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FIELD MEASUREMENT OF THE SATURATED HYDRAULIC CONDUCTIVITY OF A MACROPOROUS SOIL WITH UNSTABLE SUBSOIL STRUCTURE.

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A field method for measuring saturated hydraulic conductivity, K_s, was developed to characterize water flow in highly-weathered soils of Sitiung, Indonesia. Soils in this area are known to absorb large volumes of rainwater rapidly. However, K_s data obtained on soil cores do not corroborate field-observed rapid infiltration rates. In the field method, a constant rate irrigation was applied to a field plot, delineated to a depth of 120 cm, and bordered on the surface to contain a depth of ponded water. The rate of irrigation was sufficient to maintain the ponding depth at a constant level as well as cause water to overflow from the ponded surface. The difference between the steady-state irrigation and overflow rates was considered to be the instantaneous flux and was assumed applicable to all depths. Simultaneous tensiometric measurements of pressure head as a function of depth provided the hydraulic gradients needed for calculation of K_s using Darcy's law. Hydraulic gradients deviated considerably from unity, and soil saturation did not exceed 92% of porosity. Laboratory-measured K_s values for the stable-structured topsoil agreed well with the field data. However, those for the subsoil were 2 to 3 orders of magnitude lower than the field-measured values. The susceptibility of the subsoil to compaction during core extraction and slaking when in contact with free water appeared to be responsible for the highly reduced rates of flow in the laboratory samples. The subsoil pore structure was preserved only as long as it was overlain by the stable structured topsoil. Results suggest that measurements of water flow on small soil cores may, in some cases, be of questionable value. The field method provided accurate *in situ* data on plot-size areas. The field plot method used in this study causes minimal disturbance of the soil while the effects of sample confinement and overburden are represented fully in the measurements. (Soil Science 1998;163:841–852)

Key words: Field saturation, hydraulic conductivity, macropore flow, structural stability, confinement and overburden, hydraulic head gradient.

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SATURATED hydraulic conductivity, K_s, is an important soil hydraulic parameter that is often used to designate the internal drainability of a soil (Soil Survey Staff 1975). Drainage characteristics are important in soil classification and land use planning. Quantitative analyses of infiltration, drainage design, and seepage problems generally require data on K_s. The potential for leaching losses of fertilizers and other agricultural chemicals is generally related to the flow characteristics of the soil, of which K_s is the single most important indicator. A measured value of K_s is also normally required as a matching factor in predicting the unsaturated hydraulic conductivity function from soil water characteristic data (e.g., van Genuchten 1980; Mualem 1992).

Saturated hydraulic conductivity is generally measured on undisturbed soil cores. However, core extraction procedures and sample treatment can seriously alter soil structure (Klute and Dirksen 1986). Additionally, the effect of *in situ* overburden and confinement on water flow is not represented in a small core. Large numbers of cores are generally required to represent the spatial variability in field soils adequately.

Available field methods (e.g., Amoozegar and Warrick 1986) require the use of auger holes, thus altering the overburden and confinement of the target depth. Compaction and sealing of exposed walls of the auger holes generally cannot be avoided. It is also difficult to make measurements on individual soil horizons or soil layers. Recent advances in measurement techniques, such as the Guelph permeameter (Reynolds and Elrick 1985), disk permeameter (Perroux and White 1988), and tension disk infiltrometer (Ankeny et al. 1988) permit field measurements, but these measurements cannot be made on areas large enough to include spatial variations in soil structure. Nor do these techniques include the effects of overburden and confinement. Numerous studies have been made to evaluate and compare these and other methods under various field conditions (e.g., Dorsey et al. 1990; Gupta et al. 1993; Paige and Hillel 1993; Mohanty et al. 1994; Reynolds and Zebchuk 1996). Even though it is not uncommon to find that a particular method, in a particular study, will produce the least amount of variance or may agree well with another method, the conclusions are generally mixed from one study to another.

Many studies acknowledge the significant effects of soil texture, structure, spatial variability, and sample/site disturbance on K_s. Texture and structure effects on K_s in field soils vary both horizontally and vertically (e.g., Bosch and West 1998). Surface structure in tilled soils also often undergoes temporal changes (e.g., Messing and Jarvis 1990). Although macroporosity (such as caused by large pores, interpedal voids, root channels, or cracks) and its distribution in space has a significant effect on saturated water flow and solute transport (e. g., Bouma 1991), this feature is generally not easily quantified through analyses of small cores representing a small space.

Few attempts have been made to measure hydraulic properties under natural field conditions. Stibbe et al. (1970) used large *in situ* monoliths to evaluate K_s and reported much less variability compared with auger hole methods. They ensured complete saturation of the monolith and measured the total hydraulic head using piezometer tubes. The method of measuring the flow velocity was, however, less precise. Field soils generally do not saturate completely and uniformly. Ahuja et al. (1976) demonstrated that *in situ* measurements of the hydraulic properties at apparent field saturation can be made by using double ring infiltrometers and multiple depth tensiometers. Their measurements were made within a 30-cm-i.d. infiltration ring with a buffer zone around it. Flux estimates were obtained from the drop in the water level in the infiltrometer ring. Water was added intermittently to the infiltrometer in an attempt to maintain the ponding depth at a constant level. A fluctuating water level, however, cannot result in true steady-state conditions. In addition, lateral flow occurred at deeper depths despite the buffer zone around the infiltrometer ring. Additionally, a 30-cm-i.d. ring may not cover an area large enough to fully represent macroporosity.

