Dynamic Behavior of Soil Water Movement in a Headwater Basin

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Abstract

Temporal variation of soil moisture contents and hydraulic head profiles of soil water was measured by using the neutron moisture meter and tensiometers in a forested headwater basin. In addition to the hydrometric observations, water sampling of rainfall, throughfall, stemflow, soil water at various depths, groundwater and discharge water was performed periodically in order to make isotopic analyses of the sampled water such as deuterium and oxygen-18. The basin for the present research is underlain by volcanic rocks in Neogene and located in Nagano prefecture, central Japan. The average thickness of soil mantle is 1.63 m in the basin and 70 % of the volume of the soil mantle consists of the B horizon, which is characterized by high water retentivity.

The temporal change of soil moisture contents was notable above 1 m depth and in the vicinity of the bedrock surface. On the other hand, the soil moisture content was relatively stable except for those layers. The temporal variation of soil moisture of whole soil mantle was relatively small.

The behavior of soil water movement above 1 m depth was dynamic. The frequent appearance and disappearance of the divergent zero flux plane and the convergent zero flux plane were observed in accordance with the change of soil water flux due to rainfall events and evapotranspiration. The deepest depth where the divergent and convergent zero flux planes were formed was 1 m, and the soil water movement

was relatively stable under the 1 m depth.

The monthly isotopic ratios of deuterium and oxygen-18 of soil water above 0.5 m depth showed the values corresponding to those of throughfall. On the other hand, the monthly isotopic ratios of soil water below 1 m depth converged on the values of groundwater. It was suggested from this trend that the isotopic homogenization of soil water might occur. The isotopic homogenization of soil water could be explained by the dynamic behavior of soil water movement, that is the soil water displacement flow with consideration of mixing process which might occur around the convergent zero flux plane.

Biographical Sketch

Maki Tsujimura was born on August 5, 1965 in Tokyo, Japan. He graduated from University of Tsukuba, Ibaraki, Japan in 1988 with a B.S. major in Hydrology from the College of Natural Sciences. In March of 1990, he received a M.S. from the Doctoral Program in Geoscience at University of Tsukuba in the field of Hydrology. His M.S. thesis concerned behavior of subsurface water in a steep forested slope. In 1990 he began his Doctoral studies for development of M.S. research.

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Chapter I

Introduction

1-1. Previous studies

The most important subject in hillslope hydrology is to elucidate the water flow processes occurring in the slopes rather than the separation of runoff components in the hydrograph. Since 1970's many researches related to this subject including the field observation and the mathematical simulation have been conducted (e.g. Kirkby, 1978; Anderson and Burt, 1990). This topic has been a main theme not only in the hillslope hydrology but also in hydrological sciences, because the behavior of subsurface water depends upon the conditions such as geology, physical properties of soil mantle, vegetation and climate *in situ* (Tanaka, 1989). Therefore a deductive model can not account for the behavior of subsurface water in various fields. Similarly, a result from the observation in one field may not be applied directly to the other fields under a different condition. Dunne (1978) has reviewed the field studies of hillslope flow processes and pointed out that the processes that deliver stormflow, and the volumes and timing of their contribution varied with topography, soil properties and rainfall characteristics, and indirectly with climate, vegetation and land use.

To be concrete, a variable source area concept was presented by Hewlett and Hibbert (1967) and has been a basic concept to explain stream flow generation mechanism in the humid regions such as parts of Europe, USA and Australia (e.g. Ward, 1984; Pearce *et al.*, 1986). Pearce *et al.* (1986) have classified the results of previous studies into some categories by the prominent flows occurring in their fields on the basis of variable source area concept. These flow types include partial-area Horton overland flow (e.g. Betson, 1964), saturation overland flow (e.g. Dunne and Black, 1970a;b) and subsurface flow (e.g. Hewlett and Hibbert, 1963) which can be characterized by rapid throughflow through macropore (e.g. Mosley, 1979; 1982) and displacement of pre-event water (e.g. Hewlett and Hibbert, 1967). On the other hand, mountainous headwater basins in Japan generally consist of the steep forested slopes, therefore it is suggested that the source area can not be enlarged even under the heavy storm events on the slope with steep gradient comparing with the gentle slopes in Europe (Ohta, 1990). In other words the source area may not be variable in such a condition found in Japan. Therefore it is necessary to make case studies which cover various kinds of site conditions (Tsujimura *et al.*, 1993).

From these points of view, some previous field researches with respect to the behavior of subsurface water conducted on hillslopes in Japan are arranged in Table 1 depending on the geology of bedrock underlain. Many studies were conducted in the regions underlain by sedimentary rock in Neogene or granite; on the contrary, few researches have

Table 1 Relationship between the geology of basement rock and previous field researches on subsurface flow in the mountainous basins in Japan.

Type of basement rock		Fractional area within Japan (%)	References
Sedimentary rock	Neogene	15	Ohta *et al.*(1985a;b), Tanaka *et al.*(1988), Marui(1991), Onodera(1991)
	Mesozoic ~ Paleogene	12	
	Paleozoic	12	Onda(1993), Tsukamoto(1989)
Volcanic rock	Neogene	13	
	Mesozoic ~ Paleogene	4	Ohta & Takahashi(1986)
Plutonic rock	Neogene	1	Ishii(1990)
	Before Paleogene (Granit)	11	Kubota *et al.*(1987), Ohta(1983),Ohta *et al.*(1983) Tsukamoto(1962), Onda(1989)
Metamorphic rock		4	

(Geological classification is from Kaizuka & Chinzei, 1986) been performed in the regions underlain by volcanic rock in Neogene which are distributed widely in the middle to the north-east area of Japan. It is needed to obtain information about the subsurface flow processes by case studies in these areas.

The methods for investigation of soil water movement in the field can be classified as follows;

1) Observation of hydraulic head profiles of soil water (e.g. Ohta *et al.*, 1985a; b).

2) Measurement of temporal change of soil moisture contents (e.g. Maeda *et al.*, 1986).

3) Tracer method by using the conservative chemical components such as chloride (Cl^-), tritium (^3H), deuterium (D) and oxygen-18 (^{18}O) (e.g. Zimmerman *et al.*, 1966).

Each method has a merit and a demerit. Previous studies have not performed combinational approach which utilizes several methods at the same time. Anderson and Burt (1990) stated that the linkage between subsurface flow processes and the movement of solute through the soil-hillslope system must be investigated in future. Sklash (1990) has emphasized that there remains considerable room for research which combines natural isotopic tracer tests with hydrometric and hydrochemical approaches. From these statements it is clear that the field research on the subsurface flow must be conducted through the combination of hydrometric approach and natural isotopic tracer method.

In the headwater basins, environmental isotopes such as ^3H, D and ^{18}O have been used as effective tracers especially in separation of discharge component of hydrograph during a storm in previous studies (Fritz *et al.*, 1976; Sklash and Farvolden, 1979; Hooper and Shoemaker, 1986; Kennedy *et al.*, 1986; Sklash *et al.*, 1986; McDonnell *et al.*, 1990). In these studies, however, only the isotopic concentration of rain water and discharge water have been analyzed, and the process of variation of isotopic concentration of subsurface water in the soil mantle has not been elucidated. McDonnell (1990) has performed the continuous monitorring of three-dimensional soil moisture conditions to explain the results of isotope tracer method. This research by McDonnell is very important, because it shows an effectiveness of the combinational method of hydrometric observation and tracer method.

It has been suggested by using ^{3}H, D and ^{18}O as a tracer that the soil water movement in the flatland might be explained by the displacement flow model with diffusion. Zimmerman *et al.* (1966) concluded from the results of tracing soil water by HDO that a single rainfall labelled with isotope tracer formed a tagged layer of water blurred by diffusion effects and moved downward as a distinguishable water mass between the older rainwater below and the younger rainwater above. Kayane *et al.* (1980) and Shimada (1983; 1988) investigated soil water movement in the unsaturated zones of a volcanic ash layer, which is called the Kanto Loam formation. They have used environmental tritium as a tracer, and Shimada presented a displacement flow model with dispersion for explaining the tritium concentration profile of soil water. Saxena (1987) showed cyclic variations in ^{18}O concentration in soil water and estimated the annual recharge rate in Uppsala underlain by the post-glacial deposit, Sweden. Darling and Bath (1988) suggested that the piston displacement mechanism in unsaturated zone was inappropriate to a region where the isotope record of porewater showed no sign of seasonal cyclicity. In these studies, the piston flow (displacement flow) model was used to account for the isotopic concentration profiles of soil water. However the examination of this model in relation to the actual soil water movement has not been conducted. Previous studies using a tracer method have failed to explain the process of the formation of isotopic concentration profiles of soil water by the behavior of soil water movement evaluated by the data of hydraulic head profiles. This is an important point as detected by the present study.

As for the stable isotopic compositions of soil water in the unsaturated zone under the effect of evaporation, Barnes and Allison (1988) have reviewed the researches on this subject. Barnes and Allison (1983) have developed a mathematical model describing the shape of D and ^{18}O concentration profiles within the soil which resulted from evaporation from dry soil surface under quasi-steady state conditions. This method was then used to estimate the evaporation from the dry soil surface. Allison *et al.* (1983) examined the validity of this model by the experimental data in the columns under the steady state conditions.

Barnes and Allison (1984) have developed a model which predicted the profiles of isotopic ratio whose shape was determined with effect of evaporation from a soil under non-isothermal and quasi-steady state conditions. Fontes *et al.* (1986) investigated the stable isotope profiles in soil water to determine the loss of water by evaporation in the western Sahara where water table is at about 10 m beneath the surface. These studies have focused on the stable isotope profiles of soil water under the effect of evaporation in the arid or semi-arid regions. Unfortunately, they have paid little attention on the percolation processes of the infiltrated water. The stable isotopic composition of soil water must be explained with consideration of both effect of evapotranspiration and percolation processes in the humid regions. Few studies, however, have focused on this point previously.

The mechanism of isotopic homogenization of water in the headwater basins is also an important subject in relation to the problem as mentioned above. It is generally known that the isotopic concentration of rainfall varies considerably. On the contrary, the isotopic concentration of groundwater and discharge water always takes a value which is equal to the annual weighted mean of isotopic ratio of rainfall in a region (Mizutani, 1986; Nakai *et al.*, 1986). It is quite likely that isotopic homogenization of water may occur through the flow processes of subsurface water in the mountainous basin. However the mechanism of isotopic homogenization has been poorly understood.

Stewart and McDonnell (1991) measured the deuterium concentration values of soil water collected from 11 suction cup samplers and compared them with those of rainfall in the same period. They have presented that hydrodynamic dispersion (mixing) was dominant in isotopic homogenization process in the periodically saturated and the generally saturated soils. McDonnell *et al.* (1991) suggested that the continual downslope mixing and isotopic homogenization may occur through the flow process of soil water from upslope zone draining into the midslope zone and then into the near stream zone inter storms, and they stated that the topographic convergence of subsurface water into hollows will have a strong influence on the isotopic mixing of soil water and groundwater during the storm events. In these studies, importance of mixing in the saturated zone and effect of topography in the basin were emphasized for the isotopic homogenization of water. On the other hand, Turner *et al.*, (1987a; b) focused on the effect of water storage in the shallow groundwater on mixing of water. Turner *et al.* (1987a) measured stable isotopic composition and Cl^- concentration of rainfall, shallow and deep groundwater and streamflow in a small basin in western Australia. They showed that the response of discharge to an individual rainfall event was diminished as the season progressed and this attenuation with time was evidence of isotopic mixing and dilution of rain water of an individual rainfall event with the increasing storage of shallow groundwater. Turner *et al.* (1987b) observed the solute concentration such as D,

^{18}O, ^{3}H and Cl^- in unsaturated and saturated zones and showed that the rainfall with high variation of isotopic composition within a storm event became well-mixed during the early stages of recharge, and the variability was shown to persist in the upper 2 m of the unsaturated zone. In these researches the mixing or diffusion of water in the unsaturated and saturated zone have been considered more or less responsible for the isotopic homogenization in the basins.

Kitaoka and Yoshioka (1984) measured ^{3}H concentration of rainfall, river water and seepage water in the tunnel in Rokko mountains, Japan. They suggested that the ^{3}H homogenization of water might have occurred when the water discharged to the channel. They did not consider the mixing processes of water in the unsaturated zone. To understand the mechanism of the isotopic homogenization of water in the mountainous basin it is necessary that the hydrometric observation should be combined with the measurement of isotopic compositions of soil water.

The previous studies concerning with subsurface flow processes, especially soil water movement, occurring in the hillslopes of the headwater basins have been reviewed. The lack of information or understanding on this subject is summarized as follows;

1) There have been only limited regions where enough researches have been conducted previously on the subsurface flow processes.

2) There have been few studies conducted previously that

elucidate the linkage between the hydrometric data such as hydraulic potential of soil water and the change of stable isotopic ratios of soil water through the subsurface flow processes.

3) The mechanism of isotopic homogenization of water in unsaturated zone have not been investigated previously on the basis of detail observations of soil water movement in the headwater basins.

1-2. Objectives of the study

The present study is concerning with the soil water movement in relation to the process of formation of isotopic concentration profiles of soil water in a mountainous headwater basin underlain by the volcanic rocks in Neogene where there has been few data on this subject. The objectives of the investigation are as follows;

1) To elucidate the soil water movement during a storm event, a rainless period and in the long-time period by the observation of temporal change of soil moisture and the hydraulic head profiles of soil water.

2) To explain the process of formation of stable isotopic ratio profiles of soil water by the behavior of soil water movement.

3) To make clear the mechanism of the isotopic homogenization of water occurring in the unsaturated zone.

Chapter II

Description of study area

2-1. Geology, topography and climate

The study was conducted in a mountainous headwater basin with steep forested slopes, which is located in Nagano prefecture, central Japan (35° 54.9'N, 138° 30.2'E; Figs. 1 and 2). The basin, called the Kawakami Experimental Basin, is in University of Tsukuba Forest at Kawakami. The basin corresponds to a headwater of Misawa River which belongs to the Chikuma River system. The elevation of the basin ranges from 1500 to 1690 m, and the basin has an area of 0.14 km^2.

The basement rock of the Kawakami Experimental Basin is Meshimori-yama volcanic rock in Neogene which constitutes the surrounding mountains such as Mt. Meshimori (1670 m) and Mt. Yokoo (1818 m). Kawachi (1977) described that this basement consists of agglutinate, tuff breccia, volcanic breccia and the like. Every material is considerably weathered, and especially in lava the onion structure is remarkable.

