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COMPARISON OF SOME METHODS FOR DETERMINING OF THE UNSATURATED HYDRAULIC CONDUCTIVITY OF SOILS

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Water movement in soil profile takes place mainly in the unsaturated zone. Knowledge of diffusivity and hydraulic conductivity as a function of soil moisture content is required for the quantitative evaluation of flow processes such as infiltration, evapotranspiration and water uptake by plants [6, 8]. Changes of hydraulic conductivity at various moisture contents can be also an index of soil pore-size distribution [2].

Methods for determination of the diffusivity and unsaturated conductivity, particularly in field conditions, are laborious, very time consuming and demand complicated instrumentation.

Arya et al. [1] developed and tested a new laboratory procedure called the hot-air or evaporation method, for determination of diffusivity and unsaturated hydraulic conductivity in undisturbed soil cores. The method seems to be attractive, as it is quick, cheap and simple. Ehlers [7] found the method was effective in demonstrating the hydraulic differences between tilled and untilled loess soil. Bouma [3] also reported that the first result obtained with the method were encouraging, but more testing was necessary before a final judgement could be made.

The objective of the work as reported here is to compare the unsaturated hydraulic conductivity measurements by the hot-air method with those from the in situ crust-test procedure at the low soil moisture tensions and with the calculation method in a higher range of soil moisture tension.

MATERIAL AND METHODS

Experimental plots were located on the chernozem, cambisol, rendzina and luvisol with texture of silt loam, medium-heavy silty loam, clay loam and coarse sand, respectively (Table 1).

The diffusivity and unsaturated conductivity was determined by the laboratory method proposed by Arya et al. [1]. Soil cores with undisturbed structure of 200 ml and 5.1 cm in diameter were used. The soil cores were taken in 6 replications from the depth of 5—15 and 25—35 cm from each soil. The soil samples were saturated by capillary rise and equilibrated in a horizontal position for some days. Therefore warm air was blown over the exposed of the core to measure evaporation rates. The warm air was blown from an electric drier, equipped with an additional heating coil and autotransformer allowing to maintain a constant temperature of the blowing air in the range of 80–200°C. The soil cores were weighted every minute. After drying the soil was divided into 12—14 segments to determine soil moisture distribution. The soil was pushed out from the cylinder with a piston of the same diameter.

Preliminary tests were made to find the proper temperature and time of drying to satisfy the following presupposed conditions: cumulative evaporation must be proportional to the square root of time and the original water content of the soil at the closed end unchanged during the evaporation procedure.

In the chernozem, cambisol and rendzina these conditions were met by drying times of 15, 12 and 10 min., respectively, at the temperature of 120—130°C (Fig. 1). Only in the luvisol with the coarsest texture the original water content at the closed end was changed.

The diffusivity was calculated from the moisture distribution curves (Fig 1) obtained after evaporation as a function of volumetric water content Θ [cm^3/cm^3] at 0.02 Θ intervals according to the equation of Bruce and Klute [5]:

$$D(\theta_x) = \frac{1}{2t} \left| \frac{dx}{d\theta} \right|_{\theta_x} \int_{\theta_x}^{\theta_i} x d\theta$$

where $D(\theta_x)$ is the diffusivity [cm^2/min] as a function of water content Θ , t is the time [min], x is the distance from the evaporation surface [cm] and θ_i is the initial water content.

The boundary conditions are:

$$\begin{aligned} \Theta &= \Theta_i & x > 0 & & t = 0 \\ \Theta &= \Theta_o & x = 0 & & t > 0 \end{aligned}$$

where Θ_o is the air-dry soil moisture content.

The geometric mean of replicate D values and 95% confidence interval of the geometric mean were calculated at the University of Geettingen. Institute of Agronomy and Plant Breeding (FRG) with the method used by Ehlers [7]. The all diffusivities determined were