Lauren et al. (1988) measured the saturated hydraulic conductivity of field plots using soil volumes ranging from 884 cm^3 to 240,000 cm^3. As expected, standard deviations of the measurements for the smallest sample volume were four times those for the largest sample volume. They assumed that after 2 to 4 hours of ponding, the infiltration rate obtained from the drop in height of the ponded water level was a measure of the steady-state flux and that any variation in the ponding depth had a negligible effect on the pressure head distribution in the soil profile. These authors also assumed unit-gradient in the total head in order to equate infiltration rates with K_s for the soil profile.

The assumption of unit-gradient during constant-rate infiltration implies uniform conductivity over the entire soil profile. However, field soils generally exhibit considerable textural and structural variability with depth. Ahuja et al. (1988) found that, for both layered and uniform soils, the most serious effect of the unit-gradient assumption was in determining the saturated or near-saturated hydraulic conductivity of the topsoil. Compared with ponded infiltration experiments of the type described in this paper, the effect of unit-gradient assumption is probably not serious during the drainage phase of infiltration-drainage experiments, where surface evaporation is prevented and redistribution of water occurs under gravity (Sisson and van Genuchten 1992).

Our procedure for measuring the field-saturated hydraulic conductivity resulted from the need to better quantify water flow in the highly weathered upland acid soils in Sitiung, Indonesia. These soils absorb large volumes of rainwater rapidly (Arya et al. 1992; Dierolf et al. 1997). Rainfall in the region averages about 3000 mm per year, but even though it rains frequently, with rainstorms of 20 to 50 mm quite common, these soils seldom show signs of flooding. The highly aggregated surface soil remains friable and can be tilled soon after a rainstorm. In contrast to this field evidence, K_s data obtained on soil cores and published in local research reports (e.g., Soil Research Institute 1979; Fakultas Pertanian, Universitas Andalas 1982) indicate a much more slowly draining profile than described by Arya et al. (1992). Two other studies document rapid flow of water in Sitiung soils (Rusman 1990; Dierolf 1992). Both studies provide information about the occurrence and distribution of macropores and conclude that rapid water flow is facilitated by large pores in the subsoil.

Rapid absorption and leaching of large volumes of water has important consequences. For example, such a condition poses less risk of erosion than is commonly assumed. However, fertilizer inputs are less likely to improve soil fertility unless integrated with strategies to maximize retention of nutrients in the soil-plant system. Indiscriminate use of fertilizer and other chemicals enhances the risk of groundwater contamination. Thus, a pressing need exists for reliable methods to characterize drainability of soils in the region. Reliable quantitative information on soil water flow processes should improve the implementation of improved soil management practices.

In this paper we describe a method for measuring the apparent field-saturated conductivity that can be applied to plot-size areas with practically no disturbance of the test soil volume. The method allows for precise measurement of flux and hydraulic head gradients in a ponded field plot. A comparison of laboratory- and field-measured K_s data is also presented.

MATERIALS AND METHODS

Site Description

Field measurements were made at a site near Sitiung 1 A′ in the province of West Sumatra (approximately 1° S and 102° E). The soil at this site is classified as clayey, kaolinitic, isohyperthermic, Typic Kanhapludult. Originally primary rainforest, the site was cleared by bulldozer in the mid- 1970s to settle transmigrant farmers from Java. After several seasons of rice (*Oryza sativa* L.) cultivation, the field was left fallow and became

TABLE 1

Representative physical and hydrologic characteristics of a Typic Kanhapludult in Sitiung 1 A′, West Sumatra, Indonesia

Mean depth cm	Clay content %	Bulk density g/cm³	Particle density g/cm³	Porosity cm³/cm³	Field-saturated water content† cm³/cm³	Upper drained limit (UDL)‡ cm³/cm³	Pressure Head at UDL cm	Macropore fraction¶ % of porosity	Pore radius§ at UDL cm
3.75	71	0.91	2.61	0.651	0.568	0.395	−28	39.3	5.14×10^{-3}
15	75	0.93	2.61	0.644	0.589	0.483	−27	25.0	5.33×10^{-3}
30	75	1.09	2.73	0.601	0.541	0.570	−20	5.2	7.20×10^{-3}
45	73	1.05	2.73	0.615	0.594	0.577	−21	6.2	6.86×10^{-3}
60	73	1.03	2.73	0.623	0.582	0.578	−11	7.2	1.31×10^{-2}
75	68	1.00	2.73	0.634	0.585	0.585	−18	7.7	8.00×10^{-3}
90	68	1.01	2.73	0.630	0.603	0.596	−9	5.4	1.60×10^{-2}
105	59	1.00	2.73	0.634	0.584	0.603	−10	4.1	1.44×10^{-2}

†After the plot was irrigated and subjected to a constant depth of 4.1 cm of ponded water for 260 minutes.

