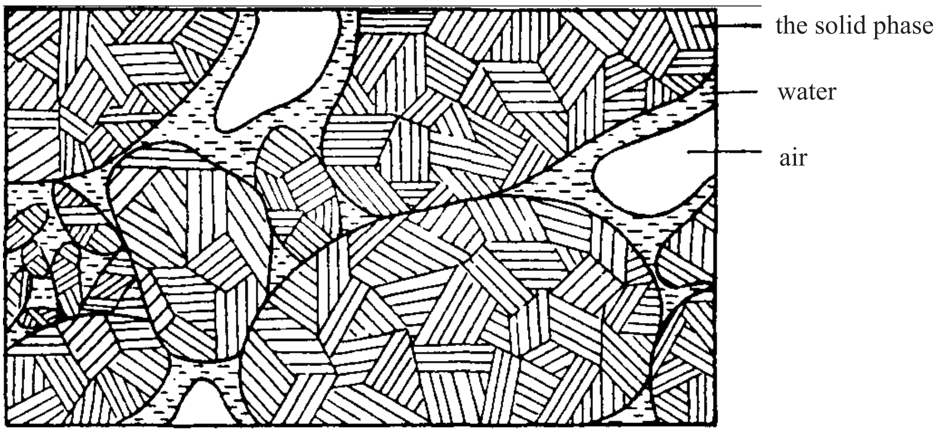

Fig. I.2. The state of field water capacity; water surrounds the solid phase particles and entirely fills the smaller pores (Thompson and Troeh, 1978).

From the practical point of view the degree of soil water availability to plants is particularly important. Generally it may be assumed that for the water to be accessible it must be bound in the solid phase with a force smaller than the plant roots' suction force (approx. 1.55 10^4 hPa) and must remain in the soil long enough for the roots to absorb it. Capillary water primarily fulfils this condition. Chemically-bound water, water in the form of ice and gravitational water is virtually useless for plants.

2. Soil water retention curve

Water contained in soil is bound with solid phase particles with a specific force called soil water potential. With the full water saturation of the soil, the potential equals zero; with the loss of water in the soil it takes negative values reaching absolute maximum values up to 10^8 J m^{-3}. The functional dependence between the soil water content and its water potential is called the soil water retention curve or the water sorption curve. Other names that are used include: static water characteristics of the soil, water retention curve or the pF curve. This curve indicates the force with which the water is bound in the soil, from the state of full saturation to a dry state (Walczak, 1977). From the retention curve for a given soil you can read the content of water accessible, in different degrees, to plants and infer about the character of water and soil relations. The course of the water retention curve depends on several factors, primarily on the texture of the soil, its compaction, aggregation, specific surface area and content of organic matter (Hillel, 1998).

The dependence between the soil water potential and the water content (moisture) is not an unequivocal function, *ie* the same value of the soil water potential may correspond to various values of soil moisture, depending on the way in which a given state was achieved (drying or wetting). This phenomenon is called hysteresis and, in the case of water bound in the soil by capillary forces, results from the irregular distribution of soil pores (Enderby, 1955; Everett, 1956; Hillel, 1998; Kaszubkiewicz, 2000; Walczak, 1977; Witkowska-Walczak, 2000; 2003b)

The purpose of this elaboration is to show:

- the status of water in the soil,
- the measurements of water content and water potential in the soil,
- the relation between water retention curves and soil compaction and aggregation,
- the pedotransfer functions for determination of water retention curves.

II. THE STATUS OF WATER IN SOIL

The complex nature of the pore space in soil and the water held therein makes it difficult to specify directly the force fields acting on the water. The description of soil-water movement on a microscopic scale depends upon not only the quantification of the force fields but also on the specification of the solid matrix geometry. These factors affect the soil-water system at all times. For specific systems some factors dominate and a particular approach can be used to describe the system in terms of these dominant factors. The overall effect of the force fields may be obtained by measuring the work required to remove an increment of water from the soil at some equilibrium state. The measured quantity when related to a reference state is known as the total potential of the soil water. It is convenient to divide the total potential into four component potentials (Boersma *et al.*, 1972; Hillel, 1971; Philip, 1960; Kutilek and Nielsen, 1994; Walczak, 1977):

- gravity potential, which relates to position in the gravitational field with respect to an arbitrary reference elevation;
- matric potential, which relates to adsorption forces between solid surfaces and water, including the effect of cohesive forces between water molecules;
- osmotic potential, which relates to forces of attraction between ions and water molecules;
- pressure potential, which relates to forces arising from unequal pressures in the gaseous phase.

Theories of water flow are based on empirically derived transport equations and well established physical principles. Hydrodynamic approaches have been used to describe liquid flow for cases in which salt and temperature effects are not significant factors. When salts are a contributing factor, diffusion and dispersion mechanisms must be incorporated into theory and the concepts of electrostatics and electrokinetics are applicable and useful. For flow systems in which temperature and salt concentrations are variable, as well as salts, the recently developed theory for the thermodynamics of irreversible processes has been applied to soil water flow. The purpose of this chapter is to describe these various approaches in detail and relate them to each other.

1. Soil water potential

If two equilibrium states of a soil system are considered, the work necessary to move water from one state to the other is a measure of the potential energy acquired in the process of transporting the water. Potential, by definition, is potential energy per unit quantity of the substance under consideration and its sign depends upon the state chosen as reference. If the quantity is measured on a mass basis, then potential is the potential energy per unit mass. And, if the reference is pure water with a flat air-water interface at the same elevation in an unsaturated soil column, the sign is negative since work must be done to remove water. This definition does not include potential energy differences due to elevation, or what is commonly referred to as gravitational potential, but it usually includes matric, osmotic and pressure potentials so that total soil water potential is the sum of the gravitational potential and soil water potential. In many cases, air pressure differences are non-existent or negligibly small and soil water potential is taken to be only the sum of the matric and osmotic potentials.

The idea of soil water potential is thermodynamical approach. The partial specific Gibbs function or chemical potential of water in a multi-component system is an expression of the capability of a unit mass of water to do work as compared to the work that an equal mass of free water at the same location could do. Hence, the potential of soil water as derived from the Gibbs function of thermodynamics likewise is negative. Gravitational potential may be derived from Newton's law of gravitation:

$$F = \frac{GmM}{x^2} \tag{II.1}$$

where: G is the universal gravitational constant, m and M are the respective masses of any two bodies, and x is the distance between their centres of mass. Within the soil profile (at a depth h), the force of attraction between the earth (mass M and radius R) and a body of water (mass m) will manifest itself through the amount of work necessary to lift the water to the soil surface. The work per mass of water Ψ_g, moved will be:

$$\Psi_g = \int_{R-h}^{R} \frac{F}{m} dx = \frac{GMh}{(R-h)R} = \frac{GMh}{R^2} \tag{II.2}$$

since: $R » h$.

By convention, the gravitational force acting on a body (mass m) at the earth's surface is:

$$F = mg \tag{II.3}$$

where: g is the gravitational acceleration. Equating equations [II.1] and [II.3], we find that:

$$g = \frac{GM}{R^2} \tag{II.4}$$

where: R replaces x for a body near the earth's surface, or that equation [II.2] is merely

$$\Psi_g = gh \qquad \text{II.5}$$

Equation II.5 forms the definition of the gravitational potential Ψ_g and shows that it is linearly related to small changes in elevation. In problems of water flow in unsaturated materials the matric potential, Ψ_M, plus the gravity potential, Ψ_g, make up what often is known as the hydraulic potential. Osmotic and pneumatic components usually are negligible for ordinary flow problems and are neglected. However, they are important in some problems.

In addition to the soil matrix (including exchangeable ions), solute concentrations, air pressure, and elevation contributing to the potential of soil water, temperature plays an important role. Since temperature is an expression of kinetic energy and equilibrium systems exist only at uniform temperature, potential concepts cannot be used. In this case thermodynamic concepts often are more useful.

1.1. Equilibrium thermodynamic approach

Equilibrium thermodynamics allows the derivation of a mathematical equation relating the soil-water potential to its four components. The equation relating, by definition, Gibbs free energy G and other state variables is:

$$G = U - TS + PV \qquad \text{II.6}$$

where: U - internal energy, T - temperature, S - entropy, P - pressure, and V- volume. The total differential equation for [II.6] is:

$$dG = dU - TdS - SdT + Pdv + VdP \qquad \text{II.7}$$

a multi-component system for,

$$dU = TdS - PdV + \sum_j \mu_j dn_j \qquad \text{II.8}$$

where: μ_j and n_j are the chemical potential and the amount (in moles or grams), respectively, of the jth component. Therefore,

$$dG = SdT + VdP + \sum_j \mu_j dn_j \qquad \text{II.9}$$

If water is one of the j components, the above equation becomes:

$$dG = -SdT + VdP + \mu_W dn_W + \sum_i \mu_i dn_i \qquad \text{II.10}$$

where: w denotes the water component. It is μ_W, which is of interest, since it is the chemical potential or the partial specific free energy, $\partial G / \partial n_W$, for soil water in the case being considered. The total differential equation relating μ_W and the common variable of interest is:

$$d\mu_W = \frac{\partial \mu_W}{\partial T} dT + \frac{\partial \mu_W}{\partial P} dP + \frac{\partial \mu_W}{\partial nW} dn_W + \sum_i \frac{\partial \mu_W}{\partial n_i} dn_i \qquad \text{II.11}$$

where: the last term is for i species of ions present in the soil water. Equation [II.11] may be simplified by recognizing that:

$$\frac{\partial \mu_W}{\partial T} = \frac{\partial(\partial G / \partial n_W)}{\partial T} = \frac{\partial(\partial G / \partial T)}{\partial n_W} = \frac{\partial S}{\partial n_W} = \bar{s}_W \qquad \text{II.12}$$

and

$$\frac{\partial \mu_W}{\partial P} = \frac{\partial(\partial G / \partial n_W)}{\partial P} = \frac{\partial(\partial G / \partial P)}{\partial n_W} = \frac{\partial V}{\partial n_W} = \bar{v}_W \qquad \text{II.13}$$

where:s_W, and v_W are the partial molar entropy or specific entropy and the partial molar volume or the specific volume, respectively. By substituting s_W and v_W into equation [II.11] and integrating from selected standard conditions (designated by superscript 0), there results:

$$\mu_W - \mu_W^0 = -\int_{T^0}^{T} \bar{s}_W \, dT + \int_{P^0}^{P} \bar{v}_W \, dP + \int_{n_W^0}^{n_W} \frac{\partial \mu_W}{\partial n_W} dn_W + \sum \int_{n_i^0}^{n_k} \frac{\partial \mu_W}{\partial n_i} dn_i \qquad \text{II.14}$$

The negative of $(\mu_W - \mu_W^0)$ is defined as the soil water potential with the standard state being T° = 298 K, P° = atmospheric pressure, n_W^0 – water content at saturation, and $n_W^0 = 0$ (zero soluble salts). The terms on the right side of equation [II.14] are in the order: temperature component, pressure component, matric component, and solute component. While convenient, the association of these names with these terms has lead to some misunderstanding. By definition, the chemical potential of soil water is the change in free energy of the whole system brought about by the addition of a small increment of water. This change cannot be assigned to a change in the energy of water alone but must also include all energy changes resulting from water-ion interactions, solid-liquid interface interaction, *etc*. The change in chemical potential of soil water with respect to soil water $\partial\mu_W \,/\, \partial n_W$, becomes, in a sense, a second derivative of the total free energy change in the system. This must be kept in mind during the physical interpretations of experimental data.

Table II.1. Relations between soil water potential units and equivalent diameter of soil pores (Walczak, 1977)

[pF]	[Bar]	[J m^{-3}]	[kJ kg^{-1}]	[hPa]	[cm H_2O]	Equivalent diameter of soil pores [μm]
0	0.001	98.1	0.1	0.98	1.0	> 1200
0.4	0.002	246	0.2	2.46	2.5	1200
1.0	0.010	981	1.0	9.81	10.0	300
1.5	0.03	3 100	3.1	31.00	31.6	100
2.0	0.10	9 810	9.8	98.10	100.0	30
2.2	0.16	15 596	16	155.96	158.5	18.5
2.3	0.20	19 620	20	196.20	199.5	15
2.7	0.50	49 050	50	490.50	501.2	6
3.0	1.00	98 100	100	981.00	1000.0	3
3.4	2.50	245 250	250	2452.50	2512.0	1.2
3.7	5.00	490 330	500	4903.30	5012.0	0.6
4.2	15.00	1 471 500	1500	14715.00	15849.0	0.2

The soil water potential in isothermal conditions is equivalent to the work accompanying the transfer of a unit mass of water from pure free water at a conventional zero level in the field of gravitation to a soil solution in a large volume of soil of a certain moisture. The unit of water potential is J kg^{-1}, but the other units, like J m^{-3}, J mol^{-1}, and erg g^{-1} are also used. Since:

$$\frac{J}{m^3} = \frac{Nm}{m^3} = \frac{N}{m^2}$$

the soil water potential can be expressed in pressure units.

During practical measurements of the soil water potential, determining the potential of the liquid phase we measure the so-called equivalent pressure or the sucking force. This is the pressure to which the soil solution should be subjected in order to stop the infiltration of pure water through semi-permeable media to the soil solution in which water has lower potential. It is also possible to subject pure water to a negative pressure. The equivalent pressure is a measure of the difference of potentials between pure water and water in soil solution. This pressure can be expressed in the following units: dyne cm^{-2}, bars (in USA), or in height of head of water in 10^{-2}m. Since pressures can adopt a wide range of values, Schofield proposed to express the equivalent pressure as the Briggsian logarithm of the equivalent pressure expressed in 10^{-2}m of head of water, and to mark it with the symbol pF. The table also presents the equivalent soil water potential values expressed in various units and their corresponding equivalent diameters of soil pores.

It should be noticed that in the above considerations the component characterizing the change in the geometry of the soil skeleton during the process of the swelling and contraction of soil occurring during its moistening and drying was not introduced.

2. The domain theory of hysteresis

All functional characteristics of the thermodynamical variables in the real process show hysteresis. For soil, the hysteresis effect is shown for example by the water potential-water content (moisture) relation. Thus this relation is nonexplicit, and the course of hysteresis is determined by the colloidal-porous properties of soil and by the ‘history’ of the system, that is by the way in which a given state was obtained. In all the statistical and dynamical investigations of water in soil the knowledge of the precise value of the water potentials and its quantity after any history of the changes of these quantities is necessary. For this purpose, apart from the boundary loops should also curves of the loops of hysteresis, the possible directions of the system inside the be investigated. The first correct attempts at solving this problem are the works of Everett and Enderby (1951, 1954, 1955, 1956) who formulated investigation methods for arbitrary hystereses.

The term hysteresis is used to describe the process by which, when any macroscopic property of the system is drawn as a function of the independent variable x oscillating between the x_1 and x_2 values, two curves are obtained (one for the growing values of x and the other for the diminishing ones), forming a loop. Everett and Enderby (1955, 1954) also present a general discussion of the phenomena of hysteresis. According to Everett (1954) all the irreversible processes fall into two classes:

- hysteresis – if all the points on both branches of the loop represent stable, reproducible values of the dependent variable,

- supersaturation – if the dependent variable can adopt non-stationary, irreproducible values.

From the point of view of thermodynamics, hystereses are characterized by temporarily independent loops at the oscillation of the independent variable between two fixed values, and supersaturation has no stable points of balance between those values. Everett assumes that the loops of hysteresis appear even at a very slow transition of the system from state to state, since there is no infinitesimal process to lead the system from one state to another. If the difference of energy between two states is very small but finite the thermal fluctuations will be sufficient for the transition of small groups of molecules over the barrier of potential separating the two states and the system can move from one state to another in a continuous way – the process is reversible, infinitesimal. If the system consists of larger groups of molecules the fluctuations cannot take it from one state to another in a continuous way and the system changes irreversibly through a series of discrete states that are not states of balance. If the groups of molecules are at the same time small enough for the macroscopic properties to be continuous functions and differ in the energies of interstate transitions, we get hysteresis in Everett's sense, otherwise we have a case of supersaturation. If the groups of molecules are so large that the transition of one group between two discrete states is noticeable in some macroscopic property of the system, this system will comprise supersaturation because of this property. If the group is so small that the energy needed for its transition between states is of the order of thermal fluctuations we have a reproducible process that does not give hysteresis. For the intermediate group sizes we get reproducible hysteresis in the sense of Everett and Enderby (1954, 1956). The group of molecules that can be in one of the two stable states in the process of hysteresis Everett calls the domain. When the state of the system which shows hysteresis changes, for in stance through a change of some thermodynamical external variable x while all the other external variables do not change, then there is no explicit relationship between the variable x and the internal dependent variable y. The variable y can adopt values equivalent to any point inside a certain area in the plane (x, y), called the area of hysteresis. This area is limited by two curves representing the sides of the system, when the variable x assumes all the values continually growing or continually diminishing, for which we get the loop of hysteresis. These curves are called the boundary curves. Through subsequent arbitrary increasing and diminishing of $-x$ it is possible to obtain any point within the loop of hysteresis, and such a change is represented by the internal curves.

The possibility of changes in the state of the system was increased because of the nonexplicit relationship between x and y, and moreover both these variables do not fully determine the thermodynamical state of the system since its properties can continually depend on the ‘history’ of the system, that is on the way in which the system has reached an arbitrary point, as for example (x_1, y_1). Schematic figures of the empting and filling processes of soil pores are shown in Fig. II.1. and II.2.

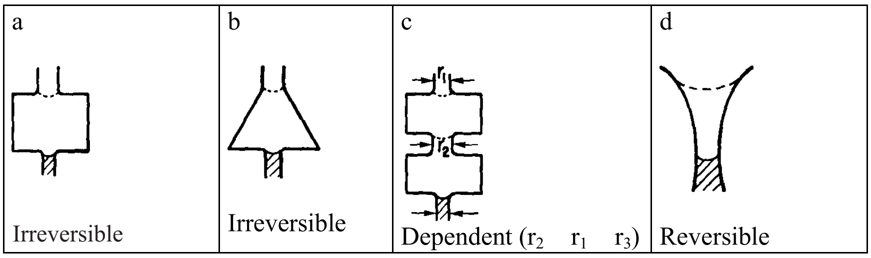

Fig. II.1. Various models of soil pores to explain reversible, irreversible and dependent domains (Iwata *et al.*, 1988).

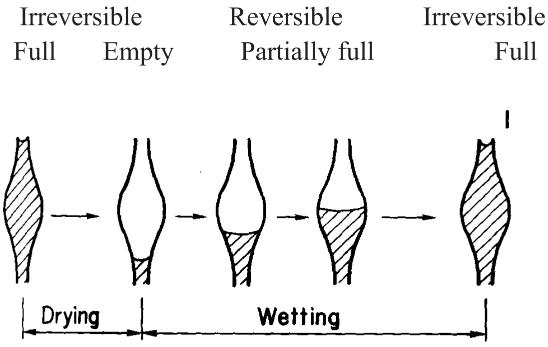

Fig. II.2. Schematic representation illustrating reversible and irreversible filling and empting of a pore (Iwata *et al.*, 1988).

In the model of domains of variables it is assumed that the system showing hysteresis can be treated (at least formally) as a complex of domains. Moreover each domain can be in state I or in state II, and the transition from one state to the other occurs at the change of the variable x. If the state I is stable for low values of x, then at an increase of x the transition of a domain to state II occurs at a certain value x_{12}, and the transition back to state I occurs at a value x_{21} that is lower than x_{12}. Further it is assumed that at least one of these changes must be irreversible in the thermodynamical sense, and the values x_{12} and x_{21} differ for particular domains. In order to reproduce the shapes of hystereses and the internal curves met in reality, it is necessary to introduce the functions of distribution, in which the difference $x_{12}-x_{21}$ is different for each domain. These functions require the introduction of at least one additional thermodynamical variable, that possesses the property of ‘memory’ and determines the future behaviour of the system with the help of its ‘history’ and allows for an explicit determination of the state of the system. The functions of ‘memory’ were introduced by Boltzmann in connection with the discussion of the phenomenon of relaxation, and now they are commonly used in the thermodynamics of irreversible processes (Coleman, 1964). The thermodynamical properties of a system of independent domains of which a part are in state I and another part in state II approximate the phenomenon of hysteresis, in the sense of Everett’s definition well. The model can be expanded to include cases when domains are in more than two stable states and are not independent, and then it describes all the phenomena of hysteresis.

For the purpose of describing such phenomena as the hysteresis of the water potential – soil moisture relationship, and also for the majority of other hystereses, the model of independent domains seems to be a sufficiently good approximation. The application of theory to the description of the processes of moistening and drying of soil and the utilizing of data obtained directly from experiments for the analytical description of the process are presented below. Soil is a porous body of a certain (continuous) geometry of pores totally or partially filled with water. The content of water can be changed, that is the soil can be dried or moistened, through the application of appropriate potential. At the increasing of potential, from the full saturation of soil with water, to a certain limit value S_{max} we get the curve *AB* (II.3).

Steering the process in the opposite direction, that is diminishing suction from S_{max} to zero, the moisture – potential relationship is described by the curve *BC.* The *AC* difference of volume, between the initial and final water content at pF 0, is equivalent to the quantity of air caught up during the change of pressure. If, after reaching point *C* we start to increase pressure, and thus dry the sample, we get curve

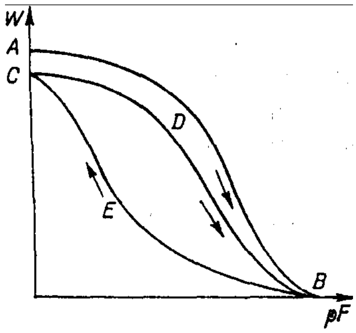

Fig. II.3. Hysteresis loop and first drying curve *AB* starting from saturation (Walczak, 1977).

CDB and after repeated wetting (diminishing of potential) we get the curve *BEC.* Analogous curves *CDB* and *BEC* (Fig. II.4.) are reproducible in subsequent cycles and are called the curves of drying and moistening respectively. The surface enclosed by these curves is the area of hysteresis. By arbitrarily increasing or diminishing the independent variable it is possible to obtain every point within the loop of hysteresis. In this way we obtain the so called internal curves, as for example the curve of level I of moistening – *DFC,* and the curve of level I of drying – *EGB,* starting respectively from the limit curves of drying and moistening, and ending at the extremities of the area of suction. The internal curves of the level II of moistening and drying, *GNE,* and *FMD*, come out from the internal curves of the level I of drying and moistening. The process can be continued in an analogous way.

