

Evaluation of rainfall infiltration characteristics in a volcanic ash soil by time domain reflectometry method

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Abstract

Time domain reflectometry (TDR) was used to monitor soil water conditions and to evaluate infiltration characteristics associated with rainfall into a volcanic-ash soil (Hydric Hapludand) with a low bulk density. Four 1 m TDR probes were installed vertically along a 6 m line in a bare field. Three 30 cm and one 60 cm probes were installed between the 1 m probes. Soil water content was measured every half or every hour throughout the year. TDR enabled prediction of the soil water content precisely even though the empirical equation developed by Topp *et al.* (1980) underestimated the water content. Field capacity, defined as the amount of water stored to a depth of 1 m on the day following heavy rainfall, was 640 mm. There was approximately 100 mm difference in the amount of water stored between field capacity and the driest period. Infiltration characteristics of rainfall were investigated for 36 rainfall events exceeding 10 mm with a total amount of rain of 969 mm out of an annual rainfall of 1192 mm. In the case of 25 low intensity rainfall events with less than 10 mm h⁻¹ on to dry soils, the increase in the amount of water stored to a depth of 1 m was equal to the cumulative rainfall. For rain intensity in excess of 10 mm h⁻¹, non-uniform infiltration occurred. The increase in the amount of water stored at lower elevation locations was 1.4 to 1.6 times larger than at higher elevation locations even though the difference in ground height among the 1 m probes was 6 cm. In the two instances when rainfall exceeded 100 mm, including the amount of rain in a previous rainfall event, the increase in the amount of water stored to a depth of 1 m was 65 mm lower than the total quantity of rain on the two occasions (220 mm); this indicated that 65 mm of water or 5.5% of the annual rainfall had flowed away either by surface runoff or bypass flow. Hence, approximately 95% of the annual rainfall was absorbed by the soil matrix but it is not possible to simulate soil water movement by Darcy's law over a long period at farm level due to the local differences in rainfall intensity.

Introduction

Many studies conducted to characterize water movement in field soils have indicated that preferential flow often occurs. Such preferential flow is divided into two types. *Bypass flow* occurs often in structured soils in which water flows rapidly through cracks, fissures and biopores. Bouma and Dekker (1978) used a methylene-blue solution as tracer and confirmed that bypass flow occurred as narrow bands on ped surfaces. *Funnel flow* or non-uniform infiltration is affected by the surface microrelief and vegetation. Saffigna *et al.* (1976) revealed that foliage interception and ridging caused non-uniform infiltration using a water-soluble dye. However, the dye-tracing method is destructive and observations cannot be repeated in the same location.

Time Domain Reflectometry (TDR) allows the monitoring of soil-water content *in situ*, so avoiding destruction of the soil profile. Using a rain simulator, Zegelin *et al.* (1992) applied water to a soil surface at a rate lower than the saturated hydraulic conductivity and measured the

increase of the soil water content by TDR. They showed that the amount of water calculated by TDR was equal to that applied within an error of $\pm 10\%$. Hence, if the increase in soil-water content in a soil profile is equivalent to the amount of rainfall, water flows uniformly into the soil matrix but, if rainfall exceeds the increase of the amount of soil-water storage, surface runoff and/or preferential flow occurs. van Wesenbeeck *et al.* (1988) used TDR to demonstrate a non-uniform infiltration rate along a line perpendicular to rows of maize crop due to leaf interception and subsequent stem flow. Parkin *et al.* (1995) measured the soil-water by installing TDR probes vertically and concluded that spatial variability of local infiltration occurred due to the occasional high-intensity pulse of rainfall.

Preferential flow leaches dissolved solutes to a greater depth than occurs by uniform infiltration into the soil matrix (Barraclough *et al.*, 1983; Roth *et al.*, 1991; Rudolph *et al.*, 1996). As a result, the opportunity for plants to absorb nutrients decreases, and possible contam-

ination of groundwater occur. Therefore, estimation of the rates of uniform infiltration and preferential flow is very important for environmentally conservative agriculture. However, the occurrence of preferential flow depends not only upon the soil type, vegetation and surface microrelief but also on the intensity and the quantity of rain, and the antecedent soil-water content.

Volcanic-ash soils, which occur in approximately half of the arable upland in Japan, are characterized by a low bulk density, high water content at -0.1 and -1.5 MPa (Warkentin and Maeda, 1980; Nanzyo *et al.*, 1993), high saturated ($>10^{-3}$ cm s $^{-1}$) and unsaturated ($>10^{-6}$ cm s $^{-1}$ at -10 kPa) hydraulic conductivities (Hasegawa *et al.*, 1994), and abundant biopores formed by decayed plant roots in the subsoil (Tokunaga, 1988).

The present study monitored the soil-water content to a depth of 1 m in an unvegetated volcanic-ash soil throughout the year and determined the effects of surface microrelief on non-uniform infiltration and the amount of matrix flow in relation to the annual precipitation.

Experimental procedures

EXPERIMENTAL SITE

Experiments were conducted in a field at the National Institute of Agro-Environmental Sciences (latitude $36^{\circ} 01' N$ and longitude $140^{\circ} 07' E$). Annual mean precipitation and temperature for 1981 to 1990 were 1219 mm and $13.0^{\circ} C$, respectively (Okuyama, 1990). Daily minimum temperatures during December to February were below $0^{\circ} C$ so that the surface soil was frozen in the morning but thawed in the afternoon.