‡Water content when downward drainage substantially decreased, while soil surface was protected against evaporation.

¶Macropore fraction = [(porosity − upper drained limit)/porosity]· 100.

§Pore radius (cm) = $(2\gamma\cos\theta)/(\rho gh)$, where γ = surface tension of water (72 g/sec²), θ = contact angle (0° for water), ρ = density of water (1.0 g/cm³), g = acceleration due to gravity (980 cm/sec²), and h = pressure head (cm H_2O).

covered with alang-alang (*Imperata cylindrica* (L.) Raeusch). Before our study took place, the alang-alang was sprayed with herbicide, and the field was plowed, harrowed, and left undisturbed for a month to settle. Undecomposed debris and roots were removed. Important physical properties of the soil are given in Table 1. Soils in the area are generally quite acid, with Al^{3+} saturation ranging from 70 to 90% and base saturation seldom exceeding 5% (Wade et al. 1988; Subajo 1988). Lime and fertilizer are necessary inputs for crop production.

Principle

An irrigation rate maintained at a level higher than the soil's intake rate must eventually result in ponding and/or runoff. Overflow should occur if the ponded water is contained within retaining borders. The overflow rate should eventually reach steady-state if the irrigation rate remains constant, lateral subsurface movement of water is prevented, and the surface soil structure is preserved. Steady-state irrigation and overflow rates also imply a steady-state flux equal to the difference in the steady-state irrigation and overflow rates. As long as steady state is maintained, the flux through any depth in the soil profile should be the same as the flux entering the soil surface. The pressure head at any depth, z, should also remain unchanged with time. Under these flow conditions, K_s at any depth, z, can be calculated from

$$(K_s)_z = [(di/dt) - (dr/dt)]/(dH/dz)_z \qquad (1)$$

where i and r are the cumulative irrigation and overflow totals during steady infiltration conditions, respectively, and H is the total or hydraulic head, given by

$$H = h + z \qquad (2)$$

where h is the pressure head and z is depth taken negative downward.

Field Procedure

A bare plot, 200 cm by 150 cm, was delineated by digging a trench around the plot to a depth of 120 cm. A double-folded plastic sheet was wrapped and stretched around the outer wall of the monolith. The upper end of the plastic sheet extended about 5 cm above the plot surface. The trench was then packed tightly with soil to eliminate air gaps between the plastic and the wall of the monolith. A water-tight wooden frame about 25 cm high, with its inner dimensions matching the outer dimensions of the test plot, was inserted 15 cm deep between the plastic sheet and the outer wall of the monolith. The soil adjacent to the wooden frame was packed tightly. The wooden frame served to define an area for irrigation, hold a depth of ponded water, and provide controlled passages for overflow. The finished internal dimension of the plot was 194 cm by 145 cm.

The plot setup, including the wooden frame and the overflow collection system, is shown in Fig. 1. The overflow collection system consisted of four gates installed on one side of the wooden frame to permit the outflow of water. Each gate was 5.9 cm deep and 10.0 cm wide. A rectangular water-tight wooden overflow box was attached to the outside of each gate. A 1-cm-diameter plastic hose drained water from the wooden overflow boxes into a large PVC pipe. The PVC pipe was placed in a sloping trench below the plot surface and used to channel the overflow into 5-L plastic containers for collection and measurement.

Three 200-L drums were used for irrigation. A 20-cm-wide strip was cut from top to bottom from each drum. The drums were then welded together so that water poured in one drum filled the entire assembly uniformly. A constant head was created by placing a bubble tube in each drum that terminated about 5 cm above the outlet at the base of one of the drums. The bubble tubes were 1-cm-i.d. glass tubes passing through tight-fitting rubber stoppers inserted in the top opening of each drum. The flow control device, made of an ordinary check valve, was installed at the drum outlet. A pointer with a stationary scale was attached to the check valve to assist in setting the valve at the desired position. The irrigation drums were placed on a table about 1 m above the ground. The outlet at the base of the drum was connected to a flow meter by a flexible plastic hose. Another plastic hose connected to the outflow end of the flow meter extended into the plot. The flow meter (with 1-L resolution) was used to measure the actual delivery of water as a function of time. Calibration of the meter showed that the actual flow was only 0.821 of the meter reading. The setup is shown in Fig. 1. Figure 2 shows the constancy of flow for various rates of discharge.

A mercury manometer-tensiometer system, as described by Arya et al. (1975), was used to measure the soil water pressure head as a function of depth and time. Tensiometers were installed vertically at nominal depths of 2.5, 5, 10, 20, 30, 40, 50, 60, 70, 80, 90, 100, and 110 cm and were arranged in several banks. Tensiometers for the 5- to 50-cm depths were in one bank, and those for the 60- to 110-cm depths were in another.