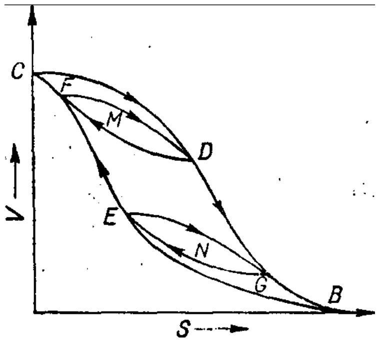

Fig. II.4. Hysteresis loop and primary scanning curves (Walczak, 1977).

The analysis of the hysteresis of the water properties of soil can be carried out 'knowing the values of suction and of water volume, that is of quantities directly measured in experiments. A domain is an element of the capillary-porous soil system and is described by three parameters: volume, suction at which it is filled with water, and suction of emptying during drying. The volume of water which comes out of soil when suction increases from 0 to S_{max} and which comes back into it when suction decreases from S_{max} to 0 can be divided into elements of volume. Each of the elements of volume preserves its identity (constant parameters), each of them is totally determined by a pair of small ranges of suction and *(jSf to jSf – dSf),* where the first of them presents the range of suction when the element comes out of the body, and the latter when the element returns to the body. Such an element is an example of an independent domain. With certain ranges of suction for different elements, each of them can be presented in a coordinate system S_e,,S_f,F where: S_e – suction during drying, S_f– suction during moistening, and F – the function of distribution (Fig. II.5).

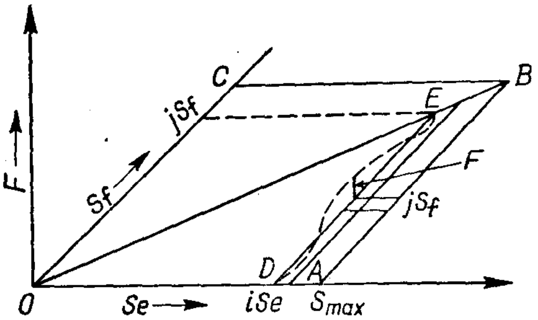

Fig. II.5. Graphical construction of the domain distribution function (pore size distribution functions) (Walczak, 1977).

For no elements is the value of S_f higher than that of S_e, so each of them must be described by a rectangle in ΔOAB. On the OB line are these elements that come out of, and return to the body at the same suction. The process of drying can be characterized by the movement of a plane perpendicular to (S_e, S_f) and parallel to S_f moving from the left hand to the right hand, side. While the process of wetting is described by the movement of a plane perpendicular to (S_e, S_f) and parallel to S_e moving from the BC line to the axis S_e. A chosen rectangle $dS_e\ dS_f$ determines the element going out when the suction increases from $_iS_e$ to $_iS_e + dS_e$, and returning when the suction diminishes from $_jS_f$ to $_jS_f - dS_f$. The volume of this element is equivalent to $F_{ij}\ dS_e\ dS_f$, which is also the volume of a V-block of a rectangular base $dS_e\ dS_f$. The function F (S_e, S_f) is non-negative in the triangle OAB, but its value is zero beyond this triangle. It is easy to notice that the change the volume of water $\frac{\partial v}{\partial _iS_e}$ coming out at a given suction (the slope of the boundary curve of drying at the point equivalent to a given value of suction and $(_iS_e)$) is equivalent to the surface of the section cut in a plane perpendicular to DE and parallel to S_f:

$$\frac{\partial v}{\partial _iS_e} = \int_{jSf}^{0} F(_iS_e, S_f) dS_f \qquad \text{II.15}$$

or

$$F_{ij} = \frac{\partial^2 v}{\partial _iS_e \partial _jS_f} \qquad \text{II.16}$$

The volume contained above the triangle of base *ODE*, which comes out of soil when potential increases from 0 to ${}_iS_e$ is:

$$V_i S_e = \int_0^{{}_iS_e} \int_0^{{}_jS_f} \frac{\partial^2 v}{\partial S_e \partial S_f} dS_e dS_f \qquad \text{II.17}$$

The domain theory of hysteresis describes the behaviour of the system in the case of dependent and independent domains. The concept of an independent domain covers domains which move from state I to state II at the same pair of values of the independent variable. The 'history' of the system does not influence the values of this variable. Actually, what influences the behaviour of a domain are the domains neighbouring it. If the behaviour of a domain depends on the state of the neighbouring domains, the domain is dependent. The state of the neighbouring domains is determined by the 'history' of the system, that is the values reached by the independent variable. Since the state of the neighbouring domains influence the behaviour of the considered domain, the pair of the values of the independent variable equivalent to its transition from state to state is a function of the maximum or the minimum values reached previously by the independent variable. The application of theory to dependent domains is practically pointless.

Soil is a porous body of a certain (continuous) geometry of pores totally or partially filled with water. The content of water can be changed, that is, the soil can be dried or moistened, through the application of an appropriate sucking force. The changes of the sucking force are accompanied, depending on the direction of these changes, by the entering or coming out of a, certain volume of water. In these considerations the sucking force will be adopted as the independent variable x, while the dependent variable will be the experimentally determined volume of water, equivalent to the volume of pores filled at the sucking force S_f and emptied at the sucking force S_e. If the same pair of sucking forces S_f and S_e, or, to be more precise, $S_f – dS_f$ and $S_e – dS_e$ is equivalent to a certain volume V, then such an element is an example of a dependent domain.

Soil pores can form string sequences. In such a case the behaviour of pores is influenced by the state in which the other pores of this sequence are. The string sequence of pores thus constitutes a dependent domain. Sequences of string pores are connected among themselves in a parallel way.

A group of string sequences forming a system of dependent domains can accidentally, if they are symmetrical, get transformed into a system of independent domains through parallel connections (Youngs, 1965). Soil consists of elementary solid particles of differentiated sizes in a wide range of values. The elementary particles usually do not occur in soil singly, but form aggregates of different sizes.

In a chosen genetic level of soil the spatial distribution of elementary particles and aggregates is subject to the laws of statistics. Let us consider an appropriately large space *ABCDEFGH* (Fig. II.6) in which there is a certain distribution of elementary particles and aggregates of different sizes. From this area we can separate smaller areas –1, 2, 3, 4. The size of these are as must be, however, sufficiently greater than the sizes of the elementary particles and aggregates, so that it is possible to determine their distribution according to their sizes.

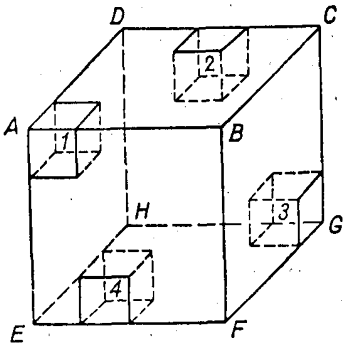

Fig. II.6. Division of soil into volume elements according to pore distribution (Walczak, 1977).

When the distribution of the elementary particles and aggregates is subject to the laws of statistics then their distribution in the sub-areas 1, 2, 3, 4 are identical as in the area *ABCDEFGH.* It is assumed that in soil this condition is fulfilled, so in this system, forming sequences of pores connected linearly, there are symmetrical sequences which, connected in a parallel way, form independent domains. For the soil water potential (pF) – moisture characteristics the theory of domains was utilized, assuming the independence of domains. This hypothesis was verified experimentally.

III. MEASUREMENT OF WATER CONTENT AND WATER POTENTIAL IN THE SOIL

1. Measurement of soil water content

The need to determine the amount of water contained in the soil arises frequently in many soil and hydrological investigations. This information is requisite for understanding the soil's chemical, mechanical, and hydrological behaviour, and the growth response of plants. There are direct and indirect methods to measure soil moisture, and several alternative ways to express it quantitatively. There is, therefore, no universally recognized standard method of measurement and no uniform way to compute and present the results. A review of the most common methods was given by Gardner (1965). Soil wetness is usually expressed as a dimensionless ratio of water mass to dry soil mass, or of water volume to total soil volume. These ratios are usually multiplied by 100 and reported as percentages by mass or by volume. We shall proceed to describe, briefly, some of the most prevalent methods for this determination.

1.1. Gravimetric method

The traditional gravimetric method of measuring the water content by mass consists of removing a sample, *eg* by augering and of determining its wet and dry weights (the latter after drying the sample to constant weight in an oven at 105°C). The gravimetric wetness (or mass wetness) is the ratio of the weight loss in drying to the dry weight of the sample (mass and weight being proportional). Occasionally, however, the mass ratio of the water to the wet soil is used. The mass (or weight) wetness on the dry soil basis (W_{md}) and the mass wetness on the wet soil basis (W_{mw}) can be converted to each other by the following equations:

$$W_{md} = \frac{W_{mw}}{1 - W_{mw}} \tag{III.1}$$

$$W_{md} = \frac{W_{md}}{1 + W_{md}} \tag{III.2}$$

To obtain the volumetric wetness θ from the gravimetric determination, a separate measurement must be made of the bulk density $_b$ and the following equation is used:

$$\theta = \rho_b \,/\, \rho_\mathrm{w}\, W_{md} \tag{III.3}$$

The measurement of bulk density, particularly in the field, is difficult and subject to errors. The gravimetric method itself, depending as it does on sampling, transporting, and repeated weighings, entails inherent errors. It is also laborious and time-consuming, since a period of at least 24 h is usually allowed for complete drying. The standard method of oven drying is also arbitrary. Some clays may still contain appreciable amounts of adsorbed water even at 105°C (Fig. III.1). On the other hand, some organic matter may oxidize and decompose at this temperature so that the weight loss may not be due entirely to the evaporation of water. The errors of the gravimetric method can be reduced by increasing the sizes and number of samples. However, the sampling method is destructive and may disturb an observation or experimental plot sufficiently to distort the results.

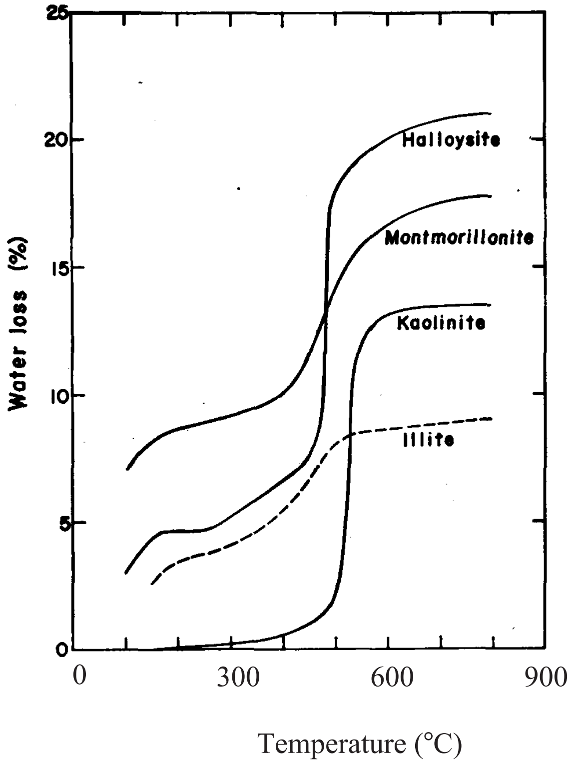

Fig. III.1. Dehydration curves of clay minerals (Marshall, 1964).

For these reasons, many workers prefer indirect methods, which permit making frequent or continuous measurements at the same points, and, once the equipment is installed and calibrated, with much less time and labour.

1.2. Electrical resistance method

The electrical resistance of a soil volume depends not only upon its water content, but also upon its composition, texture, and soluble-salt concentration. On the other hand, the electrical resistance of porous bodies placed in the soil and left to equilibrate with soil moisture can sometimes be calibrated against the water content. Such units (generally called electrical resistance blocks) generally contain a pair of electrodes embedded in gypsum (Bouyoucos and Mick, 1940), nylon, or fibreglass (Colman and Hendrix, 1949). Porous blocks embedded in the soil tend to equilibrate with the soil moisture (matric) suction, rather than with the soil-moisture content directly. Different soils can have greatly differing wetness vs. suction relationships, *eg* a sandy soil may retain less than 5% moisture at, say, 15-bar suctions, whereas a clay soil may retain three or four times as much). Hence, calibration of porous blocks against suction (tension) is basically preferable to calibration against soil wetness, particularly when the soil used for calibration is a disturbed sample differing in structure from the soil in situ. The equilibrium of porous blocks with soil moisture may be affected by hysteresis, *ie*, the direction of change. Furthermore, the hydraulic properties of the blocks (or inadequate contact with the soil) may impede the rapid attainment of equilibrium and cause a time lag between the state of water in the soil and the state of water being measured in the block. This effect, as well as the sensitivity, may not be constant over the entire range of variation in soil wetness which we may desire to measure. The electrical conductivity of most porous blocks is due primarily to the permeating fluid rather than to the solid matrix. Thus, it depends upon the electrolytic solutes present in the fluid as well as upon the volume content of the fluid. Blocks made of such inert materials as fibreglass, for instance, are highly sensitive to even small variations in salinity of the soil solution. On the other hand, blocks made of plaster of Paris (gypsum) maintain a nearly constant electrolyte concentration corresponding primarily to that of a saturated solution of calcium sulphate. This tends to mask, or buffer, the effect of small or even moderate variations in the soil solution (such as those due to fertilization or low levels of salinity). However, since gypsum is soluble, these blocks eventually deteriorate in the soil. For these and other reasons, *eg*, temperature sensitivity), the evaluation of soil wetness by means of electrical resistance blocks is likely to be of limited accuracy. Soil moisture blocks have been found to be more dependable in the drier than in the wetter range (Johnson, 1962). An advantage of these blocks is that they can be connected to a recorder to obtain a continuous record of soil-moisture changes *in situ*.

1.3. Neutron method

In recent years, this method has gained widespread acceptance for monitoring profile water content in the field. Its advantage is that it allows rapid and periodically repeated measurements in the same locations and depths of the volumetric wetness of a representative volume of soil. The instrument known as a neutron moisture meter consists of two principal parts: (a) a probe, which is lowered into an access tube inserted vertically into the soil, and which contains a source of fast neutrons and a detector of slow neutrons; (b) a sealer or ratemeter (usually battery-powered and portable) to monitor the flux of slow neutrons, which is proportional to the soil water content. The fast-neutron source may be a 2-5 millicurie mixture of radium and beryllium (which also emits hazardous y-radiation), or a mixture of americium and beryllium (with less-hazardous y-radiation).The source materials are chosen for their longevity (*eg*, radium-beryllium has a half-life of 1620 years) so that they can be used for a number of years without an appreciable change in radiation flux. The fast neutrons are emitted radially into the soil, where they encounter and collide elastically with various atomic nuclei, and gradually lose some of their kinetic energy. The average loss of energy is at a maximum when a neutron collides with a particle of a mass nearly equal to its own. Such particles are the hydrogen nuclei of water. The average number of collisions required to slow a neutron from 2 MeV to thermal energies is 18 for hydrogen, 114 for carbon, 150 for oxygen, and $9N + 6$ for nuclei of larger mass number N (Weinberg and Wigner,1958). In practice, it has been found that the attenuation of fast neutrons in the soil is proportional to the hydrogen content of the soil. The slowed (“thermal”) neutrons scatter randomly in the soil, forming a cloud around the probe. Some of these return to the probe, where they are counted by a detector of slow neutrons. The detector cell is usually filled with BF_3, gas. When a thermalized neutron encounters a ^{10}B nucleus and is absorbed, an alpha particle (the helium nucleus) is emitted, creating an electrical pulse on a charged wire. The number of pulses over a measured time interval is counted by a sealer, or indicated by a ratemeter (Gardner and Kirkham, 1952; van Bavel *et al*., 1956).

The effective volume of soil in which the water content is measured depends upon the concentration of the hydrogen nuclei, *ie*, upon the per volume wetness of the soil, as well as upon the energy of the emitted fast neutrons. With the commonly used radium-beryllium sources, the soil volume measured is in effect a sphere, which in a wet soil is perhaps 15 cm in diameter, but in a relatively dry soil may be as great as 50 cm or more (de Vries and King, 1961). This low degree of spatial resolution makes the instrument unsuitable for detection of water content discontinuities, *eg*, wetting fronts or boundaries between layers), or for measurements close to the soil surface. The relatively large volume monitored can, however, be an advantage in water balance studies, for instance, as such a volume is generally more representative of the field soil than a small sample. Methods of calibrating the neutron moisture meter were described by Holmes (1950) and Holmes and Jenkinson (1959). In most soils, it is possible to obtain a nearly linear dependence of the count rate upon the volumetric wetness of the soil. Improper use of the equipment can be hazardous. The danger from exposure to radiation depends upon the strength of the source, the distance from the source to the operator, and the duration of exposure. A protective shield is an essential component of the equipment, and it also serves as a standard absorber for checking the readings.

1.4. Gammascopic method

The gammascopic method is indirect, non-destructive method for measuring of soil water content. The gamma radiation is a highly penetrating electromagnetic form of energy, which penetrates matter. The gamma radiation attenuation method is based on absorption of monoenergetic radiation. It is adsorbed and scattered. Photoelectric absorption occurs owing to the interaction of gamma photons with electrons in the K or L shell of the absorbing atom. Absorption and scattering, the effects upon which the gamma radiation methods are based, are each dependent upon the density of the soil:

$$\rho_v = (m_w + m_s) / V_T. \tag{III.4}$$

where: m_w is the mass of water, m_s is the mass of dry soil, V_T is the total bulk volume of soil.
If the bulk density:

$$\rho_T = m_s / V_T \tag{III.5}$$

where: $_T$ is the soil bulk density,
is kept constant, the effects depend upon only the soil water content:

$$m_w / V_T = V_w / V_T = \theta \tag{III.6}$$

where: V_w is volume of water.

This method is excellent in the laboratory where the geometry is accurate, a window for collimation at the detector is available and a pulse height analyzer discards all energies below that of the original monoenergetic radiation (Nofziger and Swartzenduber, 1974).

1.5. Remote sensing method

Remote sensing occurs when the sensor is not in direct contact with the soil. Detection is performed usually from an airplane or satellite. The spectral band of electromagnetic radiation most sensitive to soil water content is utilized – wavelengths equal to and longer than those of visible radiation. The detection in visible and infrared spectra has these advantages – geometric accuracy, good resolution, easy interpretation and relatively simple procedures. When wavelength increases, those advantages disappear. The observations with visible spectrum can be made only in the daytime without cloudiness up to about 1 cm of depth. The increase of wavelength to the infrared reduces some of these disadvantages and the depth of soil layer detected is increased to a few centimetres. The roughness of the soil surface and the vegetative canopy decrease the sensitivity of this method (Baranowski, 2000; Jackson, 1993; Jackson *et al.,* 1993; Njoku and Kong, 1977; Ulaby and Elachi, 1990).

1.6. Capacitance method

Capacitance methods are based upon the measurement of dielectric constant, ε, or of an electrical parameter dependent upon ε. It is based on the values of dielectric constant. For frequencies less than 1000 MHz the dielectric constant of water is 81. Whereas for solid phase of soils it is from 4 to 8, and for air is 1. These methods belong to the nondestructive, indirect measures (Dean *et al*., 1987; Kuraż and Matousek, 1977; Malicki, 1983). The capacitor should have a cylindrical form housing the electronic circuitry with two metallic bands placed on the periphery of the cylinder at a vertical distance of 5-8 cm. The electrical field is shaped in the neighbouring soil around the electrodes to a radial distance of influence approximately 50% greater than the distance between the electrodes. Soil in the ring of influence is the dielectricum of the capacitor. The radial distance should be maximized while the depth of resolution should be minimized. Because the water in close vicinity of the soil solid particle surfaces does not have the same value of dielectric constant as for free water, calibration is required for each soil.

1.6.1. TDR method

The example of the complexity of agrophysical metrology is the reflectometric method (TDR) and monitoring equipment for porous materials moisture and electrical conductivity in situ measurement. Measuring devices produced on the licence of IA PAS (Malicki and Skierucha, 1989; 2001; Malicki *et al*., 1996) are used in many laboratories all over the world. Other companies offer similar devices and it seems that the TDR method can be the standard of moisture measurement applied not only for soil but also other porous materials like food products and building materials.

The key problem in agrophysics is water status in soil-plant-atmosphere system. The number of variables necessary to describe the water status in porous materials depends on the examined problem and is practically arbitrary. In majority of practical situations the status of water is described only by one variable – moisture. The most demanding biological material for monitoring the status of water contained in it is soil, because the soil structure is complex and its properties are not stable. Also, the response of available soil moisture sensors does not dependent on soil moisture only, it may depend on soil salinity, temperature and texture.

The TDR method for determination of porous materials (like soil, grain, wood, food products, *etc*.) moisture and electrical conductivity has many advantages over other methods like termogravimetric or neutron scattering. It provides non-invasive, fast and accurate reading of moisture and electrical conductivity (thus also salinity) from the same volume of the sample and it can be applied in automatic monitoring systems.

The TDR measurement technique takes advantage of four physical phenomena characteristic to the soil:

- in the frequency range of 1 GHz the complex dielectric constant of the soil reduces to its real value and the electromagnetic wave propagation velocity, v, in the soil can be expressed as:

$$v = \frac{c}{\sqrt{\varepsilon(\theta)}} = \frac{1}{n} = \frac{2L}{\Delta t} \tag{III.7}$$

where: c - velocity of light in free space, $\varepsilon(\theta)$ - real part of the complex dielectric constant dependent on moisture, θ - is the soil refractive index; L - length of TDR probe rods inserted into the soil; Δt - time distance between the reflections of TDR pulse from the beginning and the end of the probe rods, inserted into the soil,

- dielectric constant of the soil liquid phase has much higher value than the other soil phases, *ie* about 80 against 2-4 for the solid and 1 for the gas phase,

- relation between moisture and dielectric constant of the soil is highly correlated for most soils,

- attenuation of the amplitude of electromagnetic wave travelling along the parallel transmission line inserted into the soil depends on its electrical conductivity, σ, as:

$$\sigma[Sm^{-1}] = \frac{\sqrt{\varepsilon}}{120\pi L} \ln\left(\frac{U_{in}}{U_{out}}\right) \tag{III.8}$$

where: U_{in} and U_{out} are the amplitude of the pulse before and after attenuation caused by the pulse travel twice a distance of the probe length, L. Fig. III.2 presents the idea of simultaneous measurement soil moisture and electrical conductivity.