There is a sub-divide from 1520 to 1680 m within the basin, and this sub-divide distinguishes the basin into two sub-basins, namely north-valley and south-valley. Both valleys have the streamflow. In the north-valley the streamflow starts at the altitude of about 1550 m, and in the south-valley the beginning altitude of streamflow is

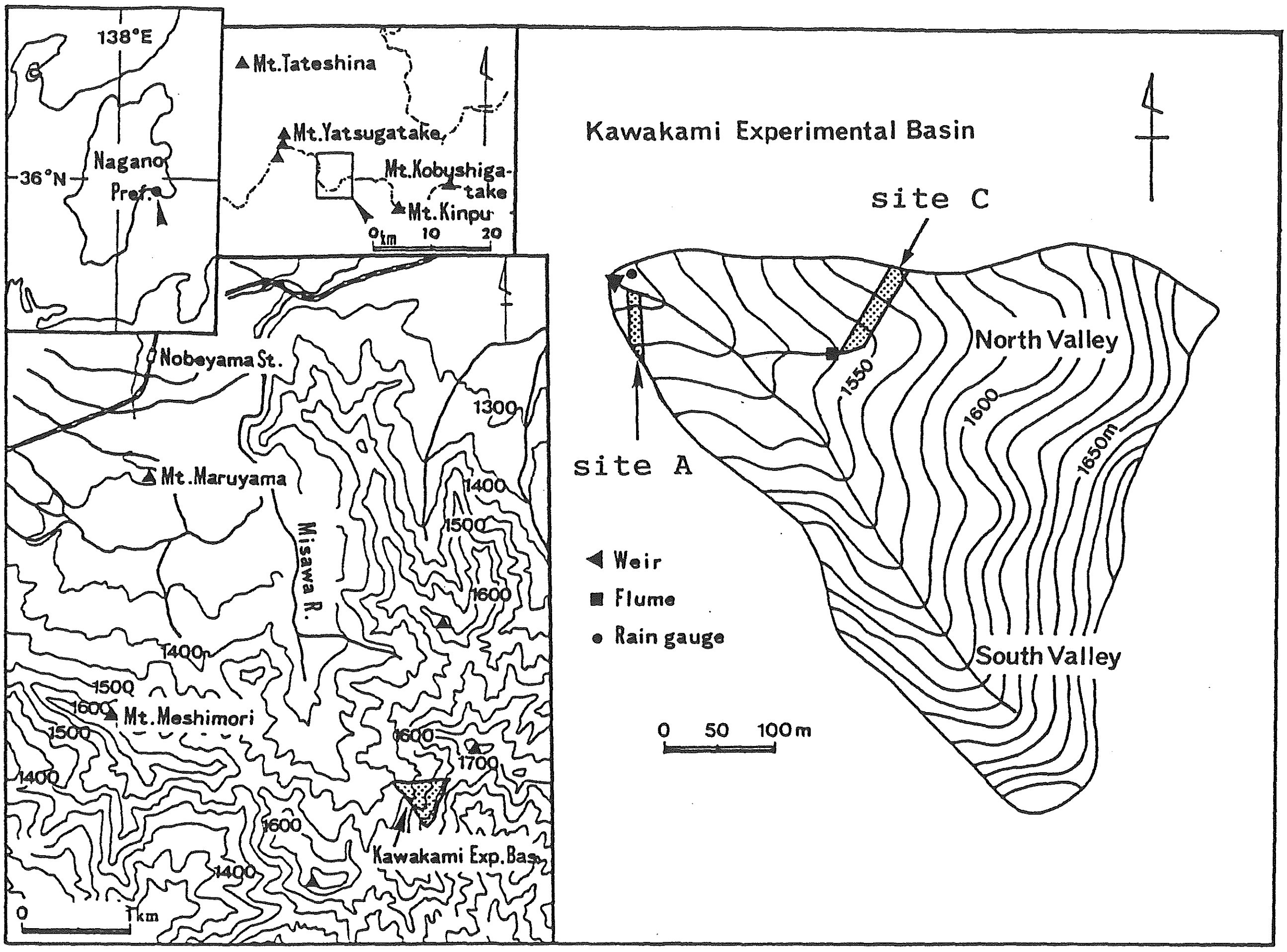


Fig. 1 Location of study area.

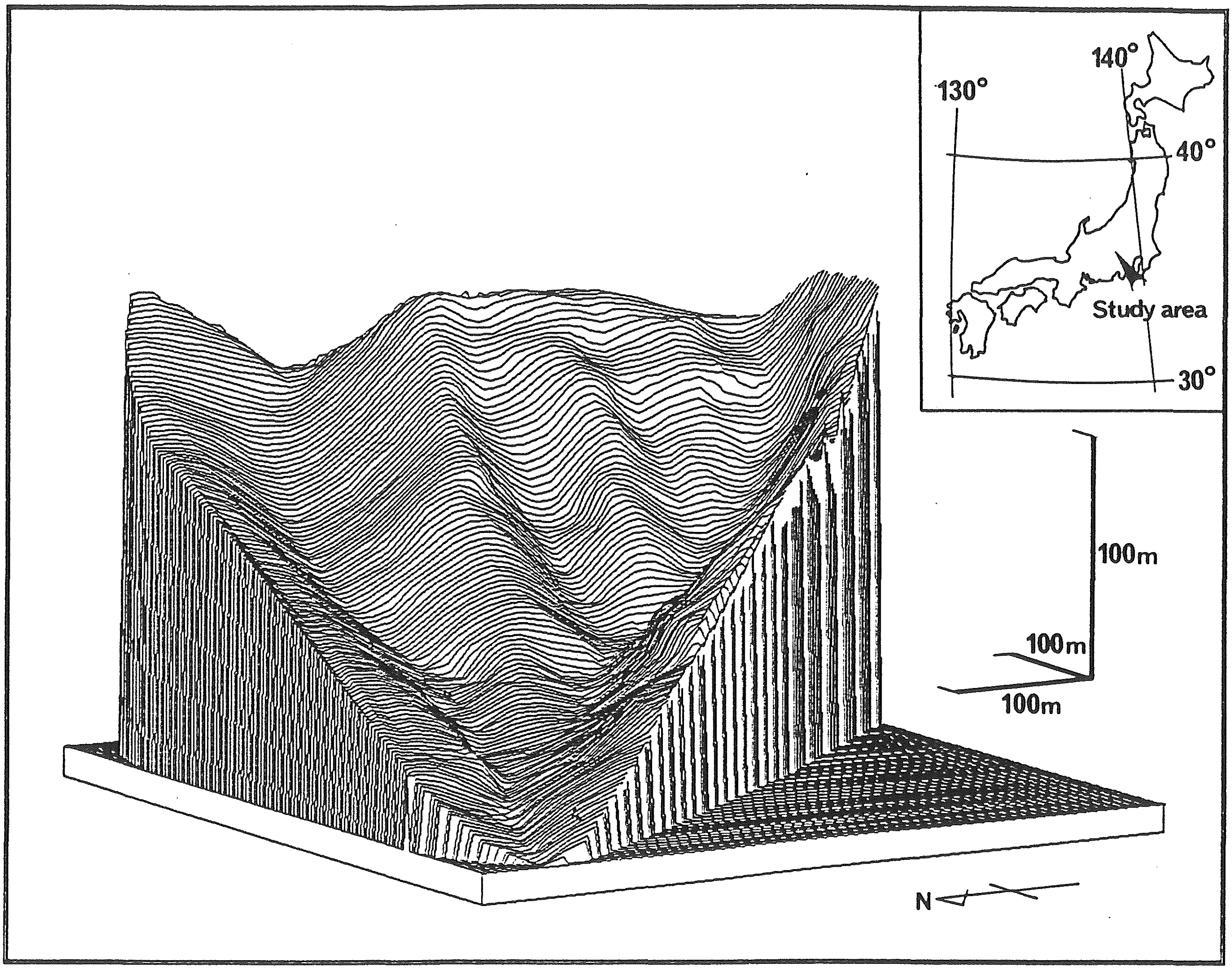


Fig. 2 Three-dimensional graphical representation of the Kawakami Experimental Basin.

about 1630 m.

The vegetation of the Kawakami basin consists of natural deciduous forest of oak and maple and of artificial larch, and the bamboo grass covers the forest floor. The spatial distribution of vegetation is shown in Fig. 3.

The region where the basin exists is under the cool climatic condition which is typical for inland area. The annual average air temperature observed at the meteorological station in the basin is 6.0 °C between 1983 and 1991, and the monthly average value of January which is the coolest month is -5.8 °C (Agricultural and Forestry Research Center, Univ. Tsukuba, 1986; 1987; 1988; 1989; 1990; 1991; 1992; 1993). Itadera (1987) observed that the thickness of frozen soil was about 0.2 m in March, 1986. Annual precipitation is from 1400 to 1500 mm. A water budget of the basin will be described in chapter 3.

2-2. Physical properties of soil mantle

To understand the thickness of soil mantle in the Kawakami basin, Oginuma (1988) conducted the sounding test at about 100 points within the basin and revealed the spatial distribution of soil mantle thickness. Sounding was made by using a cone-penetrometer, which has a cone, a weight of 5 kg, and a falling height of 0.5 m. He defined the thickness of soil mantle as the depth, within which N_{10} value was smaller than 30, where N_{10} value is equivalent to the number of impact for 0.1 m depth. The distribution of

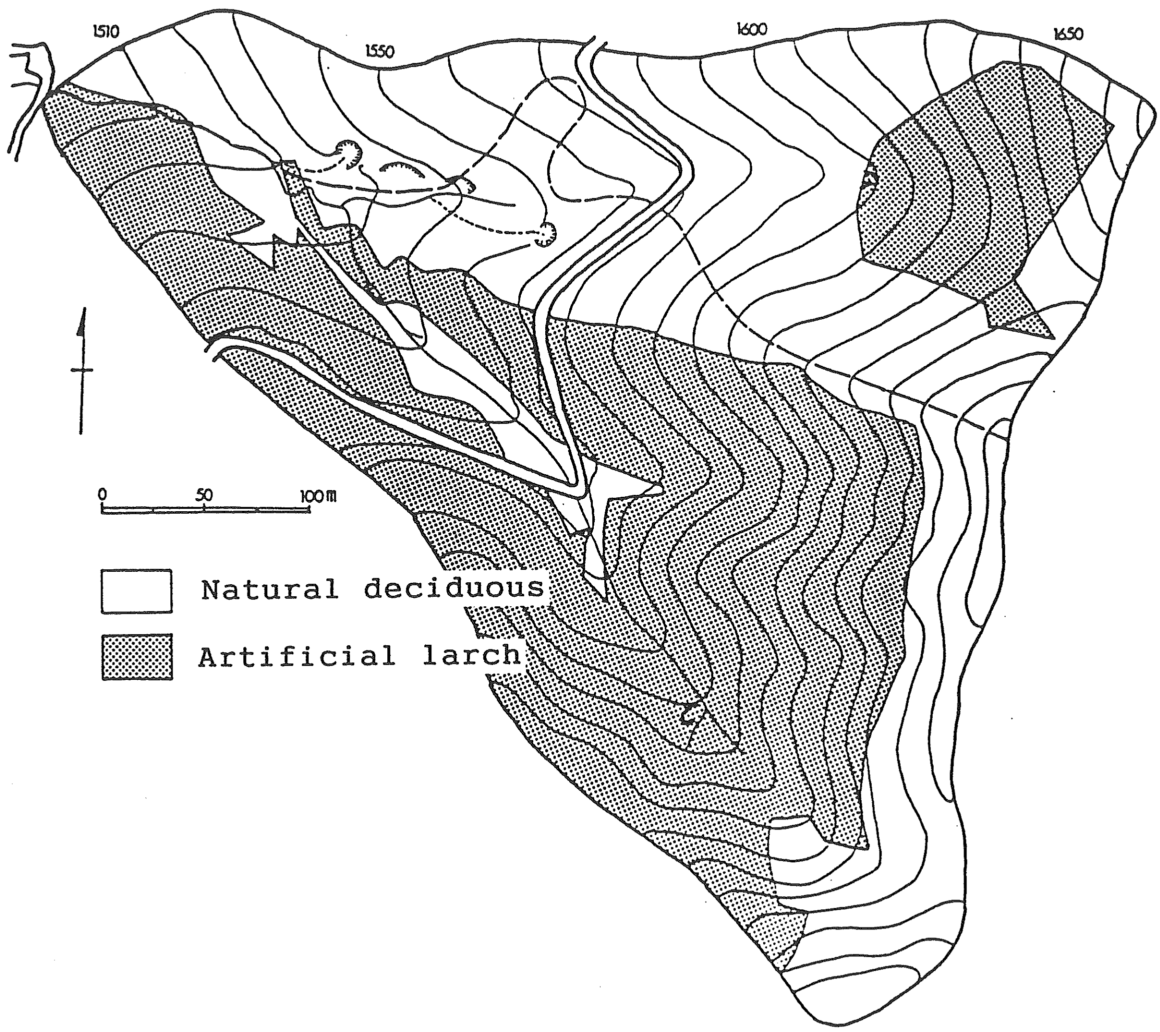


Fig. 3 Spatial distribution of vegetations.

soil thickness is shown in Fig. 4. The figure shows that the soil is thin in the upper area of the basin and relatively thick in the lower area, and especially the north-valley has the thickest soil area where the thickness reaches 6.0 m at the altitude from 1540 to 1560 m. The average value of soil thickness of the basin estimated by areal weighted mean method is 1.63 m.

To clarify the characteristics of soil mantle the soil sampling was carried out in the whole area of the basin. A lot of undisturbed soil core samples of which the volume is 100 cm^3 were taken along the line from the base of the slope to the divide, and from the surface soil to the soil just above the basement rock. The soil samples were used for some measurements of physical properties of soil water retention (drying process by suction method), saturated hydraulic conductivity (falling head method) and volumetric water content (gravimetry with drying). From these soil samplings and measurements, it became clear that the soil mantle could be classified vertically into three horizons, namely A, B and C horizons which appear to be the typical horizons for forest soil (Kawada, 1989). Figure 5 illustrates the schematic diagram of soil horizons. Each horizon has different characteristic from the view point of physical aspects which will be described below.