The soil to a depth of 200 cm is a volcanic ash, classified as a Hydric Hapludand (Soil Survey Staff, 1992), with a texture of heavy clay. The soil profile is divided into plough layer (0–20 cm), ploughsole (20–30 cm) and subsoil (>30 cm). No cracks and fissures were visible at the surface, but a large number of tubular pores formed by decayed plant roots were found in the subsoil.

The experimental site was $10\text{ m} \times 15\text{ m}$ and was kept weed-free throughout the year by hand weeding. Rainfall was collected using a tipping bucket rain gauge at the site.

SOIL WATER CONTENT

TDR

Measurements of the soil-water content by TDR and of rainfall were started on 1 December 1994 and terminated on 13 December 1995 at intervals of every 30 or 60 min. The TDR system was a Tektronix 1502B cable tester interfaced to a Campbell Scientific 21X data logger with PROMS software. The TDR probe consisted of two parallel stainless rods 5 mm in diameter and 50 mm apart. Four 1 m probes were installed vertically at intervals of 2 m along a line from north to south as shown in Fig. 1.

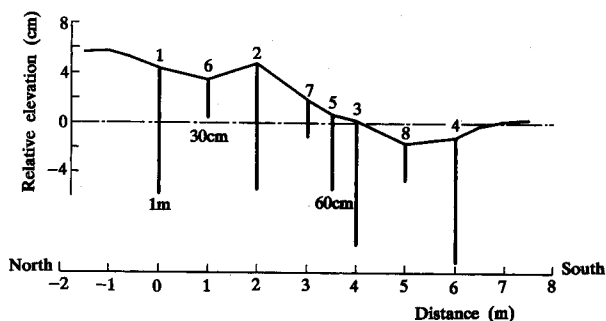


Fig. 1 Distribution of TDR probes and relative elevation of soil surface. The numbers along the soil surface indicate the probe number. Probe lengths of 1 to 4 are 1 m, 5 is 60 cm and 6 to 8 are 30 cm.

Three 30-cm probes were installed between the 1 m probes and one 60 cm probe was installed at 3.5 m from the north. Each rod was driven into the soil by hand through rod guides placed on the soil surface to keep the rods as parallel as possible. The ground surface where the individual probes were installed was surveyed on 13 April, 11 September and 13 December 1995 to confirm the elevations.

Direct soil sampling

Soil core samples were collected to convert the TDR water content by using an empirical equation (Topp *et al.*, 1980) to oven dry water content. Thin-walled steel tubes, whose inside tip diameter of 27.8 mm was 2 mm smaller than that of the tube and 110 cm long, were driven into the soil to a depth of 1 m and retrieved. Then the soil core was extracted from the tube and sectioned at every 10 cm. Nine soil cores were taken at intervals of 1 m along a line 2.0, 1.0 and 0.5 m apart from the TDR line on 20 December 1994, 19 April and 1 August 1995, and 8 soil cores were taken at the locations where TDR probes were installed and additional 2 cores were taken from around probe 1 when all the probes were withdrawn on 13 December 1995.

Dry bulk density of the plough layer was obtained on 20 January and 12 July and that of the ploughsole was obtained on 18 August 1995 using 100 ml core samplers. Dry bulk density at 1 m depth was also measured by digging a pit on 28 September 1995. Samples were taken a few meters away from the TDR line.

Results and discussion

DETERMINATION OF SOIL WATER CONTENT BY OVEN DRYING AND TDR METHODS

Volumetric soil water content of layered soil by the oven drying method is expressed as

$$\theta_d = \sum \rho_i \omega_i d_i \quad (1)$$

where θ is the average volumetric water content of a layered soil of depth $d(=\sum d_i)$, ω_i and ρ_i are the gravimetric water content and dry bulk density of the i -th layer of depth d_i , respectively.

Gravimetric water contents of 9 or 10 replicated soil core samples were averaged of each depth at 10 cm intervals and used to obtain the depth-weighted arithmetic average water content for different TDR probes. Dry bulk density of the plough layer in January was 0.79 Mg m^{-3} and increased to 0.85 Mg m^{-3} in July. In addition, the thickness of the plough layer measured with the soil core samples was not always 20 cm but fluctuated within a few cm among the samples. Dry bulk density of the plough-sole ranged from 0.80 to 0.81 Mg m^{-3} and at 100 cm depth from 0.54 to 0.56 Mg m^{-3} . To avoid complexity in using Eqn. (1), the dry bulk density of the 0 to 30 cm layer was assumed to be 0.80 Mg m^{-3} and that of the subsoil from 30 to 100 cm was 0.55 Mg m^{-3} . Equation (1), therefore, will be written for each probe as

$$\theta_{0-30} = 0.8\omega_1 \quad (2)$$

$$\theta_{0-60} = 0.8\omega_1 \times 0.5 + 0.55\omega_2 \times 0.5 = 0.40\omega_1 + 0.275\omega_2 \quad (3)$$

$$\theta_{0-100} = 0.8\omega_1 \times 0.3 + 0.55\omega_3 \times 0.7 = 0.24\omega_1 + 0.385\omega_3 \quad (4)$$

where ω_1 , ω_2 and ω_3 are the gravimetric water content of the 0–30 cm, 30–60 cm and 30–100 cm layers, respectively.