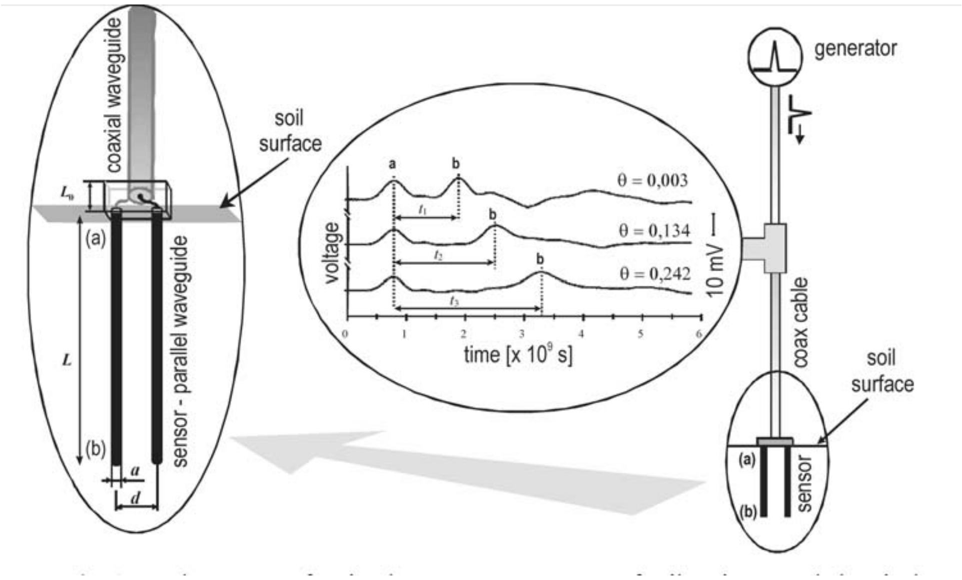

Fig. III.2. Hardware set-up simultaneous measurement of soil moisture and electrical conductivity using TDR (Malicki and Skierucha, 1989).

The TDR probe consists of two wave guides connected together: a coaxial one – the feeder, and a parallel one – the sensor, made of two parallel metal rods inserted into the measured medium. The initial needle pulse travels from the generator by the feeder towards the sensor and is registered it passes T-connector. In the connection between the feeder and the sensor, there is a rapid change in geometry of the electromagnetic wave travel path and some part of the pulse energy is reflected back to the generator, like in radar. The remaining part travels along the parallel wave guide to be reflected completely from the rods end.

The successive reflections are recorded so as to calculate the time distance between the two reflections (a) and (b). Three reflectograms (voltage as a function of time at the chosen point in the feeder) representing the cases when the sensor was placed in dry, moist and water saturated soil are presented also in Fig. III.2. The time distance, Dt, necessary for the pulse to cover the double length of metal rods in the soil increases with the soil dielectric constant, thus moisture. The reason for that is the change of electromagnetic propagation velocity in media of different dielectric constants. Also, the amplitude of the pulse at the point (b) decreases with the increase of soil electrical conductivity.

There are two approaches that allow to calculate the soil moisture from the TDR measurements: empirical and theoretical - using dielectric mixing models. The coefficients of linear regression equations fitted to empirical data points differ slightly showing that mineral composition of the soil significantly influences the TDR calibration curve where: *qgrav* is the soil moisture determined by standard thermogravimetric method ISO 11461. Reconfiguration of the calibration curve allows the calculation of q when n is measured by TDR meter. The accuracy of the TDR soil moisture measurement is about ±2% of the true thermogravimetrically determined value:

$$\theta = 0.13\sqrt{\varepsilon - 0.18} \tag{III.9}$$

The most popular calibration curve for TDR soil (mineral soils) moisture measurements was given by Topp *et al.* (1980) and is presented below:

$$\theta = -5.3\ 10^{-2} + 2.92\ 10^{-2}\varepsilon - 5.5\ 10^{-4}\varepsilon^{2} + 4.3\ 10^{-6}\varepsilon^{3} \tag{III.10}$$

The introduction in the equation the correction on soil bulk density, , that accounts for the influence of soil solid phase in the TDR readout decreases the measurement error to about ±1% giving a new calibration curve presented by:

$$\theta = \frac{\sqrt{\varepsilon}\ 0.57 - 0.58}{} \tag{III.11}$$

Applying theoretical approach the soil is usually treated as a three phase mixture, and its dielectric constant of soil, ε, can be presented as:

$$\varepsilon^{\alpha} = \theta\,\varepsilon_{w}^{\varepsilon} + (1+\phi)\varepsilon_{s}^{\alpha} + (\phi - \theta)\varepsilon_{\alpha}^{\alpha} \tag{III.12}$$

where, ε_w, ε_s and ε_α are the dielectric constants of soil as a whole, soil water, soil solids, and air, α is a constant interpreted as a measure of the soil particles geometry. On the base of the measured data collected on various soils, it has been found that for the three phases dielectric model of the soil the average value of a constant is $\phi = 0.5$.

Therefore there has been much effort put down on the soil moisture metrology and the Institute of Agrophysics PAS is a pioneer in the field TDR soil moisture measurement technique (Fig. III.3). If this is a reliable method of soil moisture measurement, it could be applicable to other biological materials with less differentiated structure.

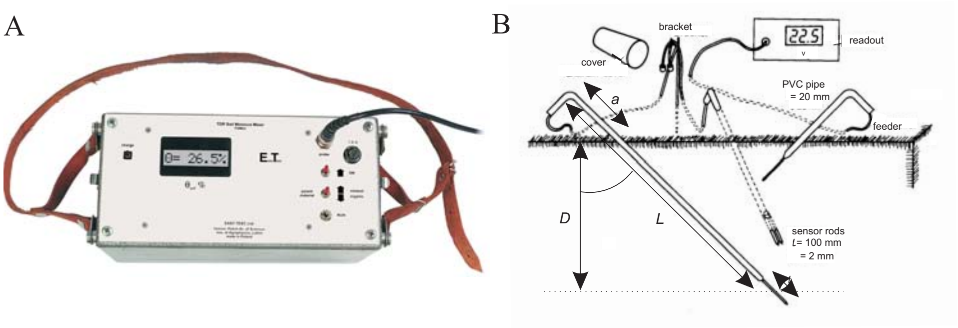

Fig. III.3. Portable field TDR soil moisture, salinity and temperature meter produced on licence of IA PAS, Lublin, Poland (Instruction, 2002); (A) and probes set-up to monitor moisture changes in soil profile (B).

The TDR method of simultaneous measurement of soil moisture and electrical conductivity (salinity) becomes more popular because of its advantages: simplicity of operation, accurate and rapid readout, usually does not need calibration, non-destructive operation, availability of portable systems, and the possibility to make automatic and to multiplex probes. Although it has also disadvantages because it requires excellent probe to soil contact, probe installation may disturb soil, and TDR meters are still expensive, it may become the standard of moisture measurement applied not only for soil but also other porous materials like food products and building materials.

2. Measurement of soil water potential

The measurement of soil water content, or wetness, though essential in many soil physical and engineering investigations, is not sufficient to provide a description of the state of soil water. To obtain such a description, the evaluation of the energy status of soil water (soil moisture potential, or suction) is necessary. In general, the two properties, wetness and potential, should each be measured directly, as the translation of one to the other on the basis of calibration curves of soil samples has been found unreliable (Stolzy *et al*., 1959). Total soil moisture potential is often thought of as the sum of matric and osmotic (solute) potentials, and is a useful index for characterizing the energy status of soil water with respect to plant water uptake. The sum of the matric and gravitational (elevation) heads is generally called the hydraulic head (or hydraulic potential), and is useful in evaluating the direction and intensity of the water-moving forces in the soil profile. Methods are available for measuring matric potential as well as total soil potential, separately or together. To measure matric potential in the field, an instrument known as the tensiometer is used. To measure soil moisture potential, freezing-point-depression measurements (Richards and Campbell, 1949), and vapour-pressure measurements of soil water by means of thermocouple psychrometers (Richards and Ogata, 1958; Rawlins, 1966; Rawlins and Dalton, 1967; Dalton and Rawlins, 1968) have been used. At equilibrium, the potential of soil water is equal to the potential of the water vapour in the ambient atmosphere. If thermal equilibrium is assured, and the gravitational effect is neglected, the vapour potential is equal to the sum of the matric and osmotic potentials, since air acts as an ideal semi-permeable membrane in allowing only water molecules to pass (provided the solutes are non volatile). At room temperature, relative humidity of the air is related to the potential (pF) by (Bolt and Frissel, 1960):

$$pF = 6.5 + \log (2 - \log R.H.) \tag{III.13}$$

where: pF = log (osmotic potential + matric potential) when these potentials are expressed as cm of water head; and *R.H.* = relative humidity. We shall now describe the tensiometer, which has won widespread acceptance as a practical device for the in situ measurement of matric suction, hydraulic head, and hydraulic gradients.

2.1. Tensiometers

The essential parts of a tensiometer are shown in Fig. III.4. The tensiometer consists of a porous cup, generally of ceramic material, connected through a tube to a manometer, with all parts filled with water. When the cup is placed in the soil where the suction measurement is to be made, the bulk water inside the cup comes into hydraulic contact and tends to equilibrate with soil water through the pores in the ceramic walls. When initially placed in the soil, the water contained in the tensiometer is generally at atmospheric pressure.

Soil water, generally at atmospheric being generally at subatmospheric pressure, exercises a suction which draws out a certain amount of water from the rigid and airtight tensiometer, thus causing a drop in its hydrostatic pressure. This pressure is indicated by a manometer, which may be a simple water- or mercury-filled U-tube, a vacuum gauge, or an electrical transducer. A tensiometer left in the soil for a long period of time tends to follow the changes in the matric suction of soil water. As soil moisture is depleted by drainage or plant uptake, or as it is replenished by rainfall or irrigation, corresponding readings on the tensiometer gauge occur. Owing to the

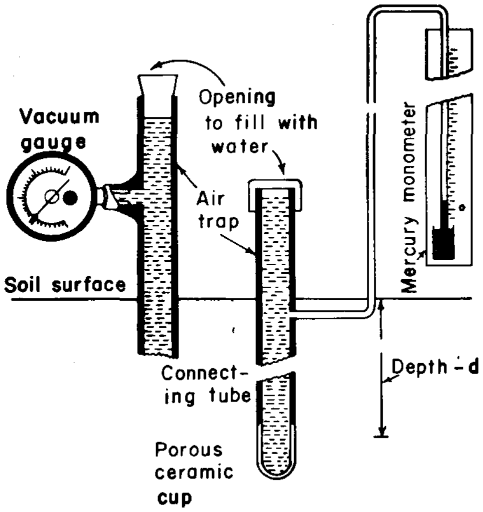

Fig. III.4. Schematic illustration of the essential parts of a tensiometer (Richards, 1965).

hydraulic resistance of the cup and the surrounding soil, or of the contact zone between the cup and the soil, the tensiometer response may lag behind suction changes in the soil. This lag time can be minimized by the use of a null-type device (Miller, 1951; Leonard and Low, 1962) or of a transducer type manometer with rigid tubing, so that practically no flow of water takes place as the tensiometer adjusts to changes in the soil matric suction. Since the porous cup walls of the tensiometer are permeable to both water and solutes, the water inside the tensiometer assumes the same solute composition and concentration as soil water, and the instrument does not indicate the osmotic suction of soil water (unless equipped with some type of an auxiliary salt sensor).

Suction measurements by tensiometry are generally limited to matric suction values of below 1 atm. This is due to the fact that the vacuum gauge or manometer measures a partial vacuum relative to the external atmospheric pressure, as well as to the general failure of water columns in macroscopic systems to withstand tensions exceeding 1 atm. Furthermore, as the ceramic material is generally made of the most permeable and porous material possible, too high a suction may cause air entry into the cup, which would equalize the internal pressure to the atmospheric. Under such conditions, soil suction might continue to increase even while the tensiometer fails to show it.

In practice, the useful limit of most tensiometers is at about 0.8 bar of maximum suction. To measure higher suctions, the use of an osmometer with a semi-permeable membrane at the wall has been proposed (Peck and Rabbidge, 1969), but the practical application of this instrument is still in the experimental stage. The limited range of suction measurable by the tensiometer is not as serious as it may seem at first sight. Though the suction range of 0-0.8 bar is but a small part of the total range of suction variation encountered in the field, it generally encompasses the greater part of the soil wetness range. Richards and Marsh (1961) have shown that in many agricultural soils the tensiometer range accounts for more than 50% (and in coarse-textured soils 75% or more) of the amount of soil water taken up by plants. Thus, where soil management (particularly in irrigation) is aimed at maintaining low-suction conditions which are most favourable for plant growth, tensiometers are definitely applicable.

Despite their fundamental and practical shortcomings, tensiometers are practical instruments, available commercially, and, when operated and maintained by a skilled worker, are capable of providing reliable data on the in situ state of soil-moisture profiles and their changes with time. Tensiometers have been found useful in guiding the timing of irrigation of field and orchard crops, as well as of potted plants (Richards and Marsh, 1961). A general practice is to place a tensiometer at one or more soil depths representing the root zone, and to irrigate when the tensiometer indicates that the matric suction has reached some prescribed value. The use of several tensiometers at different depths can indicate the amount of water needed in irrigation, and can also allow calculation of the hydraulic gradients in the soil profile (L.A. Richards, 1955). If Ψ_1, Ψ_2, Ψ_3,......Ψ_{n} are the matric suction values in centimeters of water head (in millibars) at depths d_1, d_2, d_3,.....d_n measured in centimetres below the surface, the average hydraulic gradient i between depths dn and d_{n+1} is:

$$i = [(\Psi_{n+1} + d_{n+1}) - (\Psi_n + d_n)] / (d_{n+1} - d_n) \qquad \text{(III.14)}$$

Measurement of the hydraulic gradient is particularly important in the region below the root zone, where the direction and magnitude of water movement cannot easily be ascertained otherwise.

3. Measurement of the soil moisture characteristics curves

The functional relation between soil moisture and matric potential is often determined by means of a tension plate assembly (Fig. III.5) in the low suction (< 1 bar) range, and by means of a pressure plate or pressure membrane apparatus (Fig. III.6) in the higher suction range.

These instruments allow the application of successive suction values and the repeated measurement of the equilibrium soil moisture at each potential. The maximum suction value obtainable by porous plate devices is limited to 1 bar if the soil air is kept at atmospheric pressure and the pressure difference across the plate is controlled either by vacuum or by a hanging water column. Matric potential values considerably greater than 1 bar (20 bars or even more) can be obtained by increasing

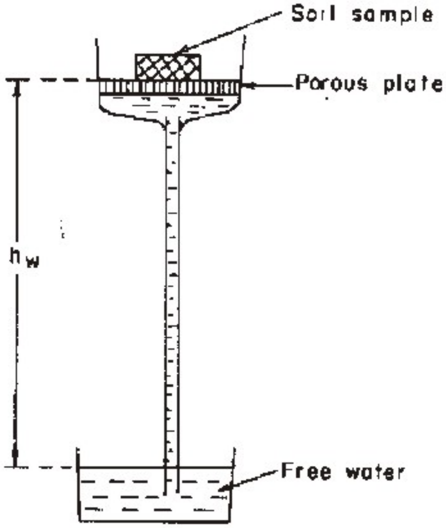

Fig. III.5. Tension plate assembly for equilibrating a soil sample with a known matric suction value. This assembly is applicable in the range of 0-1 bar only (Hillel, 1971).

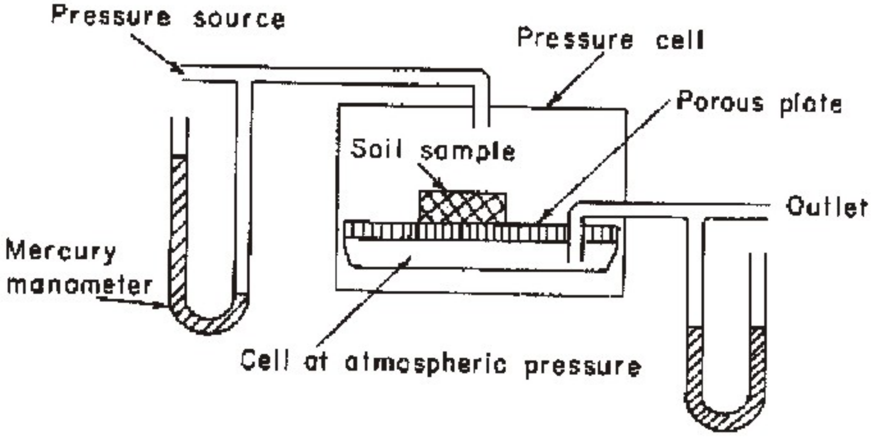

Fig. III.6. Pressure plate apparatus for moisture-characteristic measurements in the high-suction range. The lower side of the porous plate is in contact with water at atmospheric pressure. Air pressure is used to extract water from initially saturated soil samples (Hillel, 1971).

the pressure of the air phase. This requires placing the porous plate assembly inside a pressure chamber, as shown in Fig.III.6. The limit of matric potential obtainable with such a device is determined by the design of the chamber (*ie* its safe working pressure) and by the maximal air-pressure difference the saturated porous plate can bear without allowing air to bubble through its pores. Ceramic plates generally do not hold pressures greater than about 20 bars, but cellulose acetate membranes can be used with pressures exceeding 100 bars. The most popular apparatus set for measurements relation between soil water potential and water content, produced by SOILMOISTURE EQUIPMENT Company, Santa Barbara, California, USA, is shown in Fig. III.7.

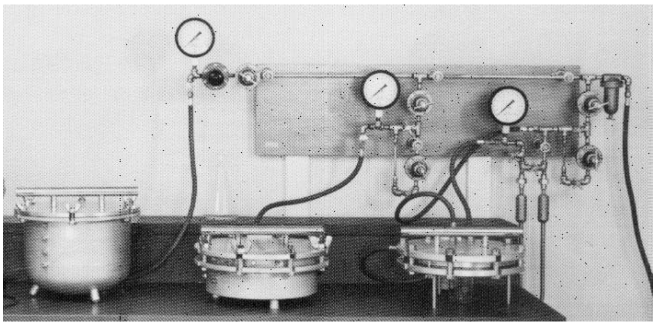

Fig. III.7. Laboratory set-up for measurements water retention curves produced by SOILMOISTURE EQUIPMENT COMP., Santa Barbara, CA, USA (Instruction, 1998).

Soil moisture retention in the low-suction range (0-1 bar) is strongly influenced by soil structure and pore size distribution. Hence, measurements made with disturbed samples, *eg* dried, screened, and artificially packed samples) cannot be expected to represent field conditions. The use of undisturbed soil cores is therefore preferable. On the other hand, soil moisture retention in the high-suction range is due primarily to adsorption and is thus correlated with the specific surface of the soil material rather than with its structure. As mentioned earlier, the soil-moisture characteristic is hysteretic. Ordinarily, the desorption curve is measured by gradually and monotonically decreasing the water content of initially saturated samples. The resulting curve, often called the soil-moisture-release curve, is applicable to processes involving drainage, evaporation, or plant extraction of soil moisture. On the other hand, the sorption curve is needed whenever infiltration or wetting processes are studied.

Two types of information are generally needed in the study of soil-water phenomena: the quantity of water contained in the soil and the energy status of soil water, *ie* soil wetness and soil moisture suction. While these can be measured independently, they are functionally related. This relationship, which is affected by hysteresis, is described by the “soil-moisture characteristic curve,” variously termed the retention, release, sorption, or desorption curve. By whatever name, it is of fundamental importance in soil physics, as it expresses the influence of structure, porosity, pore-size distribution, and adsorption on the state of soil water. This state and how it varies in the profile, in turn, determine the direction and influence the rate of soil-moisture movement and uptake by plants.

IV. SOIL WATER POTENTIAL – WATER CONTENT CHARACTERISTICS FOR DIFFERENT COMPACTION AND AGGREGATION

The application of agricultural machines loosening and compacting soil is a purposeful agrotechnical measure, aimed at the obtaining of a soil compaction optimal for the growth and development of cultivated plants. All agrotechnical measures, not always connected with the mechanical processing of soil, like for instance fertilizing, plant protection, weeding, agricultural crop transportation, are accompanied by repeated kneading of soil by the wheels and track chains of vehicles. The effects of soil kneading are considerable accompanied by repeated kneading of soil by the wheels and because of the large number of vehicles. The kneading of soil by in artificial capillary-porous bodies was carried out. For the interpretation of results the domain theory of hysteresis was applied. On the wheels and track chains of vehicles this is not purposeful and not profitable, hence the many works concerned with the investigation of the extent of this phenomenon and with seeking ways to eliminate it.

Many authors have concerned themselves with investigating the extent of soil deformation under the influence of applied pressures and the mechanism of this phenomenon. Investigations of soil have showed that the volumetric deformation of soil is the result of the compressing of gas and the filtration of gas and water. Haman (1955) investigated the process of volumetric deformation of soil in relation to the speed of deformation at different levels of soil moisture. He established that the speed of ploughing at a constant pressure does not influence the work of soil skeleton deformation and gas compression, it does, however, influence the work necessary to cause water filtration. This work is inversely proportional to speed. Direct field investigations of the influence of the speed of tool movement, pressure volumes and the initial state of soil on the extent of deformations were carried out. The investigations showed that the extent of soil deformation decreases with the increase of the speed of the vehicle and tool movement in all ranges of towing power. The extents of the volumetric deformations of soil decrease with the increase of its initial compaction, but increase with the increase of its moisture. Hence it follows that the mechanical processing of soil should be carried out at an adequately low level of its moisture. The soil compaction causes changes in the volume, dimensions and shapes of soil pores, and thus it influences the water and air properties of soil. In agrophysics many works are concerned with investigations of the soil water properties and their connections with, among other things, the specific surface, mechanical composition, and the content of organic substance. Investigations of the influence of compaction on the water characteristics of soil are not numerous.

