WATERSHED RESPONSE TO A STORM RAINFALL

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WATERSHED RESPONSE TO A STORM RAINFALL

by

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ABSTRACT

A small forested watershed with a flat valley bottom and the soils of low hydraulic conductivities was instrumented to evaluate a role of subsurface flow (subsurface stormflow and groundwater flow) in stormflow generation. Under these situations, saturation overland flow has been considered to become a predominant contributor to stormflow generation in the previous studies.

As the result of the study, however, groundwater flow was found to play a predominant role in stormflow generation. A quick response of the groundwater discharge to the rainfall was not able to be accounted for by the Darcy equation used in the customary way. The observed quick response of the groundwater discharge to the rainfall was considered to be caused by a conversion of the capillary fringe into the pressure-saturated zone.

Only the rain water brought on the lower and middle parts of the hillslope, where the groundwater body and the capillary fringe existed, could generate stormflow. The rain water brought on the upper and ridge-top parts of the hillslope gradually moved downslope and could contribute to stormflow generation induced by the forthcoming storm.
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1.1 Review of recent studies

In this study, 'stormflow' is defined as the storm-induced streamflow from the time when stream discharge begins to increase by an onset of a rainfall to the time when stream discharge is almost return to its pre-storm level after a rainfall. The period of 'storm event' is equal with that of the stormflow.

In recent years, considerable advances have been made in our understanding of stormflow generation in small forested watersheds in humid regions. Much effort has concentrated on an examination of the contributors which caused a quick response of stream discharge to a rainfall. Four contributors by which rain water appears quickly as stream discharge are now recognized by field evidences. These contributors are (I) overland flow, which consists of (1) Hortonian overland flow and (2) saturation overland flow, and (II) subsurface flow, which consists of (3) subsurface stormflow and (4) groundwater flow.

Hortonian overland flow occurs from 'partial source area' where ponding water on ground surface is generated due to saturation of surface soils by a rainfall. The size of the partial source area is controlled by the distribution of soil types. Saturation overland flow,
which is generated on 'variable source area' adjacent to a stream channel, occurs when saturation of ground surface is attained by a rise of groundwater table. In the previous studies, a stream channel was also included in the variable source area. Saturation overland flow is a combination of (a) return flow from saturated surface soils and (b) rain water that falls directly onto the variable source area. The control on the size of the variable source area is topographic and hydrological configurations of a hillslope. The third contributor is subsurface stormflow which delivers water quickly into a permanent stream channel or into an expanding intermittent channel network in the variable source area. Subsurface stormflow occurs mainly in saturated soil zones above water-impeding layers, especially in basal hillslope soils, and subsurface stormflow discharges laterally into a stream channel without entering a groundwater body. The fourth contributor is groundwater flow which discharges into a stream channel from a groundwater body.

The path by which rain water reaches a stream channel depends upon such conditions as climate, topography, soil characteristics, and vegetation. In various parts of the world, therefore, one might expect that different contributors generate stormflow or at least that the relative importance of four contributors
might vary geographically. From this viewpoint, much effort is currently paid to define the situations in which each contributor is important.

As the results of the previous studies, it is now widely accepted that Hortonian overland flow rarely occurs in undisturbed forested watersheds in humid regions (e.g., Mosley, 1979; Yasuhara, 1980).

Most of recent studies in humid regions emphasize the importance of stormflow generation on the variable source area, that is, saturation overland flow (e.g., Dunne and Black, 1970a, b; Freeze, 1974). As saturation overland flow has an extremely high flow rate and its volume is dependent on rainfall intensity, this flow is considered to wholly account for a quick response of streamflow to a rainfall.

Using the results of Whipkey (1965) and Weyman (1970), Dunne and Black (1970a, b) and Freeze (1972) doubted that subsurface stormflow, which is the flow through soils of low hydraulic conductivities, could deliver sufficient water to provide quickly significant contributions to stormflow generation. On the basis of the mathematical simulations, Freeze (1972) concluded that "only on convex hillslopes that feed deeply incised channels, and then only when the saturated hydraulic conductivities of soils are very large, is subsurface stormflow a feasible mechanism."
Present thinking has neglected a role of groundwater flow in stormflow generation (e.g., Dunne and Black, 1970a, b; Freeze, 1974). Groundwater flow is located generally in deeper soils of lower saturated hydraulic conductivities as compared with soils in which subsurface stormflow occurs. Therefore, it is not thought to be able to deliver sufficient water quickly to contribute to stormflow generation as long as the Darcy equation is used in the customary way. A hydraulic gradient in the Darcy equation has been determined traditionally based on a difference of water levels at two wells in a downslope direction. Freeze (1974) summarized the hydrologic thought as "True groundwater flow is seldom the cause of the major runoff during storms. Its primary role is in sustaining streams during low-flow periods between rainfall and snow-melt events."

Contrary to the results of the above mentioned studies, hydrograph separations based on the tracer experiments in the last decade have demonstrated that subsurface flow (subsurface stormflow and groundwater flow) could play a predominant role in stormflow generation (Martinec, 1975; Sklash and Farvolden, 1979). It is difficult, however, to conceptualize physically how subsurface flow can respond quickly enough to contribute to stormflow generation.

Jones (1971) and Mosley (1979) have indicated that subsurface flow via macropores (root channels and pipes)
in soils of low hydraulic conductivities caused a quick response of stream discharge to a rainfall. But it is doubtful that such macropores exist or continue along a long distance in soils even in all other forested watersheds.

On the other hand, Ragan (1968) and Sklash and Farvolden (1979) have suggested that rapid increases in the hydraulic head in near-stream groundwater (groundwater ridging phenomenon) might make a significant contribution of groundwater flow to stormflow generation. However, as long as the Darcy equation is used in the customary way, it is not convincing that their observed small increases in the hydraulic gradient are sufficient to account for stormflow generation quantitatively.

The quick and much contribution of subsurface flow to stormflow generation as indicated by the results of the tracer experiments is not able to be account for physically at present. Therefore, the results of the tracer experiments have been doubted and the role of subsurface flow in stormflow generation is not generally thought to be important.

1.2 Purpose of the study

Most recent studies in humid regions have given saturation overland flow to a primary role and made little of the contribution of subsurface flow in stormflow
generation. Although the process of a quick response of subsurface flow to a rainfall has not been made clear yet, there is a possibility of subsurface flow being an important contributor to stormflow generation as stressed by the results of the tracer experiments.

In recent years, Bonell and Gilmour (1978) and Yasuhara (1980) discussed that subsurface stormflow in a very permeable humus layer was most important for stormflow generation. A generation of subsurface stormflow in a humus layer has been overlooked in the previous studies. Moreover, a rapid increase of the groundwater recharge in response to a rainfall has been reported even in the fine-textured soils such as volcanic ash soils called Kanto Loam (Shimada, 1982; Kaihotsu, 1982). From these viewpoints, it is necessary to re-evaluate the role of subsurface flow in stormflow generation in forested watersheds.