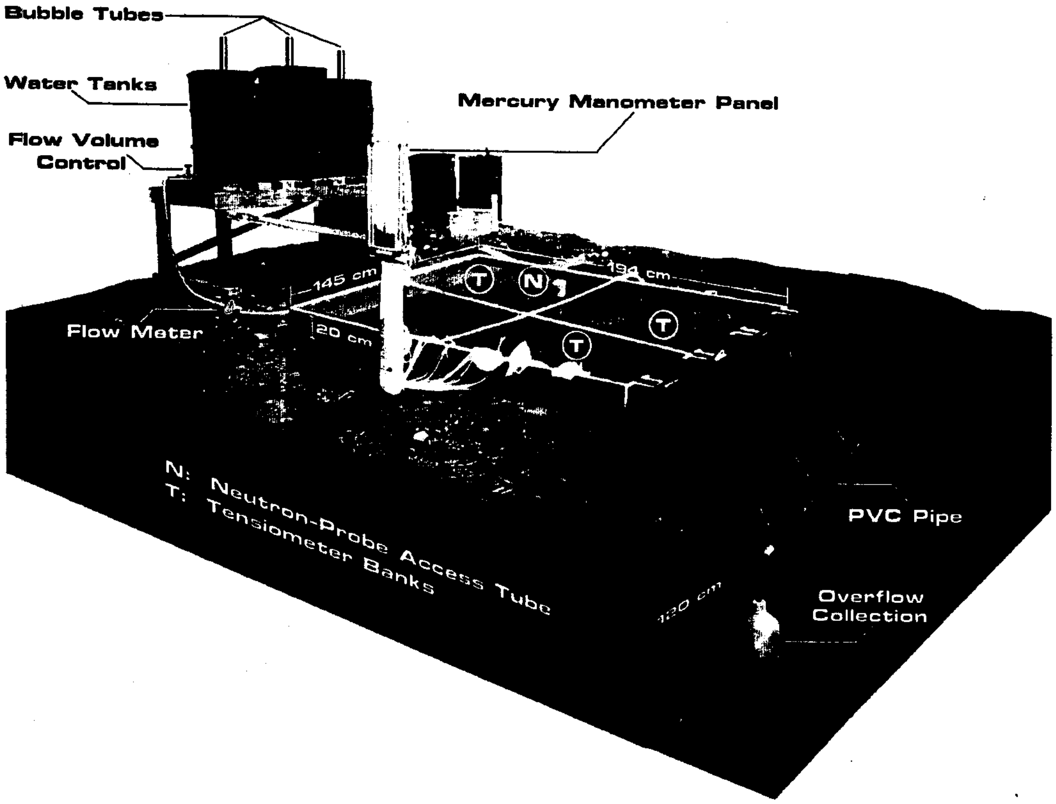

Fig. 1. Experimental setup showing the constant rate irrigation device, the containment of ponded water, the overflow collection system, the tensiometer/manometer system, and the neutron-probe access tube.

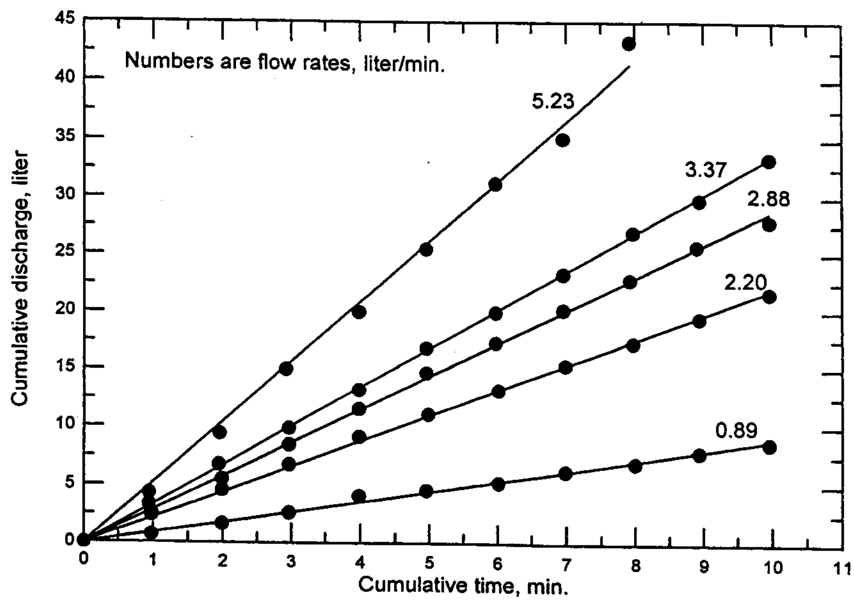

Fig. 2. Test of the constant rate irrigation system. The linear relationship between cumulative discharge and cumulative time indicates constancy of flow rate.

Two additional banks for the 20- to 50-cm depths were installed to obtain a measure of spatial variability. Two tensiometers were installed separately, 50 cm apart, for the 2.5-cm depth. The true depths of the tensiometers were determined at the end of the experiment. A neutron probe access tube installed near the plot center allowed for measurement of soil water content as a function of depth and time.