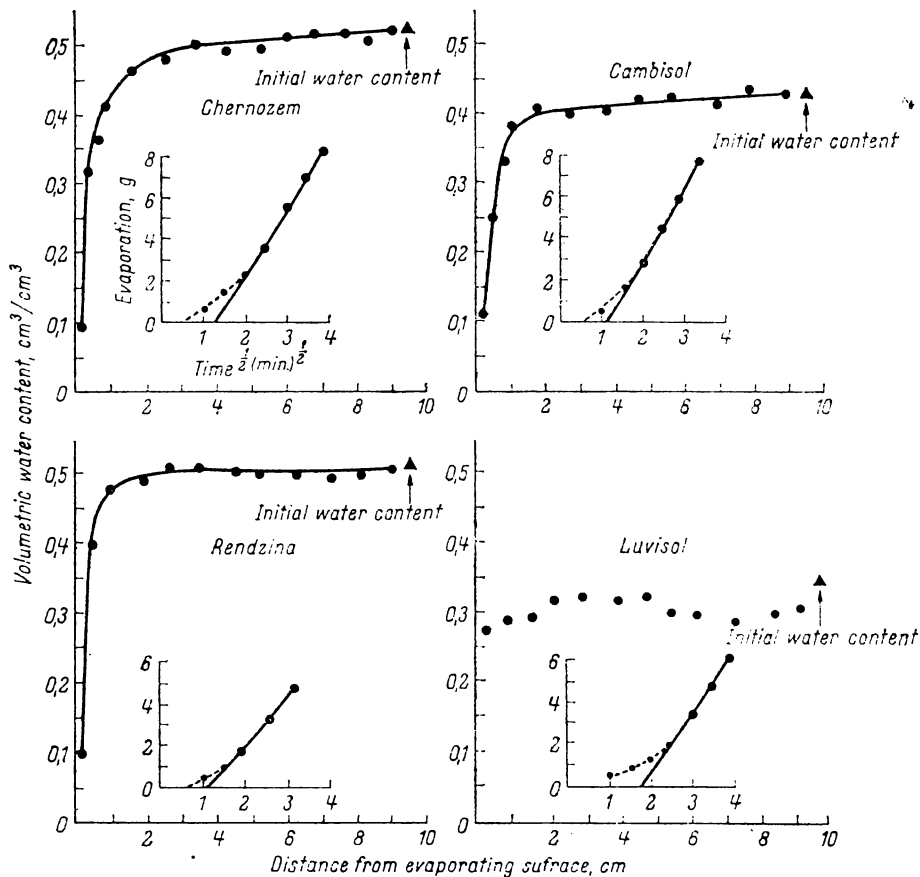


Fig. 1. Examples of moisture distribution in cores of the soils investigated in the 5—15 cm layer after evaporation. The inserts show the relationship between cumulative evaporation and the square root of time

used for computation of the best fitting $D(\theta)$ curves. These curves were used to calculate the unsaturated hydraulic conductivity k as a function of water content (θ) according to the equation:

$$k(\theta) = D(\theta) \frac{d\theta}{d\psi}$$

where $\frac{d\theta}{d\psi}$ is the slope of the moisture retention curve. For comparison with the data obtained in the crust-test procedure and calculation method k cm/day will be presented as a function of Ψ .

The second method of determining k is the in situ crust-test procedure described and tested by Bouma and Denning [4]. The me-

thod consists of a series of infiltration runs through crusts of decreasing hydraulic resistance, each of which gives one point on the hydraulic conductivity curve k (Ψ). The crusts were formed from a pre-wetted mixture composed of gypsum and sand. Each run was made in a soil pit using an excavated cylindrical column of soil with the height and diameter of 28 cm. A ring steel infiltrometer of the same diameter and 10 cm high was fitted onto the column. The water was fed from a burette with a Mariotte device maintaining a constant pressure of about 3 mm water over the crust. Soil moisture tensions were measured by tensiometers. The infiltration rate constant for a period of at least one hour, equals the hydraulic conductivity when the moisture tension gradient is zero. According to Darcy's law:

$$k = \frac{v}{i}$$

where v is the infiltration rate and i is the hydraulic gradient, which is usually close to unity, as the flow occurs only by gravity [3]. With this method the k values were measured to the soil moisture tension of 60 hPa.

The hydraulic conductivity in a higher range of the tension was calculated according to Green and Corey's method as modified by Luxmoore [9]. In the method the calculated hydraulic conductivity is obtained from soil pore-size frequency distribution and the statistical probability of continuity of pores across any plane in the soil. Two experimental values (e.g. at saturation and at some unsaturated water content) obtained with the crust-test procedure were included in computing for matching the calculated conductivity with the use of series of weighted matching factors.