The purpose of this paper is three-fold. It evaluates the importance of subsurface flow in stormflow generation based on the behavior of subsurface water (soil water and groundwater) during a storm event. Secondly, it clarifies the process which makes possible rain water or subsurface water to reach a stream so quickly and in such large quantities as to contribute to stormflow generation. Thirdly, it makes clear the role of rain water brought on each segment of the hillslope in stormflow generation.
II STUDY AREA AND OBSERVATION METHODS

2.1 Description of the study area

An experimental basin, which occupies about 2.2 ha of forested land between the 145- and 183-m elevation contours, is located in the western suburbs of Tokyo, Japan (Fig. 1). The topography is typical of a head-water basin in the Tama Hills having a steep hillslope with an angle of 25° and a flat valley bottom with an angle of 7°. The vegetation consists of sparse deciduous trees approximately 15 m high and dense bamboos 1 to 2 m high with a dense ground cover of ferns and small shrubs. Mean annual rainfall for the period between 1972 and 1981 was 1568 mm at Hachioji located 5 km north-west of the study area.

The basin is underlain by the impermeable clay or compacted clayey sand at depth of about 1 m below the ground surface. Judging from the study of Juen and Harada (1961), these clay and compacted clayey sand must be upper part or weathered materials of the Pleistocene Renkoji Alternation. By the existence of the impermeable materials below soils almost over the basin, subsurface water in the Renkoji Alternation is not considered to contribute to stormflow generation.

Soils are classified as the surface soil (humus) and the transported soil. The surface soil is interlaced
Fig. 1  Map of the experimental basin.
with old root holes, worm holes, and structural channels which seem to provide conditions that are conducive to subsurface stormflow. The transported soil, which is assumed to have been transported from upper parts of the hillslope by a action of the mass movement, consists of a mixture of Kanto Loam (volcanic ash soil) and masses of dark yellow clay.

There exist streams in northern and southern sides of the valley bottom (Fig. 1). This study investigated stormflow generation in the southern stream channel, which is 50 to 80 cm deep and 40 to 100 cm wide during a storm event. The stream bed is located on the impermeable clay mentioned above.

Within the basin, one hillslope with nearly straight contour lines was chosen for an intensive study on the relationship between behavior of subsurface water and stormflow generation (Figs. 1 and 2). Hack and Goodlett (1960) and Troeh (1964) have suggested that movement of subsurface water was controlled by shapes of hillslopes, that is, concave, convex, or straight in a plan form. They concluded that there existed no exchange of subsurface water between a straight hillslope and adjacent slopes. Consequently, only rain water brought on this experimental hillslope is expected to concern stormflow generation occurring in the lower part of the hillslope.
Fig. 2 Plan of the experimental hillslope and instrumentation.
Soil profiles in the experimental hillslope are shown in Fig. 3. Four columns show the soil profiles in the lower (TR1), middle (TR2), upper (TR3), and ridge-top (T-4) parts of the hillslope. Locations of the four profiles are shown in Fig. 2. The most parts of the soil profiles are occupied by the transported soil (TS) except for the ridge-top part of the hillslope. The transported soil is composed of much loam and a little clay in the upper parts of the soil profiles. In the lower parts of the soil profiles, however, it contains a lot of masses of clay and even some gravels in addition to the loam.

Saturated hydraulic conductivities in the experimental hillslope are also shown on right-sides of the soil profiles in Fig. 3. These values are in the order of $10^{-4}$ to $10^{-8}$ cm/s except for those of the surface soil. Similar low values of the saturated hydraulic conductivities have been reported also in the valley bottom by Sakai (1981). Because of lack of masses of clay in soils, the values in the ridge-top part (T-4) are larger than those in the other parts of the hillslope.

Soil-water characteristic curves for soil samples at various depths of the lower, middle, upper, and ridge-top parts of the hillslope are represented in Figs. 4, 5, 6, and 7, respectively. These curves will be used later to obtain volumetric water contents at various
Fig. 3  Soil profiles with saturated hydraulic conductivities and positions of troughs (TS: transported soil).
Soil-water characteristic curves for soil samples at various depths of Trench TR1.
Fig. 5  Soil-water characteristic curves for soil samples at various depths of Trench TR2.
Fig. 6  Soil-water characteristic curves for soil samples at various depths of Trench TR3.
Fig. 7  Soil-water characteristic curves for soil samples at various depths of Tensiometer plot T-4.
depths through the hillslope.

The representative extents of the groundwater body along the line of section in Fig. 2 before and at the peak discharge of the storm of Oct. 22, 1981 are indicated as A and B in Fig. 8, respectively. Although the wedge-shaped groundwater body existed in the transported soil only in the lower part of the hillslope before the storm, it extended to the middle part of the hillslope at the peak of the storm discharge.

Marui (1982) has reported that groundwater table in the most part of the flat valley bottom existed at the ground surface during a storm, where saturation overland flow was generated. Taking account of this fact and extremely low hydraulic conductivities of the soils, the situations of the experimental basin are considered to be most suitable for saturation overland flow to become a predominant contributor to stormflow generation based on the discussion of Freeze (1974).

2.2 Instrumentation

The hydrologic responses of the experimental hillslope to the storm rainfalls of Aug. 22-23, 1981 and Oct. 22, 1981 were monitored intensively using a variety of equipments as shown in Fig. 2. Streamflow was measured continuously by a 5-inch Parshall flume equipped with a water level recorder. Precipitation was measured by a
Fig. 8  Cross section of lower part of the hillslope.
20-cm diameter tipping bucket recording rain gauge located in the middle part of the valley bottom. Saturated area at the ground surface, where saturation overland flow was generated, was identified by the field observations during storm events.

Measurements of the groundwater level were made at 23 observation wells shown in Fig. 2. The depths of these observation wells are listed in Table 1. Each observation well consists of a 7.5-cm PVC pipe which is perforated at the lower 20-cm length and covered with a fine mesh screen. After PVC pipes were placed in holes hand-augered to the desired depths, the holes were backfilled with gravel and bentonite. These many wells with different depths were used to monitor generation and behavior of the saturated zone at various depths throughout the hillslope. The perennial groundwater body was observed only at Wells W1-c and W1-d in the hillslope, and the groundwater level at Well W1-c was measured automatically by a water level recorder.

Soil-water pressure heads were measured at five plots (Fig. 2) by using 30 tensiometers with 1.8-cm diameter porous cups. The depths of the tensiometers are listed in Table 2. Five plots are located each in the valley bottom, the lower, middle, upper, and ridge-top parts of the hillslope. Readings of the pressure head were made simultaneously with measurements
Table 1  Depths of observation wells.

<table>
<thead>
<tr>
<th>OBSERVATION WELL</th>
<th>DEPTH (cm)</th>
<th>OBSERVATION WELL</th>
<th>DEPTH (cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>W0 - a</td>
<td>120</td>
<td>W2 - b</td>
<td>60</td>
</tr>
<tr>
<td>b</td>
<td>190</td>
<td>c</td>
<td>95</td>
</tr>
<tr>
<td>c</td>
<td>50</td>
<td>d</td>
<td>155</td>
</tr>
<tr>
<td>W0' - a</td>
<td>110</td>
<td>W3 - a</td>
<td>25</td>
</tr>
<tr>
<td>b</td>
<td>230</td>
<td>b</td>
<td>60</td>
</tr>
<tr>
<td>WA - 3'</td>
<td>190</td>
<td>c</td>
<td>100</td>
</tr>
<tr>
<td>W1 - a</td>
<td>25</td>
<td>d</td>
<td>158</td>
</tr>
<tr>
<td>b</td>
<td>60</td>
<td>W4 - a</td>
<td>25</td>
</tr>
<tr>
<td>c</td>
<td>115</td>
<td>b</td>
<td>60</td>
</tr>
<tr>
<td>d</td>
<td>100</td>
<td>c</td>
<td>100</td>
</tr>
<tr>
<td>e</td>
<td>115</td>
<td>d</td>
<td>230</td>
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<tr>
<td>W2 - a</td>
<td>25</td>
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<td></td>
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</tbody>
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Table 2  Depths of tensiometers.