Measurements

A thin layer of rice straw was spread on the surface to protect the surface soil aggregates during irrigation. Water was initially applied from calibrated watering cans at a rate of about 4 L min^{-1}. Tensiometers in the 0- to 20-cm-depth interval responded within 4 min, and those in the 30- to 110-cm depth interval within 14 to 26 min. Watering with cans continued intermittently for 98 min. By this time, a total of 134 mm of water had been applied to the plot. Although ponding and overflow occurred well before 98 min, steady state was not reached. The constant-rate irrigation system was started 104 min after the start of irrigation. Several adjustments of the irrigation rate were necessary before steady overflow from the ponded water in the plot could be attained. The ponding depth was maintained at 4.1 cm, and watering continued until 194 min. Steady-state flow was established at about 150 min. We continued to measure the volumes of irrigation and overflow at frequent intervals during the steady-state flow process. The cumulative amount of irrigation was read from the flow meter, and any overflow during the experiment was collected in 5-L plastic containers. The drums were then refilled, and another steady-state irrigation cycle was repeated from 228 to 322 minutes.

Neutron-probe and tensiometer readings began before the start of irrigation and continued at 8- to 10-min intervals until several hours after the end of the second irrigation cycle. Thereafter, measurements were made less frequently.

Data were analyzed to determine the steady-state irrigation and overflow rates, as well as the time period over which the steady state existed. The difference between the steady-state irrigation and overflow rates was considered the steady-state flux through all depths. Pressure head readings taken during steady-state flow were used to calculate the hydraulic gradients according to Eq. (2). Hydraulic conductivities were subsequently calculated using Eq. (1).

Laboratory Procedure

Soil cores (11-cm diameter by 10-cm depth) were obtained from selected depths to compare laboratory-measured saturated hydraulic conductivities with the field data. Soil cores were extracted from an open pit adjacent to the test plot. The laboratory procedure was a variation of the constant head method outlined by Klute and Dirksen (1986). Three sets of measurements were made. For the first run, samples were trimmed on both ends and saturated overnight (15 to 20 h) before conducting the flow tests. A reasonable rate of flow was obtained only for the 10- to 20-cm-depth cores which contained large pores and stable aggregates. Flow rates for the subsoil cores were extremely low, suggesting puddling and sealing. A second run was conducted in which cores were not saturated before water flow. In addition to not being saturated before their use, cores in a third run were covered with a 1-cm layer of coarse sand on the surface to minimize the effect of puddling. This modification improved the flow rates slightly, but they still remained quite low.

RESULTS AND DISCUSSION

Field Measurements

Equation (1) is valid only for steady-state flow conditions. Additionally, the flow of water has to be strictly vertical. We had some concern that, after prolonged ponding, water might move laterally in the subsurface from the junction between the wooden frame and the plastic sheet that surrounded the plot. This did not occur as the moist clayey soil packed around the wooden frame provided a tight seal. In cases where lateral seepage may be a real problem, the plastic sheet could be sealed to the wood with an epoxy sealant. We also considered the possibility that fast (preferential) flow occurred vertically along the interface between the plastic and the soil. This type of preferential flow was minimized by ensuring a tight contact between the plastic and the outer surface of the soil monolith. Covering the sides of the soil column with plaster (e.g., Baker and Bouma 1976; Lauren et al. 1988) could be another way of eliminating the problem.

Our data show that steady-state conditions were achieved in the test plot. Figure 3 shows a plot of cumulative irrigation and cumulative overflow for two cycles of steady-state flow. Data indicate that both the irrigation and overflow rates were constant over the two measurements

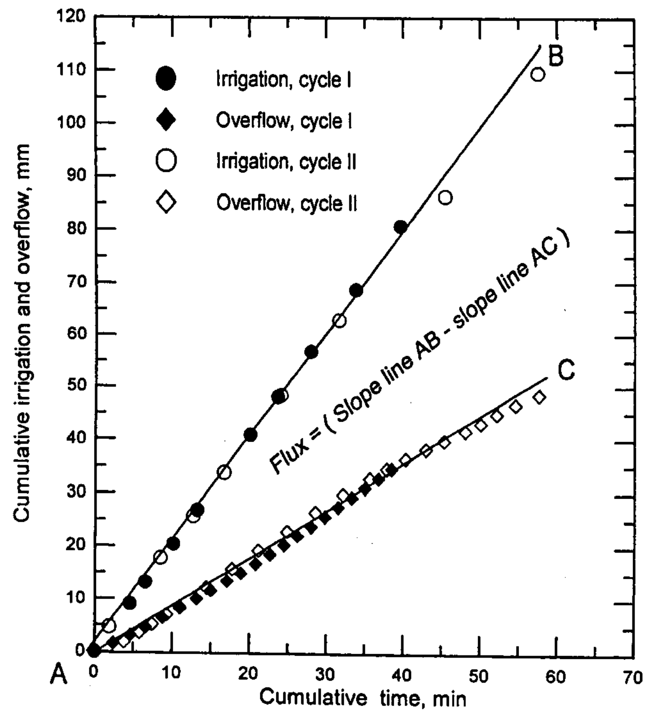

Fig. 3. Cumulative irrigation and overflow during the steady-state flow, for cycles I and II. The straight lines indicate that both the irrigation and overflow rates were constant.

cycles. A very minor deviation from constancy appeared towards the end of the second cycle. Both the cumulative irrigation and the cumulative overflow with time are well represented by straight lines. The difference between the slopes of the two lines is the flux entering the soil. The steady-state fluxes for Cycles I and Cycle II were 1.14 and 0.980 mm min^{-1}, respectively. The slight drop in the flux during the second cycle may have been a result of slow sealing of surface pores caused by constant ponding of water.