RESULTS AND DISCUSSION

Figure 1 shows that with the hot-air method the required linearity between cumulative evaporation and the square root of time was reached in chernozem and cambisol after 2—4 min, in rendzina after 1—2 min, and in the luvisol after 4—6 min. Arya et al. [1], Ehlers [7] and Opara-Nadi [10] achieved linearity after 2 min, 2,5 to 4 min and 3 do 6 min, respectively. These authors used 80 ml Hapludoll loam soil cores with the diameter of 3.65 cm and the drying temperature of about 90—100°C, 200 ml loess soil cores with the diameter of 5,4 cm and the temperature of 130°C and 500 ml soil loess cores with the diameter of 8 cm and the temperature between 185—230°C. The different external conditions could be responsible, at least partly, for the various

Table 1

Some characteristics of the soils investigated

Soils	Depth cm	Particle size /in mm/				Pore size /in, μ n/			K sat. cm/day	Density kg/m^3 10^{-3}	Organic matter content %	Specific surface m^2/g	CaCO_3 %
		1-0,1	0,1-0,02	< 0,02	< 0,002	> 30	20-3	< 3					
		%				%							
Chernozem	5-15	6	50	44	14	10,0	21,1	21,0	122,1	1,30	3,06	87,7	2,15
	25-35	6	49	45	13	9,6	20,4	19,2	85,2	1,33	2,76	86,3	0,60
Cambisol	5-15	21	31	48	8	9,3	20,0	19,3	41,0	1,40	1,48	50,7	0,15
	25-35	18	35	47	10	7,1	19,0	15,4	60,1	1,52	0,42	46,8	0,13
Rendzina	5-15	18	19	63	27	9,7	16,3	27,5	37,5	1,13	3,95	219,7	24,0
	25-35	17	21	62	29	11,1	15,0	26,4	22,5	1,18	2,87	215,0	33,7
Luvisol	5-15	76	16	8	2	17,6	13,2	10,5	201,2	1,51	1,42	29,2	0,08
	25-35	73	18	9	3	20,1	13,3	7,2	135,1	1,56	0,21	16,5	0,06

times needed for achievement linearity in the investigations mentioned. In this work 200 ml cores with the diameter of 5,1 cm and the temperature between 120—130°C were used for all soils. Therefore, the various time periods for reaching linearity in particular soils might be explained by their individual hydraulic properties and initial wetness rather than by external conditions.

The various hydraulic properties of the soils are shown in Figs. 2 and 3 and typical water distribution curves are shown in Fig. 1. The water content in the sealed ends of chernozem, cambisol and rendzina did not deviate significantly. However, in the luvisol with the coarsest texture and largest contribution of pores $> \mu\text{m}$ in equivalent diameter (Table 1) the water content was much lower. Consequently, one of the theoretical conditions of this method was not satisfied. This failure with the luvisol may be explained by the relatively high hydraulic conductivity of the sandy soil at a moisture tension near saturation and by the rapid upward movement of the water. In the luvisol the average value of k at a moisture tension of 20 hPa was 20.1 cm/day and in the remaining soils — below 4.9 cm/day. Ehlers [7] reported earlier that high hydraulic conductivity might be responsible for this effect in the 30—40 cm layer of untilled loess soil. Furthermore, after evaporation the water content range in the luvisol was much smaller than in the remaining soils and the water contents versus distance did not give a smooth curve (Fig. 1). Perhaps other sizes of core and external conditions could be sufficient to overcome these difficulties [7].

It is worth noticing that the water losses during evaporation from all soils are limited to the first two centimeters at the open end (Fig. 1). The high surface drying of the soils resulted in very steep gradients of the water content with distance near evaporating surface.

Figure 2 shows the calculated diffusivities as a function of soil moisture content. Mean values and the confidence interval were calculated for different water contents as observed in the soil columns after drying. The lowest diffusivities in the whole range of corresponding water contents were in rendzina and the highest diffusivities appeared in the wet range of cambisol.