For the TDR water content, an empirical equation between the relative dielectric constant of soil and the volumetric water content was used (Topp *et al.*, 1980). Figure 2 shows the water content calculated by Eqns (2)–(4) and the empirical equation when soil core samples were taken. Soil-water contents determined by the core sampling method (θ) were always higher than those calculated using the Topp's equation (θ_T). However, the coefficients of determination of the linear regression lines were high for all the probe lengths. It is generally recognized that the Topp's equation does not predict the water content of organic soils, fine-textured soils and low bulk density soils (Topp *et al.*, 1980; Herkelrath *et al.*, 1991; Dirksen and Dasberg, 1993). Hatano *et al.* (1995) pointed out that the empirical equation underestimated the water content of volcanic-ash soils with a low bulk density. Values of θ_T were therefore, converted to θ for all measurements by regression lines shown in Fig. 2.

ANNUAL CHANGES OF SOIL WATER CONTENT

Figure 3 shows the relationship between rainfall and soil-water content to depths of 1 m (probe 2) and 30 cm (probe 6). The soil surface where both probes were installed was relatively high (Fig. 1). Annual fluctuations of the soil surface for each TDR probe were less than 3 mm indicating the stability of the soil surface. Annual precipitation from 1 December 1994 to 30 November 1995 was 1192 mm,

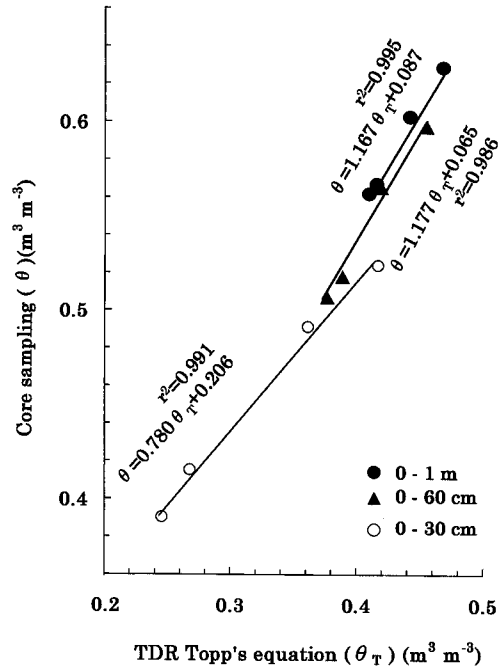


Fig. 2 Volumetric soil water content measured by the core sampling method and calculated by TDR Topp's equation. r^2 indicates the coefficient of determination.

close to the long term annual average of 1219 mm. Soil-water content to a depth of 1 m during the winter season from December 1994 to February 1995 was low and relatively constant due to the small amount of precipitation and soil evaporation. Monthly rainfall from March to July exceeded 100 mm and soil-water content in this period was high. After the end of the rainy season in the fourth week of July, the soil began to dry and the soil-water content decreased to minimum values immediately before a typhoon on September 15 saturated it (Table 1).

Average soil water content on the day after heavy rain was used to express the field capacity in this study although the physical significance of the field capacity is controversial (Hillel, 1980). Ten heavy rain events were used to determine the field capacity (Table 1). More than 640 mm of water was stored to a depth of 1 m at field capacity except at the location of probe 1 whose water content was consistently lower than that of others by 0.01 to $0.05 \text{ m}^3 \text{ m}^{-3}$. The amount of water held to a depth of 30 cm at field capacity was about 160 mm or $0.533 \text{ m}^3 \text{ m}^{-3}$. Field capacity determined in this way was fairly constant and standard deviations were lower than a few mm. Field capacities estimated by probes 2 (1 m) and 6 (30 cm) are indicated by horizontal lines in Fig. 3. Table 1 shows that the difference in the amount of water stored to a depth of 1 m between the field capacity and the driest soil condition was about 100 mm and about half of this amount occurred above 30 cm. Annual average soil water content to a depth of 1 m except for the location of probe 1 ranged from 0.60 to $0.61 \text{ m}^3 \text{ m}^{-3}$ and that to a depth of 30 cm

Table 1. Amount of storage water (mm) to depths of 30, 60 and 100 cm

Probe number	1	2	3	4	5	6	7	8
Probe length (cm)	100	100	100	100	60	30	30	30
Field capacity (mm)	630	642	647	646	363	162	156	155
SD (mm)	4	2	3	3	1	3	2	2
Driest (mm)	497	537	543	548	284	109	111	107
Wettest (mm)	666	678	693	696	389	168	167	167
Annual average (mm)	580	603	609	611	335	144	140	140