<table>
<thead>
<tr>
<th>TENSIOmeter PLOT</th>
<th>DEPTH (cm)</th>
</tr>
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<tbody>
<tr>
<td>T - 0</td>
<td>20, 40, 60</td>
</tr>
<tr>
<td>T - 1</td>
<td>10, 20, 40, 60, 80, 100</td>
</tr>
<tr>
<td>T - 2</td>
<td>10, 20, 40, 60, 80, 100, 120</td>
</tr>
<tr>
<td>T - 3</td>
<td>10, 20, 40, 60, 80, 100, 140</td>
</tr>
<tr>
<td>T - 4</td>
<td>10, 20, 40, 60, 80, 100, 140</td>
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</tbody>
</table>
of the groundwater level. Structures of the observation well and the tensiometer are described in detail by Tanaka et al. (1981).

In order to intercept subsurface flow through the hillslope, three trenches, that is, TR1, TR2, and TR3, were installed in the lower, middle, and upper parts of the hillslope, respectively. Each trench 2.0 m wide was dug down through the soils into the underlying impermeable clay or compacted clayey sand. A schematic cross section of the trench is represented in Fig. 9. In each trench, subsurface flow was collected by the 1.2-m long plastic troughs at three levels: the base of the surface soil, the base of the root zone, and the base of the transported soil. The troughs were denoted A, B, and C from the top. Positions of the troughs in each trench are indicated in Fig. 3. A metal sheet was inserted into the exposed soil surface at each trough to ensure that each drew its flow only from an expected layer of the soils. Intercepted water was led from each trough to a measuring cylinder and the volume was measured manually. Perennial seepage (groundwater discharge) was observed only at Trough TR1-C. Measurements of the groundwater discharge at Trough TR1-C have been made automatically by a 5° V-notch weir equipped with a water level recorder since April, 1982.

Atkinson (1978) described that the net of hydraulic
Fig. 9 Schematic cross section of trench.
potential on the slope was distorted by the installation of a trench and the trench received discharge from area which was not directly upslope from it. The magnitude of this distortion, however, is not known, and will presumably vary with circumstances. In this study, therefore, disturbance of the flow lines contributing to the discharge at the troughs was tried to be minimized by extending the widths of the trenches beyond the troughs by 40 cm at each side according to Mosley's (1979) recommendation.

On the other hand, the level of top of the seepage face in Trench TR1 was found to coincide always with the groundwater level at Well W1-c as shown in Fig. 8. As Well W1-c is more than 2 m aside from Trench TR1 (Fig. 2), the groundwater level observed at this well is not considered to be affected by the installation of Trench TR1. Therefore, it is not expected that undesirable changes in behavior of subsurface water due to the installation of Trench TR1 altered the measured discharge in the trench from the value before the installation of it.

A dye tracer experiment was carried out at Suction lysimeter plot S-1 in order to monitor behavior of subsurface water in the lower part of the hillslope during a storm (Fig. 10). Suction lysimeters (soil water samplers) installed at depths of 40 and 80 cm were used to extract soil water samples from four points after pouring the
Fig. 10  Plan of suction lysimeter plot.
fluorescent dye on the ground surface. The equipment, manufactured by Soilmoisture Equipment Corp. Ltd., U.S.A., is 4.8 cm in diameter and 60 cm in length. Structures and arrangements of the suction lysimeter and accessory items are displayed in Fig. 11. It took about 30 to 60 minutes to collect at least 10-mL soil water which is necessary for the fluorescence analysis. Dye concentrations in the soil water samples were determined accurately by using a spectrofluorometer of MPF-2A type manufactured by Hitachi Corp. Ltd., Japan.
pressure-vacuum hand pump, with vacuum dial gauge adapter

neoprene tube

pinch clamp

collected soil water sample

backfill

access tube

suction lysimeter (pressure-vacuum soil water sampler)

porous ceramic cup

Fig. 11 Structures and arrangements of both suction lysimeter and accessory items.
III EVALUATION OF STORMFLOW-PRODUCING CONTRIBUTORS

3.1 Role of overland flow in stormflow generation

In this section, a role of overland flow (Hortonian overland flow and saturation overland flow) in stormflow generation is discussed. Many storms of total rainfalls more than 30 mm, which storms generated considerable stormflows regardless of antecedent soil conditions, were recorded in the course of the study between June 1981 and August 1982. Among these storms, intensive observations were carried out during and after two representative storms of Aug. 22-23, 1981 and Oct. 22, 1981. The former was a 77.0-mm storm following an extremely dry antecedent condition and the latter was a 172.5-mm storm following a relatively wet antecedent condition.

Peak rainfall intensities of the two storms were 7.5 mm/h and 27.5 mm/h, which rainfall intensities induced the peak storm discharges of 2.1 l/s and more than 20 l/s at the 5-inch Parshall flume, respectively (Figs. 12 and 13). The 172.5-mm storm was the largest one and storms of total rainfalls more than 77.0 mm were observed three times in the course of the study.

The volumes of subsurface water held before the two storms were obtained by using the readings of the pressure head at Tensiometer plot T-1 and the soil-water characteristic curves shown in Fig. 4. They were 532.0 mm.
Fig. 12  Responses of streamflow and groundwater discharge to the storm of Aug. 22-23, 1981.
Fig. 13  Responses of streamflow and groundwater discharge to the storm of Oct. 22, 1981.
at 1200 h Aug. 20 and 557.0 mm at 0100 h Oct. 22 in the soil profile of the 115-cm thickness. The hydrologic responses during and after these two heavy storms under the different antecedent conditions are mainly discussed in this study.

There were no generations of Hortonian overland flow all over the basin even at the peak rainfall intensities of the two storms. Hortonian overland flow, therefore, was not found to be a contributor to stormflow generation in this experimental basin as often reported in other humid forested basins.

As the rainfall continued, groundwater table rose up and formed saturated area at the ground surface in the valley bottom. Both return flow (upward flow to the ground surface) and direct precipitation on the saturated area generated saturation overland flow. It has been already reported that upward flow to the ground surface existed in the valley bottom at the peak discharge of the heavy storm (Tanaka, 1982). The area, however, was restricted in the small saturated part of the valley bottom and noticeable amounts of return flow emerging from the saturated ground surface were not observed even at the peak discharges of the two monitored storms. Judging from these facts, return flow was not a major component of saturation overland flow in the experimental basin.
The maximum extent of the saturated area at the peak discharge of the 77.0-mm storm (0340 h Aug. 23, 1981) was mapped in Fig. 14. The saturated area also contains a stream channel. Since saturation overland flow from the valley bottom was not able to reach the stream as is obvious from Fig. 14, the saturated area contributing to stormflow generation was only the stream channel.