Hydraulic conductivity as a function of moisture tension is presented in Fig. 3. In both layers of all soils investigated the hydraulic conductivities measured by the hot-air method were lower than those determined by the crust-test procedure to the soil moisture tension of 60 hPa. The degree of agreement was different in particular soils. In the layer of 5—15 cm the lowest differences were in chernozem, somewhat larger in cambisol and much higher in rendzina. At the depth of

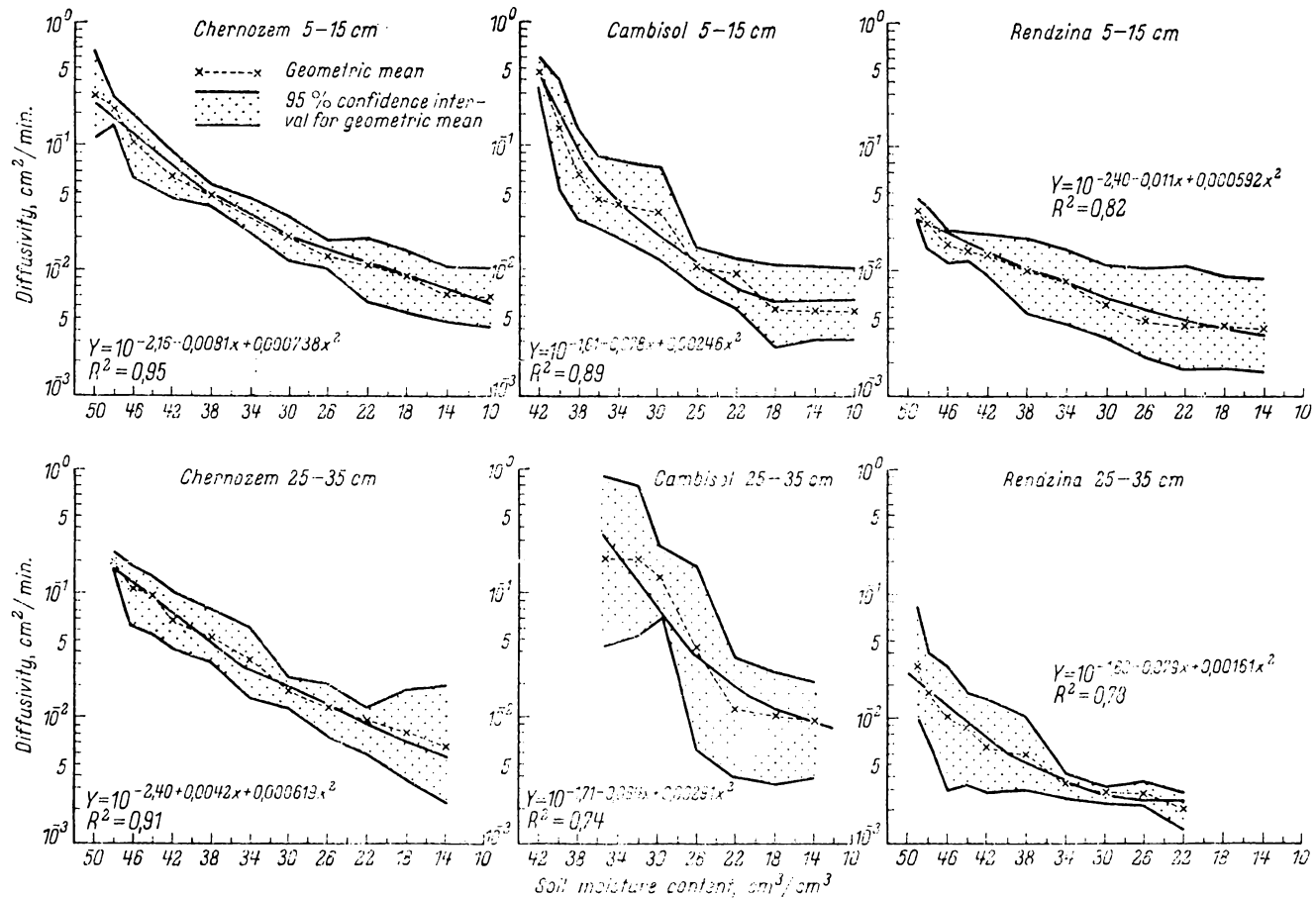


Fig. 2. Diffusivities as a function of volumetric water content in two layers of the soils investigated