Some works, through investigating the influence of compaction on the growth and development of plants, are concerned with the determination, indirectly, of the changes in air - water relations in soil (Dołgow and Modina, 1969; Iwanow and

Korobova, 1969; Słowik, 1970). This method has an obvious advantage, since it connects the compaction of soil with the final effect that is interesting from the agricultural point of view, that is with the productivity of soil. The authors of these papers established that the compaction of soil has a considerable influence on the growth and development of plants and that it depends on the kind of soil and plant. Also direct investigations of the water properties of soil of different compaction were carried out. In laboratory conditions the changes of water characteristics, pore volume and their distribution according to size were investigated (Lebron et al., 2002; Oniszczenko and Miczurin, 1971; Rovdan et al., 2002; Sokołowska, 1967; Walczak et al., 2001, 2002a, Witkowska-Walczak, 2003a).

In the investigations of the water properties and their connection with compaction the effect of hysteresis and the influence of compaction of the extent of this effect were neglected. Works concerned with the effect of hysteresis base of which the distribution of pores was calculated, with the irregularities of their shapes taken into consideration. The investigations showed that the effect of hysteresis should be considered in order to obtain an explicit determination of the moisture-sucking force relation in capillary-porous bodies. Poulovassilis (1962) carried out investigations on the process of drying and moistening the capillary-porous medium in the range of the sucking force values up to 30 cm of water column. He found that the domain theory describes the process investigated by him with reasonable accuracy. Topp and Miller (1966) carried out a similar experiment and obtained negative results, because they used material of a narrow range of pore sizes. This experiment is discussed by Poulovassilis (1969) who stated that the domain theory, being statistical, can describe the process well only when the investigated body has a range of pore sizes which is as wide as possible. Childs and Poulovassilis (1971) carried out investigations of the sucking force-moisture effect of hysteresis for pure sand. Experimental verification of the domain theory confirmed its applicability in describing the process in question.

1. Influence of soil compaction on the course of the soil water potential - water content (moisture) characteristics

For the Mollic Gleysol of different compaction soil water potential (pF)-moisture characteristics were determined during the process of drying. The water retention curves are presented in two versions since the characteristics have different courses in both cases. Their different courses are caused by the fact that in the case of expressing moisture in kg kg^{-1}(%, mass) the amount *dt* water retained by a unit of soil mass at a given sucking force is determined. The amount of water in this case depends only on the changes in the volume of pores and their distribution. In the case of expressing moisture in m^{3} m^{-3} (%,vol.) the change in the amount of soil in a unit volume depending on its compaction is also considered. The water potential-moisture

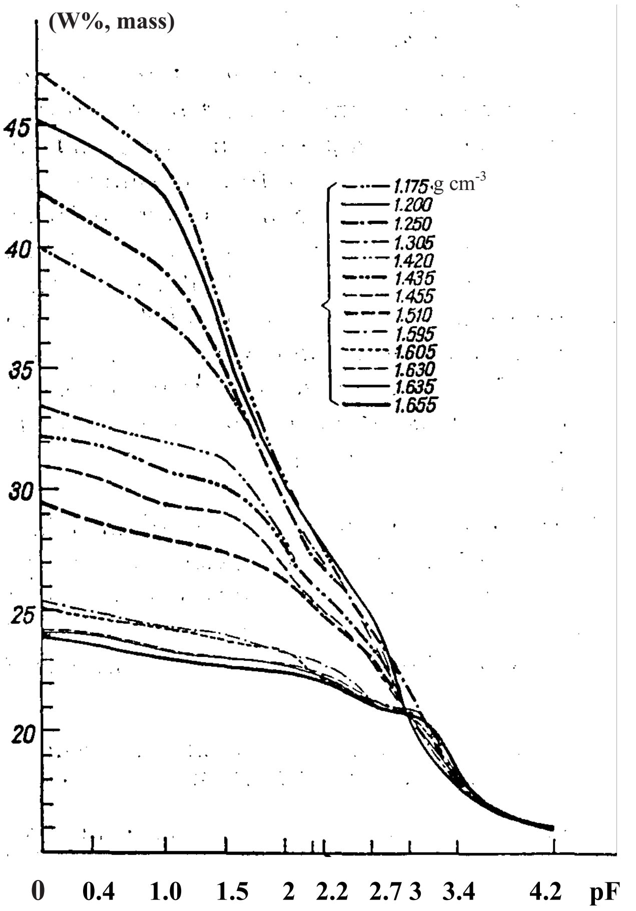

Fig. IV.1. Soil water potential (pF) – moisture (W%, mass) characteristics for soil of different compaction (Walczak, 1977).

characteristics expressed in %,mass are presented in Fig. IV.1. The courses of the curves show the changes in the amount of water bound with different forces in a unit of soil mass. The courses of the characteristics of both soils are clearly determined by their compaction. The changes of moisture (W%,vol.)with compaction are shown in Fig. IV.2.

The amount of water bound with low forces in a unit volume of soil mass diminishes with the increase of compaction. The above relationships prove the considerable influence of the increase of compaction on the quantity of pores of large diameters. The courses of the characteristics indicate that compaction occurs first of all as a result of the diminishing of the volume of pores of diameters bigger than

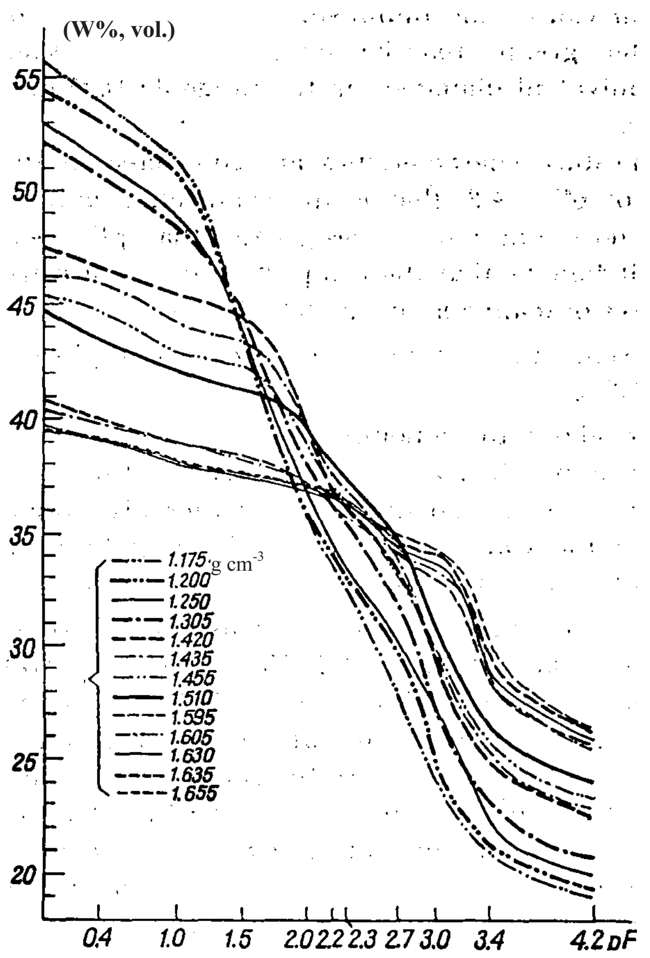

Fig. IV.2. Water retention curves for soil of different compaction (W%,vol.) (Walczak, 1977).

a certain value d_m, and the d_m value of pore diameters decreases with the increase of compaction. Since the compaction is caused by the pressures affecting soil it is possible to state that the increase of the pressures causes irreversible volumetric deformations through destroying pores of decreasing diameters that is that the mechanical resistance of the system of soil particles forming small pores is greater. Compaction causes the diminishing of the volume of pores of equivalent diameters greater than 3 10^{-6} m and the increasing of the volume of pores of equivalent diameters in the range 3 10^{-6}m - 0.2 10^{-6}m in a unit of mass.

The pF-moisture characteristics are connected in the range of error at the value of pF 4.2, that is the compaction does not influence the amount of water bound with forces greater than pF 4.2 in a unit of soil mass. Hence it follows that the compaction does not cause changes of the volume of pores of diameters below 0.2 10^{-6}m.

Knowledge about the amount of water bound with different forces contained in a unit soil mass is particularly useful from the agricultural point of view, since the possibilities of water absorption by a plant are determined by the water quantity and binding force in the volume of soil embraced by the root system. The water content in a unit soil mass at different sucking forces depends on the compaction of soil. The amount of water contained in a unit volume, bound with low forces, diminishes with the increase of compaction, while the amount of water bound with high forces increases with the increase of compaction. In the range pF 1.5 to pF 2.7, which is equivalent to diameters 100 10^{-6} to 6 10^{-6} m, the characteristics cross, that is the amount of water, contained in a unit volume of soil, bound with forces of values within the enumerated ranges, depends on the compaction of soil only to a slight degree. The increase in the amount of water bound with higher forces, like for instance pF 4.2, is caused by the quantity of dry mass of soil contained in a unit volume, increasing with the increase of the compaction of soil. It can be foreseen that further increase of sucking forces would cause the diminishing of the water content in soil. At high values of sucking forces the water content will tend to 0, and thus characteristics will tend asymptotically to the axis of abscissae.

Fig. IV.3. presents the relationship between the content of water in a unit mass of soil, bound with different forces, and the compaction of soil. The curve for pF 0 presents the relationship between the compaction of soil and its maximum water capacity. The curve for pF 0.4 presents the relationship between the compaction of soil and its capillary water capacity. The maximum and capillary capacities of soil decrease with the increase of the compaction of soil.

The influence of compaction on the amount of water bound with different forces decreases with the increase of the values of the forces. different compaction The amount of water contained in a unit of mass, bound with forces higher than pF 3 depends on compaction only to a slight degree.

Having the initial data concerning the compaction of soil and its moisture or sucking force, we can, on the basis of the diagram, determine the value of compaction after reaching, which there must occur filtration of the water in the process of compaction.

In Fig. IV.3 an example of the determination of this value compaction is presented. Adopting as the initial data a soil compaction of = 1.250 g cm^{-3} and a moisture of W = 36.3%, mass, a straight line parallel to the axis of the abscise for this value of moisture should be drawn. This straight line crosses the curve equivalent to pF 0 at the compaction value = 1.365 g cm^{-3}. At this value of compaction the

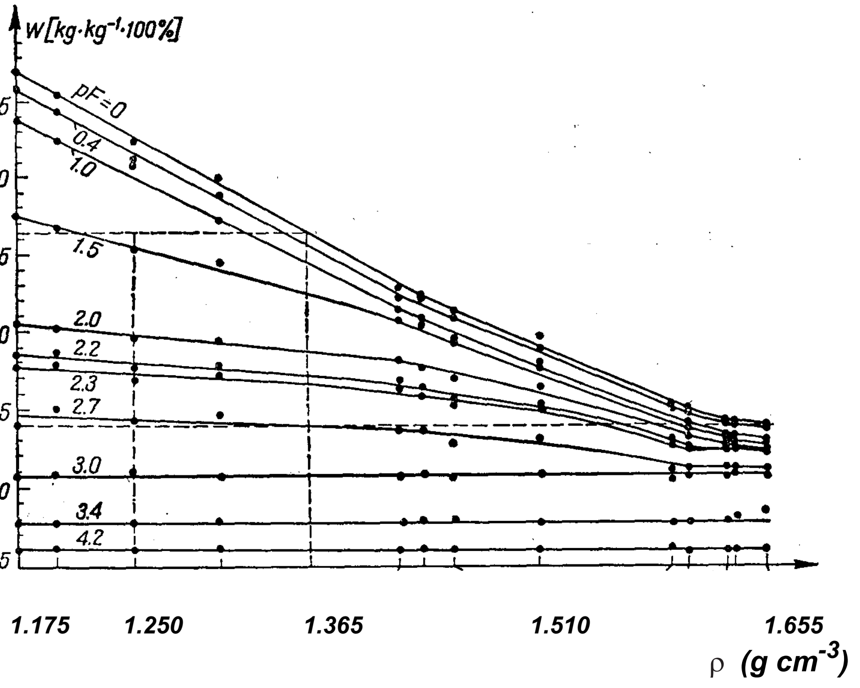

Fig. IV.3. Amount of water of different bonding energy in the unit soil mass of different compaction (Walczak, 1977).

maximum water capacity is 36.3%, mass, so further compaction must be accompanied by infiltration of water. During compaction in the range = 1.250 g cm^{-3} to = 1.365 g cm^{-3}, there occurs filtration of gas, and also filtration of water resulting from the compression of gas can occur. From Fig. IV.3 it is possible to determine the characteristic moisture, and in this case it is 24%, mass. If the moisture of the investigated soil is lower than that quoted above compaction to the value $_{max}$ does not have to be accompanied by filtration of water. There is 'a high probability of the occurrence of such a process, mainly at low compactions, since the values of moisture are so low that only gas can filtrate.

Fig. IV.4 presents the relationship between the amount of water bound with different forces, contained in a unit volume of soil, and the compaction of soil.

The amount of water bound with forces equivalent to the values pF 0, 0.4, 1.0, 1.5 decreases with the increase of compaction. From the diagram it follows that for compactions in the range = 1.420-1.510 g cm^{-3} the maximum amounts of water provided through overhead irrigation can be retained, since the values of the field water capacities are in this range the greatest. The amount of water bound with higher forces increases with the increase of the compaction of soil.

On the basis of Fig. IV.4 it is possible to determine the parts of the soil volume that are taken up by the solid, liquid, and gas phases. The part of the volume, expressed in percentages, that is taken up by- the solid phase is calculated for a chosen compaction by deducting the moisture in percentages from 100% (the moisture is equivalent to the maximum water capacity - pF 0). At a partial saturation with water the part of the volume, that is taken up by the gas phase is the difference the maximum water capacity (pF 0) for a chosen compaction and the part of the volume that is taken up by the liquid phase, that is the moisture. Thus diagrams enable the determination of the air-water relations for any compaction of soil at different values of sucking forces.

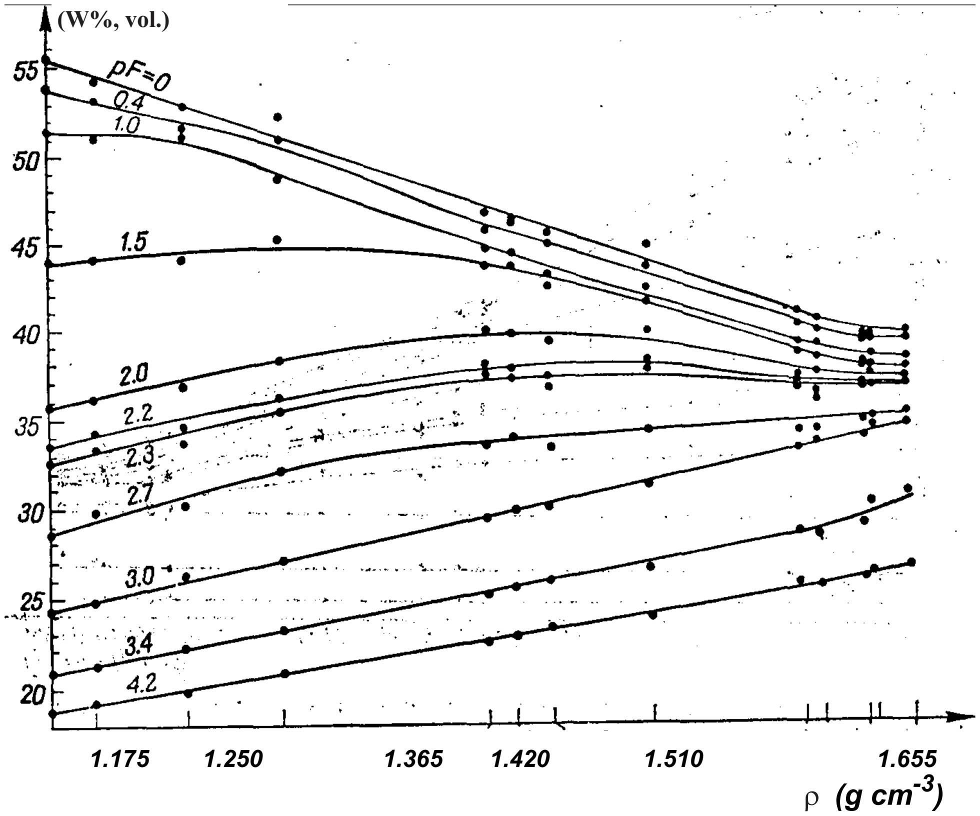

Fig. IV.4. Amount of water of different bonding energy in the volume of soil with different compaction (Walczak, 1977).

If it is known that in the soil a certain sucking force predominates for the longest periods of time, then it is possible to determine the optimum compaction for the most-profitable air-water conditions.

From the agricultural point of view the most interesting thing is the information concerning the amount of water contained in soil that is accessible to plants. The relationship between the amount of water accessible to plants at different pF values and the compaction of soil is presented in Fig. IV.5.

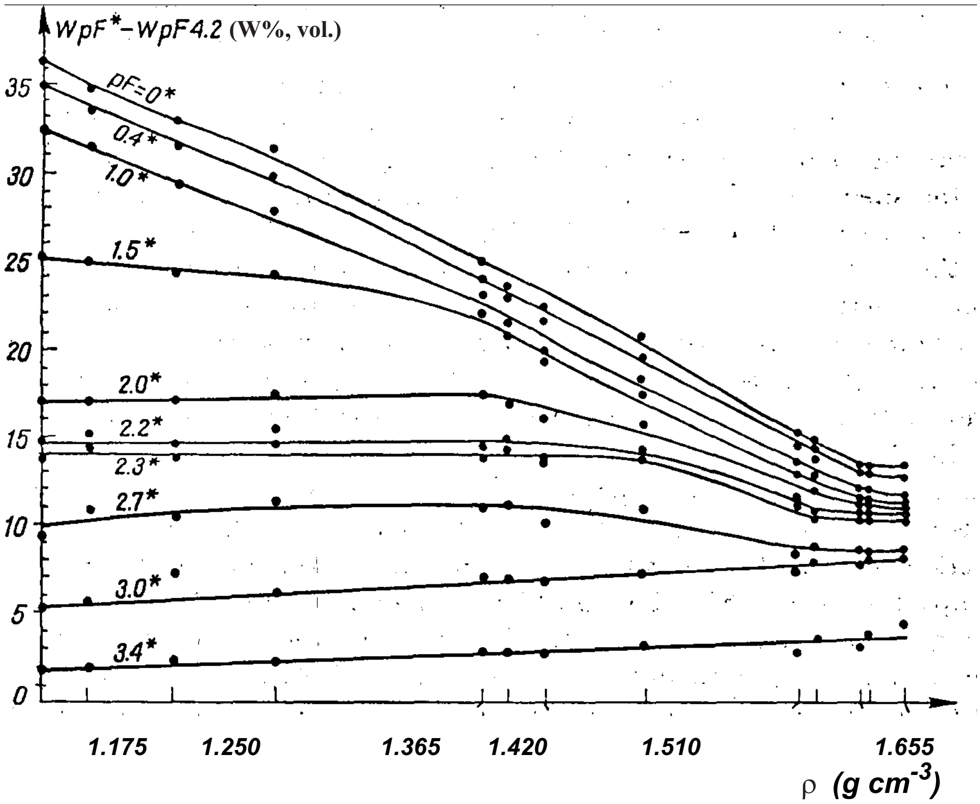

Fig. IV.5. Amount of water available for plants versus soil compaction for different pF values, (WpF* moisture corresponding to the value pF – 0-3.4, WpF 4.2 moisture corresponding to the value pF – 0-4.2) (Walczak, 1977).

The curves were obtained by deducting the value of moisture at pF 4.2 from the value of moisture for the particular pF values. The moisture at values pF 0, 0.4, 1.0, 1.5 decreases with the increase of compaction. The amount of water contained in a unit of volume of soil at values pF 2.0, 2.2, 2.3 and 2.7 does not depend on compaction to a certain its value, that is, in the case of the investigated soil, it decreases from ρ = 1.510 g cm^{-3}. The amount of water that is accessible to plants increases with difficulty with the increase of the compaction of soil. From the diagrams it follows that the exceeding of the above-quoted values of the compaction of soil can particularly negatively influence the growth and development of plants, since it causes a decrease in the field water capacity with a simultaneous decrease in the total porosity. The worsening of the air-water conditions of soils together with the increase of compaction in caused also by the increase of the proportion of water accessible only with difficulty to plants from the total amount of water in soil.

1.1. Investigations of the effect of the soil water potential-moisture hysteresis and calculation of the distribution of pores in soil

For a soil of different compactions investigations of the affect of hysteresis in the range pF 0 to pF 2.7 have been carried out according to the earlier quoted methodology. According to the demands of the theory the limit curves and the first order curves of moistening were determined experimentally (Fig. IV.6). The investigations indicated that the soil water content (pF)-moisture relationship is a non-explicit function. The differences of moisture for the same pF value depending on the way in which a given state was achieved (drying or moistening) can be as much as 10% of the value of moisture, which can lead, if this effect is not considered, to relative errors of up to 20%. In Fig. IV.7 the loops of hysteresis for the maximum, medium, and minimum compactions are presented as an example. Let us consider a case when there occurs the contact of two layers of soil, one of low compaction ρ = 1.175 g cm^{-3} and low moisture, and another of high compaction ρ = 1.655 g cm^{-3} and high moisture. During the stabilizing of the state of balance, the layer of low compaction will be dried, while that of high compaction will be moistened. After achieving a state of balance there will occur a leveling of the sucking forces of the two layers. If for example the state of balance is achieved for pF 1, then we can find out from the diagram that, the layer of the compaction ρ = 1.175 g cm^{-3} has a moisture W=51% (approximately) and the layer of compaction ρ = 1.655 g cm^{-3} has a moisture W=39% (approximately). In the opposite case, that is when a layer of low compaction ρ = 1.175 g cm^{-3} is moistened and that of high compaction ρ = 1.655 g cm^{-3} is dried, a state of balance for pF 1 is achieved when the compacted layer has a moisture of 39% and the loose layer a moisture of 42%. Hence it follows that distribution of moisture in a soil profile is considerably influenced by the differences in the compactions of the layers and by the effect of hysteresis.