The peak rate of direct precipitation on the saturated area (stream channel), the other component of saturation overland flow, is therefore obtained as follows:

\[ 21.0 \text{ m}^2 \times 7.5 \text{ mm/h} = 157.5 \text{ l/h} \]

where 21.0 m\(^2\) is the observed maximum extent of the stream channel and 7.5 mm/h is the peak rainfall intensity. Assuming that there is an approximate equilibrium between the rainfall intensity and the streamflow, the calculated peak rate of 157.5 l/h is negligible in comparison with the peak storm discharge of 7560.0 l/h (2.1 l/h). Moreover, this rate is found to be too small to play a significant role in forming even a rising limb of the storm hydrograph in Fig. 12.

Supposing that the maximum extent of the stream channel did not change throughout the storm, the total volume of direct precipitation on the stream channel
Fig. 14  Extent of saturated area at 0340 h Aug. 23, 1981 at peak discharge of the 77.0-mm storm.
is calculated as follows:

\[ 21.0 \text{ m}^2 \times 77.0 \text{ mm} = 1.6 \text{ m}^3 \]

where 77.0 mm is the total rainfall of the storm of Aug. 22-23. The estimated total volume of 1.6 m$^3$ is also negligible compared with that of the total storm discharge of 69.0 m$^3$ from the time when streamflow began to increase (1400 h Aug. 22) to the time when streamflow was almost return to its pre-storm level (1200 h Aug. 24) in the bottom graph of Fig. 12. Moreover, as the direct precipitation on the saturated area disappeared suddenly after the rainfall, it could not participate in forming a falling limb of the storm hydrograph at all.

Although the maximum extent of the saturated area during the storm of Oct. 22, 1981 was not observed, the saturated area at 0200 h Sept. 12, 1980 at the peak discharge of the 194.5-mm storm was mapped in Fig. 15 by Sakai (1981). The total rainfalls and the peak rainfall intensities of both storms were nearly equal. Therefore, the maximum extent of the saturated area at the peak discharge of the storm of Oct. 22, 1981 is considered to be almost equal to that of Sept. 12, 1980. Taking account of micro-reliefs in the valley bottom, the extent of the saturated area which contributed to stormflow generation at the 5-inch Parshall flume is estimated as 356.0 m$^2$.

The peak rate and the total volume of direct
Fig. 15  Extent of saturated area at 0200 h Sept. 12, 1980 at peak discharge of the 194.5-mm storm.
precipitation on the saturated area are calculated in the same way as the case of the storm of Aug. 22-23, 1981, that is,

\[ \text{356.0 m}^2 \times 27.5 \text{ mm/h} = 9790.0 \text{ l/h} \]

and

\[ \text{356.0 m}^2 \times 172.5 \text{ mm} = 61.4 \text{ m}^3 \]

where 356.0 m² is the maximum extent of the saturated area, 27.5 mm/h is the peak rainfall intensity, and 172.5 mm is the total rainfall of the storm of Oct. 22. The peak rate of direct precipitation of 9790.0 l/h is too small to account for the peak storm discharge of more than 72000 l/h (20 l/s). Moreover, this rate is not sufficient to form even a large part of a rising limb of the storm hydrograph in Fig. 13. Compared with the total storm discharge of more than 852 m³ from 0700 h Oct. 22 to 1200 h Oct. 24, 1981 in Fig. 13, the total volume of direct precipitation of 61.4 m³ is not found to play a significant role in stormflow generation.

Judging from the results of the previous studies, the experimental basin was thought to be in the most suitable situations for saturation overland flow to become a predominant contributor to stormflow generation. However, the role of it in stormflow generation was found to be negligible or to be unimportant as mentioned
above.

3.2 Role of subsurface flow in stormflow generation

In this section, a role of subsurface flow (subsurface stormflow and groundwater flow) in stormflow generation is discussed. Subsurface stormflow rates measured at Troughs TR1-A and TR1-B in Trench TR1 are represented in Fig. 20 (at the peak discharge of the storm of Aug. 22-23) and Figs. 29 to 32 (during and after the storm of Oct. 22). Subsurface stormflow, which was observed to drip along some roots from unsaturated soil surface in the trench, was generated only just before and after the peak of the storm discharge. Moreover, in spite of an existence of the very permeable humus layer, subsurface stormflow rates were extremely too small to contribute to stormflow generation quantitatively.

There existed no perched saturated zone also in the valley bottom because of lack of a distinct impeding layer above the groundwater table. A significant contribution of subsurface stormflow to stormflow generation was, therefore, not expected from the valley bottom.

Consequently, the role of subsurface stormflow in stormflow generation was found to be negligible in the experimental basin. This result means that the other component of subsurface flow, that is, groundwater flow
must play a significant role in stormflow generation.

Response of the groundwater discharge at Trough TR1-C to the storm of Aug. 22-23, 1981 following a prolonged dry period is shown in Fig. 12 with the storm hydrograph at the 5-inch Parshall flume. The groundwater discharge was collected from the 120-cm wide saturated zone in the trench. It is noticeable that the shapes of two hydrographs are very similar. The groundwater discharge began to increase abruptly from 2330 h Aug. 22 after more than the 40-mm rain was falling and it decreased rapidly after the rainfall just in the same way as the streamflow. Especially, the most striking feature is a close coincidence of the times showing the peaks. In spite of the hillslope consisting of the soils with very low hydraulic conductivities, the groundwater discharge attained the peak soon after that of the rainfall intensity in the same way as the streamflow.

The hydrographs induced by the storm of Oct. 22, 1981 following a wet antecedent condition are represented in Fig. 13. Just like the case of the storm of Aug. 22-23, 1981, the shapes of two hydrographs are very similar. Both the groundwater discharge at Trough TR1-C and the streamflow attained the peaks around 2200 h Oct. 22 soon after the peak of the rainfall intensity. It is noticeable that the groundwater discharge began to increase quickly from its initial flow rate of 0.08 l/h.
to 0.11 1/h at 0800 h Oct. 22 when only the 4-mm rainfall was brought.

It was found that the groundwater discharge to the stream was so quick as to participate in generation of the storm hydrograph in response to the rainfall. In the next place, a quantitative role of groundwater flow in stormflow generation is discussed.

The peak groundwater discharges at Trough TR1-C (1.2 m wide) of 8.1 1/h at 0410 h Aug. 23 and 433.0 1/h at 2215 h Oct. 22 are selected as typical events in the basin. Length of an upstream reach from the 5-inch Parshall flume is about 37.0 m. Therefore, the groundwater discharges from both sides of the 37.0-m long upstream reach to the stream are calculated as follows:

$$8.1 \text{ 1/h} \times (37.0 \text{ m} \times 2) / 1.2 \text{ m} = 499.5 \text{ 1/h}$$

and

$$433.0 \text{ 1/h} \times (37.0 \text{ m} \times 2) / 1.2 \text{ m} = 26701.7 \text{ 1/h}$$

These peak rates of the groundwater discharge brought to the stream are 3.2 and 2.7 times larger than those of the direct precipitation on the saturated area and can account for about 6.6 % and 37 % of each peak storm discharge, respectively.