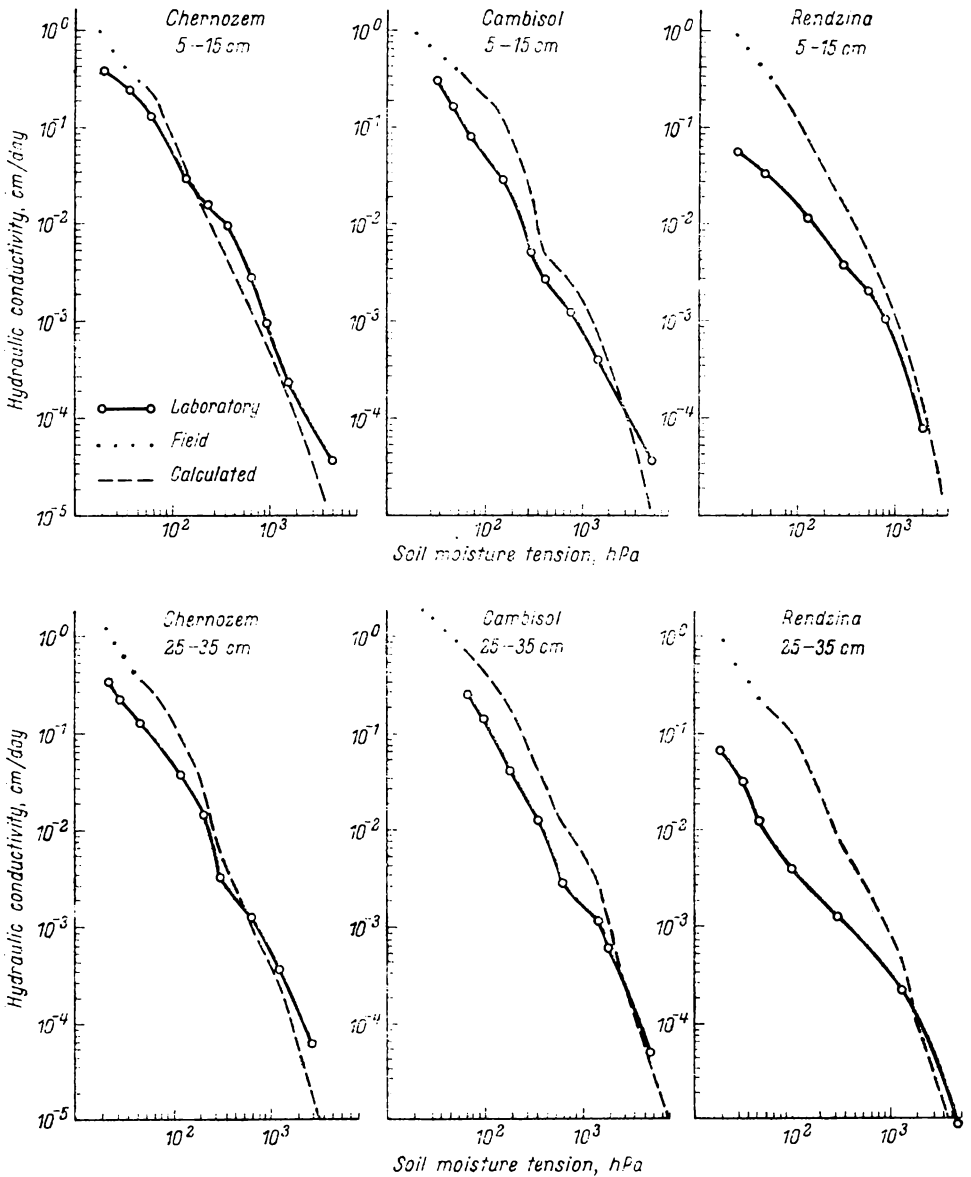


Fig. 3. Hydraulic conductivity as a function of moisture tension in two layers of the soils investigated

25—35 cm the differences in k results, as in the upper layer, were lowest in the chernozem and largest in rendzina. However, in cambisol the results obtained by the hot-air method were barely from soil moisture tension of 70 hPa, thus beyond the tension range, for which the k values were determined by the crust-test procedure.