The surface soil layer between 0 and 0.4-0.5 m depth in average which shows a black or dark brown color belongs to the A horizon (humus layer) which is rich in organic matter and has relatively high hydraulic conductivity of the order

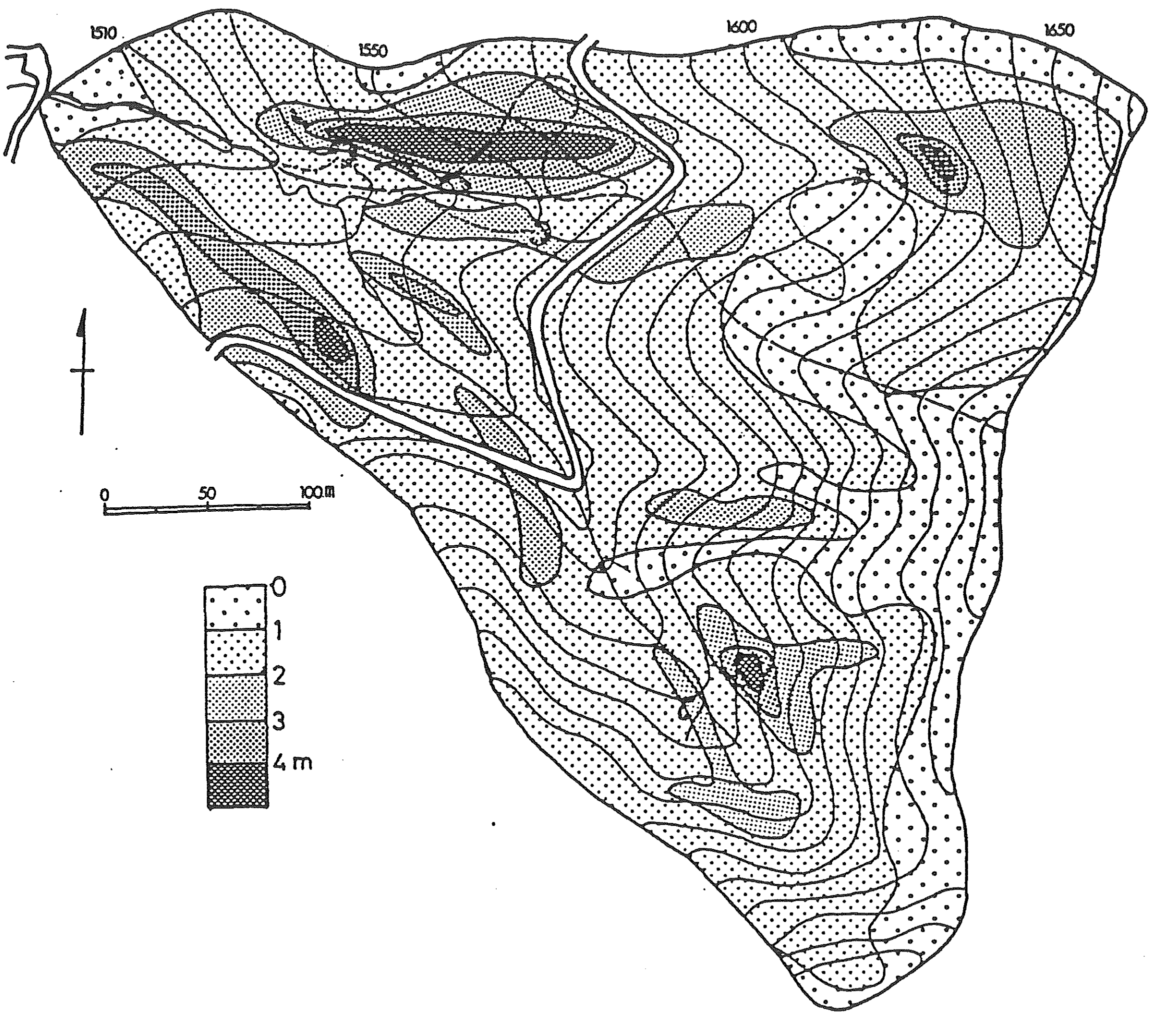


Fig. 4 Spatial distribution of soil mantle thickness. (add the data to Oginuma, 1988)

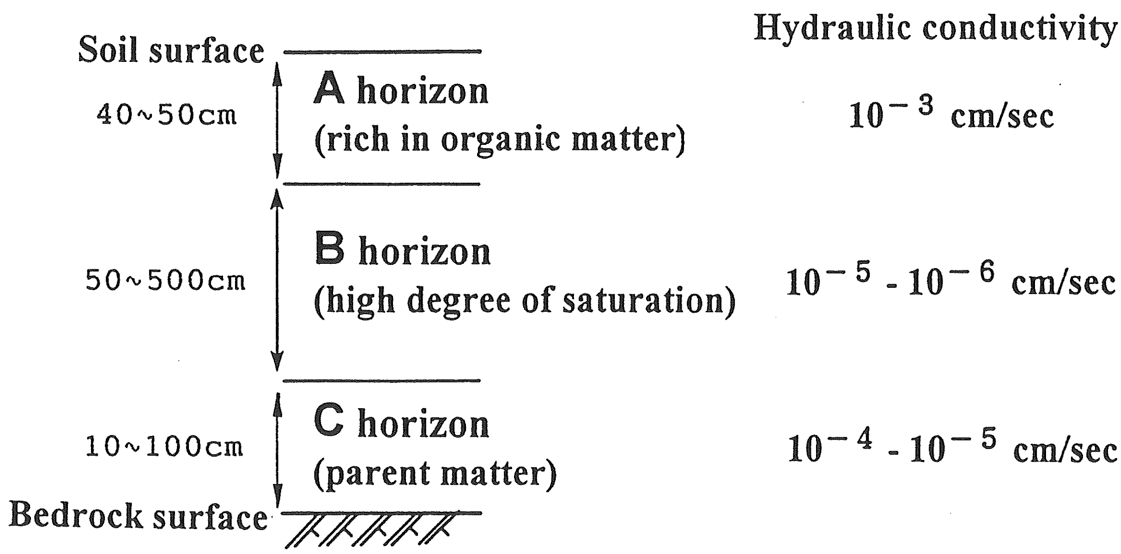


Fig. 5 Schematic diagram of soil horizon.

of 10^{-3} cm/sec. The soil water retention curve in Fig. 6 indicates that the soil moisture is relatively high and the air-entry value is low (below -10 cmH_2O), and that the upper boundary of saturated and unsaturated capillary water zone is defined clearly.

The B horizon (illuvial horizon) exists below the A horizon. It is highly weathered, clayey and light brown colored. Oginuma (1988) indicated that the relationship between the thickness of B horizon and that of soil mantle showed a positive correlation, and the thickness of B horizon corresponds to 68.8 % of the thickness of soil mantle in average. Therefore the B horizon appears to play an important role in soil water movement. The saturated hydraulic conductivity of the horizon is relatively low of 10^{-5} cm/sec, and the lowest (10^{-6} cm/sec) value was observed at the upper boundary of the horizon. The soil water retention curves show (Fig. 7) that the soil moisture is relatively high and the air-entry value is also high (above -31.6 cmH_2O = pF 1.5), and that the upper boundary of capillary water zone is not well defined. It is observed from this figure that the variation of soil moisture with change of pressure head of soil water is remarkably small. This characteristic is important to consider the soil water movement.

The horizon between the lower limit of B horizon and the bedrock surface is C horizon, which is the parent material. The material of bedrock is observed a little in the horizon. The average thickness of the C horizon is

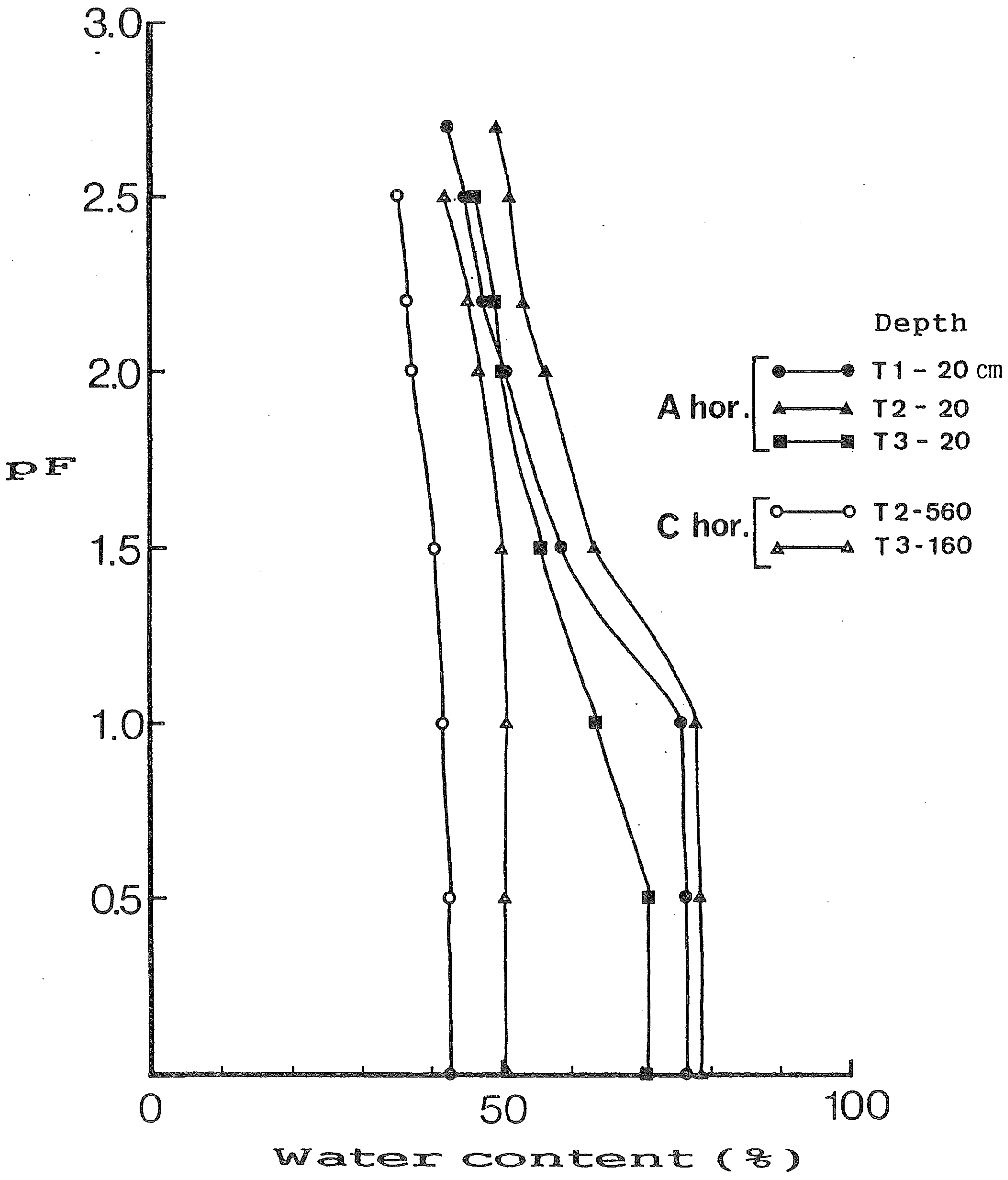


Fig. 6 Soil water retention curves for A and C horizons.

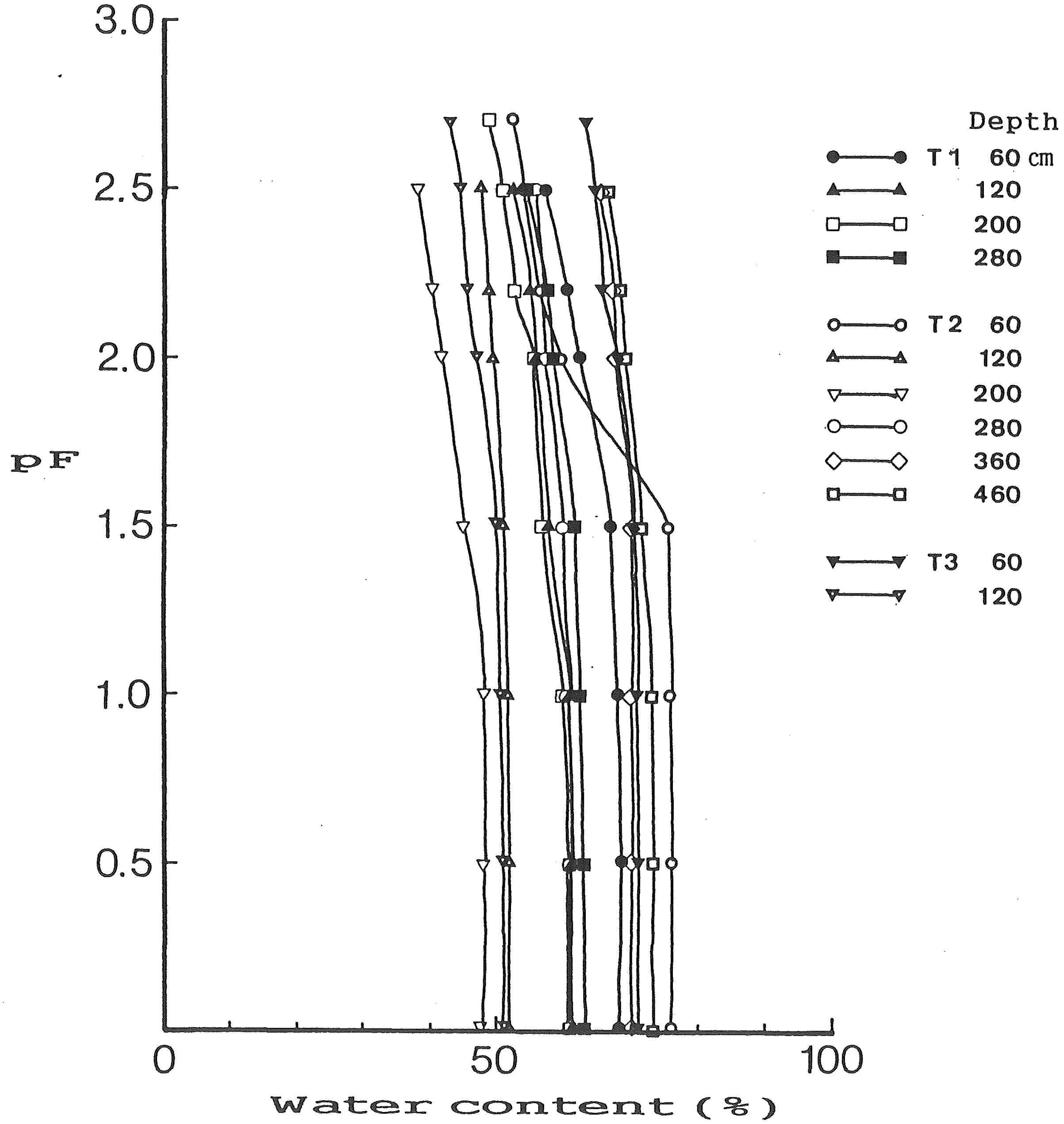


Fig. 7 Soil water retention curves for B horizon.

around 0.4 m, though it is quite variable from site to site. The hydraulic conductivity is of the order of 10^{-4} to 10^{-5} cm/sec. The soil moisture is relatively low, and the upper boundary of capillary water zone is not very clear.

Figure 8 illustrates the spatial distribution of soil water storage measured by gravimetry method with drying of more than 500 samples at about 90 sites in the basin. The data base includes newly obtained samples from north-valley and those of south-valley obtained by Oginuma (1988). A comparison with Fig. 4 shows that the spatial trend of soil water storage agrees well with that of soil mantle thickness. Moreover the B horizon occupies about 70 % of soil mantle as mentioned above. These facts suggest that the B horizon plays a considerably important role in the percolation and storage of soil water, and the physical characteristics of B horizon soil appear to be spatially homogeneous.