On the other hand, the total groundwater discharges to the stream during the storm events of Aug. 22-24 and
Oct. 22-24 are calculated respectively as follows:

\[ 85.9 \, \ell \times (37.0 \, \text{m} \times 2) / 1.2 \, \text{m} = 5.3 \, \text{m}^3 \]

and

\[ 3840.0 \, \ell \times (37.0 \, \text{m} \times 2) / 1.2 \, \text{m} = 236.8 \, \text{m}^3 \]

where 85.9 \ell and 3840.0 \ell are the total groundwater discharges measured at Trough TR1-C during each storm event. The periods for calculating the total groundwater discharge are identical with those for the total storm discharge in Section 3.1. The calculated total volumes of the groundwater discharge are 3.3 and 3.9 times larger than the total volumes of saturation overland flow and can account for 7.7 % and 27.8 % of each total storm discharge.

These percentages may be considered to be so small for determining groundwater flow as a predominant contributor to stormflow generation. However, it must be noticed that the location of Trench TR1 is about 5 m upslope from the stream. The groundwater body in the hillslope increased its thickness and upslope length considerably just near the stream as shown in Fig. 8. Moreover, test borings indicated that the extents of the groundwater body in the concave hillslope (west side of the experimental hillslope) and in the valley bottom were larger than that in experimental hillslope.

Much more contribution of groundwater to stormflow
generation was expected from these thicker and larger groundwater bodies contacting with the stream. In addition to this fact, there existed no significant contributors to stormflow generation except for groundwater flow as mentioned before. Therefore, there could be no problem for determining groundwater flow as a predominant contributor to stormflow generation quantitatively. Contrary to the results of the many previous studies such as Dunne and Black (1970a, b) and Freeze (1974), it was found that groundwater flow played a predominant role in stormflow generation even in the basin with the soils of extremely low hydraulic conductivities.
IV PROCESS OF QUICK RESPONSE OF GROUNDWATER DISCHARGE TO RAINFALL

4.1 Changes in groundwater level and groundwater discharge

It was made clear in Chapter III that groundwater flow could play a predominant role in stormflow generation. In this section, the process which makes possible for groundwater to reach the stream so quickly and in such large quantities as to contribute to stormflow generation is discussed based on relationships between the behavior of groundwater and the groundwater discharge at Trough TR1-C.

Figure 16 represents the hydrologic response in the lower part of the hillslope during and after the storm of Aug. 22-23, 1981. Changes in the groundwater level at Well W1-c located near Trench TR1 were not similar to those in the groundwater discharge. The maximum height of the groundwater table, which was attained at least at 1600 h Aug. 24 more than a day after the rainfall, was not corresponding to the peak of the groundwater discharge.

As represented in Fig. 17, the same phenomenon was also observed in the case of the storm of Oct. 22, 1981 following a wet antecedent condition, although changes in the groundwater level and the groundwater discharge were almost similar. Namely, the groundwater discharge
Fig. 16 Hydrologic response in lower part of the hillslope to the storm of Aug. 22-23, 1981.
Fig. 17  Hydrologic response in lower part of the hillslope to the storm of Oct. 22, 1981.
continued to decrease rapidly from 1600 h Oct. 23 to 1000 h Oct. 24 in spite of the fact that the groundwater level at Well W1-c was almost constant.

It was, therefore, made clear that changes in the groundwater level were not completely coincident with those in the groundwater discharge. This fact raises a doubt as to an application of the Darcy equation in the customary way to an evaluation of the groundwater discharge.

This problem becomes more obvious by taking account of changes in the groundwater level at Well W0-c just near the stream and expressing the groundwater discharge in terms of the flux as shown in Fig. 18. Figure 18 shows changes in the measured and calculated fluxes of the groundwater discharge at Trough TR1-C during and after the storm of Oct. 22, 1981. Both fluxes have dimensions of volume per unit time per unit area ($m^3/h/cm^2$).

The measured flux of the groundwater discharge is obtained by dividing the measured groundwater discharge at Trough TR1-C by an area of the seepage face in Trench TR1. The width of the seepage face was always 120 cm and its vertical length was assumed to be equal to the thickness of the groundwater body at Well W1-c. It must be noticed here that no observable concentrating flows such as pipe flows existed at the seepage face even at
Fig. 18  Comparison between measured flux at Trough TR1-C and calculated flux by the Darcy equation used in customary way during and after the storm of Oct. 22, 1981.
the peak of the groundwater discharge.

The calculated flux is obtained by using the Darcy equation in the customary way, that is,

\[ q = k \cdot i \]  

(1)

where \( q \) is the calculated flux of the groundwater discharge, \( k \) is the constant of proportionality, or the saturated hydraulic conductivity of the soils, and \( i \) is the hydraulic gradient. The saturated hydraulic conductivity of \( 2.3 \times 10^{-6} \) cm/s used for the calculation is the value obtained at the depth of 70 cm in Trench TR1 (Fig. 3). As the groundwater table rose to near the depth of 70 cm only at the peak discharge of the storm of Oct. 22 (Fig. 17), use of this value for the calculation throughout the storm event results in overestimated groundwater fluxes. According to the customary way, the hydraulic gradients are calculated only based on the observed groundwater levels at Wells W1-c and W0-c in a downslope direction.

Changes in the measured flux of the groundwater discharge are found to be closely coincident with changes in the rainfall intensity. On the other hand, changes in the calculated flux obtained by using the Darcy equation in the customary way are not coincident with changes in the rainfall intensity. The calculated flux begins to decrease with an increase of the rainfall intensity.
because the hydraulic gradient decreases due to a more rapid rise of the groundwater level at Well W0-c than that at Well W1-c.

The measured flux is obviously much larger than the calculated flux. The difference is more than 4 orders of magnitude at the peak of the groundwater discharge.

As structures of a soil sample might be destroyed at the time of sampling, the saturated hydraulic conductivity at the depth of 70 cm before sampling can be larger than \(2.3 \times 10^{-6}\) cm/s. No large structures such as pipes and cracks were, however, observed in the transported soil in Trench TR1. Therefore, the actual saturated hydraulic conductivity at the depth of 70 cm is not considered to be up to 4 orders of magnitude larger than the conductivity used for the calculation.

These results indicate that the Darcy equation used in the customary way can not account for the measured groundwater discharge at all and extremely underestimates the actual contribution of groundwater flow in stormflow generation. One reason why a role of groundwater flow in stormflow generation has been neglected in the previous studies such as Freeze (1974) is due to an application of the Darcy equation in the customary way to an evaluation of the groundwater discharge.

It must be noticed here again that there existed no observable concentrating flows in the seepage face
in Trench TR1. This fact means that the observed quick response of the groundwater discharge to the rainfall can be caused without pipe flows. The result is contrary to the conclusions of Jones (1971) and Mosley (1979). Therefore, a much larger 'hydraulic gradient' is necessary for generating a quick response of the groundwater discharge to the rainfall.

4.2 Changes in soil-water pressure head and groundwater discharge

The behavior of soil water is considered to make a contribution to a quick response of the groundwater discharge to the rainfall. From this viewpoint, the behaviors of subsurface water in the lower part of the hillslope during two storm events will be discussed in this section based on the relationships between changes in the soil-water pressure head and the groundwater discharge.