Figure 4 shows observed soil water pressure heads at selected depths as a function of time from 150 to 325 min after the start of irrigation. The pressure heads were essentially constant with time during the two cycles. Slight fluctuations (2 to 3 cm at most) may have resulted from small errors in manual reading of the mercury manometers or from abrupt changes in atmospheric pressure (common in the afternoons in Sitiung). Fluctuations in atmospheric pressure were not accounted for when computing soil water pressure heads. We conclude that pressure heads at various depths indicate the existence of steady-state flow conditions. Although most of the depths showed positive heads, the soil at the 110-cm depth maintained a small negative pressure throughout the irrigation cycles. These patterns indicate layers of different hydraulic conductivity in the soil profile.

Preirrigation pressure heads between replicated tensiometers showed little variability (−1 to −3 cm) when the soil was drier. At full apparent saturation, however, spatial variations of 3 to 9 cm were observed. The observed variations are attributed, in part, to variability in the true depths of the replicated tensiometers. For a given target depth, actual tensiometer depths differed by 1 to 3 cm. We attribute the remainder of the variability to planar variations in saturation levels and saturated conductivity.

Our data suggest that uniform saturation probably does not occur even under constant ponding. Water content distributions with depth (Fig. 5) show that the soil below the 30-cm depth was very near maximum field saturation even before the start of irrigation. The addition of water raised the water content of the top 30 cm of the soil profile, but only insignificant amounts of water were added in the deeper layers. Amounts of

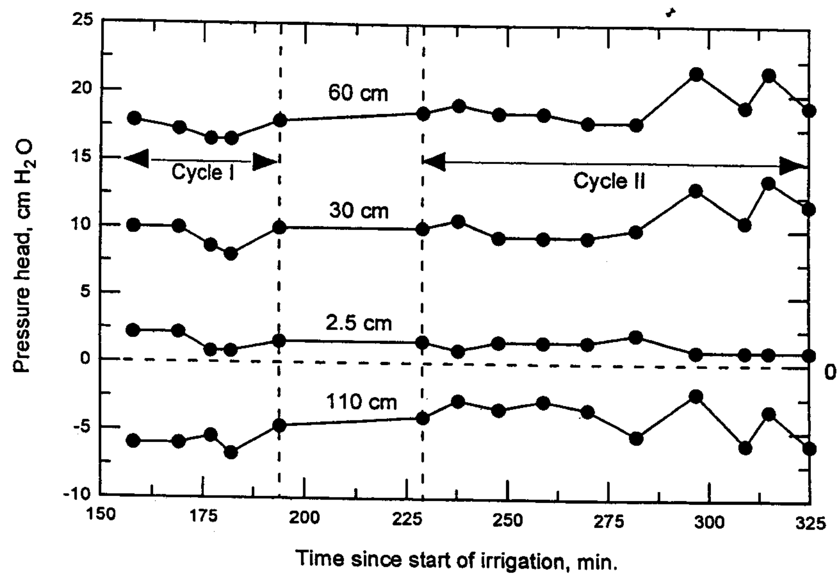

Fig. 4. Soil water pressure head as a function of time at several depths during steady-state flow.

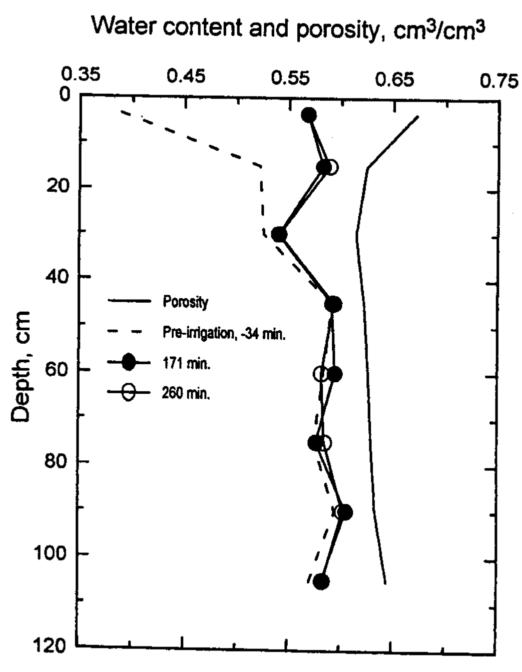

Fig. 5. Porosity profile and water content distribution with depth prior to irrigation and during steady-state flow at 171 min (cycle 1) and at 260 min (cycle 2).

water added to the plot up to 171 and 260 minutes after the start of irrigation were 245 and 366 mm, respectively. A mass balance for the 0 to 112.5-cm-deep profile, at the end of 260 minutes, showed that only 30.4 mm of the 366 mm applied water was actually retained in the soil (Fig. 5). The remainder had either overflowed or drained through the soil profile. Between 171 and 260 minutes, 121 mm of water was applied. Of this amount, only 0.3 mm was retained in the soil profile. Calculations show that, 94.3 mm of irrigation entered the soil in 89 min, whereas 26.4 mm overflowed. These results confirm the high flow rates in Ultisols and Oxisols of Sitiung.