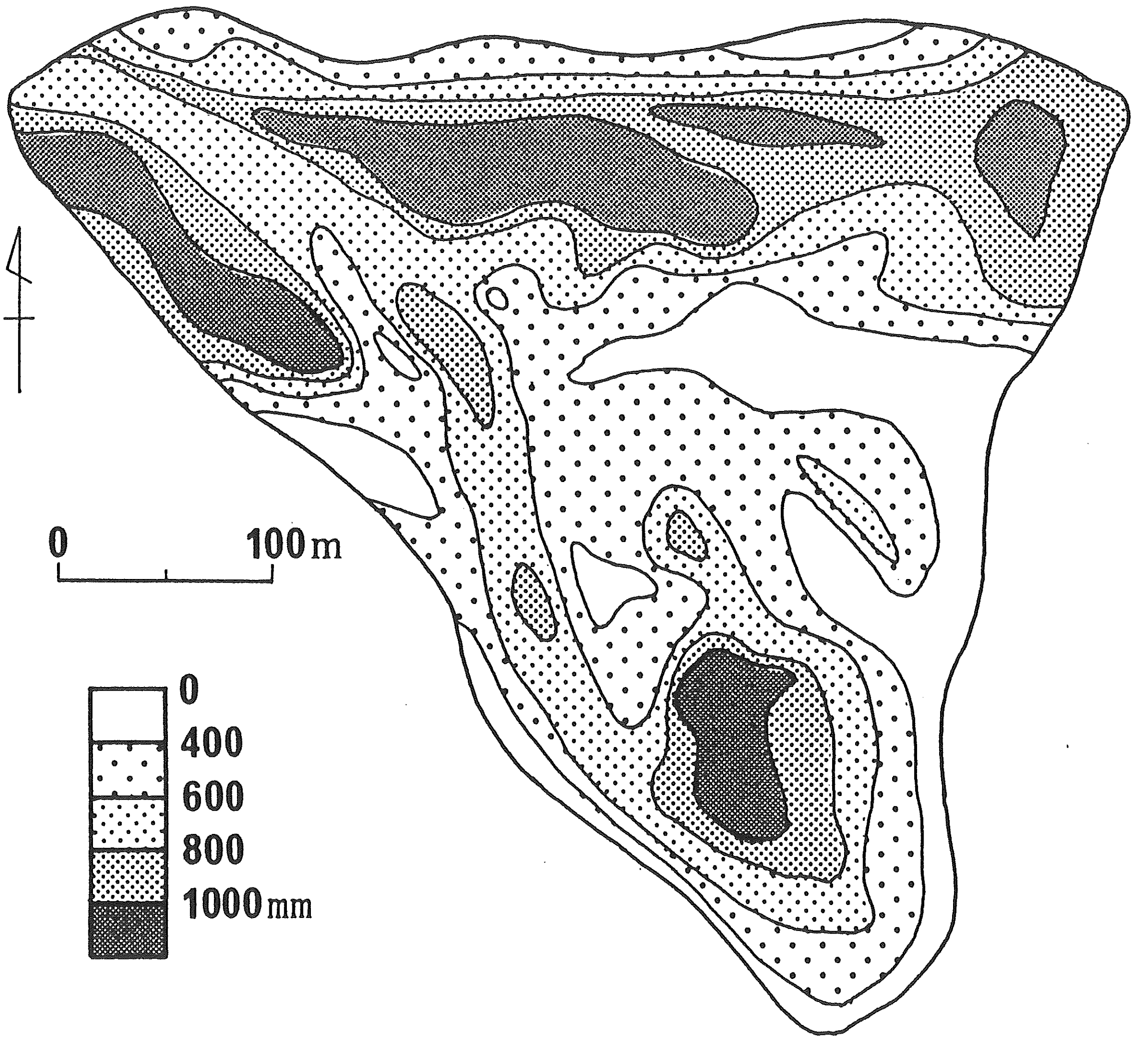


Fig. 8 Spatial distribution of soil moisture storage (units of depth) calculated by summing up the soil moisture contents of all soil samples. (add the data to Oginuma, 1988)

Chapter III

Methods

3-1. Field observations

Discharge and precipitation

In the Kawakami Experimental Basin, runoff has been observed continuously by a concrete weir with 90° V-notch since November, 1985 (Fig. 1). The water level was recorded continuously and was converted into the discharge by Numachi, Kurokawa and Fuchizawa's formula (Japan Society of Civil Engineers, 1971). Air temperature, precipitation, wind direction, wind speed and duration of sunshine have been observed at the meteorological station near the weir. As for precipitation, the observation can not be performed from January through March every year because of freezing of the instrument. So the precipitation data observed at a weather station of AMeDAS (Automated Meteorological Data Acquisition System operated by Japan Meteorological Agency) 4 km apart from the basin was used during this period.

Calibration of the instruments

The temporal change of soil moisture contents was measured by a neutron moisture meter (SX-8n type; Japan ETL co. ltd.). The source of the probe is californium-252 (50 μCi). The calibration of neutron moisture meter was conducted by comparing its readings with gravimetrically determined soil moisture of the samples taken at the mid-slope of site C (Fig. 1). Furthermore the count ratio values of water and of a completely dry sand (Toyoura Standard Sand; median diameter = 0.17 mm) were added to the data base obtained above, which allowed a determination of better calibration curve shown in Fig. 9.

The tensiometers with water pressure sensors (KDC-S5; KONA SYSTEM co. ltd.) were used for the observation of the pressure head of soil water for the long-time period. Some researchers have pointed out that the tensiometer with water pressure sensors tends to be affected by ambient air temperature (e.g. Tanaka *et al.*, 1993). So the effect of temperature on the tensiometer with pressure sensors was examined. At first under the atmospheric pressure the output values at various temperature were examined. Then under the various suction conditions the output of the sensor against various temperatures was tested. Figure 10 shows the output values plotted against the temperature under atmospheric pressure. The output values are plotted against pressure head regulated under some temperature conditions as shown in Fig. 11. The maximum variation of output values of sensor was 10 mV, which corresponds to 10 cmH_2O in suction. In conclusion this experiment showed that the effect of temperature on the tensiometer with pressure sensors is negligibly small and it is accurate enough for observations of soil water pressure head in the fields, because the variation of temperature during the period of observation in the field is about 30 °C at most.

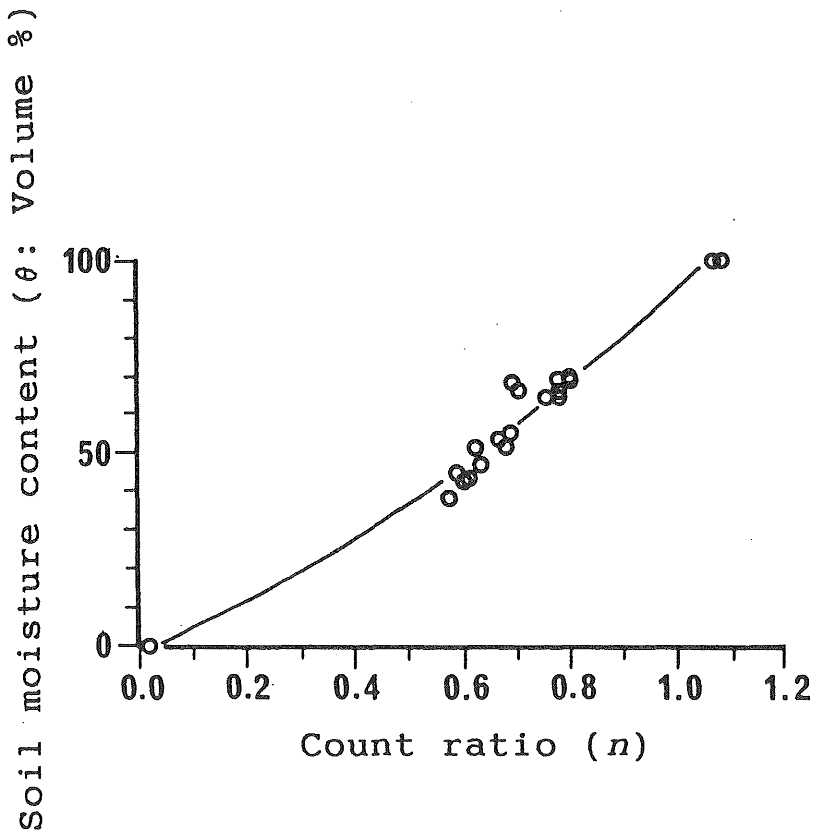


Fig. 9 Calibration curve for neutron moisture meter.
$\theta = -1.41 + 59.98n + 33.58n^2$

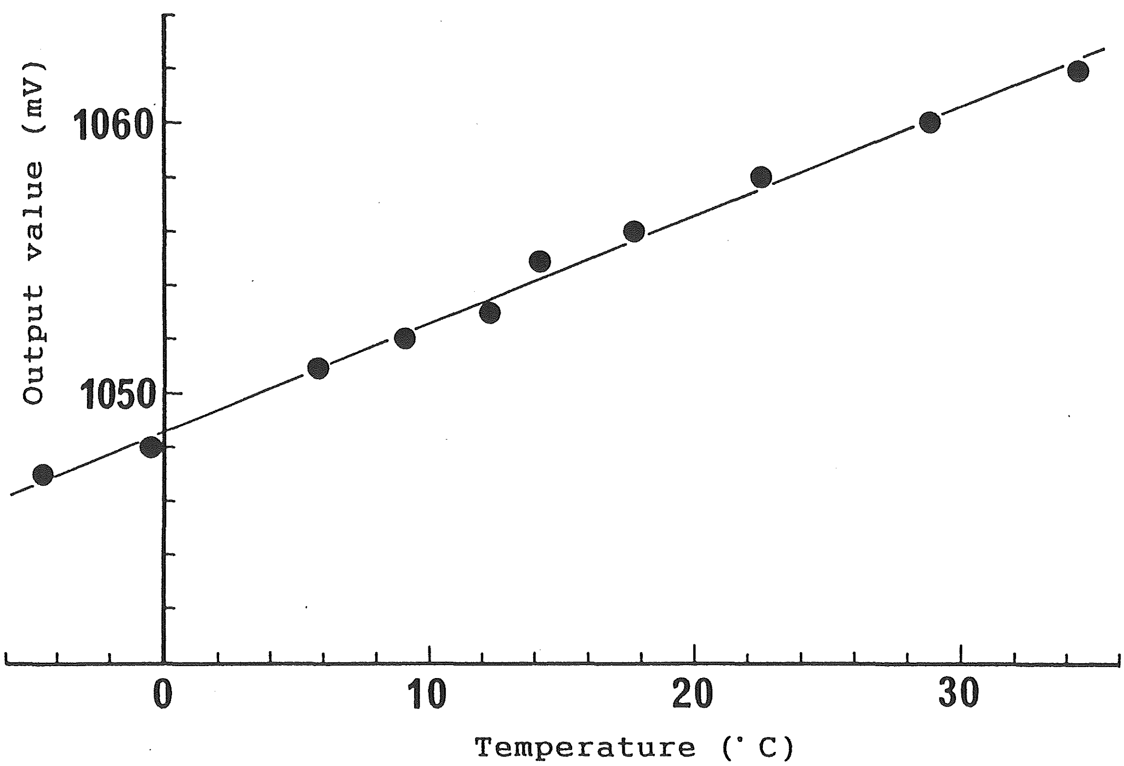


Fig. 10 Output values of pressure sensor plotted against the temperatures under air pressure.

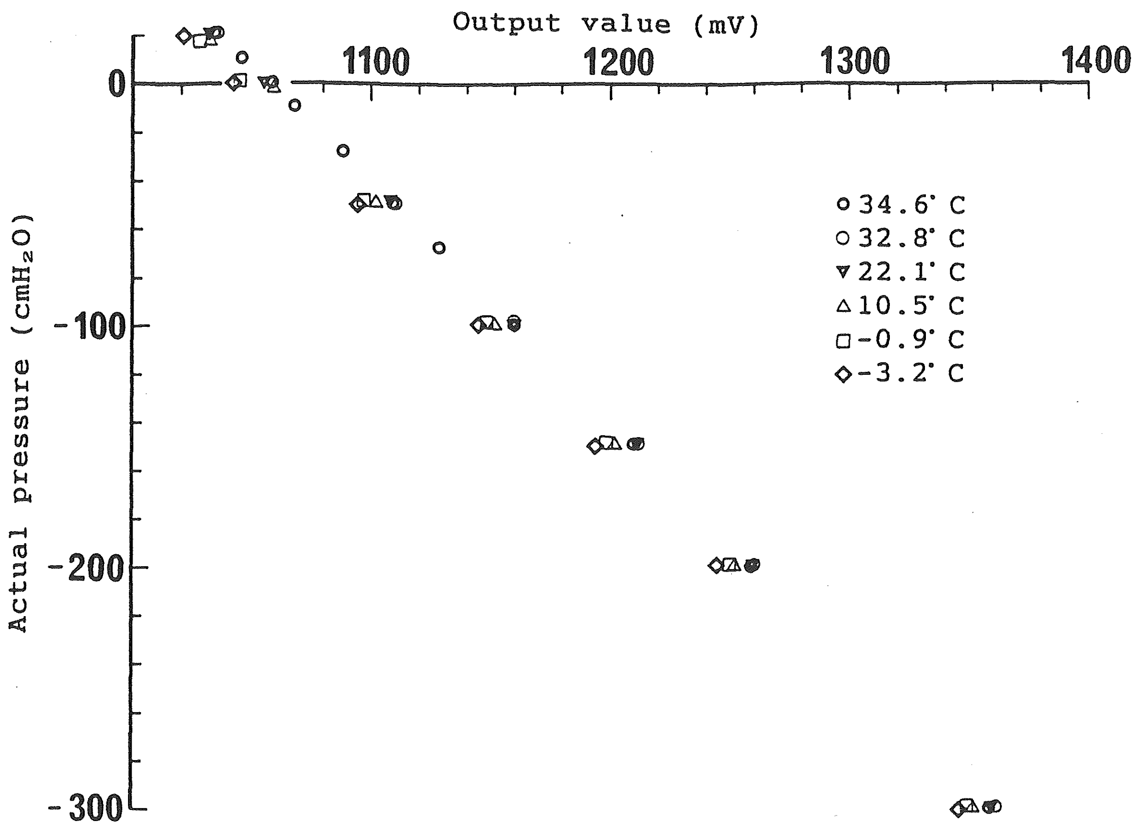


Fig. 11 Output values of sensor against pressure regulated under some temperature conditions.

Hydrometric observations

The hydrometric observations were performed at site C. The instrumentation is shown in Fig. 12. The temporal change in soil moisture contents was measured by the neutron moisture meter at the lower-, middle- and upper-points of the slope at site C. The access tubes for the neutron probe were installed down to the bedrock surface. The depth of each tube was 3.8 m at the lower-slope, 5.9 m at the mid-slope and 1.6 m at the upper-slope. The measurements were performed once in a month from April, 1991 through March, 1992. In this period, a few of measurements were conducted during a storm event between 20th and 23th in August, 1991. One minute measurement of count ratio was carried out at 0.2 m intervals from surface to the bottom of the access tube twice at each level.