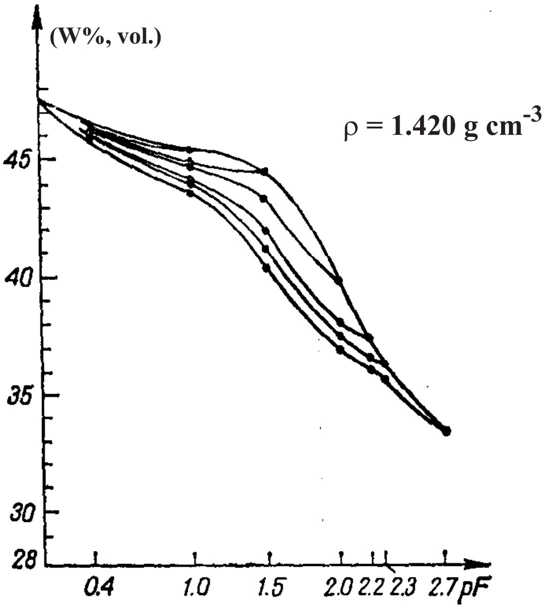

Fig. IV.6. Example of hysteresis loop and primary wetting curves (Walczak, 1977).

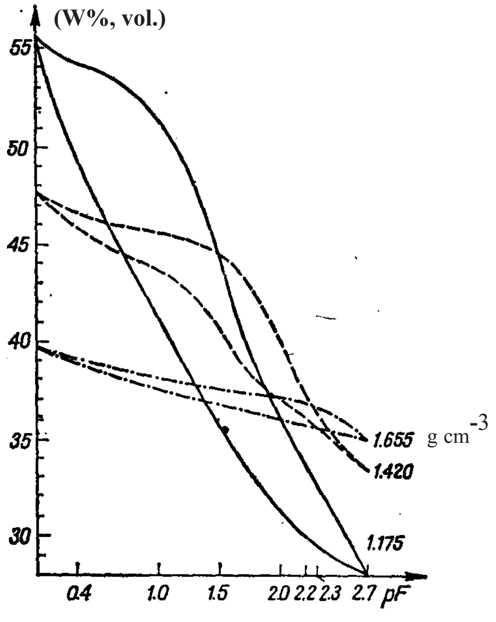

Fig. IV.7. Loops of pF-moisture hysteresis for maximum, middle and minimal soil compaction (Walczak, 1977).

The extent of the effect of hysteresis is indicated by the sizes of the loops of hysteresis. The relationship between the area of a loop of hysteresis limited by the limit curves and the limit curve of drying and the subsequent first order interial curves of moistening is presented in Fig. IV.8. From the diagram it follows that the area of a loop of hysteresis diminishes with the increase of compaction. The curves of the areas of the hysteresis-compaction loops have shapes similar to a parabola up to the value ρ = 1.305 g cm^{-3} and beyond this value of compaction this relationship is linear. This effect can probably be explained by the destruction of the inter-aggregate pores at a compaction of soil of the above-quoted value, while at a further compaction only the destruction of the inside-aggregate pores occurs.

A computer program (Kaniewska and Walczak, 1974) which enables the calculation of the diagram of distribution (Table IV.1) is analogous to the diagram of the function of distribution presented in Fig. IV.8. On the axis of abscissae the sucking forces in the process of drying are marked, and on the axis of ordinates the sucking forces in the process of moistening. The values of the figures in the diagram are equivalent to the volume of water which comes out of the soil during drying or enters the soil during moistening at certain values of sucking forces (S_e of drying and S_f of wetting).

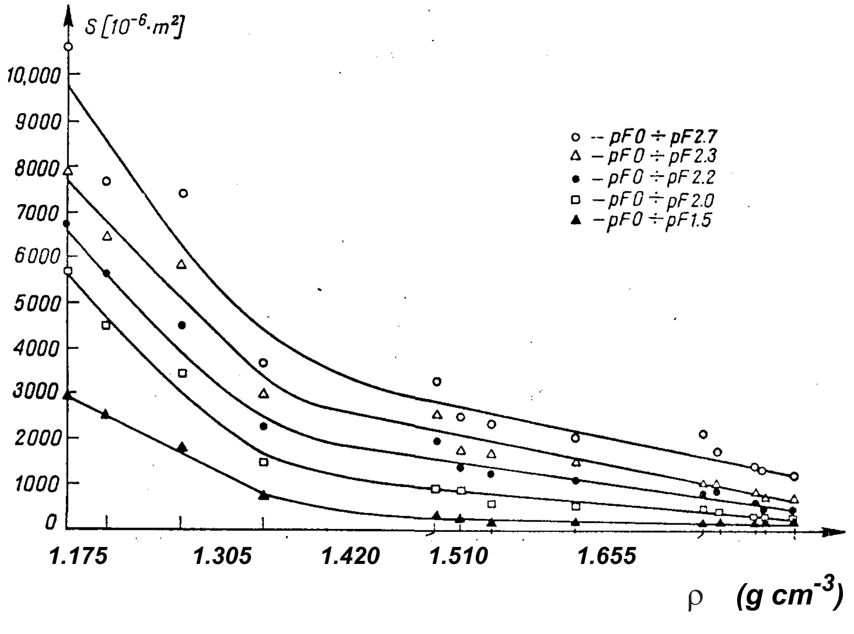

Fig. IV.8. The influence of soil compaction on the area of hysteresis loop (Walczak, 1977).

The diagram allows us to determine the amount of water in soil after any changes of sucking force. The using of the diagram in order to determine explicitly the pF-moisture relationship can be shown on the example of Table IV.1.

Table IV.1. Example of domain distribution function (pore size distribution function) (Walczak, 1977)

CE ρ= 1.175 dmax	pF wetting							
6	2.7							1.41
15	2.3						.13	.26
18.5	2.2					.70	.00	.59
30	2.0				3.05	.93	.00	.92
10.0	1.5			2.52	3.15	.00	.00	.00
30.0	1.0		1.81	3.54	.95	.53	.57	.14
120.0	0.4	1.62	.75	1.26	.93	.22	.25	1.14
	0	0.4	1.0	1.5	2.0	2.2	2.3	2.7 pF drying
		1200	300	100	30	18.5	15	6 d min

Let us assume that in the initial state the soil will be fully saturated with water. The process of drying can be expressed by the movement of a straight line perpendicular to the axis of abscissae in the direction of the increasing pF values, for example to pF 2.2. The volume of water which came out of soil in this process is equivalent to the sum of volumes expressed in rectangles, covered by the movement of the perpendicular. Let us then assume that the sucking force decreases to the value pF 1. In such a case the process of moistening at a change of the sucking force from pF 2.2 to pF 1 can be expressed by the movement of a straight line perpendicular to the axis of ordinates in the range pF 2.2 to pF 1. Then the amount of water which entered the sample is equivalent to the sum of volumes expressed in rectangles covered by the movement of this perpendicular. After two such processes the volume of water which left the sample is equal to the sum of volumes expressed in rectangles contoured by the line.

To the sucking forces the following pore diameters can be ascribed: to the value S_e – the minimum pore diameter, to the value S_f – the maximum pore diameter. The minimum pore diameter determines the drying, the maximum one determines the moistening. Thus the diagram gives the distribution of pores, considering also the irregularity of their shapes. Diagrams are difficult to compare because of their containing a great number of figures precisely characterizing the distribution of pores considering the irregularity of their shapes. The comparison of diagrams can be made through the synthesizing of the results, but this denies us the advantage of the distribution of pores with the help of diagrams. In this work no exact analysis of the changes in the distribution of pores as the result of the changes of compaction was performed, since this was not its objective. As an example only the changes of the content of some chosen groups of pores caused by the changes of compaction were determined (Fig. IV.9).

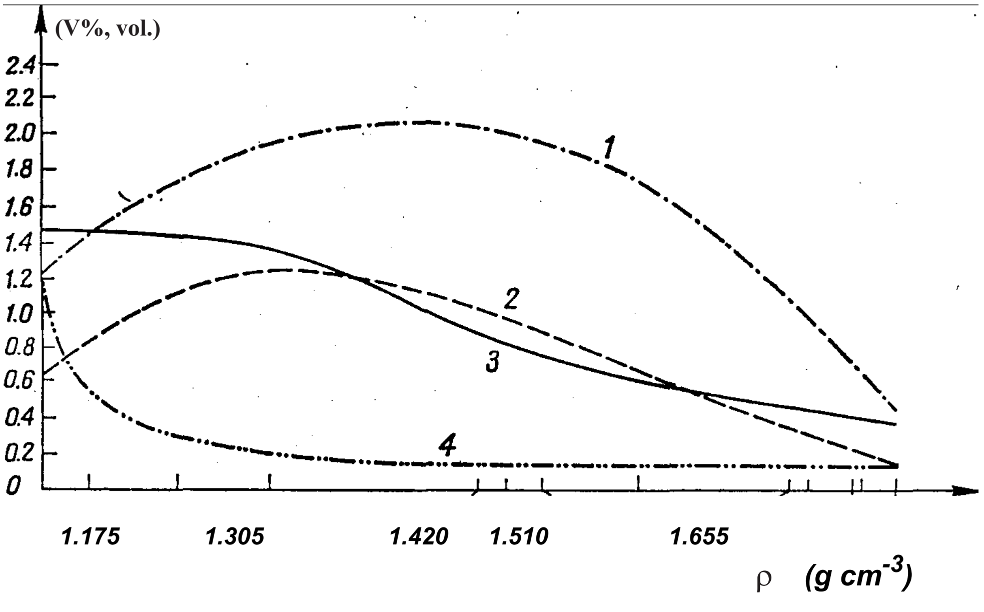

Fig. IV.9. Participation of pores of chosen groups as the function of soil compaction. Amounts of pores in given ranges, diameters: 1– d_{min} (15-610^{-6} m),d_{max}(15-610^{-6}m); 2 – d_{min} (30-18.510^{-6}m), d_{max} (30-18.510^{-6}m); 3 – d_{min} (120010 $^{-6}$m), d_{max} (120010^{-6} m); 4 – d_{min} (15-610^{-6}m), d_{max} (120010^{-6}m) (Walczak, 1977).

In Fig. IV.9 it can be seen that pores of smaller diameters and more regular shapes are created as a result of the decrease in the amount of pores of the greatest and of the irregular diameters. It should be emphasized that in investigating the distribution of pores from the curves of drying only, the pores of the groups I and IV are considered identical. The investigation of the effect of hysteresis and the calculation of the diagrams of distribution enables their differentiation.

1.2. The calculation of the energy expenditure and dissipation of energy at changes of soil moisture

The water in soil is bound, that is, it has a negative potential energy. According to the definition of soil water potential the process of drying is accompanied by work. The soil water potential is a differential thermodynamical function. The value of the potential in a soil saturated with water and the value of the potential of pure free water, being on the same conventional level in the gravitational field, are practically equal. That is why the potential of water does not allow for the determination of the differences between the energy states of water in saturated soil and of free water. The difference in the energy states can in this case be determined with the help of the so-called integral potential of water. The integral potential of water in soil - e - is described as work related to a unit volume of water, which must be performed in order to move all the water in the soil at a given moisture - w - to free water at a conventional zero level. According to the definition the integral potential of water in soil can be calculated with the formula:

$$e = \frac{\int_0^w \mu dw}{\int_0^w dw} \qquad \text{IV.1}$$

The value of the integral potential of soil water can also be calculated in chosen ranges of moisture:

$$e = \frac{\int_{w_1}^{w_2} \mu dw}{\int_{w_1}^{w_2} dw} \qquad \text{IV.2}$$

The integral potential of soil water can be described as the mean expenditure of energy per unit volume of water, accompanying the process of moisture changes - w - in a chosen range. In this work the mean expenditure of energy per unit of the volume of water, accompanying the change of moisture from the full saturation of soil with water to the moisture value W(%,vol.) equivalent to pF 4.2. The calculations were performed according to the computer program prepared for the calculations of the dissipations of energy occurring during the changes of moisture. The dependence of the integral potential of soil water on compaction is presented in Fig. IV.10. It follows from the diagrams that the mean expenditure of energy in the process of drying the soil from full saturation with water to a moisture equivalent to pF 4.2 increases with the increase in the compaction of soil. In the case of the investigated soil for compactions about ρ = 1.510 g cm^{-3} there occurs a very clear increase of the mean energy of water binding. It is possible to conclude that the exceeding of this value of compaction is particularly negative from the point of view of the accessibility of water to plants. The increase in the mean energy of water binding with the increase in the compaction of soil suggests that the influence of this effect on the work connected with the total deformation of soil should be considered, since the mean energy of water binding in the investigated ranges of compaction for the investigated soils increases many times. Considering the expenditures of energy in the processes of moistening and drying, attention should be paid to the fact that these processes show hysteresis, so a part of the energy is subject to dissipation.

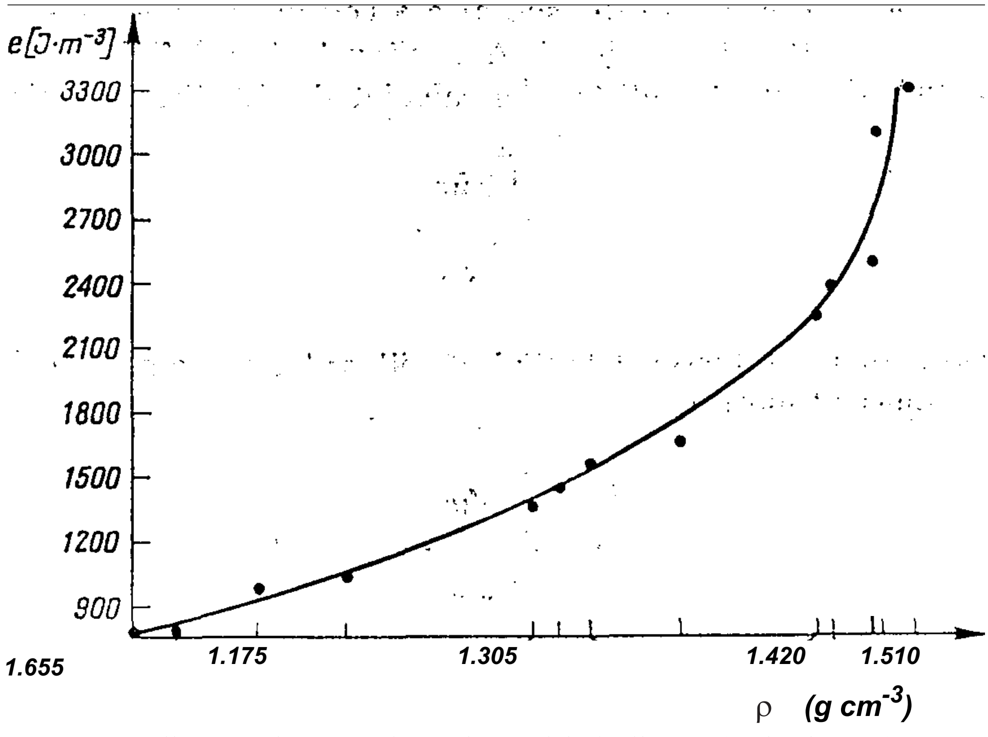

Fig. IV.10. Soil compaction versus integral potential of soil water (Walczak, 1977).

Using the computer program (Walczak and Kaniewska, 1975) the dissipation of energy in the processes of drying and moistening was calculated (Fig. IV.11). The given values are expressed in J per 1 m^3 of soil, which is the value of the energy that undergoes dissipation at a change in moisture of 1 m^3 of soil caused by the changes in the sucking force in the given in Fig. IV.11 cycles.

The dissipation of energy in the drying-moistening process decreases with the increase in the compaction of soil. This fact can be explained by the diminishing of the effect of hysteresis with the increase of compaction. It also follows from the curves that the dissipation of energy clearly increases with the increase of the values of the sucking forces in the cycle, and it can achieve a value of the order of 50% of the expenditure of energy accompanying the process of drying in this cycle.

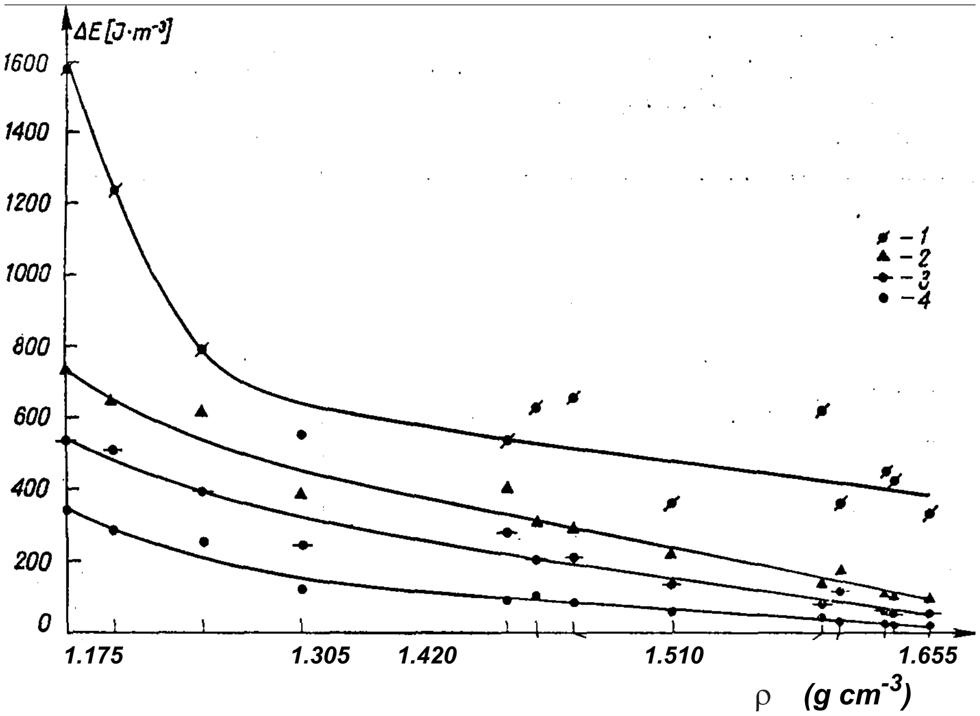

Fig. IV.11. Energy dissipation during drying-wetting cycles for different soil compaction. 1 – pF 0 - pF 2.7 - pF 0, 2 – pF 0 - pF 2.3 - pF 0, 3 – pF 0 - pF 2.2 - pF 0, 4 – pF 0 - pF 2 - pF 0 (Walczak and Kaniewska, 1975).

The results obtained for the investigated soils (brown soil derived from loess and black earth proper derived from medium loam) allow us to draw the following conclusions:

- the compaction of soil clearly influences the course of the soil water potential-moisture characteristics;
- the amount of water bound with forces $0<pF<3$ contained in a unit mass of soil decreases with the increase in compaction, while the amount of water bound with forces $3<pF<4.2$ increases slightly with compaction, the amount of water bound with forces $pF<1.7$ (approximately) in a unit mass of soil diminishes with the increase in compaction, while the amount of water bound with forces in the range about 1.7<pF<about 2.7 depends on compaction only to a slight degree, the amount of water bound with forces $pF>2.7$ (approximately) increases with compaction;
- the increase in compaction causes the diminishing of the amount of water easily accessible to plants ($pF<3$) and the increase in the amount of water accessible to plants with difficulty ($3<pF<4.2$), the increase in the compaction of soil does not influence the amount of water inaccessible to plants ($pF > 4.2$) contained in a unit mass of soil, but it causes the increase of its amount in relation to the unit volume of soil;
- the changes of the courses of the water characteristics indicate that the proportion of pores of large diameters decreases with the increase in compaction, while the number of pores of small diameters increases;
- compaction did not cause, however, the formation of pores of diameters $< 0.2\ 10^{-6}$ m,
- the soil water potential-moisture characteristics show a very clear effect of hysteresis;
- the effect of hysteresis diminishes with the increase in the compaction of soil;
- the theory of hysteresis for independent domains describes processes of drying and moistening of soil;
- the investigating of the effect of hysteresis enables the calculation of the distribution of soil pores according to their sizes with the consideration of the irregularity of their shapes;
- the energy effects accompanying the changes of soil moisture depend on its compaction;
- the expenditures of energy per unit volume of water accompanying the process of drying in the range pF 0-4.2 increase with the compaction of soil;
- in the closed cycles: drying-moistening of soil, the effect of dissipation of energy is considerable and it can achieve values of the order of 50% of work accompanying the process of drying in the cycle;
- the effect of dissipation of energy diminishes with the increase in the compaction of soil;

- the effect of dissipation of energy increases when in the cycle greater values of the sucking forces occur;
- the different compaction of soil and the effect of hysteresis have considerable influence on the distribution of moisture in a soil profile.

2.Influence of soil aggregation on the course of the soil water potential-water content (moisture) characteristics

Soil aggregation directly influences soil compaction, therefore its porosity and compactness, and what is the most important pore distribution function in soil. Thereby the structure is responsible for the amount and distribution of water in a soil profile, the exchange between soil and atmospheric air, shaping of a temperature field in soil, soil water conductivity, resistance to agricultural tools as well as directly influences the amount of water that can be used by plants in physiological processes. (Gummatow and Pachepsky, 1991; Hamblin, 1982; Weisskopf *et al*., 1999). It was stated that soil porosity increases with the degree of soil aggregation and what improves especially is the relation of macropores to micropores which conditions the equilibrium of aerobic and anaerobic processes taking place in soils. Moreover, the self-compaction ability diminishes, thereby facilitating the growth of roots and diminishing resistance to agricultural tools (Abrol and Palta,1979; Barzegar *et al*., 1997; Chan *et al*., 1994; Dechnik and Dębicki, 1976; Dechnik and Lipiec, 1976; Domżał, 1983; Droogers *et al*., 1997; Tamboli *et al*., 1964; Witkowska-Walczak, 1981; Witkowska-Walczak and Walczak, 1999; Wu *et al*., 1990).

The influence of differentiated soil aggregation on static soil water characteristics, *e.g.,* water retention curves and hysteresis effect, was analyzed by Witkowska-Walczak (2000) in her model investigations for 8 soils. Some of the results of her investigations are presented below.