A comparison of the observed water content distributions with the porosity profile in Fig. 5 indicates that soil saturation at all depths remained well below porosity despite having positive pressures in most of the profile (Fig. 4). Saturation values, even under ponded conditions, did not exceed 92% of porosity as calculated from bulk density and particle density data. Incomplete saturation appears to be quite common under field conditions. Chong et al. (1981) reported an average of 85% saturation in Hawaiian Oxisols. Saturation values obtained on small soil cores are generally higher than saturation values observed under field conditions.

Table 2 presents field-measured values of the saturated hydraulic conductivities, K_s. An average value was calculated for those depths for which tensiometers were replicated. The other estimates pertain to a single column of tensiometers. Results in Table 2 indicate a slight decrease in the K_s-values from the first to the second cycle, caused primarily because the value of the constant rate flux for the second cycle (0.98 mm min^{-1}) was slightly less than that for the first cycle (1.14 mm min^{-1}). Mean conductivities in the surface layer (0–5 cm depth) were in the range of 2 to 4 cm h^{-1}. Elsewhere they varied from slightly more than 3.5 to more than 9 cm h^{-1}. These values of K_s are relatively high and help explain why water flows so rapidly in the Ultisols and Oxisols of Sitiung.

Data in Table 2 also indicate that the soil profile has layers with different conductivities. The variations in conductivity with depth cannot be determined if a unit potential gradient is assumed for the entire profile, as is often done (e.g., Lauren et al. 1988). If a constant steady-state flux exists in the entire soil profile (as was the case in the present study), then a unit gradient implies that all soil layers have the same field-saturated conductivity. This was not the case in our study, as indicated by the distributions of the hydraulic gradients at two times (Fig. 6). Data show that unit gradients existed only in the 5 to 10-, 30 to 50-, and 70 to 80-cm-depth intervals. Elsewhere, hydraulic gradients showed considerable deviation from unity. The assumption of a unit gradient would have resulted in soil profile K_s values of 6.84 cm h^{-1} (Cycle I) and 5.88 cm h^{-1} (Cycle II). Although these values are similar to profile averages indicated in Table 2, they over- and underestimate the conductivities of several layers substantially. Hence, if only water flow through the entire profile were of interest, then the assumption of unit gradient would probably be appropriate for this soil. However, if one were interested in characterizing and understanding water flow through the various soil layers, then the unit gradient assumption would be inappropriate.

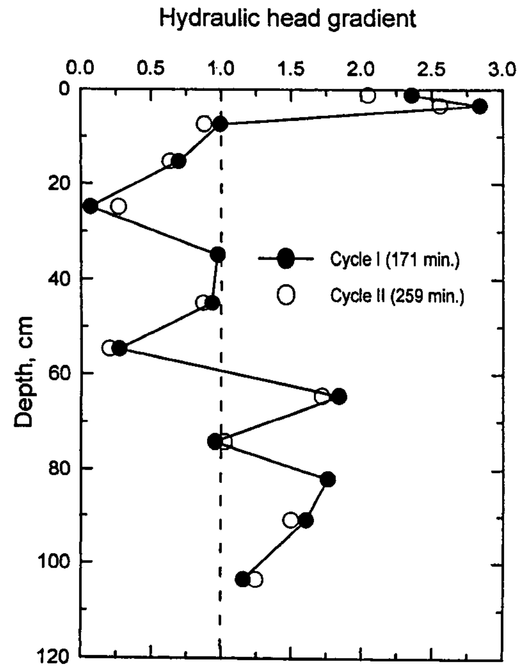

Fig. 6. Distribution of hydraulic head gradients in the soil profile during steady-state flow at 177 min (cycle I) and 259 min (cycle II).

TABLE 2

Field-measured saturated hydraulic conductivities of a Typic Kanhapludult in Sitiung, Indonesia

Depth increment	Saturated hydraulic conductivity, K_s cm h^{-1}		
cm	Cycle I†	Cycle II†	Mean
0–3	4.24	3.54	3.89
3–5	2.57	2.31	2.44
5–10	6.96	6.40	6.68
10–20	9.84	8.63	9.24
20–30	9.24	8.10	8.67
30–40	6.48	5.52	6.00
40–50	7.01	5.86	6.44
50–70	8.73	8.68	8.71
70–80	6.90	5.78	6.34
80–90	3.99	3.38	3.69
90–100	4.26	3.88	4.07
100–110	5.44	4.70	5.07
Profile mean	6.31	4.57	5.94

†Field conductivities for each cycle are means of five values corresponding to five successive pressure head profiles (Fig. 4) measured during the steady-state flow.

Sitiung soils are generally fine-textured, with clay contents ranging from 50% in the surface to 70 to 80% in the subsoil (Soil Research Institute 1979; Subagjo 1988). The surface layer in these soils generally consists of large and highly stable aggregates, which are thought to be responsible for the rapid water flow. Our measurements have shown that the subsoil horizons actually have higher K_s values despite their higher clay content. Higher conductivities in the clayey subsoil are attributed to macropores (Rusman 1990; Dierolf 1992; Arya et al. 1992). It appears that integrity of subsoil macropores is preserved only when the subsoil is overlain by a stable-structure topsoil. The presence of continuous macropores in the subsoil has real and implied effects on drainage, water recharge, and nutrient transport and accumulation.