SD: standard deviation of field capacity

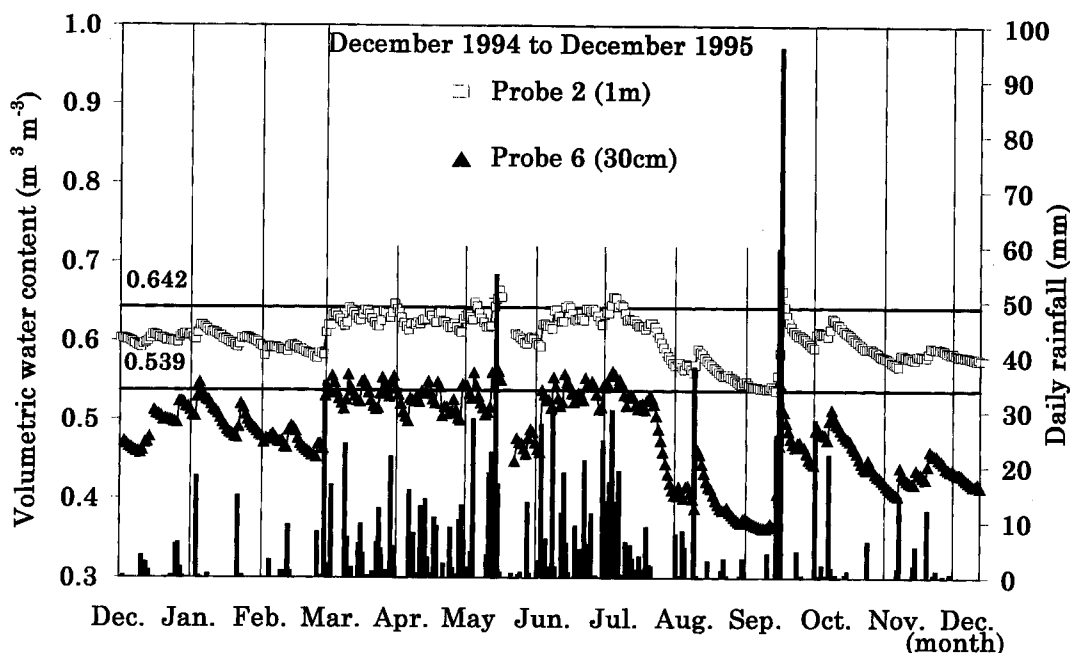


Fig. 3 Annual changes of soil water content to depths of 30 cm and 1 m. The horizontal lines and numbers along them indicate the soil water content at field capacity in each layer.

from 0.47 to 0.48 $\text{m}^3 \text{m}^{-3}$ indicating that spatial variability was not appreciable even though the elevation of the soil surface was not identical among the probes. The low values recorded consistently by probe 1 were difficult to explain but could be partly due to soil disturbance in the past or imperfect contact between soil and probe.

INCREASE OF AMOUNT OF STORAGE WATER BY RAINFALL

The amount of water stored in the soil as a result of rainfall is influenced by the initial soil-water content and the intensity and quantity of rainfall. When the soil is dry and the intensity of rainfall is low, rainwater infiltrates uniformly into the soil matrix increasing the soil-water content. Until the wetting front reaches a depth of 1 m, the

increase of the amount of water stored in the profile is equal to the amount of rainfall. When the intensity of rainfall exceeds the infiltration rate of water into the soil, surface runoff to the depressions occurs and the amount of water infiltrating there may exceed the amount of rainfall. Furthermore, water ponded on the surface is likely to flow rapidly through macropores bypassing the soil matrix. If bypass flow occurs through the profile, the amount of water infiltrating exceeds the increase in the amount of stored water.

It has been assumed that rainfall in the range between 0.1 and 10 mm was absorbed completely by the soil matrix. The total amount of such rain was 223 mm, 18.7% of the annual precipitation. When the rainless period exceeded four hours, rain events were classified separately. The 36 instances when rainfall exceeded 10 mm were classified

Table 2. Classification of rainfall type

Type of Rainfall	Frequency	Sum (mm)
<10mm h ⁻¹ to dry soil	25	524
>10mm h ⁻¹ to dry soil	5	134
<10mm h ⁻¹ to wet soil	4	90
Large amount	2	221
	36	969

into four types (Table 2) to determine whether the amount of rainfall was equal to the increase in the amount of stored water. Probe 2 (1 m) was used mainly to investigate the infiltration characteristics where the field capacity was 642 mm.

LOW INTENSITY RAINFALL TO DRY SOIL

Figure 4 shows the changes in the amount of stored water and cumulative rainfall with time for two rain events. The amount of water stored to a depth of 1 m before the first rain event was 583 mm, 60 mm less than field capacity (Table 1). The total amount of 54 mm rain applied to the soil increased the element of stored water by 56 mm. Considering that a 1% error of 1 m TDR probe is equivalent to 10 mm water, the increase of stored water esti-

mated the amount of rainfall fairly well. Figure 4 also indicates that deep percolation beyond 1 m depth did not occur after the first rain event. Soil water content measured with TDR was lower in the daytime, but daily changes of the soil water content between March 2 and 4 were less than 0.004 m³ m⁻³. These fluctuations may be due to the low relative dielectric constant of water at higher soil temperature as pointed out by Zegelin *et al.* (1992) and Pepin *et al.* (1995).

In 25 out of 36 rain events, the intensity was lower than 10 mm h⁻¹ and the initial amount of stored water was below field capacity. The total amount of such rain was 524 mm, 44.0% of the annual total. Figure 5 shows the relations between rainfall measured by the rain gauge and that estimated by the increase of the amount of stored water using probes 2 and 4 for 25 rain events. The plots of probe 4 were more scattered than those of probe 2. Linear regression analysis indicates that amount of rainfall could be predicted well by the TDR method because of the small standard error of the gradients as shown in Table 3. The difference in the gradient among the probes cannot be explained readily.