The observations of the temporal and spatial variation of hydraulic head of soil water were performed during and after a storm event in July and August, 1989 at site C. The tensiometers with mercury manometers were used for observation of pressure head at lower-, middle- and upper-points of the slope (Fig. 12). The depths of porous cups at each plot are listed in Table 2 and the schematic diagram of tensiometers is shown in Fig. 13. To estimate the gradient of pressure head along the direction of soil surface, a pair of tensiometer with the porous cup at the same depth was installed at each plot as shown in Fig. 13. An intensive manual observation of pressure head by using these

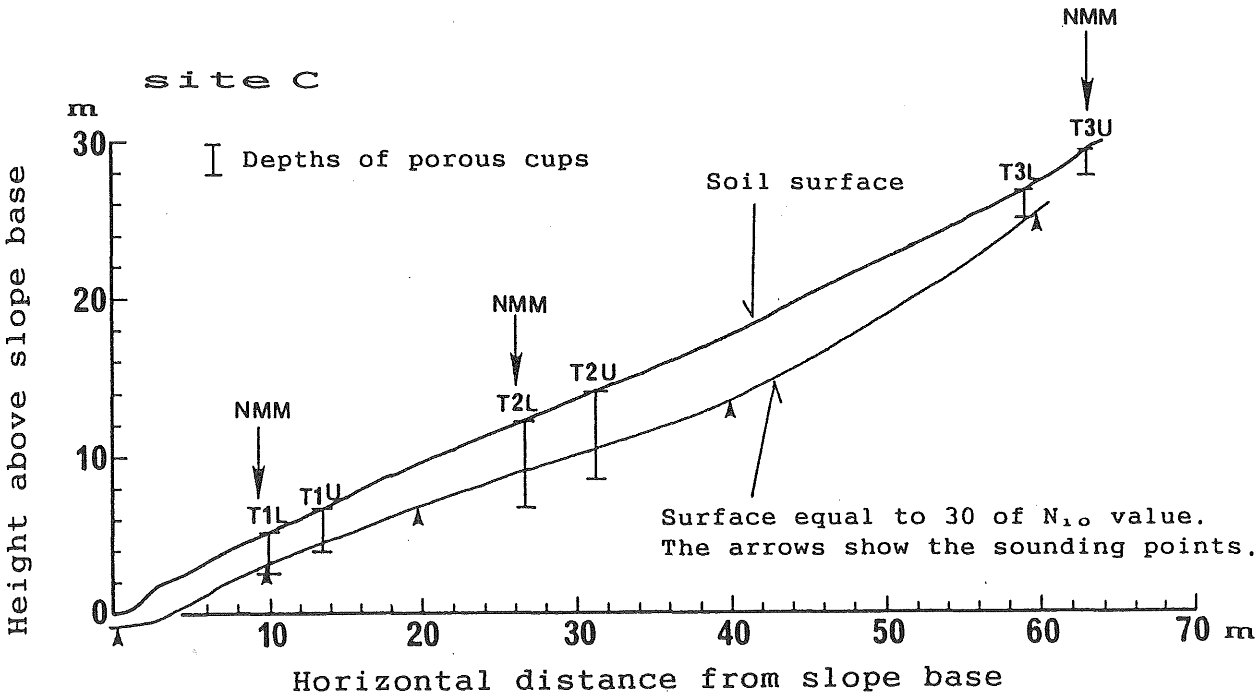


Fig. 12 Instrumentation for the observations of soil moisture and hydraulic head at site C.
T1L,T1U..: Setting points of tensiometers,
NMM: Measurement points of soil moisture by the neutron moisture meter.

Table 2 Depths of tensiometers with mercury manometers.

T1L,U	T2L,U	T3L,U
0.2m	0.2m	0.2m
0.6	0.6	0.6
1.2	1.2	1.2
2.0	2.0	1.6
2.8	2.8	
	3.6	
	4.6	
	5.6	

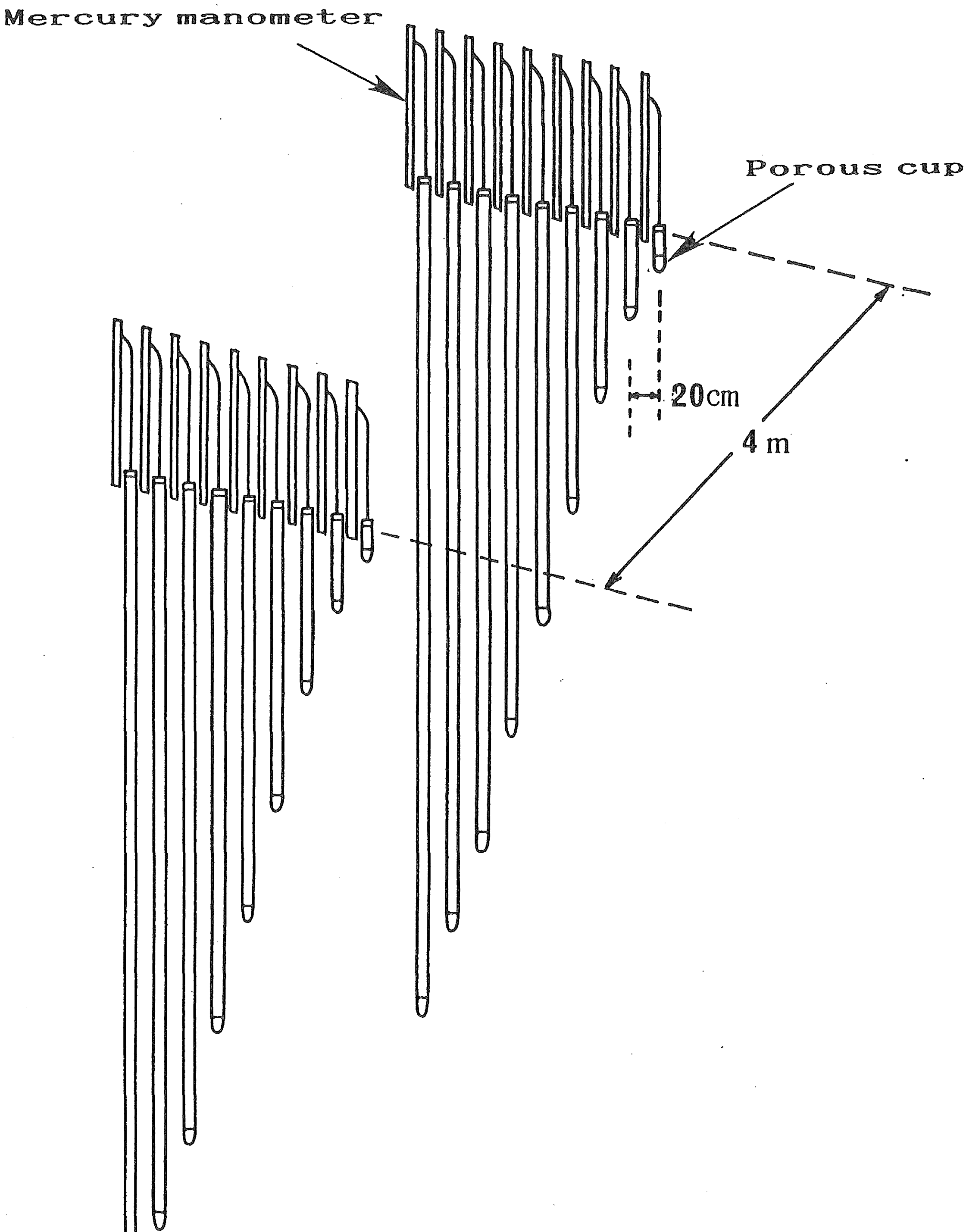


Fig. 13 Schematic diagram of tensiometers.

tensiometers was conducted from 7th through 21th, July (during storms) and from 5th through 17th, August (during a rainless period) in 1989.

Observation of the hydraulic head in the long-time period and water samplings

The observation of hydraulic head profiles in the long-time period and water sampling were performed at sites A and C (Fig. 1). The instrumentation for the measurement is illustrated in Fig. 14. The schematic illustration of the detailed measurement system at both sites is shown in Fig. 15. The pressure head of soil water was measured by tensiometers with water pressure sensors located at depths of 0.1, 0.3, 0.5, 0.7, 1.0 and 1.5 m. The outputs from the pressure sensor were recorded by a data logger at 30 minutes interval continuously from August through December in 1991.

As for water sampling, rainfall and throughfall were collected by a 210-mm diameter funnel with a small ball for prevention of evaporation (Shimada and Sanjo, 1987) and stored in polyethylene tanks. Stemflow was collected by vinyl sheets bound around the trunk (Suzuki *et al.*, 1979) and stored in polyethylene tanks. The water samples in these tanks were collected once in a month for isotopic analyses. Also, total amount of throughfall and stemflow were evaluated at the same time.

Soil water was sampled by a suction sampler which consists of a porous cup with a diameter of 17 mm, a glass collecting bottle and a connecting tube with a diameter of 2

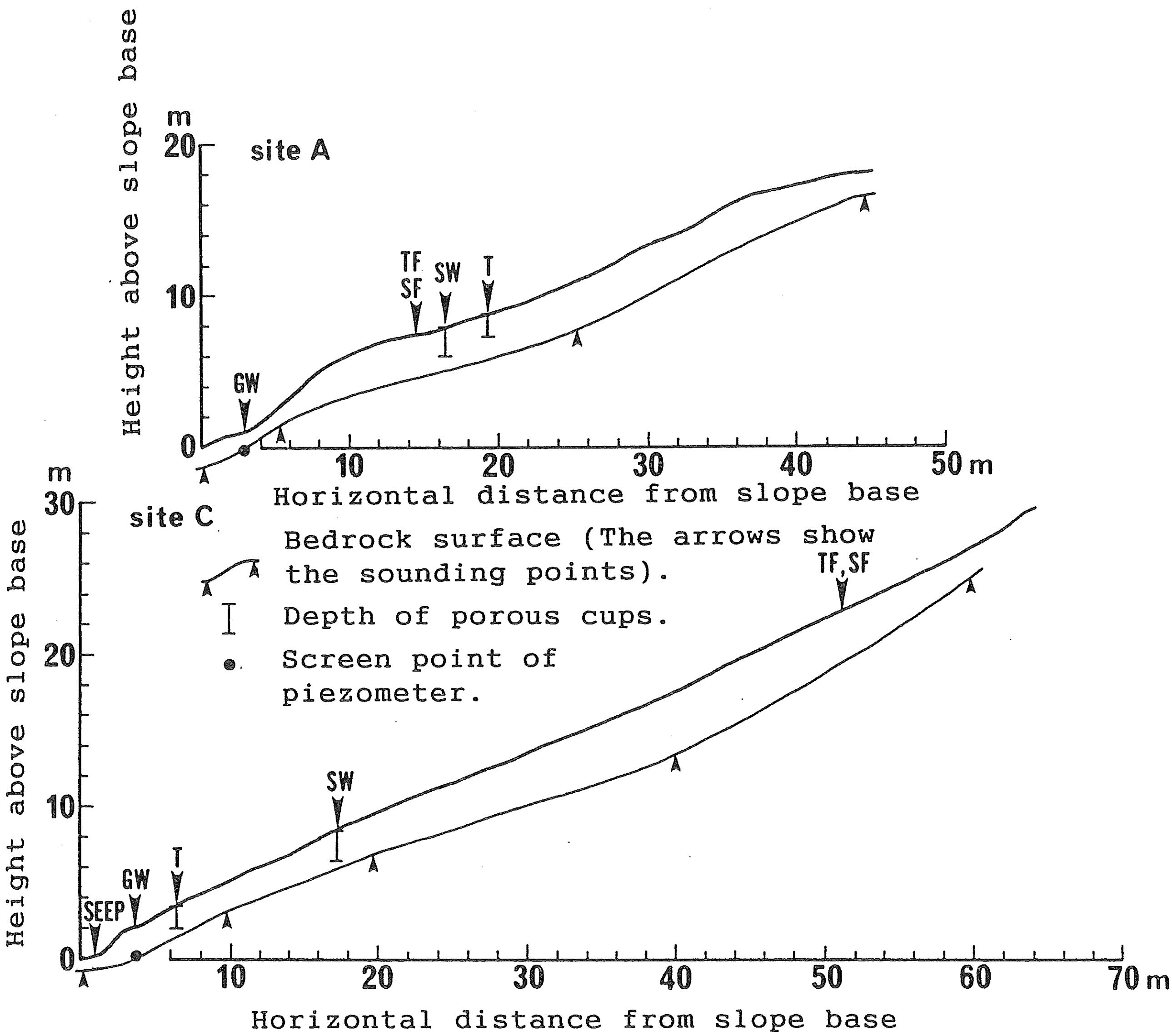


Fig. 14 Instrumentation for hydrometric observations and water sampling at sites A and C.
T: Tensiometers; SW: Suction sampler; TF: Throughfall sampler; SF: Stemflow sampler; GW: Piezometer for sampling of groundwater; SEEP: Sampling point of seepage water.

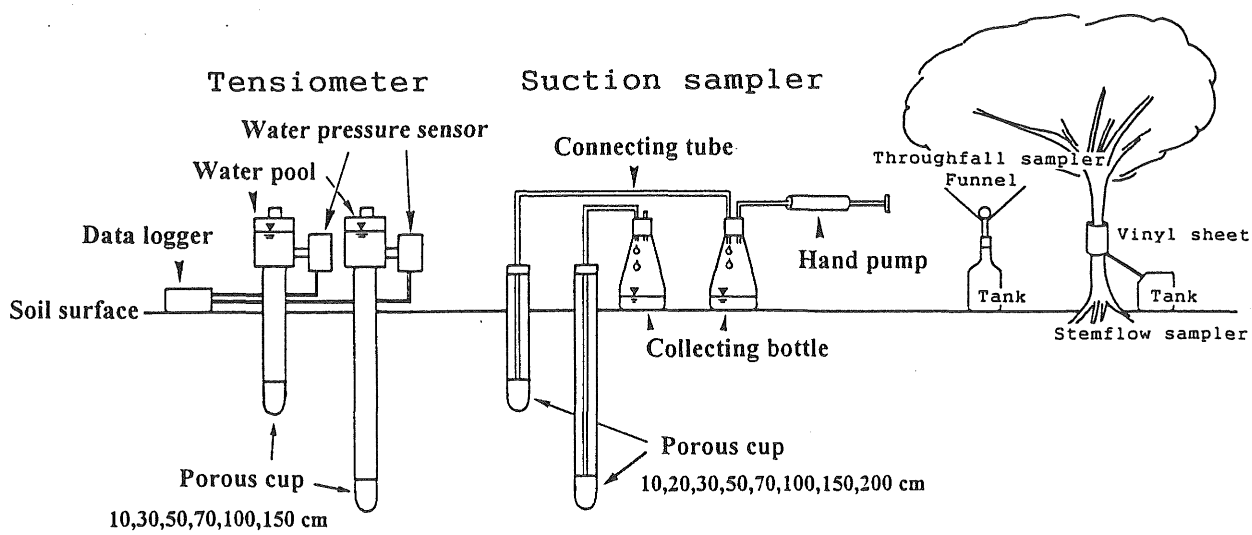


Fig. 15 Schematic illustration of the instrumentations for hydrometric observations and water sampling at sites A and C.

mm (Fig. 15). A set of the device was installed at depths of 0.1, 0.2, 0.3, 0.5, 0.7, 1.0, 1.5 and 2.0 m at each site. To extract soil water a small hand pump was connected to the device and a 60 to 70 kPa suction (corresponding to -612 cmH_2O to -714 cmH_2O) was applied to the bottle and tube. Soil water extracted through the porous cup was collected in the bottle via the tube. The soil water sampled by this method is a mixture of water which has a matric potential between 0 and -714 cmH_2O that is under pF 2.85. The water of matric potential under pF 2.85 corresponds to gravitational water and part of suspended water. Therefore the extracted water was probably active in percolation and evapo-transpiration processes.