Changes in the pressure head profile at Tensiometer plot T-1 during the storm of Aug. 22-23, 1981 are indicated in Fig. 19 with changes in the groundwater level and the groundwater discharge. The line of pF = 1.8 (soil-water pressure head is -60 cmH₂O) is drawn as a boundary between gravitational water and suspended water according to the study of Tsukamoto (1966).
Fig. 19  Changes in soil-water pressure head profile at Tensiometer plot T-1 during the storm of Aug. 22-23, 1981.
The storm was brought on a very dry antecedent condition shown by the pressure head profile (1) in Fig. 19. The pressure head above the 40-cm depth increased with time due to the rainfall and finally became nearly zero, and the transmission zone was formed at 0340 h Aug. 23 just before the peak of the groundwater discharge (profile (3) in Fig. 19). There were, however, no considerable changes in the pressure head at the depth of 60 cm from an onset of the rainfall.

Figure 20 represents distributions of (a) the soil-water pressure head through the hillslope and (b) the subsurface flow rate at each trough at 0340 h Aug. 23. The light dotted part indicates a zone of subsurface water with the pressure head of 1.3 (-20 cmH₂O) < pF ≤ 1.8 (-63 cmH₂O), where water can move easily by a gravitational force. The striped and dark dotted parts indicate a zone of subsurface water with the pressure head of 0 < pF ≤ 1.3 and a zone with the pressure head of pF = 0, that is, the groundwater body, respectively. Subsurface water in the zone with the pressure head less than pF = 1.3 can move quite easily by a gravitational force (Tsukamoto, 1966). It is clear from Fig. 20 that there existed a dry zone between the transmission zone above the 40-cm depth and the groundwater table in the lower part of the hillslope even just before the peak of the groundwater discharge.
Fig. 20  Distributions of soil-water pressure head through the hillslope at 0340 h Aug. 23, 1981 (at peak of storm discharge) with subsurface flow rates at troughs.
Figure 21 indicates changes in the pressure head profile at Tensiometer plot T-1 during the storm of Oct. 22, 1981 following a wet antecedent condition. Changes in the groundwater discharge and the groundwater level are also shown in the Fig. 21. Just in the same as the case of the storm of Aug. 22-23, 1981, the groundwater discharge increased quickly in response to an increase of the rainfall intensity till 1630 h Oct. 22 (about 5 hours before the peak of the groundwater discharge). Nevertheless, any considerable changes in the pressure head at the depth of 60 cm were not observed from an onset of the rainfall (profile (4) in Fig. 21).

The result implies that a quick response of the groundwater discharge to the rainfall occurred without much rain water percolating below the 60-cm depth. Therefore, groundwater existed before the storm was expelled quickly and in large quantities to the stream in response to the rainfall.

Although the pressure head increased to nearly zero throughout the soil profile at 1030 h Aug. 23 (profile (4) in Fig. 19), the groundwater discharge was smaller than that at 0340 h Aug. 23 (profile (3) in Fig. 19; just before the peak of the groundwater discharge). This phenomenon is considered to be due to a decrease of the rainfall intensity (Fig. 16). Even after the rainfall, the pressure head at the depth of 60 cm was larger than
Fig. 21 Changes in soil-water pressure head profile at Tensiometer plot T-1 during the storm of Oct. 22, 1981.
that at 0340 h Aug. 23 and the soils were very wet throughout the soil profile as shown in Fig. 16. In spite of suitable conditions for the downward movement of soil water, the groundwater discharge after the rainfall was much smaller than that during the storm. The above mentioned facts indicate that the groundwater discharge is controlled by the rainfall intensity.

The discussion on a significant role of the rainfall intensity in the groundwater discharge is also supported by changes in the pressure head and the groundwater discharge after the storm of Oct. 22, 1981. Although the soils were still very wet after the rainfall (Fig. 17), the groundwater discharge decreased extremely. This decrease of the groundwater discharge is assumed to be due to a cessation of the rainfall.

4.3 Results of dye tracer experiment

On the basis of changes in the pressure head and the groundwater discharge, Section 4.2 discussed that a quick response of the groundwater discharge to the rainfall could be caused without much rain water percolating below the depth of 60 cm. In order to confirm the assumed behavior of subsurface water directly, a tracer experiment was carried out during the storm of Oct. 22, 1981.

A fluorescent dye tracer (Sulpho rhodamine B of 25 g diluted by the 500-mℓ water) was poured 2.3 m upslope
from Trench TR1 at 0840 h Oct. 22 (Fig. 10), when the groundwater discharge increased already from a pre-storm value. Sulpho rhodamine B has been recommended by Smart and Laidlaw (1977) for the field experiment because of its high resistance to the adsorption by soil particles and low background fluorescence in the nature.

The results of the dye tracer experiment are represented in Fig. 22. Unfortunately, data of the dye concentration were not obtained at Suction lysimeter S2-b because of a failure of the installation.

The dye concentration at Suction lysimeter S1-a (at the depth of 40 cm) attained its peak value at about 7 to 10 hours (or 1510 h to 1810 h Oct. 22) after the pour of the dye tracer as shown in Fig. 22. Most of the dye tracer could move down only 40 cm during more than 7 hours. In spite of this fact, the groundwater discharge increased considerably from a pre-storm value at 1510 h. The result obtained in Section 4.2 was, therefore, confirmed by the dye tracer experiment.

A part of the dye tracer reached already Suction lysimeter S1-b (at the depth of 80 cm) before 1300 h Oct. 22. This fact suggested an existence of substantial rapid flow to the groundwater table. On the other hand, the dye tracer was not detected at Suction lysimeter S2-a (at the depth of 40 cm) even at 1810 h Oct. 22. The result indicated that the movement of soil water might be
Fig. 22 Results of dye tracer experiment.
restricted only in a vertical direction.

4.4 Regression analysis for explanation of groundwater discharge

It was discussed in Sections 3.2 and 4.2 that the groundwater discharge increased quickly in response to an increase of the rainfall intensity and it decreased suddenly after a cessation of the rainfall. The groundwater discharge was, therefore, considered to be dependent mainly on the rainfall intensity. In order to clarify a relationship between the groundwater discharge and the rainfall intensity statistically, a regression analysis is conducted in this section.

Because of malfunctions of the equipments, completely reliable data were obtained only for twelve storms in the course of the study. However, these storms covered a wide range of the hydrologic conditions.

Twelve peak rainfall intensities (p; mm/h) and peak fluxes of the groundwater discharge (q; m²/h/cm²) are used for the analysis because a time-correspondence between these two quantities is distinct. The peak rainfall intensity is in a range of 4.0 mm/h ≤ p ≤ 27.5 mm/h. The peak flux of the groundwater discharge is obtained by dividing the peak groundwater discharge measured at Trough TR1-C by an area of the seepage face in Trench TR1.
A relationship between the peak rainfall intensity and the peak flux of the groundwater discharge is shown in Fig. 23 and the regression equation is expressed as follows:

\[
\ln q = 0.18 p - 0.19 \quad \begin{cases} 
  r = 0.98 \\
  r^2 = 0.96 
\end{cases}
\]  (2)

In this single regression, the rainfall intensity is found to be very important as an explanatory variable, explaining 96% of the variance of the peak flux of the groundwater discharge.

The 'antecedent rainfall indexes (\(\theta; \text{mm}\))' for twelve storms are calculated at Tensiometer plot T-1 by adding the volumes of subsurface water in the soil profile of the 115-cm thickness before the storms to the rainfalls brought before the peak groundwater discharges. Soil water contents at various depths are obtained by using the soil-water characteristic curves in Fig. 4.