HIGH INTENSITY RAINFALL

In 6 rain events, the intensity exceeded 10 mm h⁻¹. The instance of a large amount of rainfall brought by a typhoon, will be discussed later and, was excluded here.

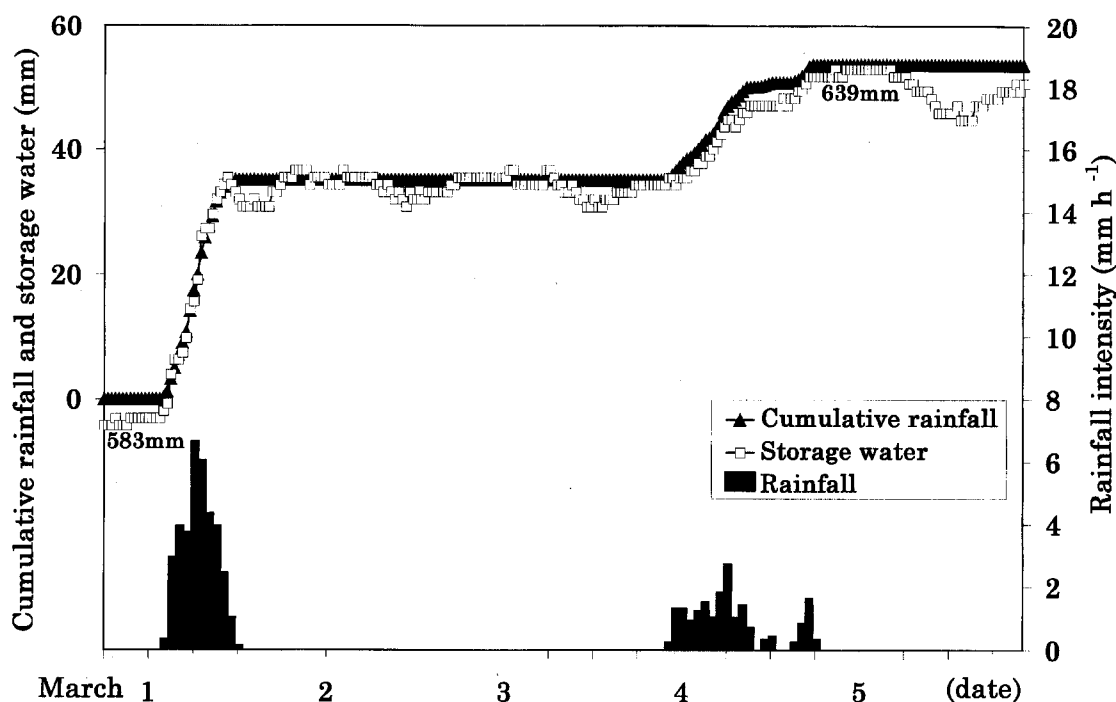


Fig. 4 Cumulative rainfall and increase of the amount of storage water to a depth of 1 m associated with low intensity rains on dry soil (probe 2). Figures indicate the amount of storage water.

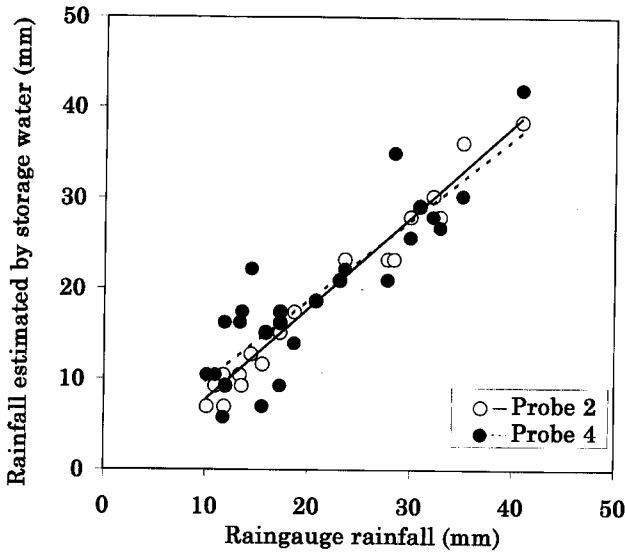


Fig. 5 Regression between rain gauge rainfall and rainfall estimated by the increase of the amount of storage water.

Soil-water content after the rain was lower than the field capacity except for the rain event on 1 July. Figure 6 shows an example of the cumulative rainfall and the increase of the amount of stored water measured with probes 2 and 4 on 3 and 4 June. A total rainfall of 30.1 mm at a maximum intensity of 16 mm h⁻¹ fell but the amount

Table 3. Linear regression between rain gauge rainfall and the amount of storage water to a depth of 1 m determined by TDR

Probe	Linear regression
1	$y = (0.90 \pm 0.10)x + 0(\pm 2)$
2	$y = (1.01 \pm 0.04)x - 2(\pm 1)$
3	$y = (0.96 \pm 0.07)x + 0(\pm 1)$
4	$y = (0.90 \pm 0.10)x + 1(\pm 2)$

x = rain gauge (mm)
 y = TDR (mm)

of stored water did not exceed the field capacity. Probe 4 installed at the lower elevation absorbed 1.5 times more water than probe 2 installed at the higher elevation.