Laboratory Measurements

Laboratory measurements were made on three sets of soil cores, as described in the Materials and Methods section. The rate of flow for the cores from the 10- to 20-cm depth was comparable to field flow rates. These cores contained numerous large pores and stable aggregates. Flow rates for the subsoil cores were extremely low compared with field flow rates, suggesting puddling and sealing of pores. Not saturating the cores before the flow tests or providing a layer of sand on the surface to protect the pores from sealing improved the flow rates, but they still remained quite low.

Field- and laboratory-measured K_s values for selected depths are compared in Table 3. Data show that laboratory-measured conductivities for the subsoil were 360 to 860 times lower than the field values when the cores were saturated before flow and the surface was left unprotected. They were 5 to 12 times lower than the field values when cores were not saturated before water flow and the surface of the core was protected with sand. It is quite apparent that the core samples did not represent the field pore structure; conductivities obtained on small cores could not explain the rapid flow that occurs in the Ultisols and Oxisols of Sitiung.

Data on water flow and macropore distribution (Arya 1992; Dierolf et al. 1997) suggest that water flow in this and similar soils in Sitiung takes place primarily through well defined macropore channels. The fraction of macropores decreases from 25 to 40% in the top 15 cm of the soil to 4 to 8% in the subsoil (Table 1). However, the decrease in macropore volume with depth is accompanied by an increase in the macropore size. Larger values of K_s in the subsoil are consistent with the observed increase in macropore size with depth (Table 1).

CONCLUSIONS

The proposed method measured field saturated hydraulic conductivities in our experiment adequately. The unstable nature of the subsoil made the field procedure a necessity. The main advantage of the procedure is that large plots can be used with virtually no disturbance of the soil. Measurements can be made on a natural structure soil profile with effects of *in situ* overburden and confinement fully represented, thus making it possible to evaluate the relative importance of sample confinement and overburden on saturated water flow. The problems of sample compaction, fracturing, puddling, and sealing associated with small cores and auger holes can be eliminated. A plot-size sample should also better represent the field. Although relatively large, the core samples in our study had a cross-sectional area of only 95 cm². The field plot, on the other hand, had a cross-sectional area of about 28,000 cm² and represented a soil volume about 300 times that of the soil cores.

Our field- and laboratory-measured conductivity data (Table 3) indicate that stability of the pore structure is an important parameter affecting water flow in Sitiung soils. Results of this study also show that field behavior cannot be studied effectively using small soil cores. Data on porosity, macropore fraction, and macropore size (Table 1) suggest that a core of 11-cm diameter should

TABLE 3

Comparison of field- and laboratory-measured saturated hydraulic conductivities of a Typic Kanhapludult in Sitiung, Indonesia

Depth increment cm	K_s field† cm h⁻¹	K_s laboratory cm h⁻¹		
		Run 1‡	Run 2¶	Run 3§
10–20	9.24	10.33 (n = 6, SD = 9.97)		
20–30	8.67	0.024 (n = 6, SD = 0.012)	0.70 (n = 5, SD = 0.66)	
40–50	6.44	0.008 (n = 5, SD = 0.007)	0.71 (n = 6, SD = 0.79)	1.51 (n = 4, SD = 1.32)

†Taken from Table 2.

‡Cores saturated before flow, surface not protected.

¶Cores not saturated before flow, surface not protected.

§Cores not saturated before flow, surface protected with coarse sand.

n = number of cores; SD = standard deviation.

contain about 9000 macropores. However, the fact that these pores did not produce the expected flow rates suggests compression, puddling, and sealing of pores. Swelling is not suspected to have much of an effect on water flow in highly weathered soils because of the predominantly kaolinitic mineralogy. The observed absence of flooding, even under heavy rainfall, suggests that pore structure in the subsoil is preserved only when it is overlain by a stable structured topsoil. In addition, the effects of confinement, overburden, and incomplete saturation are not included in measurements made on isolated small cores. We suspect that the problem of pore structure stability and failure of soil cores to represent *in situ* flow behavior may be quite general. The field method developed for the Sitiung study could greatly improve the accuracy and reliability of K_s measurement in other soils as well. Data for K_s obtained with soil cores are frequently used in solute transport research and for soil management decisions. Soil cores may not represent field conditions correctly, in which case environmental and soil management problems may not be diagnosed and mitigated properly.

Although no technical constraints prevent their use, larger plots do require extra personnel, material, and time. However, the increased accuracy and reliability of data should outweigh these considerations. Measurements are also made under conditions of natural field saturation. Although complete saturation of porosity was not attained, saturation values obtained in our investigation are probably the maximum that would occur in similar Ultisols and Oxisols in Sitiung. Incomplete saturation under field conditions, caused primarily by air entrapment, is a general phenomenon (e.g., Hillel 1980; Bruce and Luxmoore 1986). For these reasons it seems appropriate that only the maximum measurable conductivity and the associated saturated or satiated water content, under apparent field saturation, should be used in field applications of saturated-unsaturated flow theory.

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