The volumes of subsurface water held in the soil profile before twelve storms ranged from 532.0 mm (1200 h Aug. 20, 1981) to 610.0 mm (1000 h Apr. 17, 1982). A range of the antecedent rainfall index was determined as 578.0 mm \(\leq \theta \leq\) 675.0 mm. Therefore, it was not until at least the 46.0-mm rainfall was brought on the driest antecedent condition that only the rainfall intensity could explain changes in the groundwater discharge.
Fig. 23 Relationship between peak rainfall intensity and corresponding peak flux of groundwater discharge.

\[ \ln q = 0.18p - 0.19 \]
\[ r = +0.98 \]

\( p \): peak rainfall intensity (mm/h) and
\( q \): peak flux of groundwater discharge (m\(^3\)/h/cm\(^2\)).
4.5 Process of quick response of groundwater discharge to rainfall

On the basis of the results in Sections 4.1 to 4.4, the process of a quick response of the groundwater discharge to the rainfall is discussed in this section. Judging from the pressure head profile (3) in Fig. 19 and profile (4) in Fig. 21, much air was considered to be entrapped in a vicinity of the 60-cm depth (between the transmission zone and the groundwater table).

A air-entry value in the soil-water characteristic curve for drying at the depth of 90 cm in Trench TRL is -20 to -30 cmH$_2$O (Fig. 4). On the basis of the definition of Bouwer (1978, p. 25~29), a height of the capillary fringe above a falling groundwater table is expected to be 20 to 30 cm in the lower part of the experimental hillslope.

Bouwer (1966) concluded that a height of the capillary fringe above a rising groundwater table was about one-half as high as that above a falling groundwater table. A height of the capillary fringe above a rising groundwater table is, therefore, assumed to be 10 to 15 cm in the experimental hillslope. This height of the capillary fringe is confirmed by the pressure head profiles (1) to (3) in Fig. 19.

Namely, much air was found to be entrapped at the depth of 60 cm just above the capillary fringe. The entrapped air could not escape easily to the ground.
surface by an existence of the transmission zone and the air was thought to become compressed with the successive rainfall. An increase of the pneumatic pressure of the entrapped air could break an equilibrium of forces acting on water in the capillary fringe in response to the rainfall intensity and could convert the capillary fringe into the groundwater body. As the result, the groundwater discharge increased quickly without much rain water percolating below the 60-cm depth. A possibility that the entrapped air readily converts the capillary fringe into the groundwater body has been discussed by Horton and Hawkins (1965) and Kaihotsu (1982) also in the column experiments.

The pressure heads increased with time due to the rainfall and finally became nearly zero throughout the soil profiles at 1720 h and 2130 h Oct. 22 (profiles (5) and (6) in Fig. 21). The capillary fringe attained from the groundwater table to the ground surface at these times. The rainfall broke an equilibrium of water in the capillary fringe throughout the soil profile easily and made the groundwater discharge increase considerably as shown in Fig. 13.

The quick conversion of the capillary fringe into the pressure-saturated zone is expected to be dependent on the rainfall intensity. In consequence, the groundwater discharge responded quickly to the rainfall
intensity.

Although the capillary fringe extended to near the ground surface even after the rainfalls judging from the values of the pressure head in Fig. 16 and 17, the groundwater discharge decreased suddenly after the rainfalls. These decreases of the groundwater discharge are due to the fact that an equilibrium of water in the capillary fringe was not broken easily because of cessations of the rainfalls.

The groundwater discharge was only 0.08 l/h at 2330 h Aug. 22 even after more than the 40-mm rainfall was brought on an extremely dry antecedent condition (Fig. 12). The phenomenon is explained in the following way.

As the soils were very dry before the rainfall (pressure head profile (1) in Fig. 19), much rain water was necessary for generating the transmission zone which prevented the entrapped air escaping to the ground surface and compressed the entrapped air. A degree of the compression of the entrapped air was not thought to be sufficient to break an equilibrium of water in the capillary fringe until more than the 40-mm rainfall was brought.

Only the 4-mm rainfall which was brought on the pressure head profile (1) in Fig. 21 made the groundwater discharge increase quickly from 0.08 l/h to 0.11 l/h within about 1 hour after an onset of the storm of Oct. 22,
1981 (Fig. 17). A quick increase of the groundwater discharge in response to the rainfall with such a weak intensity is supposed to be caused by the following process.

As the soils were in a wet condition at an onset of the storm (pressure head profile (1) in Fig. 21), the capillary fringe in small idealized successive pores was expected to reach near the ground surface as in the model proposed by Horton and Hawkins (1965). An equilibrium of capillary water in these pores was broken easily even by the 4-mm rainfall. The existence of substantial rapid flow to the groundwater table discussed in Section 4.3 supports the above assumption. The groundwater discharge, therefore, increased soon after an onset of the rainfall.

In order to verify the above mentioned process of a quick response of the groundwater discharge to the rainfall, calculations of the water balance are conducted in the experimental hillslope. There were few changes in the pressure head profile at Tensiometer plot T-1 between 1720 h and 2130 h Oct. 22 (profiles (5) and (6) in Fig. 21). Unfortunately, the pressure head profile at 2400 h Oct. 22 was not measured. Judging from the fact that the 44.5-mm rainfall was brought in such a short period between 2130 h and 2400 h, the pressure head profile at 2400 h was considered to be nearly equal to the profiles at 1720 h and 2130 h. On the other hand, the pressure
head profiles at Tensiometer plot T-2 are shown in Fig. 24. They were not expected to change considerably between 1630 h and 2400 h by the same reason for the case of Tensiometer plot T-1. Therefore, changes in the water content, that is, storages of the rain water at Tensiometer plots T-1 and T-2 between 1720 h and 2130 h Oct. 22 and between 2130 h and 2400 h Oct. 22 were thought to be negligible.

Rough calculations of the water balance are carried out in the lower and middle parts of the hillslope. The results are shown in Fig. 25. For instance, the 58.5-mm rainfall was brought per unit area of the ground surface and the total groundwater discharge of 1065 l was generated at Trough TR1-C between 1720 h and 2130 h Oct. 22. All of the 58.5-mm rain water had to contribute to the groundwater discharge at Trough TR1-C within this period since there were no considerable changes in the water content in the lower and middle parts of the hillslope as described above.

As represented in Fig. 25, the area § A contributing to the groundwater discharge between 1720 h and 2130 h Oct. 22 is, therefore, obtained by dividing the total groundwater discharge of 1065 l by both the total rainfall of 58.5 mm and the length of Trough TR1-C of 120 cm. The area § B contributing to the groundwater discharge between 2130 h and 2400 h Oct. 22 is calculated in the
Fig. 24 Changes in soil-water pressure head profile at Tensiometer plot T-2 during the storm of Oct. 22, 1981.
Fig. 25  Estimated areas contributing to stormflow generation during the storm of Oct. 22, 1981.
same way.

The two areas A and B are well coincident with the observed extents of the groundwater body at 1630 h and 2130 h Oct. 22, respectively. The result indicates that only the rain water brought on the lower and middle parts of the hillslope, where the groundwater body and the capillary fringe existed during the storm, could contribute so quickly and in such large quantities to the groundwater discharge at Trough TR1-C. It is, therefore, proper to consider that a conversion of the capillary fringe into the pressure-saturated zone caused a quick response of the groundwater discharge to the rainfall.