As the hydraulic conductivity of the surface soil at -3 to -4 cm matric potential measured in the laboratory was about 80 mm h⁻¹ (Hasegawa *et al.*, 1994), formation of a soil crust by the impact of rain drops must cause surface runoff under high-intensity rainfall resulting in the infiltration of more water in the lower parts such as in the case of probes 3 and 4. This phenomenon was recognized by Parkin *et al.* (1995) using a rainfall simulator. They stated that the spatial variability in local infiltration rates may be due to an occasional high-intensity pulse of rainfall leading to temporary ponding and redistribution of water at the soil surface. Under vegetation, Saffigna *et al.* (1976),

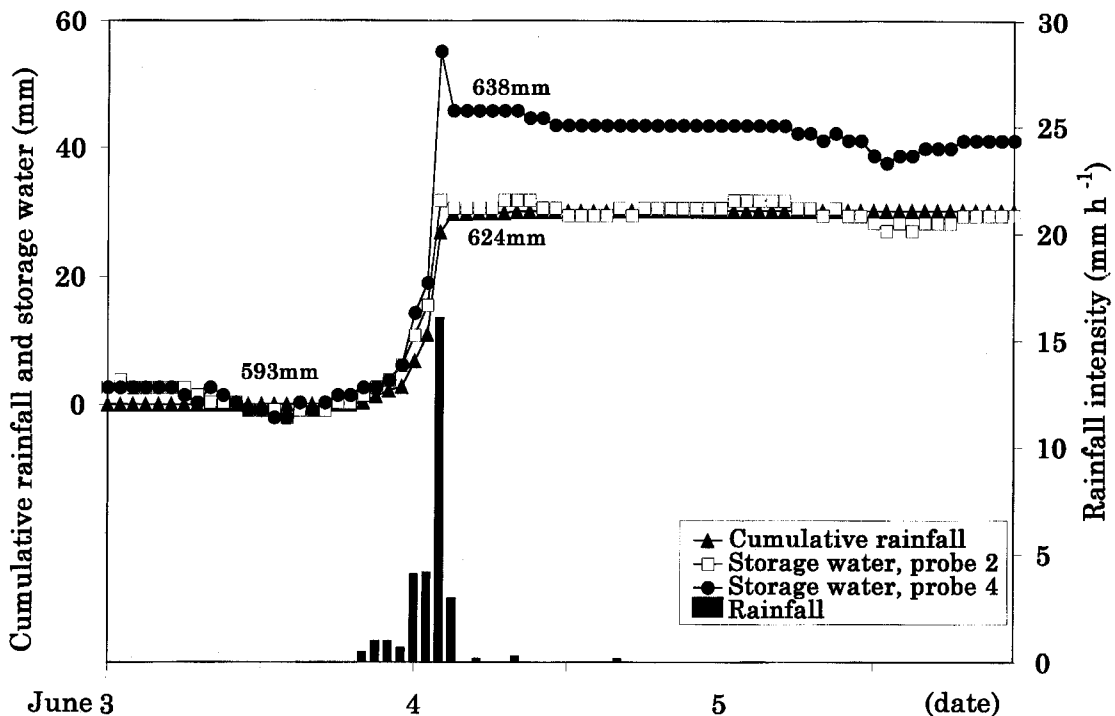


Fig. 6 Cumulative rainfall and increase of the amount of storage water to a depth of 1m associated with high intensity rain. Figures indicate the amount of storage water.

based on dye-tracing, reported that foliage interception and ridging of a potato field caused non-uniform infiltration and van Wesenbeek *et al* (1988) used measurements of the soil water content by TDR showing that the rainfall recharge was different between row and inter-row of maize crops because of leaf interception and subsequent stem flow. Non-uniform infiltration caused by ponding at depressions and/or canopy interception may occur under field conditions even though no surface runoff occurs.

Table 4 summarizes the amount of water stored to depths of 30 cm and 1 m at high intensity rainfall. Probes 2 and 6 were installed at higher elevations whereas probes 4 and 8 were installed at lower elevations as shown in Fig. 1. The standard deviation of the mean indicates large spatial variation among the probes, of the order of 20 to 30% in the values of the water stored. Lower parts received 1.4 to 1.6 times more rainfall than the higher parts. Furthermore, most of the rainfall was absorbed and stored in the surface 30 cm layer, and bypass flow might not occur in the sub-soil. Although water movement through the soil matrix obeys Darcy's law, it is difficult to simulate soil water movement by solving a flow equation at farm level due to local differences in water application rates by rainfall. It is also important to note that the annual average soil water contents in the case of probes 2, 3 and 4, as shown in Table 1, were similar even though the amount of water infiltrating varied considerably under high intensity rainfall.