The interdependence between the soil water potential and moisture for aggregates 0.25-0.5 and 1-3 mm of 8 various soils is shown in Fig. IV.12.

It can be noted from Fig. IV.12 that the soil samples created from aggregates 0.25-0.5 mm show a high level of moisture, invariable practically for a given soil, in the range pF 0-pF 1.5. It fluctuates in a very wide range and with pF0 is from 35% for soils No.1 and 3 to 60% for soils No. 6 and 8, and with pF 1.5 from 28% for soil No. 2 to 54% for soil No.8. When the soil water potential exceeds pF 1.5, moisture drops drastically to pF 2.7 reaching the values of 10% for soil No. 2 and 22% for soil No. 7 respectively. With the potential corresponding to pF 4.2, moisture fluctuates from 3% for soil No.3 to 12% for soil No. 8. Retention curves of the samples initially created from aggregates of a diameter 1-3 mm take an absolutely different course. Soil water potential – moisture characteristics of aggregates 1-3 mm show a drastic decrease with pF between 0-0.4. With pF over 0.4 there was a systematic continuous decrease

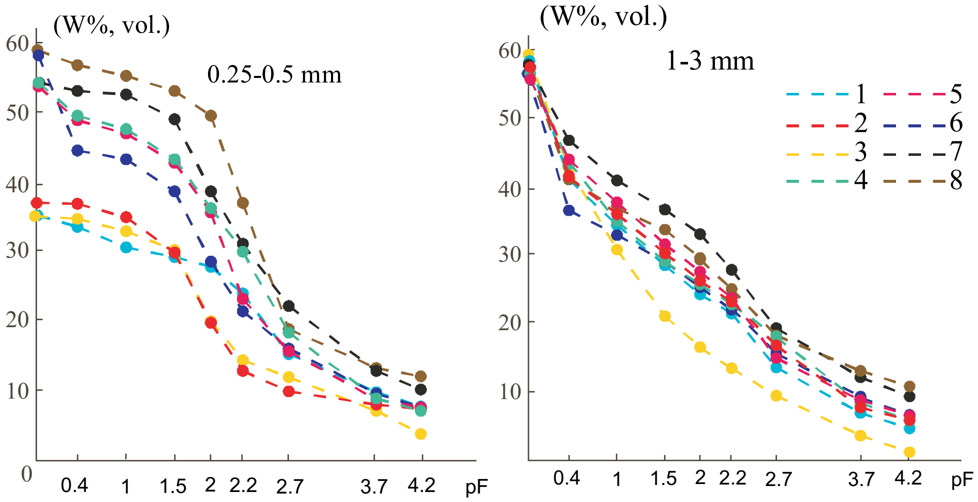

Fig. IV.12. Soil water potential (pF) – moisture (W%, vol.) characteristics in drying process for different monoagregate samples. 1. Orthic Podzol, 2. Orthic Podzol, 3. Eutric Cambisol, 4. Calcaric Cambisol, 5. Haplic Luvisol, 6. Orthic Luvisol, 7. Haplic Phaeozem, 8. Stagnogleyic Phaeozem (Witkowska-Walczak, 2000).

of water content in the samples created from aggregates of a diameter 1-3 mm. The differences of soil moisture at given values of soil water potential for the same aggregates did not exceed 12%, with the exceptions brought about by the course of a retention curve for soil No. 3. To sum up, it can be concluded that the course of soil water potential-moisture characteristics is different for aggregates of a diameter 0.25-0.5 mm and 1-3 mm. The samples created from aggregates of a diameter 0.25-0.5 mm show high water content for low soil water potentials in the range from pF 0 to 1.5.-2.2, then they decrease drastically to pF 2.7 and much more gently to pF 4.2. On the other hand, the samples created from aggregates of a diameter 1-3 mm, show a drastic moisture decrease with pF between 0-0.4 and then their water content gently diminishes to pF 4.2. Such a course of retention curves proves the fact that with the increase of aggregate size the number of big pores in samples increases ($\phi > 18.5$ 10^{-6} m) and with aggregates of a diameter over 1mm it is the number of very big pores that increases ($\phi > 1200 \cdot 10^{-6}$ m). The number of small pores ($\phi < 0.2 \cdot 10^{-6}$ m) is not dependent on aggregate size but on the type of soil. Their greatest number was observed for black earth (soils No. 7 and 8) from 14.2 to 9.8% m^3m^{-3}, and the least number for brown soil (No. 3) from 3.2 to 1.9% m^3m^{-3}.

On the base of the data obtained, the level of the hysteresis effect of soil water potential- moisture characteristics for selected pF values, *ie* 0; 0.4; 1; 1.5; 2 and 2.2 (W), for all aggregate samples of the soils investigated was calculated. The values calculated are shown in Fig. IV.13.

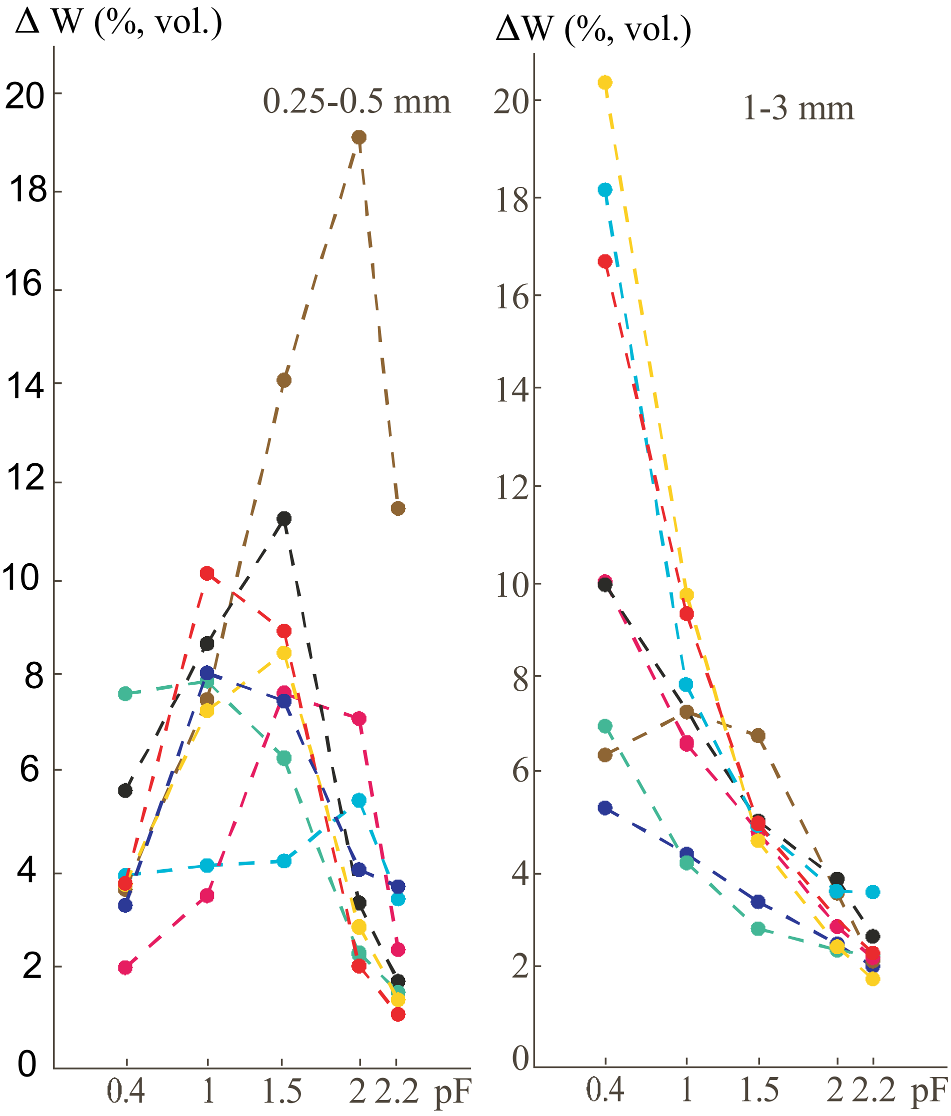

Fig. IV.13. Hysteresis effect (W) of soil water potential-moisture characteristics at chosen pF values from samples created from aggregates 0.25-0.5 and 1-3 mm (Witkowska-Walczak, 2000). Explanations as in Fig. IV.12.

Aggregates 0.25-0.5 mm displayed the minimum values of the hysteresis effect at pF 2.2 – from 1 to 3.5 %, with the exception of soil No. 8 which reached the value of 11.5% at pF 2.2. The level of the hysteresis effect for this fraction displayed an increasing tendency to pF1-1.5-2, depending on the type of soil and at times reached up to 8-12% whereas the maximum value of 19% was observed for soil No. 8 at pF 2. Aggregates 1-3 mm were characterized by the decrease of the hysteresis effect level with the increase of the soil water potential value. The only exception was the fraction 1-3 mm of soil No. 8. The decrease of the hysteresis effect level for aggregates 1-3 mm of soils Nos 1,2 and 3 was significant, with 18-20.5% at pF 0.4 to under 3.5% at pF 2.2.

For the samples created from aggregates 0.25-0.5 and 1-3 mm the hysteresis effect was significant, which proves the fact that the aggregates create structures with a great number of pores of irregular shapes. The existence of a great number of pores of irregular shapes in soil consequently brings about major differences in the amount of water bound by soil in the drying-moistening process at the same potential values depending on the way it was achieved. It means that, as opposed to the moistening process, different soil moisture is obtained in the drying process due to different pore groups filling or emptying themselves in a different way. The results obtained prove the fact that in the case of soils of differentiated aggregation there is a great number of pores of irregular shapes, which, while calculating the soil water content, may lead to some relative errors of up to 50%.

V. PEDOTRANSFER FUNCTIONS FOR DETERMINATION OF WATER RETENTION CURVES

Water retention is a basic hydrophysical characteristic of soil, described by the dependence between soil water content and soil water potential. Knowledge of the soil water potential - soil water content characteristics is necessary for studying water availability for plants, plant water stress, infiltration, drainage, water conductivity, melioration and solutes movement in the soil. The spatial distribution of water characteristics in the soil is also an important factor in the investigations of consequences due to climate change.

The determination of the soil water potential - soil water content characteristics is time and labour consuming as well as it requires the use of expensive special equipment. This is the reason that intensive work has been done for over twenty years on formulation of algorithms - models, which enable to determine the soil water retention curves on the base of other soil physical properties routinely measured in laboratories (Rajkai and Varallyay, 1989; Williams *et al*., 1992). The following soil properties are most frequently taken into consideration for the estimation of soil water retention curves: particle size distribution or percentage contribution of particular granulometric fractions, organic matter content and bulk density (Gupta and Larson, 1979; Pachepsky *et al*., 2002; Rawls and Brakiensiek, 1982; Rajkai and Varallyay *et al*., 1989). In some instances granulometric distribution is considered as the only parameter (Ahuja *et al*., 1985; Haverkamp and Parlange, 1986; Husz, 1967). Additionally the soil particles density (Arya and Paris, 1981), soil structure and mineralogical composition of clays are used (Nimmo, 2002; Williams *et al*., 1992). For estimation of the water retention curve the particular measured values of water characteristics are sometimes used, *ie*, water content under complete saturation, water content at chosen soil water potential values and amount of water available for plants (Carsel and Parrish, 1988; Rawls and Brakensiek, 1982; Rawls *et al*., 2001). More and more frequently fractals and artificial neural nets are used in modelling. The comparison of the agreement of water retention curve courses obtained in laboratory and predicted from different models has been presented in papers.

Investigations concerning the influence of the chosen soil physical and chemical parameters on the course of water retention and water conductivity curve began in Poland in the 1970's. What was mainly investigated at that time were the quality relations, *ie.* interdependences between particular granulometric fractions content and the amount of water at different soil water potential values. Turski *et al*. (1974; 1975) concluded that the colloidal clay fraction has significant positive influence on the maximum hygroscopicity and the content of water unavailable for plants. Domżał (1979; 1983) described the influence of colloid s content and fine fractions on water retention curve courses, stressing at the same time that the influence of soil compaction on water retention depends on the type of soil, *ie* its grain size distribution, particularly in the range pF 2-2.7 (100-500 hPa). Zawadzki (1970) observed significant influence of soil texture on water retention curve courses and on the size of potential useful retention, taking into account in the first place the fraction of soil particles of a diameter smaller than 0.02 mm. He also proposed a formulae for calculating water content (W%,v/v) at pF 2 (100 hPa) as:

$$W_{pF2} = 0.667\ P + 7.54$$

where: P- total porosity and obtained the correlation coefficient $r = 0.965$.

Zawadzki, Michałowska and Stawiński (1971; 1974) presented a formulae for calculating water content at pF 4.2 (15 000 hPa) as:

$$W_{pF4.2} = 0.528\ S + 1.37$$

where: S – soil surface area ($m^2\ g^{-1}$).

Trzecki (1974) presented equations for calculating field water capacity (WPP), starting moment for plant wilting point (WPHWR) and permanent plant wilting point (WTWR) as functions of 6 soil granulometric fractions and organic compounds content for arable layers as:

$$WPP = 0.0188x_1 + 0.0879x_2 + 0.240x_3 + 0.296x_4 + 0.649x_5 + 0.316x6 + 2.34x_7$$

$$WPHWR = -0.0213x_1 - 0.0338x_2 + 0.115x_3 + 0.451x4 + 0.513x_5 + 0.323x_6 + 2.25x_7$$

$$WTWR = 0.00121x_1 - 0.00868x_2 + 0.0488x_3 + 0.0737x_4 + 0.0485x_5 + 0.142x_6 + 1.25x_7$$

and for subarable layers as:

$$WPP = 0.0157x_1 + 0.091x_2 + 0.284x_3 + 0.353x_4 + 0.105x_5 + O.603x_6$$

$$WPHWR = -0.000227x_1 + 0.0205x_2 + 0.0395x_3 + 0.303x_4 + 0.260x_5 + 0.524x_6$$

$$WTWR = 0.00193x_1 + 0.243x_2 + 0.0111x_3 + + 0.0262x_4 + 0.193x_5 + 0.272x_6$$

where: x_1 - weight % of the fraction of 1.0 - 0.1 mm, x_2 - 0.1 - 0.05 mm, x_3 - 0.05 - 0.02 mm, x_4 - 0.02- 0.006 mm, x_5 - 0.006-0.002 mm, x_6- < 0.002 mm, x_7- weight % of organic matter.

According to numerous investigations, the influence of soil compaction on soil water retention curve courses is extremely significant but, at the same time, conditioned by the type of soil and its specific surface area. It was documented that the increase of the solid phase compaction brings about the decrease of water capacity in the range pF 0-pF 2 (1-100 hPa). It also leads to the increase of the amount of strongly bound water (pF > 3.4) (2 500 hPa), including the amount of water at pF 4,2 (15 000 hPa) by 0.2-0.3%, v/v in sandy soils and 4-6%v/v in the heaviest clay loams (Domżał, 1983). Walczak (1977; 1984) investigated the relation of soil compaction with the retention curve course both in the drying and wetting process. He indicated that soil compaction significantly influences the level of hysteresis effect of the soil water potential-moisture relationship, particularly for low soil compaction values (Kaniewska and Walczak, 1974; Walczak and Kaniewska, 1984). For a soil of different compaction investigations of the affect of hysteresis in the range from pF 0 to pF 2.7 (1-500 hPa) have been carried out according to the earlier quoted methodology. According to the demands of the theory the limit curves and the first order curves of wetting were determined experimentally. The investigations indicated that the pF-moisture relationship is a non-explicit function. The differences of moisture for the same pF value depending on the way in which a given state was achieved (drying or wetting) can be as much as 10% of the value of moisture, which can lead, if this effect is not considered, to relative errors of up to 20%.

The role of humus in estimating soil water retention is usually related to its hydrophysical properties and its influence on soil microstructure. Direct relations between humus and moisture, at the chosen soil water potential values, are hard to capture, however, as a factor modifying soil physical properties, humus is usually taken into account being one of the parameters influencing water retention. Dobrzański, Dechnik and Stawiński (1972) indicated a close relation between soil surface area measured by water vapour and humus compounds (Fig.V.1).

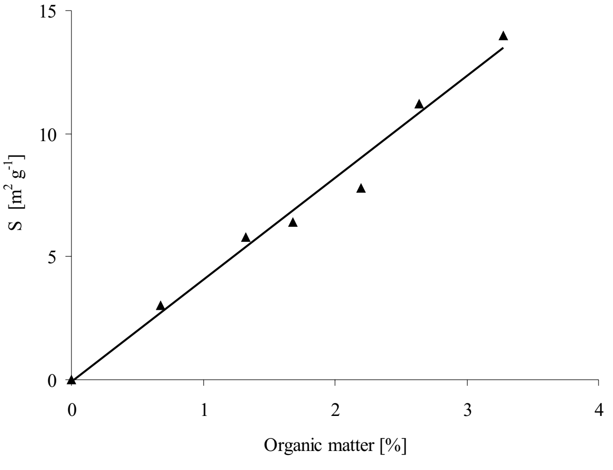

Fig. V.1. Relationship between organic matter and specific surface area (Dobrzański *et al.*, 1972).

Using model soil material (39 genetic levels of 14 different soils) Walczak (1984) investigated the influence of a complex of parameters characterizing the physical state of soil solid phase on the water retention curve courses, taking into account the hysteresis effect. He took into account:

- the specific surface area as an indicator specifying both the development of soil particles surface and the quality of clay minerals,
- grain size distribution as a characteristic of the soil particles spatial configuration aiming at finding a complex numerical indicator of soil texture significantly related to the water retention curve courses, (particles geometric surface, particles distribution indicator (F) (Giesel *et al.*, 1972), statistically central particles diameter, and statistically mean particles diameter),
- compaction as an indicator of soil particle density, illustrating changes in total porosity, and of pore distribution according to size,
- humus content as as indicator of the amount of organic matter influencing soil particle binding.

After analysis of the investigated relationships he proposed the model of retention curve based on the following equations of multiple regression:

$$\theta_p = b_0 + b_1 Y_1 + b_2 Y_2 + b_3 Y_3,$$

for water potential values in the range from pF 0 to pF 2.7 (1-500 hPa) and

$$\theta_p = b_0 + b_1 Y_1,$$

for water potential values higher then pF 2.7 (500hPa),
where: $_p$ is the predicted water content (g g^{-1}), Y_1 - the specific surface area ($m^2 g^{-1}$), Y_2 - the mean weight diameter of particles (mm), Y_3 - the bulk density (g cm^{-3}) and the parameters b_0, b_1, b_2, b_3 are the regression coefficients.

For particular soil water potential values (pF) in the drying process, regression equations have the following shape:
and in the wetting process the following shape:

$$\theta_p(pF0) = 91.60 + 0.060 \cdot Y_1 - 0.93 \cdot Y_2 - 43.77 \cdot Y_3$$

$$\theta_p(pF1) = 79.79 + 0.072 \cdot Y_1 - 1.34 \cdot Y_2 - 37.36 \cdot Y_3$$

$$\theta_p(pF1.5) = 73.66 + 0.084 \cdot Y_1 - 7.75 \cdot Y_2 - 33.40 \cdot Y_3$$

$$\theta_p(pF1.6) = 71.37 + 0.085 \cdot Y_1 - 10.47 \cdot Y_2 - 31.83 \cdot Y_3$$

$$\theta_p(pF2) = 58.86 + 0.091 \cdot Y_1 - 24.64 \cdot Y_2 - 23.02 \cdot Y_3$$

$$\theta_p(pF2.2) = 49.46 + 0.102 \cdot Y_1 - 29.23 \cdot Y_2 - 17.52 \cdot Y_3$$

$$\theta_p(pF2.3) = 45.93 + 0.101 \cdot Y_1 - 30.13 \cdot Y_2 - 15.69 \cdot Y_3$$

$$\theta_p(pF2.7) = 31.07 + 0.110 \cdot Y_1 - 23.59 \cdot Y_2 - 9.72 \cdot Y_3$$

$$\theta_p(pF3.7) = -1.81 + 0.212 \cdot Y_1$$

$$\theta_p(pF4.2) = -1.74 + 0.175 \cdot Y_1$$

$$\theta_p(pF2.3') = 38.23 + 0.104 \cdot Y_1 - 27.09 \cdot Y_2 - 12.44 \cdot Y_3$$

$$\theta_p(pF2') = 44.80 + 0.103 \cdot Y_1 - 26.66 \cdot Y_2 - 15.63 \cdot Y_3$$

$$\theta_p(pF1.6') = 53.33 + 0.096 \cdot Y_1 - 20.05 \cdot Y_2 - 20.82 \cdot Y_3$$

$$\theta_p(pF1.5') = 55.53 + 0.094 \cdot Y_1 - 17.65 \cdot Y_2 - 22.21 \cdot Y_3$$

$$\theta_p(pF1') = 64.46 + 0.088 \cdot Y_1 - 10.13 \cdot Y_2 - 28.06 \cdot Y_3$$

The proposed model enables to calculate moisture on the base of the knowledge of three chosen physical properties of the soil solid phase, both in the drying and wetting process. Therefore, it takes into account the hysteresis effect. The obtained correlation coefficients are included in the range $0.94<R<0.98$. Introducing, for the first time, the specific surface area into the multiparametric model enables to determine within its framework the influence of mineralogical composition and the number of loam particles on the water retention curve courses. Grain size distribution, expressed by the statistically mean particles diameter, and particles density measure, expressed by soil density, together with the specific surface area form a physically justified group of parameters determining soil water properites.

The structure of the Walczak's model is similar to Gupta and Larson and Rawls and Brakiensiek models. These models are commonly used for estimation of water retention curve on the base of the knowledge of the soil solid phase parameters. The water retention model of Gupta and Larson is based on the following multiple regression equation:

$$\theta_p = a_1 X_1 + a_2 X_2 + a_3 X_3 + a_4 X_4 + a_5 X_5$$

where: θ_p (m^3m^{-3}) is a predicted water content, X_1 - the percentage content of sand fraction, X_2 - the percentage content of silt fraction, X_3 - the percentage content of clay fraction, X_4 - the percentage content of organic C, X_5 - the bulk density (g cm^{-3}), while parameters a_1, a_2, a_3, a_4 and a_5 are the regression coefficients.