Groundwater was collected directly from the piezometer with a single screen. Discharge water was sampled at the weir. All water samplings were performed once in a month from May through November in 1991, and the samples were preserved in the vials of 20 ml volume.

3-2. Stable isotope analyses of water

Stable isotope variation of water is expressed as a ratio of the less abundant isotopes such as D and ^{18}O against the more abundant isotopes such as ^{1}H and ^{16}O respectively (Faure, 1986). In general, the stable isotopic compositions are reported in terms of difference of $^{18}O/^{16}O$ and D/H ratios relative to a standard called SMOW (Standard Mean Ocean Water). This standard was defined by Craig (1961) with reference to a large volume of distilled water distributed by the National Bureau of Standards in the United States (NBS-1), such that:

$$D/H\ (SMOW) = 1.050\ D/H\ (NBS\text{-}1) \qquad (3.1)$$

$$^{18}O/^{16}O\ (SMOW) = 1.008\ ^{18}O/^{16}O\ (NBS\text{-}1) \qquad (3.2)$$

The isotopic compositions of deuterium and oxygen-18 of a sample are expressed as per mil differences relative to SMOW:

$$\delta D = \left[\frac{(D/H)_{sample}}{(D/H)_{SMOW}} - 1 \right] \times 1000 \qquad (3.3)$$

$$\delta^{18}O = \left[\frac{(^{18}O/^{16}O)_{sample}}{(^{18}O/^{16}O)_{SMOW}} - 1 \right] \times 1000 \qquad (3.4)$$

Stable isotopic ratios of D and ^{18}O in water samples were analyzed at the Institute for Study of the Earth's Interior, Okayama University.

For deuterium analysis, liquid water sample is deoxydized by a modified granulated zinc reduction method (e.g. Coleman *et al.*, 1982) into H_2 gas before measurement. This method reduces a sample of 5 μl water into H_2 gas over 0.2 g of granulated zinc in a reaction tube under the temperature of 450 °C for about 30 minutes. Sample H_2 gas is then introduced into inlet system and D/H ratio is determined using a HITACHI RMS-HD mass spectrometer.

For oxygen-18 analysis, a CO_2 equilibration method is used for the preparation of a sample. Forty small cups in which there is 5 ml of water sample are set into the automatic sample preparation system. After the CO_2 gas is introduced into the all cups, the system shakes the cups for 4.5 hours to make equilibrium condition between CO_2 gas and the water sample in each cup. Then $^{18}O/^{16}O$ ratio of the equilibrated CO_2 gas is successively measured by automatic mass spectrometer, VG MICROMASS 903. Chiba *et al.* (1985) describes more details on this procedure.

Chapter IV

Water budget

The water budget of a basin is given by the following equation,

$$P = Q + ET + \Delta S \qquad (4.1)$$

where P is the precipitation, Q is the discharge, ET is the evapotranspiration and ΔS is the change of soil water storage. The ΔS can be neglected in consideration of the annual water budget. The annual water budget of the Kawakami basin from 1986 through 1991 is shown in Table 3. The annual runoff ratio ranges from 50 to 60% in the Kawakami basin. In the following sections it will be discussed in more detail about each component of water budget equation to understand hydrological characteristics of the basin.

4-1. Rainfall, throughfall and stemflow

In this section the problem of input to the basin will be discussed with emphasis on the throughfall and the stemflow.

As mentioned above monthly volumes of throughfall and stemflow were measured at sites A and C. Figure 16 shows the monthly throughfall plotted against the monthly rainfall measured at weather station of AMeDAS between May, 1991 and

Table 3 Annual water budget of the Kawakami Experimental basin.

Year	Precipitation mm	Runoff mm	Loss mm	Ratio of runoff %
1986	1151	607	544	52.7
1987	1191	>387	<804	
1988	1484	821	663	55.3
1989	1712	1058	654	61.8
1990	1358	>503	<855	
1991	1837	988	849	53.8

(From Matsutani *et al.*, 1993)

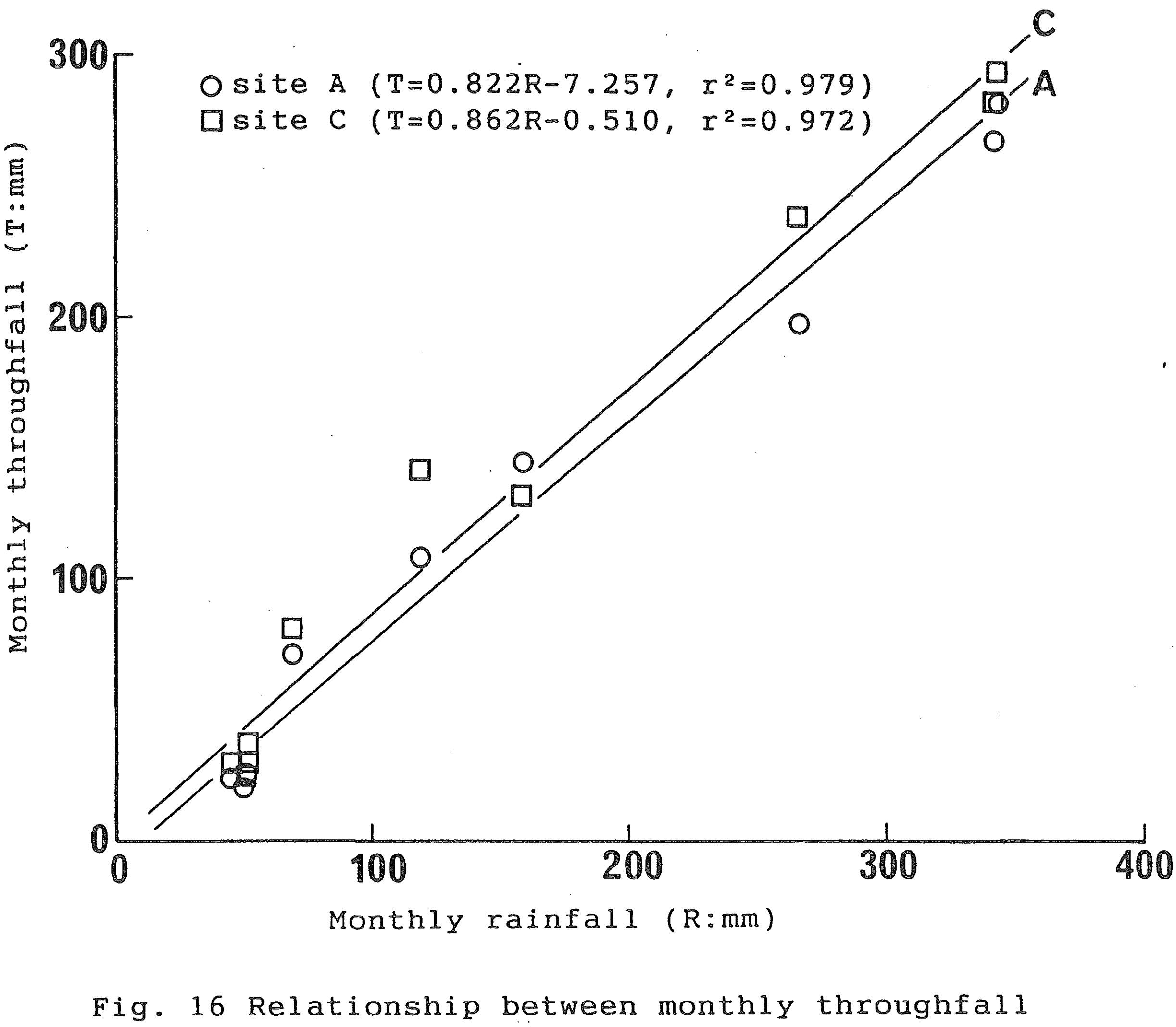


Fig. 16 Relationship between monthly throughfall and rainfall from May, 1991 through February, 1992.

February, 1992. The figure clearly shows that the average ratio of throughfall against the rainfall was 82 % at site A (larch forest) and 86 % at site C (oak forest). Nakano (1976) mentioned that the throughfall accounted for 70 to 95 % of the rainfall in the deciduous forest and 70 to 75 % in the artificial larch forest generally in Japan. Thus the volume of throughfall in the Kawakami basin appears to be a standard value in the Japanese forests.

A careful attention has to be paid in consideration of input process of the stemflow to the soil surface (Tanaka *et al.*, 1991). In general, the amount of stemflow is presented as a quotient per canopy projected area. The stemflow of the basin calculated by this method is only about 1 % of rainfall (Fig. 17). However the stemflow infiltrates the soil surface as a point source around the tree base. Tanaka *et al.* (1991) showed the empirical relationship between the diameter of a tree base (x) and the radius of infiltration area of stemflow (y) from their field observations as follows:

$$y = 25.07 \ln x - 34.92 \qquad (4.2)$$

The result of recalculation of stemflow by using this equation shows that the stemflow per actual infiltration area was 20 % of rainfall at site A and 27 % at site C (Fig. 17). Therefore, a large amount of water converges and infiltrates around the tree base. However in consideration of average input to the soil surface the situation is a

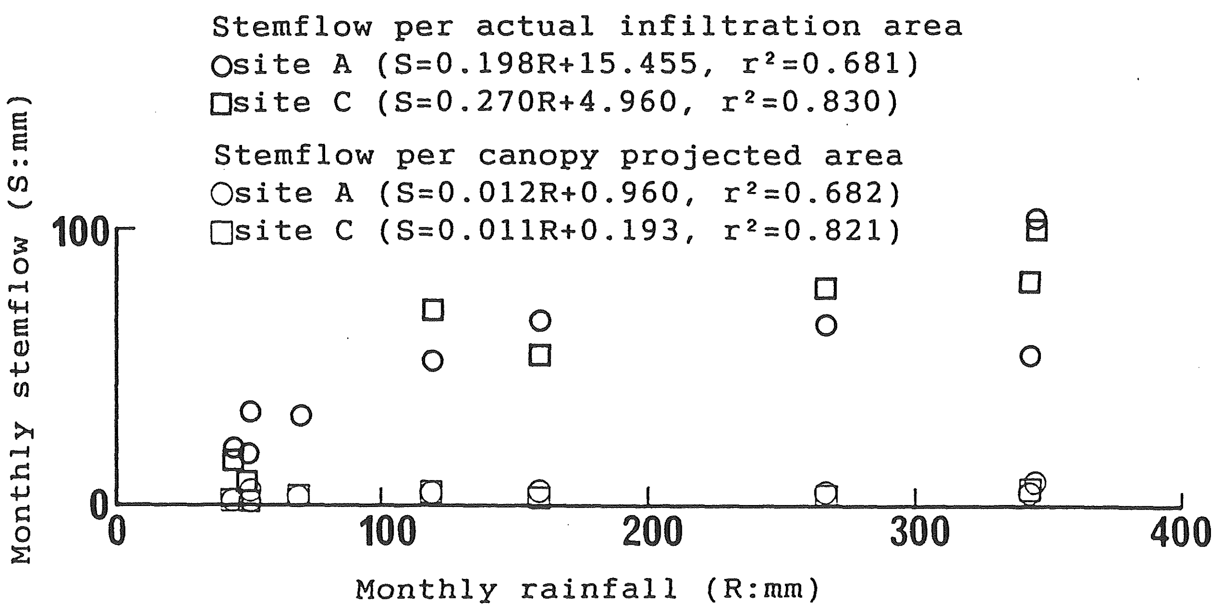


Fig. 17 Relationship between monthly stemflow and rainfall from May, 1991 through February, 1992.

little more complicated. For this purpose the stemflow per actual infiltration area for each tree must be accumulated within the observation plot, and the accumulated volume have to be divided by the area of the plot. The stemflow recalculated by this method is almost the same as the value per canopy projected area. In this study the average input to the forest floor in the plot is more important, and thus the throughfall can be considered as a major component as an input to the forest floor and the stemflow is negligibly small in this quantitative consideration.

In conclusion the rainfall interception in the basin was 13 to 17 % on average and this appears to be within the range of standard value in Japanese forests reported by Nakano (1976) and Suzuki *et al.* (1979).

4-2. Runoff characteristics

In order to understand the runoff characteristics of the Kawakami Experimental Basin quantitatively and to compare with other basins, a simple runoff separation of quick flow in hydrograph by Hewlett and Hibbert (1967) was conducted. This method has been popular for its simple procedure and is probably good enough to understand the runoff characteristics. Moreover this method has been used by many researchers so it is possible to compare the results with those obtained in other basins. In Hewlett and Hibbert method the quick flow (QF) is defined as a component of hydrograph which is separated by an increasing discharge line having the gradient of 0.05 cubic feet per second per square mile per hour (corresponding to 0.5467 $l/sec/km^2/hr$).

The runoff analysis was performed in July and August, 1989. In this period the intensive hydrometric observation was also conducted as described in section 5-1. The list of results are shown in Table 4 and the hydrographs in which QF was separated are illustrated in Fig. 18. The QF ratio defined as the QF per total amount of rainfall ranges from 2.5 to 28.5 %, and the average ratio is 10 %. The QF ratio in various basins by some researchers are listed in Table 5 for comparison. The QF ratio in the Kawakami basin shows relatively low value and the variation of the ratio is small. Hirose *et al.* (1993) compared the runoff characteristics of small basins underlain by different bedrocks and suggested that the amount of quick flow depended on the thickness of soil mantle to a certain degree. It appears that the QF ratio of the basins with a thin soil mantle varies in a wide range (e.g. Mosley, 1979 and Pearce *et al.*, 1986: Maimai in New Zealand). On the other hand, the QF ratio of the basins with a thick soil mantle tends to be low and is not so variable (e.g. Onda, 1989; Hirose *et al.*, 1993). The QF ratio observed in the Kawakami basin appears to be result of the thick soil mantle cover and the high water retentivity of the soil.