LOW INTENSITY RAINFALL TO WET SOIL

When soil to a depth of 1 m was wet due to previous rainfall and low intensity rainfall continued for a long period of time, deep percolation below 1 m occurred. Figure 7 illustrates a measurement by probe 2 during the period from 3 to 9 July. Soil water reached field capacity on the afternoon of 3 July. The increase in the amount of water stored after 4 July did not coincide with the cumulative rainfall due to percolation beyond 1 m. The rate of deep percolation below 1 m for 24 hours after noon on 6 July may be calculated as 0.7 mm h⁻¹ if evaporation from the soil surface is neglected. Rainfall on 8 July was also underestimated by TDR due to deep percolation. A total of 90.2

mm of rain falling on wet soil during the year was not estimated by TDR. However, the intensity of such rainfall was apparently low enough to enable water to be absorbed by the soil matrix. Bypass flow was less likely to occur under such conditions.

LARGE AMOUNT OF RAINFALL

There were two large rainfall events in the year. In one, 65.4 mm of rain fell on from 15 and 16 May following 40.8 mm on 12 and 13 May. In the other, 155.2 mm fell as a result of a typhoon on 16 and 17 September. Figure 8 shows the cumulative rainfall and the amount of water stored to a depth of 1 m measured by probe 2 when the typhoon struck. As 14 September was the driest day of the year, a large amount of water was absorbed by the soil. The amount of water stored increased with the cumulative rainfall up to 678 mm. This value is considered to be the maximum water stored to a depth of 1 m and succeeding rainfall must flow away by surface runoff or percolate as bypass flow without appreciable increase of the soil water content. The amount of stored water then decreased rapidly after the end of rainfall by deep percolation.

The difference between the amount of rainfall and the amount of storage water was 40 mm. In the case of the 65 mm rainfall on 15 and 16 May, the difference was 25 mm. Therefore, a total of 65 mm of rain did not increase the soil water content. Surface runoff might carry away a certain amount of rain. Under near saturation conditions, bypass flow is likely to occur through continuous macropores whose volume is very small compared to the pore volume in the soil matrix.

The amount of bypass flow was, definitely, less than 65 mm or 5.5% of the annual rainfall when the surface runoff was subtracted. This must be characteristic of the volcanic-ash soil unlike other fine textured soils. Little shrinkage occurs and, in any event, the lack of long dry spells impedes the development of large drying cracks. A soil matrix having high saturated and unsaturated hydraulic conductivity conducts infiltrating water rapidly and has a chance to absorb water flowing through the macropores into the soil matrix. These must be major reasons that

Table 4. Increase of the amount of storage water by high intensity rainfall

Date	Rainfall Intensity		Amount of storage water at locations where probes (1 to 8) were installed (mm)							
	(mm)	(mm h ⁻¹)	1	2	3	4	6	7	8	Mean
3 to 4 June	30.1	16.0	39	32	40	44	28	28	40	35.9 ± 6.4
1 July	24.5	16.0	21	22	32	33	19	19	29	25.0 ± 6.1
10 August	37.9	33.1	40	36	67	78	—	44	59	54.0 ± 16.7
14 to 15 Sept.	25.4	10.6	36	23	32	32	20	20	23	26.6 ± 6.6
7 to 8 Nov.	15.9	11.1	33	15	30	27	16	15	28	23.4 ± 7.8
Total	133.8		169	128	201	214	—	126	179	169.5 ± 36.5

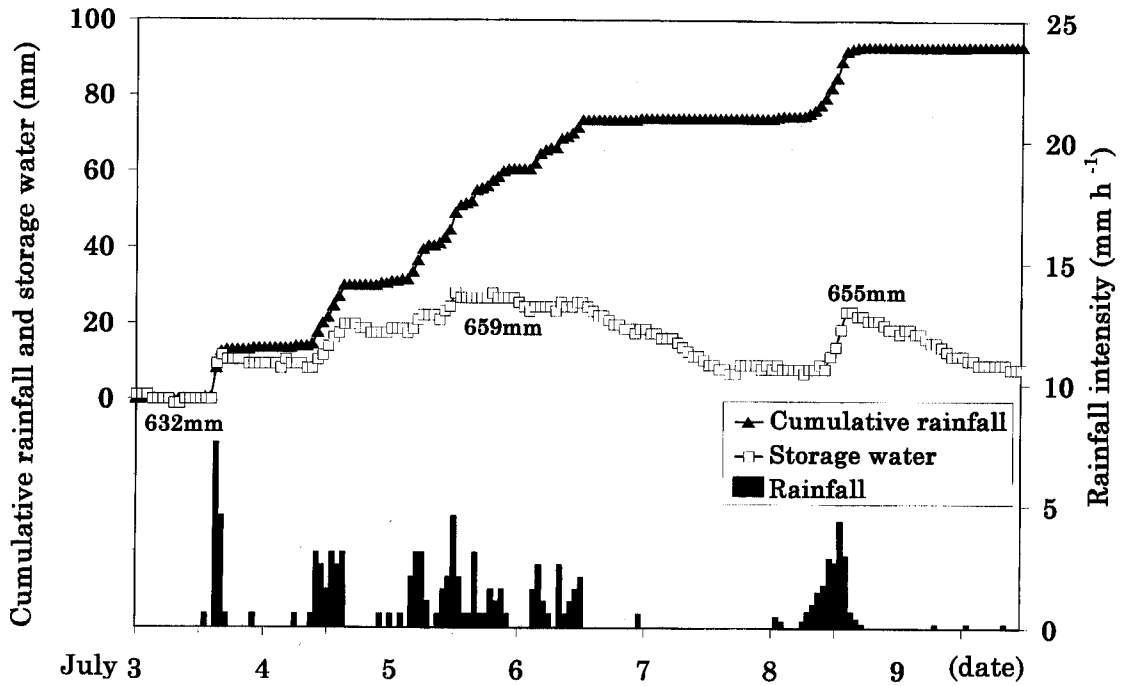


Fig. 7 Cumulative rainfall and increase of the amount of storage water to a depth of 1m associated with low intensity rain on wet soil (probe 2). Figures indicate the amount of storage water.