In the water retention model of Rawls and Brakensiek, which is a modification of Gupta and Larson model, the following equation of multiple regression is used:

$$\theta_p = a_0 + a_1X_1 + a_2X_2 + a_3X_3 + a_4X_4 + a_5X_5$$

where: θ_p (m^3m^{-3}) is a predicted water content, X_1 - the percentage content of sand fraction, X_2 - the percentage content of silt fraction, X_3 - the percentage content of clay fraction, X_4 - the percentage content of organic C, X_5 - the bulk density (g cm^{-3}), while parameters a_o, a_1, a_2, a_3, a_4 and a_5 are the regression coefficients.

Walczak *et al*. (2002) compared above mentioned models on 10 different soils. The water content values for the investigated soils, predicted using the presented models have been compared with water content values measured. The comparison was done by the analysis of correlation parameters between measured - $_m$ and predicted - θ_p soil water content values using each of the models. Fig. V.2 presents measured values of soil water content versus predicted soil water content using the analysed models. For each model the regression equation was determined between water content values predicted using the analysed models and measured water content values:

$$\theta_{Gupta} = 0.11160 + 0.72197\,\theta_{measured}$$

$$\theta_{Rawls} = 0.06762 + 0.88028\,\theta_{measured}$$

$$\theta_{\ Walczak} = -0.0289 + 0.91\,\theta_{measured}$$

The results of performed analysis are presented in Table V.1. The models of Gupta and Larson and Rawls and Brakensiek have very similar structure. They differ only by a free factor in the equation of multiple regression. As independent variables, percentage content of sand, silt and clay, percentage content of organic C and bulk density are used in these models. The use of percentage content of sand, silt and clay in the regression equation seems to be incorrect from statistical point of view, because these quantities are linearly dependent and their sum equals 100%. In Walczak's model, the particle size distribution has been replaced with one parameter - the mean weight diameter of particles, the content of organic C is neglected, the specific

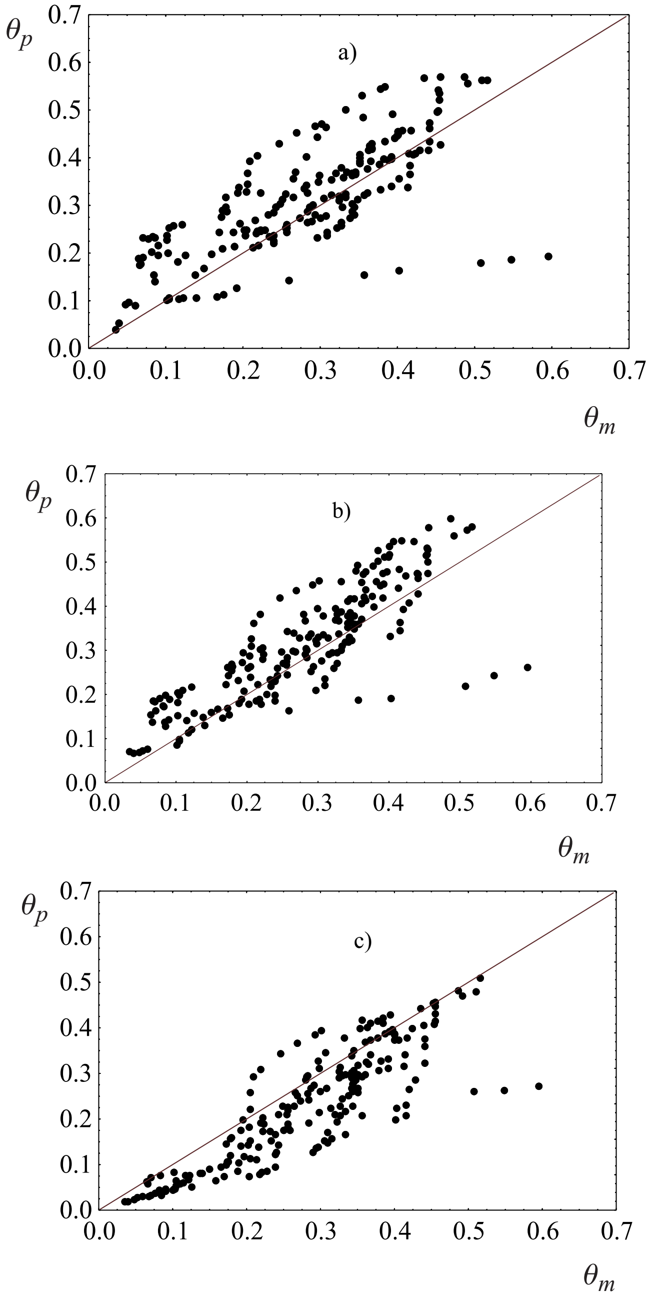

Fig. V.2. Predicted water content (θ_p) versus measured water content (θ_m) values according to models: a – Gupta and Larson; b – Rawls and Brakiensiek and c – Walczak (Walczak *et al.*, 2002b).

Table V.1. Parameters of the multiple linear regression to estimate drying and wetting water retention curves (Walczak *et al*., 2002b)

	Drying water retention curve				Wetting water retention curve			
pF	b0	b1	b2	b3	b0	b1	b2	b3
0	91.60	0.06	-0.93	-43.77	nd	nd	nd	nd
1	79.79	0.07	-1.34	-37.36	64.46	0.09	-10.13	-28.06
1.5	73.66	0.08	-7.75	-33.40	55.53	0.09	-17.65	-22.21
1.6	71.37	0.09	-10.47	-31.83	53.33	0.10	-20.05	-20.82
2	58.86	0.09	-24.64	-23.02	44.80	0.10	-26.66	-15.63
2.2	49.46	0.10	-29.23	-17.52	nd	nd	nd	nd
2.3	49.93	0.10	-30.13	-15.69	38.23	0.10	-27.09	-12.44
2.7	31.07	0.11	-23.59	-9.72	nd	nd	nd	nd
3.7	-1.81	0.21	0	0	nd	nd	nd	nd
4.2	-1.74	0.18	0	0	nd	nd	nd	nd

surface area and the bulk density are used, which statistical significance is evident in the potential range from pF 0 to pF 2.7 (1-500 hPa). Above the potential equal pF 2.7 (500 hPa) in this model, the specific surface area plays the significant role and the bulk density can be neglected. The performed statistical analysis (Table V.2) leads to the conclusion that from the analysed models based on the linear multiple correlation between soil water content and the chosen parameters of soil solid phase at chosen soil water potential, Walczak's model describes the course of the actual retention curve with the smallest estimation error. This is confirmed by the highest value of correlation coefficient (R = 0.8658), the smallest standard error of estimation (SEE =0.0633) and the highest value of Snedecor (F = 593) and (t = 24.35) coefficient which speak for the best correlation between soil water content values predicted from the model and measured water content values. On the base of performed laboratory investigations and soil water retention characteristics calculated using chosen models as well as statistical analysis, it was stated that Walczak's model, including the mean

Table V.2. Correlation coefficients (R), standard errors of estimation (SEE), Snedecor coefficients (F) and (t) coefficients for the analysed models (Walczak *et al*., 2002b)

Model	R	SEE	F	t
Gupta and Larson	0.7352	0.0795	232	15.26
Rawls and Brakensiek	0.8239	0.0723	418	20.46
Walczak	0.8658	0.0633	593	24.35

weight diameter of soil particles, specific surface area and bulk density, is the best from the studied models for describing the real courses of soil water retention curves. It was confirmed by the highest value of correlation coefficient, the smallest standard error of estimation and the highest values of Snedecor and (t) coefficients. These data assure us of the best correlation between soil water content predicted from this model and measured water content values.

Witkowska-Walczak *et al*. (2002) proposed a model with elements of soil structure based on Walczak's model. It was assumed that a possibility exists to check a hypothesis that incorporating the parameters describing the soil aggregate structure into the Walczak's model would increase its predictive ability because the aggregation decides about a number of pores and their shapes, and especially about a content of big pores. The available soil material *ie* soil samples (8 different soils and 6 aggregate fractions) with domination of one aggregate fraction, gives only a chance to recognise the direction of the investigations to be followed to estimate the retention curves of the soils with natural structure. For the statistical analysis, the multiple linear regression method was used and calculations were performed using statistical software STATISTICA. The following input parameters were assumed:

- soil bulk density (BD) as a parameter informing about a maximum soil porosity and thus directly influencing the ability of water accumulation,
- specific surface area (SSA) determined using the method of water vapour adsorption, informing about the soil mineralogical composition,
- percentage contribution silt fraction (SILT) and clay fraction (CLAY) as parameters of soil granulometric distribution,
- quantity (%) of water-stable aggregates of diameters 5-10 mm (F5-10), 3-5 mm (F3-5) and 1-3 mm (F1-3) as parameters of soil aggregate structure.

It should be emphasized, that percentage contribution of sand fraction was not incorporated as an input parameter, because it is linearly correlated with silt and clay fractions considered in the model, what would be improper from the statistical point of view. The same refers to the percentage contribution of the fraction of aggregates with diameters smaller than 1 mm, not being considered here. In the investigation, the gravimetric moisture content referring to a specific value of soil water potential (pF) in the process of drying was assumed as the dependent variable. This choice was prompted by the fact that in the regression analysis soil bulk density was used as an independent variable, which is a function of volumetric soil water content. Finally the shape of the proposed model is:

$$\theta_p = a_o + a_1\,BD + a_2\,SSA + a_3 CLAY + a_4\,SILT + a_5\,F(1\text{-}3) + a_6\,F(3\text{-}5) + a_7\,F(5\text{-}10)$$

where: θ_p - is the predicted water content (g g^{-1}), BD – bulk density (g cm^{-3}), SSA – specific surface area (cm^2 g^{-1}), CLAY – percentage content of clay (%), SILT – percentage content of silt (%), F(1-3) – percentage content of water resistant aggregates of 1-3 cm diameter (%), F(3-5) – percentage content of water resistant aggregates of 3-5 cm diameter (%), F(5-10) – percentage content of water resistant aggregates of 5-10 cm diameter (%). Parameters a_0, a_1, a_2, a_3, a_4, a_5, a_6 and a_7 are the regression coefficients. For particular soil water potential values (pF) in the drying process, regression equations have the following shape:

$$\theta_p\,(pF0) = 109.6 - 54.3\,BD - 0.022\,SSA + 0.186\,CLAY - 0.074\,SILT +$$

$$+ 0.055\,(1-3) - 0.012\,F(3-5) - 0.007\,F(5-10)$$

$$\theta_p\,(pF0.4) = 83.9 - 38.9\,BD + 0.172\,SSA - 0.275\,CLAY + 0.130\,SILT +$$

$$- 0.145\,F(1-3) - 0.210\,F(3-5) - 0.181\,F(5-10)$$

$$\theta_p\,(pF1) = 60.7 - 27.2\,BD + 0.201\,SSA - 0.185\,CLAY + 0.186\,SILT +$$

$$- 0.202\,F(1-3) - 0.148\,F(3-5) - 0.146F(5-10)$$

$$\theta_p\,(pF1.5) = 34.5 - 15.6\,BD + 0.203\,SSA - 0.126\,CLAY + 0.320\,SILT +$$

$$- 0.161\,F(1-3) - 0.070\,F(3-5) - 0.113\,F(5-10)$$

$$\theta_p\,(pF2) = 25.5 - 13.0\,BD + 0.158\,SSA - 0.161\,CLAY + 0.378\,SILT +$$

$$- 0.106 - F(1-3) - 0.050 - F(3-5) - 0.082 - F(5-10)$$

$$\theta_p\,(pF2.2) = 20.3 - 10.5\,BD + .0102\,SSA - 0.106\,CLAY + 0.372\,SILT +$$

$$- 0.064\,F(1-3) - 0.038\,F(3-5) - 0.041\,F(5-10)$$

$$\theta_p\,(pF2.7) = 15.4 - 7.9\,BD + 0.014\,SSA - 0.035\,CLAY + 0.252\,SILT +$$

$$- 0.006\,F(1-3) - 0.036\,F(3-5) - 0.019\,F(5-10)$$

$$\theta_p\ (pF3.7) = -5.86 + 4.47\,BD + 0.131\,SSA + 0.026\,CLAY + 0.090\,SILT + \\ -0.019\,F(1-3) + 0.037\,F(3-5) + 0.006\,F(5-10)$$

$$\theta_p\ (pF4.2) = -5.50 + 4.09\,BD + 0.146\,SSA + 0.034\,CLAY + 0.00025\,SILT + \\ -0.015\,F(1-3) + 0.008\,F(3-5) + 0.005\,F(5-10)$$

The results of the conducted statistical analysis of the comparison of the model without the elements of the soil aggregate structure (Model I) and a model with the elements of the soil aggregate structure (Model II) has been presented in Table V.3.

Table V.3. Results of statistical analysis of the comparison of Model I and II (Witkowska-Walczak *et al.*, 2002)

Water content ($g\ g^{-1}$) (dependent value) referring to log ψ	Statistically significant parameters in:		R^2	
pF	Model I	Model II	Model I	Model II
0	BD, CLAY	BD, CLAY	0.9530	0.9573
0.4	BD, SSA, SILT	BD, SSA, CLAY, SILT, F(1-3), F(3-5), F(5-10)	0.7408	0.9031
1	BD, SSA, SILT	BD, SSA, SILT, CLAY, F(1-3), F(3-5), F(5-10)	0.6982	0.8800
1.5	SSA, SILT	BD, SSA, SILT, F(1-3)	0.7570	0.8399
2	SSA, SILT	BD, SSA, SILT, F(1-3)	0.7913	0.8333
2.2	SSA, SILT	BD, SSA, SILT	0.8146	0.8343
2.7	BD, SILT	BD, SILT	0.7943	0.8008
3.7	BD,SSA, SILT	SSA, SILT	0.7470	0.7591
4.2	BD, SSA	BD, SSA	0.8759	0.8826

All the statistical analyses were performed with significance level value =0.005. The results of the statistical analysis confirm the hypothesis that including soil aggregation parameters into the models of soil content prediction under different soil water potential, significantly increases the predictive ability of the models, what is expressed by the increase of the value of the determination coefficient R^2. For the water content referring to pF 0.4 (2.5 hPa), the statistically significant parameters are contents of water-resistant aggregates of fractions 1-3, 3-5, 5-10 mm together with bulk density, specific surface, the contents of silt and clay fractions. Their incorporation into the model caused the increase of the R^2 value from 0.7408 to 0.9031. A similar increase of R^2 value from 0.6982 to 0.8800 was notified for pF 1 (10 hPa). In these both cases Model I contained only soil bulk density, specific surface and silt content as statistically significant parameters. In the case of pF 1.5 (31 hPa) and pF 2 (100 hPa) in the Model I only specific surface and silt content occurred as statistically significant, while in Model II additionally soil bulk density and the content of the aggregates of 1-3 mm fraction appeared as statistically significant. This caused the increase of R^2 value for pF 1.5 (31 hPa) from 0.7570 in Model I to 0.8399 in Model II and for pF 2 (100 hPa) – from 0.7913 to 0.8333. For the other soil water potentials, the incorporation of the elements of the soil aggregation did not show any significant correlation with soil water content determination, it only changed the number of significant parameters for the determination of water content at pF 2.2 (155 hPa) and pF 3.7 (5 000 hPa). This change referred in both cases to the bulk density. Its occurrence in the model increased R^2 value from 0.8146 to 0.8343 for pF 2.2 (155 hPa) and from 0.7470 to 0.7591 for pF 3.7 (5 000 hPa). For the other soil water potential values, i.e. pF 0 (1 hPa), pF 2.7 (500 hPa) and pF 4.2 (15 000), the same statistically significant parameters occurred in both models. The minimum increase of R^2 value, even when the statistically significant parameters changed or not, were caused by the higher number of incorporated parameters in the analyzed models. In the case when the statistically significant aggregate fractions appeared, the increases of R^2 values were considerable and reached even 0.18. It is quite characteristic that for pF 0.4 (2.5 hPa) and pF 1 (10 hPa), the fractions of 1-3, 3-5 and 5-10 mm were statistically significant whereas for pF 1.5 (31 hPa) and pF 2 (100 hPa) significant was only the aggregate fraction of 1-3 mm.

The results of the statistical analysis of the comparison of the investigated models lead to the following conclusions:

- when estimating the soil water content referring to the water potential of pF 0 (1 hPa) using both models, the similar correlation coefficient values were obtained and the statistically significant parameters were soil bulk density and the content of silt fraction; the soil structure is not revealed when using the Model II because at soil saturation with water (pF 0) all the pores are filled with liquid phase and there are no phase interactions;

- in the case of estimating the water content referring to water potentials of pF 0.4 (2.5 hPa) and pF 1 (10 hPa) with the use of the Model II the soil structure revealed considerably by a significant increase of correlation coefficient value; percentage contents of aggregate functions of 5-10, 3-5 1-3 mm were statistically significant parameters; it can be explained by the fact that in the created inter-aggregate structures, the three phase interactions start to dominate;
- in the case of estimating soil water content referring to the water potentials of pF 1.5 (31 hPa) and pF 2 (100 hPa) with the use of the Model II, the percentage content of aggregate fraction 1-3 mm appeared as a statistically significant parameter and a considerable increase of the correlation coefficient value was observed; interaction between phases in inter-aggregate spaces created by this aggregate fraction became predominant;
- in case of estimating the water content referring to soil water potentials of pF 2.2 (155 hPa) and pF 4.2 (15 000 hPa), *ie* for an amount of water available for plants stored in medium-size pores, the similar values of correlation coefficients were obtained; parameters of the soil structure incorporated into the Model II stopped to reveal, what proves that smaller and smaller pores are occupied with water and it is stronger bound with soil solid phase; bigger pores, number of which is directly connected with occurring of the inter-aggregate spaces, gradually stop to play a role in the process of water binding.

Gnatowski (2001) investigated the statistical relations between Mualem-van Genuchten's equations parameters describing the water retention curve course, and water conductivity coefficient for 87 peat soil samples and their elementary physical properties (ρ - density (g cm^{-3}) and P - ashes (% a.s.m.)). In order to guarantee the physical interpretation of the results obtained, he made the following transformations:

$$L^* = \frac{L+10}{10 - L}$$

$$\theta_r * = \frac{\theta_r + 0.05}{0.50 - \theta_r}$$

$$K_s^* = \ln(K_s)$$

$$\alpha^* = \ln(\alpha)$$

$$\theta_s * = \ln(\theta_s)$$

$$n^* = \ln(n - 1.1)$$

where: θ_s – saturation moisture ($\mathrm{cm^3\,cm^{-3}}$), θ_r – residual moisture ($\mathrm{cm^3\,cm^{-3}}$), α ($\mathrm{hPa^{-1}}$), n (-), L (-) – model parameters. As a result of the statistical analysis conducted the following regression equations were obtained:

$$\theta_s^* = 0.138468 + 0.0108598\,P - 0.762692\,\rho - 0.111439\,\ln(P)$$

$$\theta_r^* = 96.7386 - 436.177\,\rho + 0.016193\,P^2 + 867.459\,\rho^2 + 27.6022\,\ln(\rho)$$

$$\begin{aligned}\alpha^* = {} & 114.697 - 0.659748\,P - 238.113\,\rho + 4.08889\,\rho^{-1} + 3.63416\,\ln(P) + \\ & + 60.6568\ln(\rho) + 2.14613\,P\,\rho\end{aligned}$$

$$n^* = 8.03053 + 0.650829\,P - 39.3579\,\rho + 134.476\,\rho^2 - 6.28299\,\ln(P)$$

$$\begin{aligned}K_s^* = {} & 1179.23 - 0.462353\,P - 3324.32\,\rho + 22.454\,\rho^{-1} + 3648.58\,\rho^2 + \\ & + 4.49555\,\ln(P) + 484.505\,\ln(\rho)\end{aligned}$$

$$L^* = 4.29603 + 1.45987\,\ln(P) + 2.89744\,\ln(\rho) - 1.08517\,P\,\rho$$

For the above equations the following correlation coefficients were obtained R =0.90 for θ_s*, R=0.82 for θ_r*, R=0.79 for α*, R=0.63 for n*, R=0.65 for K_s* and R=0.80 for L*.

VI. RECAPITULATION

The soil is a heterogeneous, polyphasic, particulate, disperse and porous system, in which the interfacial area per unit volume can be very large. The disperse nature of the soil and its consequent interfacial activity give rise to such phenomena as adsorption of water and chemicals, ion exchange, adhesion, swelling and shrinkage, dispersion, flocculation and capillarity. The three phases are represented in the soil, *ei* solid phase – soil matrix (mineral and organic); the liquid phase – soil water or soil solution and the gaseous phase – soil atmosphere. The organization of the solid components of the soil determines the geometric characteristics of the pores spaces in which water and air are retained and transmitted. Finally, soil water and air vary in composition both in time and in space. The relative proportions of these three phases in the soil vary continuously and depend on weather, vegetation and management

Throughout this elaboration we considered the soil from the point of soil physics, which can be described as the branch of soil science dealing with the physical properties, especially static hydrophysical properties, of the soil, as well as their measurements and prediction.

The fundamental study of soil physics aims at achieving a basic understanding of the mechanisms governing the behaviour of the soil and its role in the biosphere. On the other hand, the practice of soil physics aims at the proper management of the soil by means irrigation, drainage, soil and water conservation, tillage, aeration and the regulation of soil heat.

We hope that this elaboration which shows the complicated nature of soil-water system allows to understand better the processes and interactions in the soil. Descriptions of the water status in the soil, relationship between soil water potential and water content during drying and wetting processes (hysteresis effect), the ways of the measurements of soil water potential and moisture, the influence of compaction and aggregation on the water characteristics of soil as well pedotransfer functions, should be help for everybody who wants to know the involved character of the phenomena occurring in the soil.

Soil and water are the two fundamental resources of our natural environment as well as of our agriculture. For this reason, it has become increasingly important to deeper and disseminate knowledge of the properties and behaviour of the soil-water system in relation to climatological conditions, plant growth and the hydrological cycle.