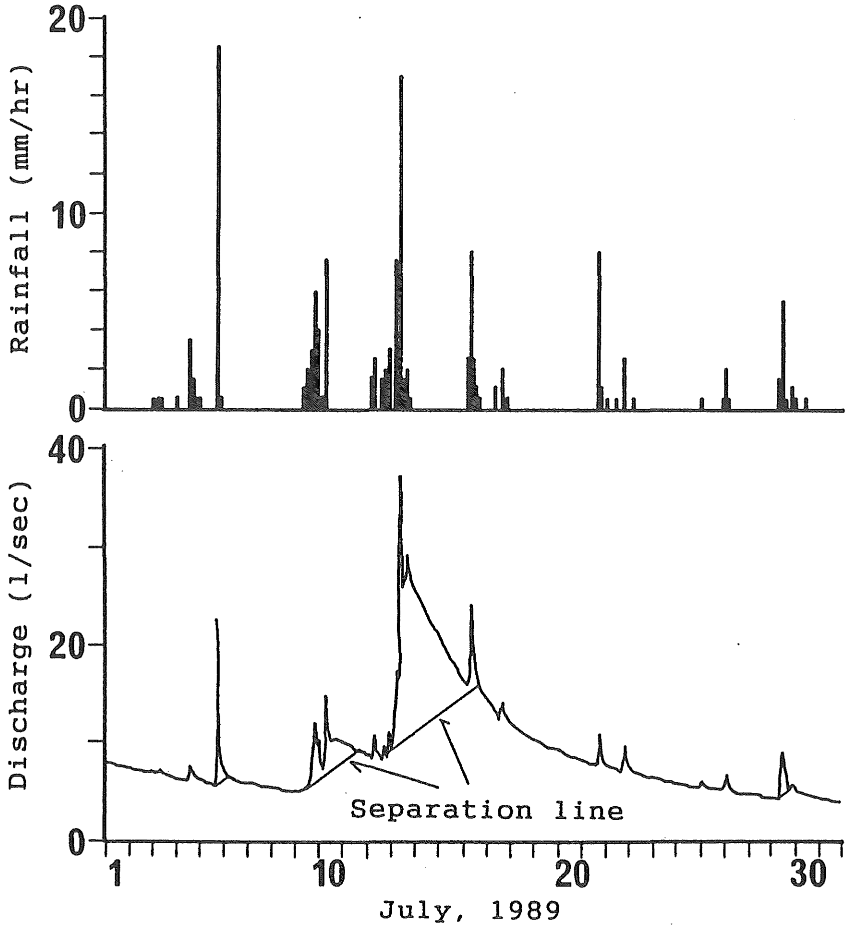
4-3. Temporal variation of soil moisture contents

In this section, the temporal change of soil moisture

Table 4 Separation of quick flow in hydrograph by Hewlett and Hibbert (1967) method in July and August, 1989.

Period	P(mm)	QF(mm)	QF/P(%)	ID(l/sec)
Jul.5/16:00-6/9:00	19.0	0.8	4.1	5.5
Jul.9/16:00-11/16:00	52.0	3.4	6.6	5.5
Jul.13/3:00-16/16:00	72.0	20.4	28.4	9.2
Jul.21/16:00-20:00	9.0	0.3	3.3	7.9
Jul.29/8:00-19:00	20.0	0.5	2.5	4.1
Aug.5/24:00-8/12:00	63.0	6.5	10.3	5.3
Aug.27/6:00-28/12:00	67.3	3.1	4.7	2.4

P: Total amount of rainfall within a period.
QF: Quick flow.
ID: Initial discharge rate.



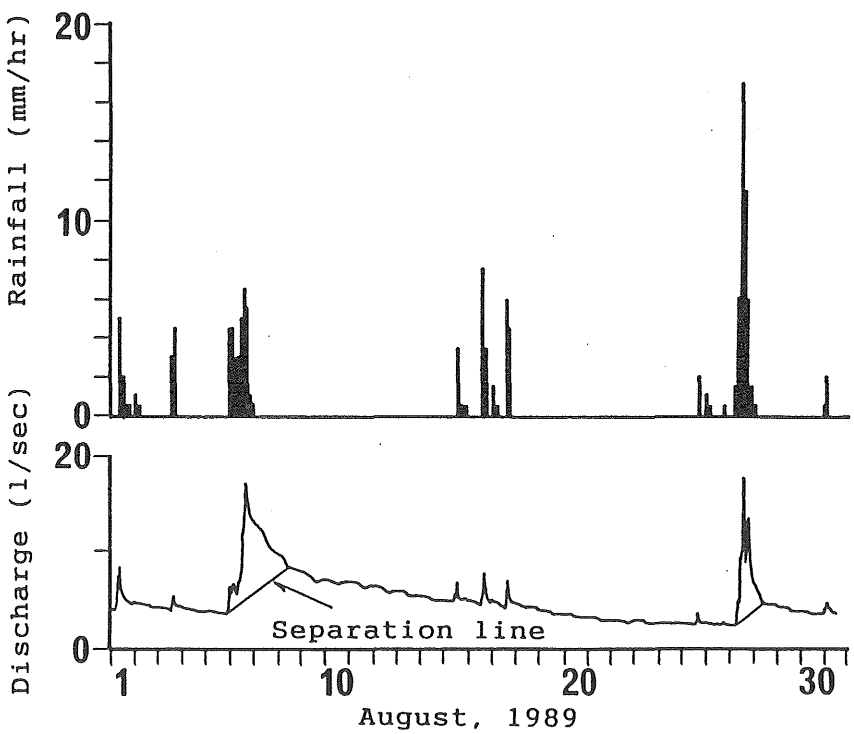


Fig. 18 Quick flow separation in hydrograph by Hewlett and Hibbert (1967) method.

Table 5 QF ratios in previous studies.

References	Study area	Geology	Soil thickness(m)	QF/P (%)
Harr(1977)	Oregon, USA	Volcanic breccia	1~2	23~42
Mosley(1979) Pearce *et al.* (1986)	Maimai, New Zealand	Pleistocene conglomerate	0.6	3~72
Ohta & Takahashi (1986)	Iwate, Japan	Tertiary	0.6~1	<38
Kubota *et al.* (1987)	Shiga, Japan	Weathered granite	0.75	<27
Onda(1989)	Obara, Japan	Fine-grained granite	2~5	5~6
		Course-grained granite	1	40
Hirose *et al.* (1993)	Abukuma, Japan	Granite	1	1.5~7.8
		Granodiorite	2	1.6~3.8
		Gabbro	2~3	0.9~7.0
		Limestone	<0.5	0.2~0.3

contents for a year will be discussed to understand the soil water movement from the view point of the soil moisture. The temporal variations of soil moisture contents measured by the neutron moisture meter are illustrated as the isopleths in Fig. 19 at the upper, middle and lower points of the slope from April, 1991 through March, 1992. It can be pointed out that the range of temporal change of soil moisture contents was usually as small as 5 % by volume in the maximum, with a few exceptions when the saturated zone being formed above the bedrock at the middle and lower points.

In the lower-slope, the soil moisture content above the 1 m depth changed notably, however under the 1 m the soil moisture was quite stable. Above the bedrock surface the saturated zone was observed, because the groundwater in the vicinity of the valley bottom expanded up to this point in September and October.

In the mid-slope, the depth where the soil moisture varied considerably was above 1 m. Under the 1 m the variation was not very notable, and especially from 2.5 to 3.5 m the soil moisture was stable. On the contrary, from 3.5 to 5 m depth the change was relatively large. And periodically, above the bedrock surface a saturated zone was observed.

In the upper-slope, the variation of soil moisture over the 0.7 m depth was noticeable and under this depth it was relatively stable. Although, this point is near the divide, a considerable decrease of soil moisture was not observed

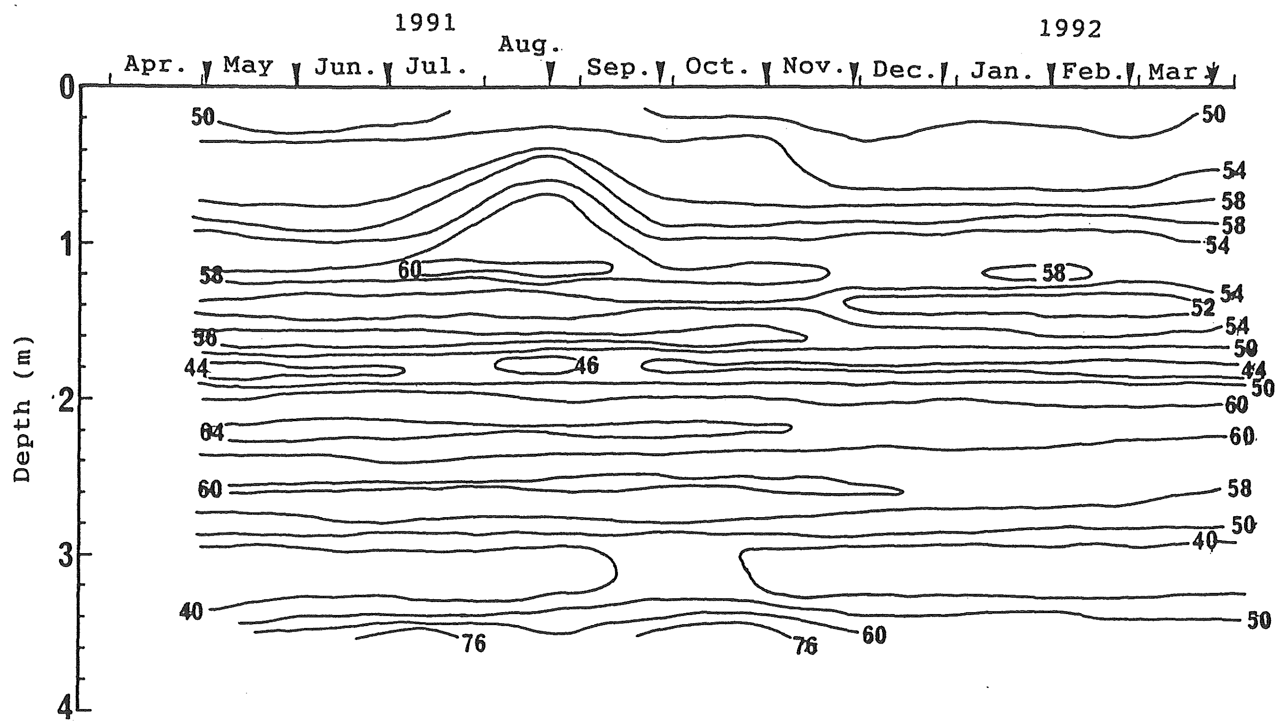


Fig. 19-a Isopleth of soil moisture contents (volume %) at the lower-slope (the arrows show the dates of measurements).

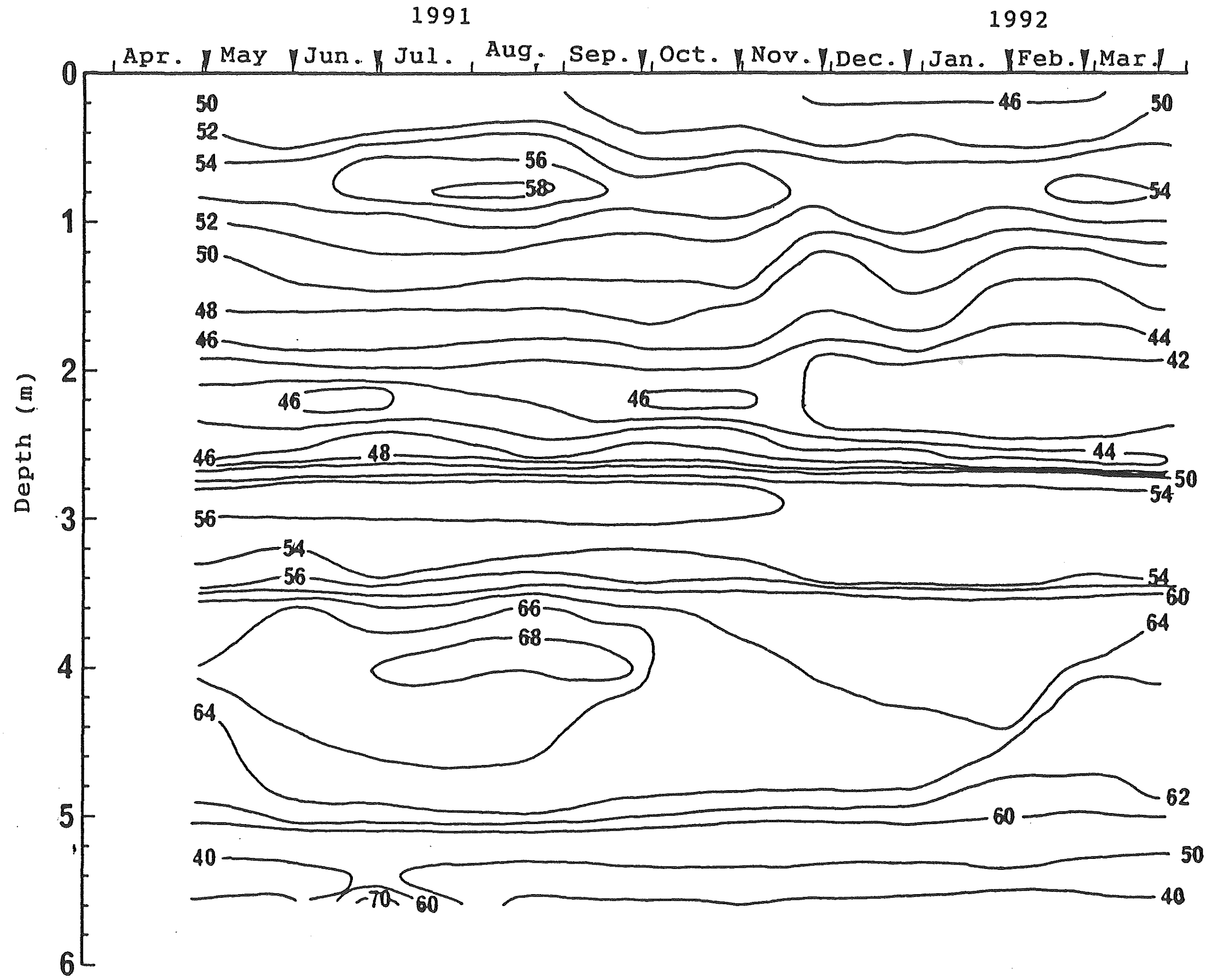


Fig. 19-b Isopleth of soil moisture contents at (volume %) the mid-slope (the arrows show the dates of measurements).

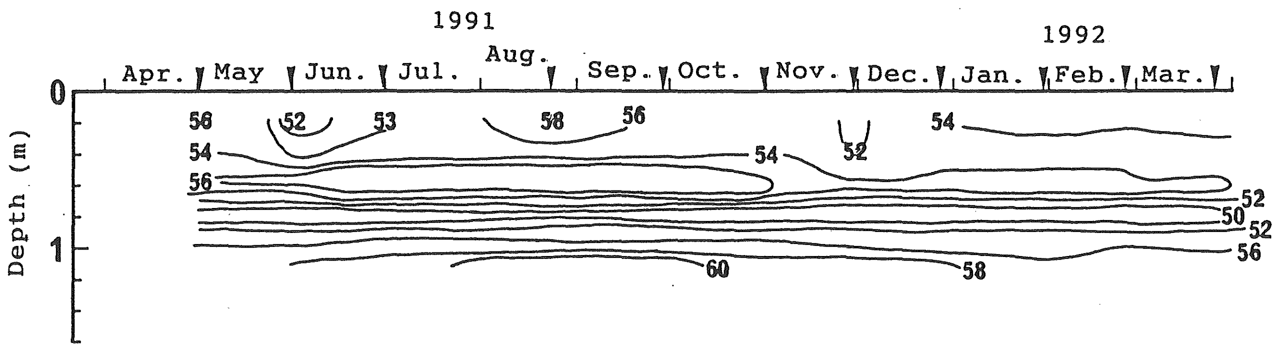


Fig. 19-c Isopleth of soil moisture contents at (volume %) the upper-slope (the arrows show the dates of measurements).

even under the dry conditions.