At the tension range > 60 hPa agreement between k values by the hot-air method and calculated values was also different in different soils (Fig. 3). The best agreement in the moisture tension range corresponding to useful effective retention was found for the chernozem. In the cambisol at the same tension range the results obtained by the hot-air method were slightly lower compared to the calculated values. However, in the rendzina the k values from the hot-air method were significantly lower compared to those from the calculated procedure up to tension of about 300 hPa. However, in the higher tension range the $k(\Psi)$ curves rapidly close together.

It may be concluded that degree of agreement of $k(\Psi)$ obtained with the methods tested depends on the kind of soil and its properties. In chernozem and cambisol soils showing rather uniform pore-size distribution, the results obtained by the laboratory hot-air method compare well with those from the crust-test procedure and the calculation method. However, in the rendzina, showing the lowest diffusivity and the largest contribution of pores $< 3 \mu\text{m}$ in equivalent diameter, the k results from the hot-air method were markedly lower than those obtained by the crust-test and calculation methods (to the tension of 300 hPa). In the luvisol, in which the contribution of coarse pores $< 30 \mu\text{m}$ was large the initial water content at the closed end of soil column was changed when evaporation started.

Consequently, it precluded application of the hot-air method to that soil. The application of this method was also not possible on quartz powder with non-uniform pore-size distribution [10].

In judging the usefulness of the methods in terms of labour consumption and the apparatus required, it might be stated that for the determination of k values in a wide tension range the hot-air method is most suitable, as it is rapid, cheap and demands no special instrumentation.



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СРАВНЕНИЕ НЕКОТОРЫХ МЕТОДОВ ОПРЕДЕЛЕНИЯ ВОДОПРОВОДИМОСТИ ПОЧВ ПРИ НЕПОЛНОМ НАСЫЩЕНИИ

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Резюме

Сравнивали величины коэффициента водопроводимости к при неполном насыщении почв полученные при использовании нового и быстрого лабораторного метода эвапорации с результатами этой водопроводимости полученными по полевому методу при применении гипсово-песчаный корок с сосущим давлением до 60 гПа и по расчетному методу с высшими значениями сосущего давления. Исследования проводились на черноземе, бурой почве, рендзине и палевой почве. Наиболее сходные значения к полученные с помощью сравниваемых методов были установлены в черноземе и бурой почве характеризующихся сравнительно однородным распределением пор в соответствии с их размерами. В рендзине характеризующейся самой малой зернистостью и самым высоким участием малых пор с эквивалентным диаметром $< 3 \mu\text{m}$, значения полученные по методу эвапорации были заметно меньше значений полученных с помощью остальных методов. В паковой почве характеризующейся крупнозернистой гранулометрией и самым высоким участием крупных пор $> 30 \mu\text{m}$, не было соблюдено предварительное теоретическое условие эвапотранспирационного метода, что сделало невозможным ее применение к этой почве.

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PORÓWNANIE NIEKTÓRYCH METOD OZNACZANIA PRZEWODNICTWA
WODNEGO GLEB PRZY NIEPEŁNYM NASYCENIU

Zakład Agrofizyki PAN w Lublinie

Streszczenie

Porównano wartości współczynnika przewodnictwa wodnego k przy niepełnym nasyceniu gleb, uzyskane nową i szybką metodą laboratoryjną ewaporacji z wynikami tego przewodnictwa otrzymanymi metodami: połową z zastosowaniem skorup gipsowo-piaskowych przy ciśnieniu ssącym do 60 hPa oraz obliczeniową przy wyższych wartościach ciśnienia ssącego. Badania przeprowadzono na następujących glebach: czarnoziemie, glebie brunatnej, rędzinie i glebie płowej. Najlepszą zgodność wartości k uzyskanych porównywanymi metodami otrzymano w czarnoziemie i glebie brunatnej, charakteryzujących się dość jednorodnym rozkładem porów według ich wymiarów. W rędzinie, wykazującej najdrobniejsze uziarnienie i największy udział porów małych o średnicy równoważnej $< 3 \mu\text{m}$, wartości k uzyskane metodą ewaporacji były wyraźnie mniejsze niż wartości uzyskane pozostałymi metodami. W glebie płowej, odznaczającej się gruboziarnistym rozkładem granulometrycznym i największym udziałem porów dużych $> 30 \mu\text{m}$, nie został spełniony wstępny warunek teoretyczny metody ewaporacyjnej, co uniemożliwiło jej zastosowanie do tej gleby.

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