As mentioned above, the behavior of water in the capillary fringe plays a predominant role in a quick response of the groundwater discharge to the rainfall. A conversion of the capillary fringe into the pressure-saturated zone is not able to be detected directly by a traditional equipment because the conversion may occur in a short time. However, the effect of this conversion on the groundwater body must be incorporated in the hydraulic gradient in the Darcy equation. On the other hand, as vertical flow predominates in the capillary fringe during the storm event, the hydraulic gradient is to be determined in a vertical direction along a flow line. Therefore, a new hydraulic gradient, which is necessary for explaining a quick response of the
groundwater discharge to the rainfall by the Darcy equation, has to be defined by taking account of the behavior of water in the capillary fringe.
CONTRIBUTION OF UPPER AND RIDGE-TOP PARTS OF THE HILLSLOPE TO STORMFLOW GENERATION

5.1 Subsurface flow rates in trenches

Chapter IV discussed that only the rain water brought on the lower and middle parts of the hillslope, where the groundwater body and the capillary fringe existed, could contribute to stormflow generation. Roles of the rain water brought on the upper and ridge-top parts of the hillslope in stormflow generation are discussed in this chapter.

Changes in the principal subsurface flow rate measured at each trough during and after the storm of Aug. 22-23, 1981 are represented in Fig. 26. Peak flow rates throughout the hillslope are shown by arrows in Fig. 20. The subsurface flow rate at Trough TR1-C is equal to the groundwater discharge in the previous discussions.

Even at the peak of the storm discharge, considerable amounts of subsurface flow were observed only at Troughs TR1-C and TR3-A. Subsurface flow at Trough TR3-A in the upper part of the hillslope was found to be derived from the partially saturated soil surface in Trench TR3. It was generated only during the rainfall with a high intensity and disappeared suddenly after a cessation of the rainfall as shown in Fig. 26.

Subsurface flow rate at Trough TR3-A did not
Fig. 26  Subsurface flow rates in trenches during and after the storm of Aug. 22-23, 1981.
relate to that at Trough TR2-A. Moreover, no water tables were observed at Wells W2-a and W3-a (the most shallow wells in the middle and upper parts of the hillslope). Subsurface flow at Trough TR3-A, therefore, could not directly take part in stormflow generation occurring in the base of the hillslope. The local existence of subsurface flow has also been reported by Betson and Marius (1969) in an upper part of the hillslope of North Carolina, U.S.A.

The subsurface flow rates induced by the storm of Oct. 22, 1981 are indicated in Figs. 27 to 32. Subsurface flow at Trough TR3-A appeared soon after an onset of the rainfall because the soils were in a wet antecedent condition, and it continued just after the peak rainfall.

As is obvious from Fig. 30, however, the subsurface flow rate at Trough TR3-A did not correspond to that at Trough TR2-A even at the peak of the storm discharge. Moreover, the observed flow rate was extremely small in comparison with the flow rates at Troughs TR1-C and TR2-C. Therefore, subsurface flow in the upper part of the hillslope was not considered to directly contribute to stormflow generation occurring in the base of the hillslope.

In consequence, the rain water that was brought on the upper and ridge-top parts of the hillslope, where
Fig. 27 Subsurface flow rates in trenches during and after the storm of Oct. 22, 1981.
Fig. 28 Distributions of soil-water pressure head through the hillslope at 0100 h Oct. 22, 1981 (before the storm) with subsurface flow rates at troughs.
Fig. 29  Distributions of soil-water pressure head through the hillslope at 1630 h Oct. 22, 1981 with subsurface flow rates at troughs.
Fig. 30  Distributions of soil-water pressure head through the hillslope at 2130 h Oct. 22, 1981 (at peak of storm discharge) with subsurface flow rates at troughs.
Fig. 31  Distributions of soil-water pressure head through the hillslope at 1630 h Oct. 23, 1981 (after the storm) with subsurface flow rates at troughs.
Fig. 32  Distributions of soil-water pressure head through the hillslope at 1030 h Oct. 31, 1981 with subsurface flow rates at troughs.
no saturated zone was generated even at the peak of the storm discharge, could not directly contribute to stormflow generation.

5.2 Changes in soil-water pressure head and water content through the hillslope

Distributions of the soil-water pressure head before the storm of Oct. 22, 1981 are shown in Fig. 28. The explanations for three categories of pF have been described in Section 4.2. Before the storm, the hillslope was wetter in the downslope part than in the upslope part of the cross section. As the rainfall continued, the hillslope became wetter through the cross section (Fig. 29) and extremely wet at the peak of the storm discharge (Fig. 30). After the storm, the soils became drier with time as shown in Figs. 31 and 32. More decreases in the pressure head took place in the ridge-top and upper parts of the hillslope. As time goes on, the soils are expected to become wetter in the downslope part than in the upslope part of the hillslope just in the same as the condition before the storm.

Figure 33 indicates changes in the volumetric water content between 2130 h Oct. 22 (at the peak of the storm discharge) and 1030 h Oct. 31 (about 8 days after the storm). The soil-water characteristic curves in Figs. 4 to 7 were used for a conversion of the pressure head
Fig. 33 Changes in volumetric water content between 2130 h Oct. 22 and 1030 h Oct. 31, 1981.
measured at each depth by the tensiometer into the volumetric water content. Decreases in the water content with time were most marked in the ridge-top and upper parts of the hillslope.

It is considered that unsaturated downslope flow made the upslope part of the hillslope dry and the downslope part of the hillslope wet. As described in Section 5.1, the rain water brought on the upper and ridge-top parts of the hillslope was not important for the event of stormflow generation. On the basis of the above results, however, the rain water brought on the upper and ridge-top parts of the hillslope could contribute to stormflow generation induced by the next storm, keeping the lower and middle parts of the hillslope in a wet condition after the storm. Such a wet condition is suitable for a quick response of the groundwater discharge to even the rainfall with a weak intensity as discussed in Section 4.5.
VI CONCLUSIONS

The study was conducted in the experimental basin with a flat valley bottom and the soils of low hydraulic conductivities. The groundwater table in the valley bottom rose up to the ground surface during the storm, and generated saturation overland flow. Under these situations, a role of subsurface flow (subsurface stormflow and groundwater flow) in stormflow generation has been regarded to be negligible in the previous studies. As the result of the study, the following conclusions were drawn.

(1) Contrary to the results of the previous studies, groundwater flow plays a predominant role in stormflow generation. The groundwater appears so quickly and in such large quantities in the stream as to contribute to stormflow generation.

(2) A quick response of the groundwater discharge to the rainfall is not able to be accounted for by the Darcy equation used in the customary way.

(3) A quick response of the groundwater discharge to the rainfall is considered to be caused by the break of an equilibrium of forces acting on water in the capillary fringe, that is, by a conversion of the capillary fringe into the pressure-saturated zone.

(4) The area contributing to stormflow generation is
restricted in the lower and middle parts of the hillslope, where the groundwater body and the capillary fringe exist during the storm. The rain water brought on the lower and middle parts of the hillslope breaks an equilibrium of forces acting on water in the capillary fringe and contributes quickly to stormflow generation.

(5) The rain water brought on the upper and ridge-top parts of the hillslope gradually moves downslope and keeps the lower and middle parts of the hillslope in a wet condition. As the result, the groundwater discharge increases rapidly in response to even the rainfall with a weak intensity at a beginning of the forthcoming storm.
BIBLIOGRAPHY