REFERENCES

Abrol I.P., Palta J.P., 1970. A study of tthe effect of aggregate size and bulk density on moisture retention characteristics of selected soils. Agrochimica, XIV, 2-3, 157-165.

Ahuja L.R., Naney J.W., Williams R.D., 1985. Estimating soil water characteristics from simpler properties or limited data. Soil Sci. Soc. Am. J., 49, 1100-1105.

Arya L.M., Paris J.F., 1981. A physicoempirical model to predict the soil moisture chaaracteristics from particle size distribution and bulk density data. Soil Sci. Soc. Am. J., 45, 1023-1030.

Baranowski P., 2000. Modelowe badania wykorzystania temperatury radiacyjnej roślin do oceny ewapotranspiracji aktualnej. Rozprawa doktorska. Maszynopis. IA PAN, Lublin

Barzegar A.R., Nelson P.N., Oades J.M., Rengasamy P., 1992. Organic matter, sodicity and clay macro-aggregate distribution and structural stability of the surface soil in a humid climate. Soil & Till. Res., 23, 361-372.

Boersma L., Cary J.W., Evans D.D., Ferguson A.H., Gardner W.H., Hanks R.J., Jackson R.D., Kemper W.D., Miller D.E., Nielsen D.R., Uehara G., 1972. Soil water. Eds D.R.Nielsen, R.D. Jackson, J.W. Cary, D.D. Evans, Am. Soc. Agronomy Soil Sci, Madison, WI.

Bolt G. H., Frissel M. J., 1960. Thermodynamics of soil moisture. Neth. Jour. Agr. Sci., 8, 57.

Bouyoucos G.J., Mick A.H., 1940. An electrical resistane method for the continuous measurement of soil moisture under field conditions. Michigan Agr. Exp. Sta. Tech. Bull., 172.

Buckman H., Brady N.C., 1971. Gleba i jej właściwości. PWRiL, Warszawa.

Carsel R.F., Parrish R.S., 1988. Developing joint probability distributions of soil water retention characteristics. Water Resour. Res., 24, 755-769.

Chan K.Y., Heenan D.P., Ashley R., 1994. Seasonal changes in surface aggregate stability under different tillage and crops. Soil & Till. Res. ,28, 301-314.

Coleman B. D., 1964. Thermodynamics of materials with memory. Arch. Rat. Mech. Anal., 17, 19-25.

Colman E.A., Hendrix G.B., 1945. Moisture and energy conditions during downward entry of water into moist and layered soils. Soil Sci.Soc. Am. Proc., 9, 3-11.

Connolly R.D., 1998. Modelling effects of soil structure on the water balance of soil-crop systems: a review. Soil & Till. Res., 48, 1-19.

Crawford J.W., Matsui N., Young I.M., 1995. The relation between the moisture-release curve and the structure of soil. European J. Soil Sci., 46, 369-375.

Dalton F.N., Rawlins S.L., 1968. Design criteria dfr Peltier effect thermocouple psychrometers. Soil Sci., 105, 12-17.

Dean T., Bell J.P., Baty A.J., 1987. Soil moisture measurement by an improved capacitance technique. J. Hydrology, 93, 67-78.

Dechnik I., Dębicki R., 1976. Czynniki zaskorupiania gleb oraz metody przeciwdziałania temu zjawisku. Problemy Agrofizyki, 21, 1-86.

Dechnik I., Lipiec J. 1976. Dynamics of moisture and water-resistance of aggregation of soils cultivated with active implements. Polish. J. Soil Sci., VIII, 2, 117-124.

De VriesJ., King K.M., 1961. Note on the volume of influence odf a neutron surface moisture probe. Can. J. Soil Sci., 41, 253-257.

Dobrzański B., I. Dechnik and J. Stawiński, 1972. Correlation between the soil surface-area and humus compounds in the soil. Polish. J. Soil Sci., 2, 99-105.

Dołgow S.I., Modina S.A., 1969. About Some Rules of Relationship between Plant Yield and Soil Compaction. Tieoria Obrobotki Poczw, Gidrometizdat Press, Moskwa.

Domżał H., 1979. Impact of soil compaction on water useful and unuseful for plants, Roczniki Gleboznawcze, XXX, 3, 45-57.

Domżał H., 1983. Compaction of the solid phase and its role in the formation of the water-air properties of soil. Zeszyty Problemowe Postępów Nauk Rolniczych, 220/II, 137-154.

Droogers P., Meer van der F., Bouma J., 1977. Water accessibility to plant roots in different soil structures occuring the same soil type. Plant & Soil, 188, 83-91.

Enderby A. J., 1955. The domain model of hysteresis. Trans. Faraday Soc., 51.

Enderby A. J., 1956. The domain model of hysteresis. Trans. Faraday Soc., 52.

Everett D. H., Whitton W. J., 1952. A general approach to hysteresis. Trans. Faraday Soc., 48.

Everett D. H., 1954. A general .approach to hysteresis. Trans. Faraday Soc., 50.

Everett D. H., 1955. A general approach to hysteresis. Trans. Faraday Soc., 51.

Gardner W.H., 1965. Water content. In: Methods of Soil Analysis, 82-127, Amer. Soc. Agron., Monograph, 9.

Gardner W.R., Kirkham D., 1952. Determination of soil moisture by neutron scattering. Soil Sci., 73, 391-401.

Gliński J., Walczak R., Witkowska-Walczak B., 1989. Soil water problems in Poland. Agrokemia as talajtan. 38/3-4, 563-577.

Giesel W., Renger M., Strebel O., 1972. Berechnung des kapillaren Aufstiegs aus derGrundwasser in der Warzelraum unter stationaren Bedingungen, Z. Pflanzenern. U. Bodenkd., 132, 1, 9-18.

Gnatowski T., 2001. Retention and hydraulic properties of peat and moorh soils from Biebrza River Valley, PhD Thesis, Agricultural Academy of Warsaw.

Gummatow N.G., Paczepski J.A., 1991. Sowremiennyje predstawlienia o strukturie poczw i strukturo-obrazowanii. Dinamika i Faktory. Izd. AN ZSRR. Puszczinskij Naucznyj Centr, Institut Poczwowiedienija i Fotosinteza, 1/24.

Gupta S.C., Larson W.E., 1979. Estimating soil water retention characteristics from particle size distribution, organic matter content and bulk density. Water Resour. Res., 15, 1633-1635.

Hakansson I., Lipiec J., 2000. A review of the usefulness of relative bulk density valuea in studies of soil structure and compaction. Soil and Till. Res., 53, 2, 71-85.

Hamblin A.P., 1982. Soil water behaviour in response to changes in soil structure. J. Soil Sci., 33, 375-386.

Hamblin A.P., 1991. Sustainable agricultural systems: What are the appropriate measures for soil structure? Austr. J. Soil Res., 29, 709-715.

Haman J., 1955. The influence of plough speed and soil moisture on the work of its volume deformation. Annales UMCS, E, X, 8-19.

Haverkamp R., Parlange J.Y., 1986. Predicting the water retention curve from particle size distribution. Soil Sci., 142, 325-339.

Haverkamp R., Reggiani P., 2002. Physically based water retention prediction models. In: Methods of soil analysis. Madison, WI, SSSA, 762-777.

Hillel D., 1971. Soil Water. Physical Principles and Processes. Academic Press, New York-London.

Hillel D., 1998. Environmental Soil Physics. Academic Press, San Diego-London-Toronto.

Holmes J.W., 1950. Calibration and field use ofthe neutron scattering method of measuring soil water content. Austr. J.Appl. Sci., 7, 45-58.

Holmes J.W., Jenkinson A.F., 1959. Techniques for using the neutron moisture water. J.Agr. Eng. Res., 4, 100-109.

Husz G., 1967. Ermittlung der pF-Kurve aus der Textur mit hilfe von multiplen Regression. Zeitschrift. Pflanzenern. Bodenk., 116, 23-29.

Iwanow P.K., Korobowa L.I., 1969. Tieoria Obrobotki Poczw. Gidrometizdat Press, Moskwa.

Iwata S., Tabuchi T., Warketin B.P., 1988. Soil-Water Interactions. Mechanisms and Applications. M. Dekker, New York-Basel.

Jackson T.J., 1993. Measuring surface soil moisture using passive microwave remote sensing. Hydrologic Processes, 7, 139-152.

Jakson T.J., Le Vine D.M., Griffis A.J., Goodrich D.C., Schmugge T.J., Swift C.T., O"Neall P.E., 1993. Soil moisture and rainfall estimation over a semiarid environment with the ESTAR microwave radiometer. IEEE Trans. Geosci., Rem. Sens., 31, 836-841.

Johnson A.I., 1962. Methods of measuring soil moisture in the field. U.S.Dept. Interior, Geological Survey Water Supply, paper 1619-U.

Kaca E., 1996. Zintegrowana gospodarka wodna zlewni rolniczej. /w/ "Ochrona i zróżnicowany rozwój środowiska wiejskiego". Wyd. SGGW, 104-109.

Kaniewska, J., Walczak R., 1974. A numerical interpretation of the hysteresis phenomenon in the drying and wetting processes of soil, Polish J. Soil Sci., VII, 63-70.

Kaszubkiewicz J., 2000. Modele histerezy wodnej gleb. Wyd. AR we Wrocławiu, 1-134

Kosugi K., 1994. Three-parameter lognormal distribution model for soil water retention. Water Resour. Res., 30, 4, 891-901.

Kuraż V., Matousek J., 1977. A new dielectric soil moisture meter for fiied measurement of soil moisture. Bull. ICID, 26, 1, 76-79.

Kutilek M., Nielsen D. R., 1994. Soil Hydrology. Catena Verlag, Cremlingen-Destedt.

Lebron I., Suarez D.L., Schaap M.G., 2002. Soil pore size and geometry as a result of aggregate size distribution and chemical composition. Soil Sci., 163(3), 165-172.

Leonard R.A., Low P.F., 1962. A self-adjusting nullpoint tensiometer. Soil Sci. Soc. Am. Proc., 26, 123-125.

Lwowicz M.I., 1965. Wodnyj balans i poczwiennyj pokrow. Poczwow., 9, 43-55.

Malicki M., 1983. A capacity meter for investiagtion of soil moisture dynamics. Zesz. Probl. Post. Nauk Roln., 220/II, 201-214.

Malicki M., Skierucha W., 1989. A manually controled TDR soil moisture meter operating with 300 ps rise-time needle pulse. Irrigation Sci., 10, 153-163.

Malicki M., Plagge R., Roth C., 1996. Reduction of soil matric effect on TDR dielectric moisture determination by accounting for bulk density or porosity. European J. Soil Sci., 47, 3, 357-366.

Marshall C.E., 1964. The Physical Chemistry and Mineralogy of Soils. Wiley, New York.

Mbagwu J.S., Auerswald K., 1999. Relationship of percolation stability of soil aggregates to land use, selected properties, structural indices and simulated rainfall erosion. Soil & Till. Res., 50, 197-206.

Miller R.D., 1951. A technique for measuring tension in rapidly changing systems. Soil Sci., 72, 291-301.

Milly P.C.D., 1994. Climate, soil water storage, and the average annual water balance. Water Resour. Res., 30, 7, 2143-2156.

Mioduszewski W., 1996. Gospodarowanie zasobami wodnymi w krajobrazie rolniczym. W: "Ochrona i zrównoważony rozwój środowiska wiejskiego". Wyd. SGGW, 91-103.

Nimmo J.R., 2002. Property transfer from particle and aggregate size to water retention. In: Methods of soil analysis. Madison, WI, SSSA, 777-782.

Njoku E.G., Kong J.A., 1977. Theory for passive microwave remote sensing of near-surface soil moisture. J. Geophys. Res., 82, 3108-3118.

Nofziger D.L., Swartzendruber D., 1974. Material content of binary physical mixtures as measured with a dual energy beam of gamma-rays. J. Appl. Phys., 45, 5443-5449.

Oades J. M., Waters A.G., 1991. Aggregate hierarchy in soils. Austr. J. Soil Res., 29, 815-828.

Oniszczenko W.S., Miczurin B.N., 1971. The influence of soil compaction on water properties of soils. Pocvov., 5, 12-24.

Quirk J.P., Murray R.S., 1991. Towards a model for soil structure behaviour. Austr. J. Soil Res., 29, 829-867.

Pachepsky Y.A., Timlin D.J., Rawls W.J., 2001. Soil water retention as related to topographic variables. Soil Sci. Soc. Am. J., 65, 1787-1795.

Peck A.J., Rabbidge R.M., 1969. Design and performance of an osmotic tensiometer for measuring capillary potential. Soil Sci., Soc. Am. Proc., 33, 196-202.

Philip J.R., 1960. Absolute thermodynamic function in soil-water studies. Soil Sci., 89, 51-73.

Poulovassilis A., 1962. Hysteresis of pore-water, an application of the concept of undependent domains. Soil Sci., 93, 171-195.

Poulovassilis A., Childs E.C., 1971. The hysteresis of pore-water: the nondependance domains. Soil Sci., 112, 201-231.

Rawlins S.L., 1966. Theory for thermocouple psychrometers used to measure water potential in soil and plant samples. Agr. Meteorol., 3, 293-310.

Rajkai K., Varallyay G., 1989. Estimating soil water retention from simpler soil properties by regression techniques. In: van Genuchten M.Th., Leij F.J. & Lund L.J. (Eds.). Indirect methods for estimating the hydrailic properties of unsaturated soils. Proc. Intern. Workshop “Indirect methods for estimating the hydraulic properties of unsaturated soils”. USDA, ARS, Riverside, California, Oct., 11-13.

Rawlins S.L., Dalton F.N., 1967. Psychrometric measurement od soil water potential without precise temperature control. Soil Sci., Soc. Am. Proc., 31, 297-301.

Rawls W.L., Brakensiek D.L., Saxton K.E., 1982. Estimation of soil water properties. Trans. ASAE. 25, 1316-1320.

Rawls W.J., Brakensiek D.L., 1982. Estimating soil water retention from soil properties. J. Irrigation and Drainage. Div. ASCE. 108, 166-171.

Rawls W.L., Gish T.J., Brakensiek D.L., 1991. Estimating soil water retention from soil physical properties and characteristics. Adv. Soil Sci., 16, 213-234.

Rawls W.J., Pachepsky Y.A., Shen M.H., 2001. Testing soil water retention estimation with the MUUF pedotransfer model using data from the United States. J. Hydrology, 251(3/4), 177-185.

Richards L.A., 1960. Advances in soil physics. Trans. Int. Cong. Soil Sci., Madison, 7th, I, 67-69.

Richards L.A., Campbell R.B., 1949. The freezing point of moisture in soil cores. Soil Sci. Soc. Am. Proc., 13, 70-74.

Richards L.A., Ogata G., 1958. A therocouple for vapor pressure measuremnt in biological and soil systems at high humidity. Science, 128, 1084-1090.

Richards S.J., 1965. Soil suction measurementswith tensiometers. In: Methods of Soil Analysis. 153-163, Amer. Soc. Agron., Monograph 9.

Richards S.J., Marsh A.W., 1961. Irrigation based on soil suction measurements. Soil Sci., Soc. Am. Proc., 25, 65-69.

Rovdan E., Witkowska-Walczak B., Walczak R., Sławiński C., 2002. Changes in the hydrophysical properties of peat soils under anthropogenic evolution. Int. Agrophysics, 16, 219-226.

Russell E.W., 1971. Soil structure: its maintenance and improvement. J. Soil Sci., 22, 2, 137-151.

Shainberg I. Levy G.J., Levin J. Goldstein D., 1997. Aggregate size and seal properties. Soil Sci., 162, 470-478.

Słowik K., 1970. The influence of kneeding action of machines on physical properties of soil and on apple trees growth. Prace Inst. Sadownictwa, 14, 29-45.

Sokołowska B.A., 1970. O roli płotnisti pocwy w uswojenii dostupnoj włagi rastieniami. Sbornik Trudow AFI, Leningrad, 22, 32-41.

Stolzy L.H., Weeks L.V., Szuszkiewicz T.E., Cahoon G.A., 1959. Use of neutron eqipment for estimating soil suction. Soil Sci., 88, 313-316.

Tamboli P.M., Larson W.E., Amemiya M., 1964. Influence of aggregate size on soil moisture retention. Iowa Acad. Sci., 71, 103-108.

Thompson L.M., Troeh F.R., 1978. Soil and soil fertility. Mc Graw-Hill, London.

Topp G.C., Miller F.E., 1966. Hysteretic moisture characteristics and hydraulic conductivities for glass-bead media. Soil Sci. Soc. Am. Proc., 30, 28-4.

Topp G.C., Davis A.P., Annan A.P., 1980. Electromagnetic determination of soil water content. Water Resour. Res., 16, 574-582.

Topp G.C., Reynolds W.D., Green R.E., (Eds). 1992. Advances in Measurement of Soil Physical Properties. SSSA Special Publ., No. 30, SSSA, Madison, WI.

Trzecki S.,1974. Determination of water capacity of soils on the basis of their mechanical composition. Roczn. Glebozn., XXV, 33-41.

Turski R., H. Domżał and A. Słowińska-Jurkiewicz, 1974. Impact of clay and humus content on maximal hygroscopity of loess soils. Roczn. Glebozn., XXV, 3, 85-99.

Turski R., H. Domżał, A. Słowińska-Jurkiewicz and W. Martyn, 1975. Influence of coloidal clay, CaCO3 and humus content on water retention, plasticity and swelling of Polish rendzinas. Roczn. Glebozn., XXVI, 3, 35-43.

Turski R., Słowińska-Jurkiewicz A., Hetman J., 1999. Zarys gleboznawstwa. Wyd. AR, Lublin.

Ulaby F.T., Elachi C., 1990. Radar Polarimetry for Geoscience Applications. Artech House, London.

Van Bavel C.H.M., Underwood N., Swanson R., 1956. Soil moisture measurement by neutron moderation. Soil Sci., 82, 29-41.

Vorosmarty C.J., Moore III B., Melillo J. M., Peterson B.J., 1989. Continental scale models of water balance and fluvial transport. Global Biogeochemistry Cycles, 3, 241-265.

Vyn T. J., Raimbault B.A., 1993. Long-term effect of five tillage systems on corn response and soil structure. Agron. J., 85, 1074-1079.

Walczak R., 1977. Model investigations of water binding energy in soils of different compaction. Zesz. Probl. Post. Nauk Roln., 197, 11-43.

Walczak R., 1984. Modelowe badania zależności retencji wodnej od parametrów fazy stałej gleby. Problemy Agrofizyki, 41, 1-69.

Walczak R., Kaniewska J., 1975, A method for calculating enery effects and dissipation of energy in changes of soil moisture. Polish J. Soil Sci., VIII, 101-107.

Walczak R., Witkowska-Walczak B., Sławiński C., 2001. Charakterystyki hydrofizyczne gleb brunatnych i płowych Polski wytworzonych ze skał masywnych. Acta Agrophysica, 57, 159-168.

Walczak R.., Rovdan E., Witkowska-Walczak B., 2002a. Water retention characteristics of peat and sand mixtures. Int. Agrophysics, 16, 161-165.

Walczak R., Witkowska-Walczak B., Sławiński C., 2002b. Comparison of correlation models for the estimation of the water retention characteristics of soils. Int. Agrophysics, 16, 79-82.

Weinberg A.M., Wigner E.P., 1958. The Physical Theory of Neutron Chain Reactors. Univ. Chicago Press, Chicago, Illinois.

Weisskopf P., Zihlmann U., Wiermann C., Horn R., Anken T., Diserens E., 1999. Influence of ploughing on soil structure. Agrarforschung, 6, 293-296.

Williams J., Prebble R.E., Williams W.T., Hignett C.T., 1983. The influence of texture, structure and clay mineralogy on the soil moisture characteristics. Austr. J. Soil Res., 21, 15-22.

Williams R.D., Ahuja L.R., Naney J.W., 1992. Comparisons of methods to estimate soil water characteristics from soil partcle size distribution, bulk density and limited data. Soil Sci., 153, 172-184.

Witkowska-Walczak B., 1980/1981. The influence of soil aggregation on its water retention. Annales UMCS, XXXV/XXXVI, 13, B, 209-225.

Witkowska-Walczak B., 2000. Wpływ struktury agregatowej gleb mineralnych na ich hydrofizyczne charakterystyki. Acta Agrophysica, 30, 1-96.

Witkowska-Walczak B., Walczak R., 1999. Water characteristics and evaporation of soil aggregates formed from the different soil. J. Hydrology Hydromech., 47, 6, 417-429.

Witkowska-Walczak B., Walczak R., Sławiński C., 2002. Correlation model for water retention prediction with soil structure parameters, Polish J. Soil Sci., XXXV/1, 1118.

Witkowska-Walczak B., 2003a. Właściwości hydrofizyczne agregatów gleby brunatnej wytworzonej z piasku słabogliniastego. Acta Agrophysica, 84, 167-176.

Witkowska-Walczak B., 2003b. Hysteresis effect of water retention curves for different aggregates of mineral (Haplic Luvisol) and organic (Gleyic Phaeozem) soils. Polish J. Soil Sci., XXXVI, 1, 13-19.

Wu L., Vomocil J.A., Childs S.W., 1990. Pore size, particle size, aggregate size and water retention. Soil Sci. Soc. Am. J., 54, 952-956.

Youngs E.G., 1960. The hystersis effect in soil moisture studies. 7th Cong. ISSS, Madison, WI, I-4, 108-113.

Zawadzki S., 1970. The influence of mechanical composition of soils on soil moisture retention. Polish J. Soil Sci., III, 1, 11-16.

Zawadzki S., 1999. Gleboznawstwo. PWRiL, Warszawa.

Zawadzki S. Michałowska K., 1974. The relationship between pF-values and soil surface area. Polish J. Soil Sci., VII,1, 4751.

Zawadzki S., Michałowska K., Stawiński J., 1971. The application of surface area measurements of soils for determination of the content of water unavailable for plants. Polish J. Soil Sci., IV, 2, 89-92.

Adress of Authors

Barbara Witkowska-Walczak
Ryszard Walczak
Cezary Sławiński

Institute of Agrophysics
Polish Academy of Sciences
ul. Doświadczalna 4
P.O. Box. 201
20-290 Lublin 27
Poland

e-mail: agrof@demeter.ipan.lublin.pl