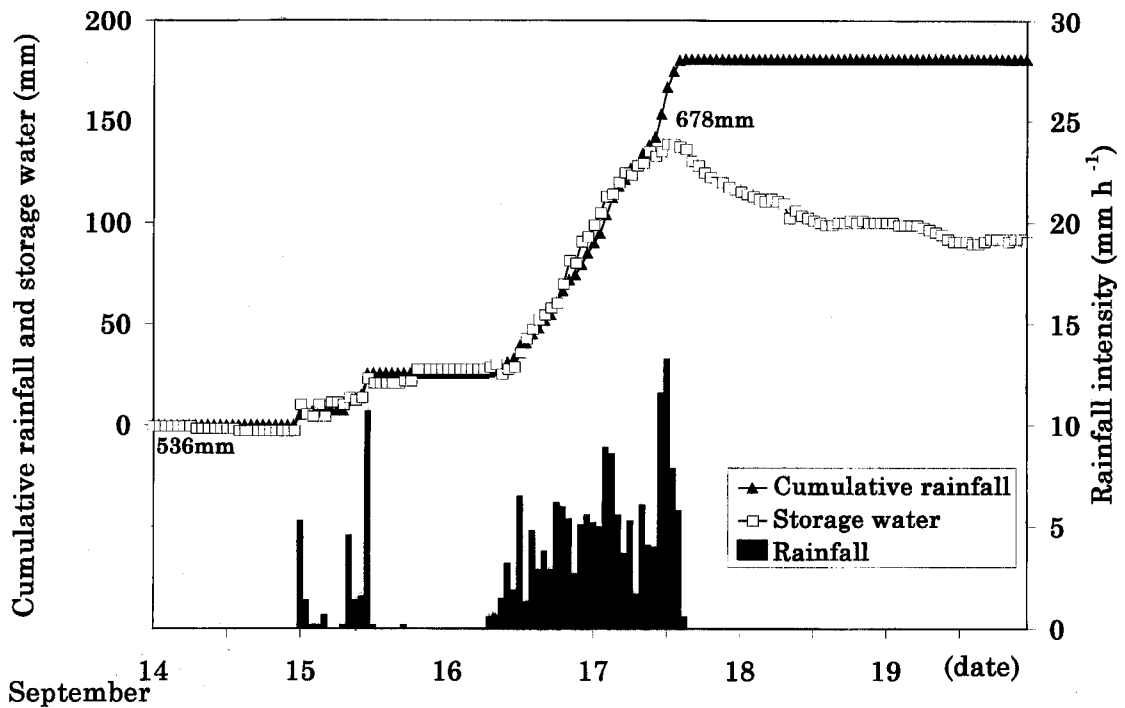


Fig. 8 Cumulative rainfall and increase of the amount of storage water to a depth of 1m associated with a large amount of rain (probe 2). Figures indicate the amount of storage water.

bypass flow seldom occurred for the volcanic ash soil. Deep percolation is influenced strongly by macropores bypassing the soil matrix as reviewed by Beven and German (1982). However, quantitative estimation of matrix and bypass flow to the annual rainfall has not been well investigated. Bouma and de Laat (1981) assumed that bypass flow occurred at a rate of 20% for a daily precipitation exceeding 10 mm and that it accounted for 16% of annual rainfall in a field with swelling clay soils. The results obtained here in a volcanic ash soil were very different from those of swelling clay.

Conclusion

Soil water conditions and movement were studied using the TDR method in a bare field consisting of volcanic ash soil. Storage water to depths of 30, 60 and 100 cm, monitored by TDR, enabled analyses of the soil water conditions such as field capacity, water stored in the root zone and intensity and duration of dry spells. Comparison of the amount of rainfall obtained by TDR and a rain gauge enabled characterization of the rainfall infiltration. Low intensity rain falling on dry soil increased the amount of water stored in proportion to the cumulative rainfall measured with a rain gauge. When the intensity of rainfall was higher than 10 mm h⁻¹, non-uniform infiltration influenced by the surface micro-relief, caused more water to infiltrate in depressions than in convex parts due to temporal ponding. Non-uniform infiltration was considered to influence the depth of penetration of rainwater, but the annual average soil water contents were not affected by the surface micro-relief. A large amount of rainfall exceeding 100 mm including previous rainfall must cause surface runoff and/or bypass flow. Even though flooding water percolated through macropores, the contribution of bypass flow was estimated to be less than 5.5% of the annual rainfall. Unlike other clayey soils, the high hydraulic conductivity of the soil matrix of the volcanic ash soil must suppress bypass flow. One year's monitoring of soil water proved that non-uniform infiltration was common but bypass flow was exceptional for the soil in this study.

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