Temporal change in the total soil moisture contents of whole soil layer in the mid-slope where the change in soil moisture was the most considerable is shown in Fig. 20. The difference of soil moisture content between June and February was 4.1 % by volume, i.e. 226 mm/5500 mm. Because the average thickness of soil mantle is 1.63 m in the basin, this change of soil moisture of 4.1 % corresponds to 67 mm in the seasonal water budget. This suggests that the soil moisture variation is relatively small and the soil moisture content is stable in the basin, though there are a few exceptions for the soil layer above the 1 m depth and on the surface of the bedrock. This is caused mainly by the physical properties of the soil mantle, especially in the B horizon. Because the change in soil moisture content with change in suction of soil water is quite small, and it may cause less temporal variation of soil moisture content in the basin.

4-4. Estimation of evapotranspiration by using the short-time period water budget method

It is an important subject to estimate actual evapotranspiration rate in a basin for evaluating not only the meteorological condition of the basin but also the actual infiltration rate into the soil mantle. The evapotranspiration of a forested basin can be distinguished into three components, i.e. interception loss, transpiration

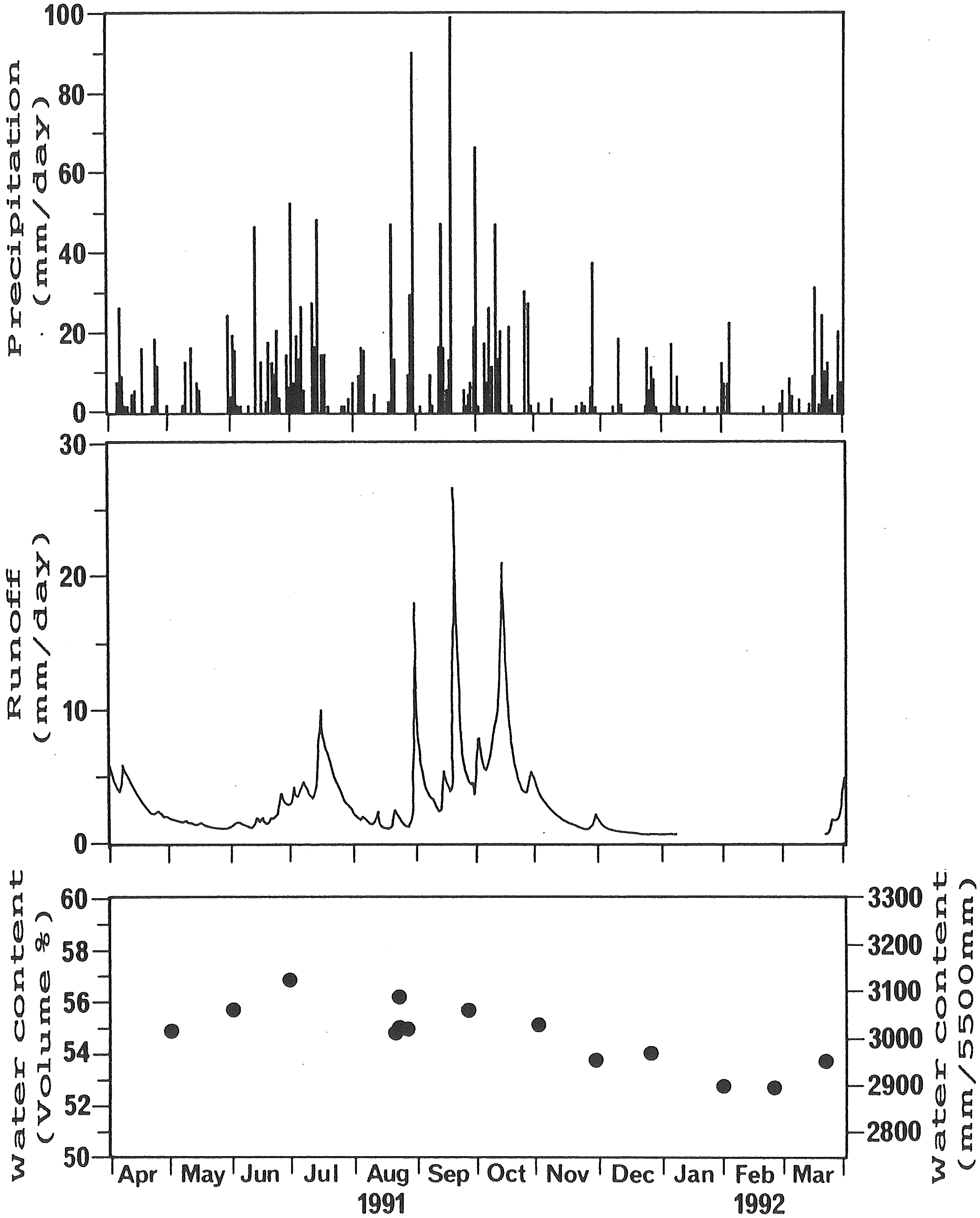


Fig. 20 Temporal change of soil moisture content of total soil mantle at the mid-slope. (units of volumetric water content: % and depth: mm)

from the vegetation and evaporation from the forest floor (Tani, 1989). Today's measurement technology dose not allow the evaluation of each component with enough accuracy. Therefore it is probably the best alternative strategy to elucidate the total amount of evapotranspiration from the water budget data.

The short-time period water budget method was developed by Linsley *et al.* (1958) and applied to some basins (e.g. Hamon, 1961; Takase and Maruyama, 1978; Suzuki, 1980; 1985). The principle of the method can be summarized as follows:

$$S(t) = f[q(t),\ dq/dt,\ a,\ b, \ldots] \qquad (4.3)$$

where $S(t)$ is the water storage in a basin, $q(t)$ is the discharge rate, a, b,... are the parameters to be determined for each basin. If the basin has the same value of $q(t)$ and dq/dt at time t_1 and time t_2, the water storage $S(t_1)$ and $S(t_2)$ should also be the same. Assuming $\Delta S(t)=0$, the water budget calculation can be conducted for the period from t_1 to t_2 under this condition. Evapotranspiration E is given as:

$$E = P - Q = \int_{t1}^{t2} p(t)\,dt - \int_{t1}^{t2} q(t)\,dt \qquad (4.4)$$

where P and Q are the total amount of rainfall and runoff within the period from t_1 to t_2 respectively, $p(t)$ is the rainfall intensity, and $q(t)$ is the discharge rate.

To select the appropriate water budget period between t_1 and t_2, the following criteria (modified Suzuki (1985) method) were used;

1) both t_1 and t_2 should be the last day of three or more consecutive days without rainfall,

2) daily discharge on day t_1 and that on day t_2 should agree within 2 %,

3) the period between t_1 and t_2 should be in the range of 8 and 120 days.

The procedure of 1) is needed for excluding the effect of quick flow. To omit the period under 8 days is necessary for avoiding the variation of evapotranspiration rate, and the period over 120 days is not suitable for the objective of the present study.

After applying the procedure mentioned above, seasonal and daily evapotranspiration values of the basin are calculated from 1989 through 1991 by the eqn. (4.4).

Figure 21 shows the change of evapotranspiration calculated from 1989 through 1991 together with the hydrograph and the hyetograph. Because the soil between surface and about 0.3 m depth is frozen from January through March and the effect of meltwater on runoff characteristics usually remains until the beginning of April, it is not possible to apply the eqn. (4.4) to that period. From the figure it is clear that the daily evapotranspiration rate changes seasonally. The rate in July and August is about 3.0 mm/day and that of May, October and November is about 2.0 mm/day on average. In 1990 which can be characterized by

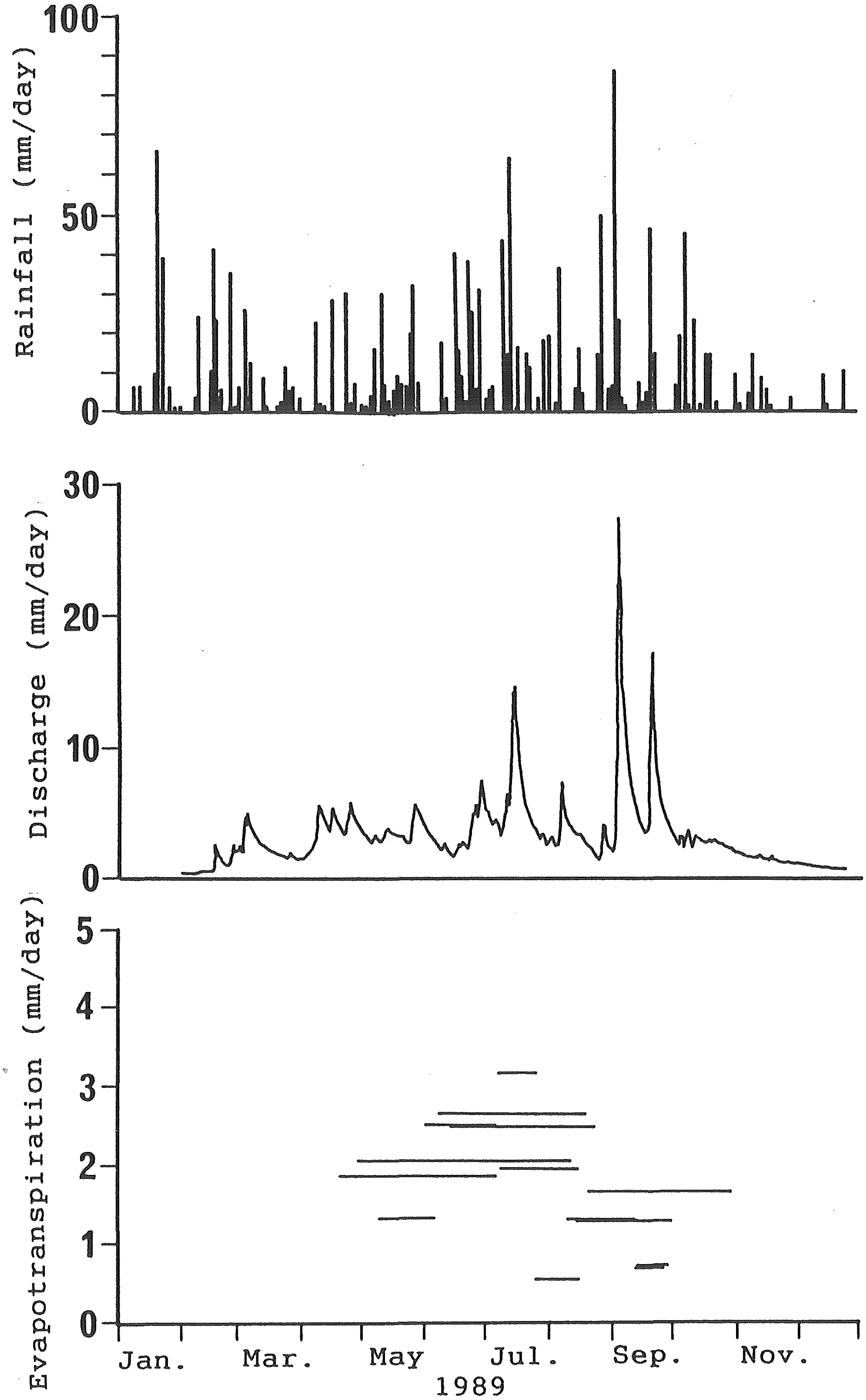


Fig. 21-a Change of daily evapotranspiration rate in 1989

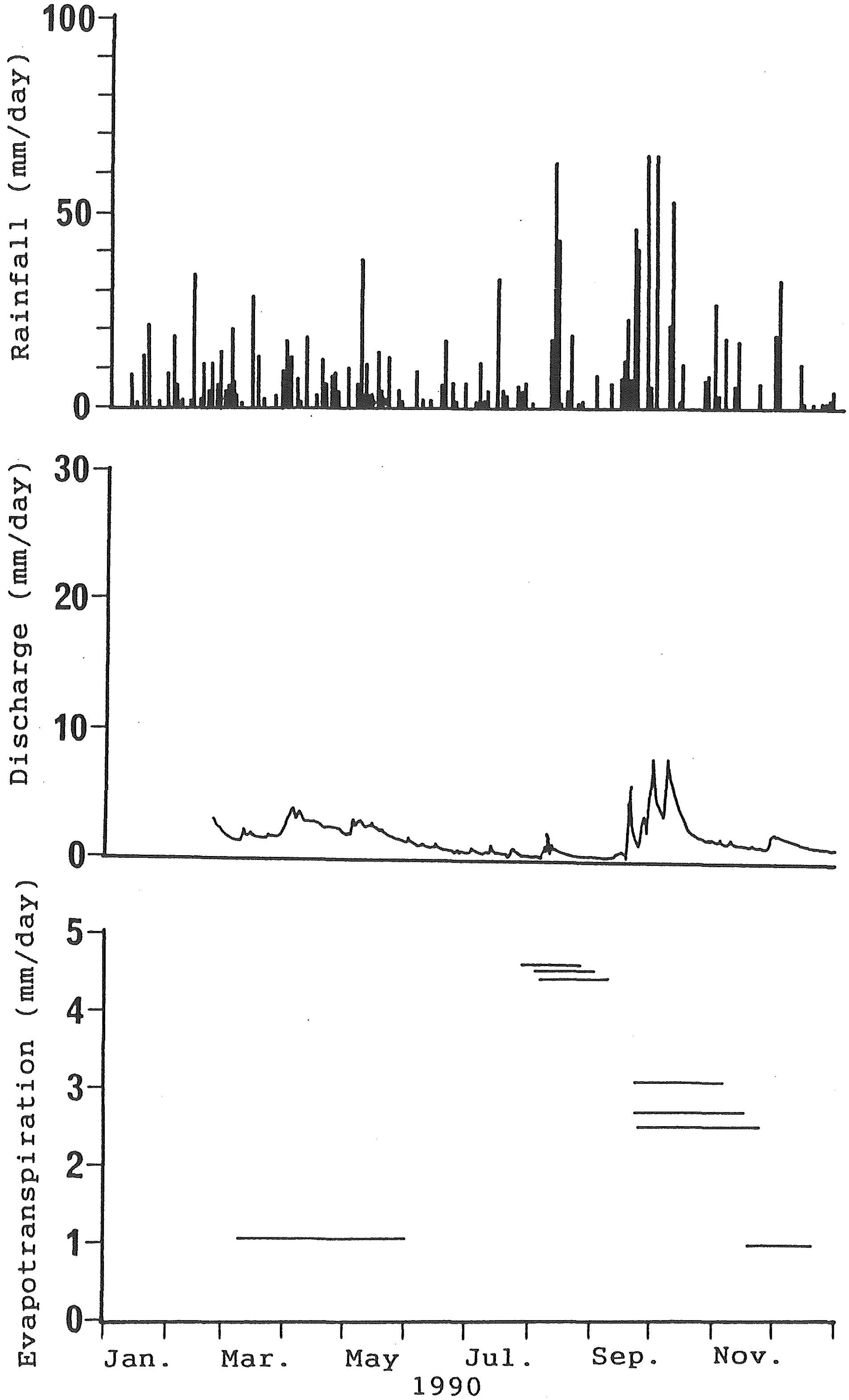


Fig. 21-b Change of daily evapotranspiration rate in 1990

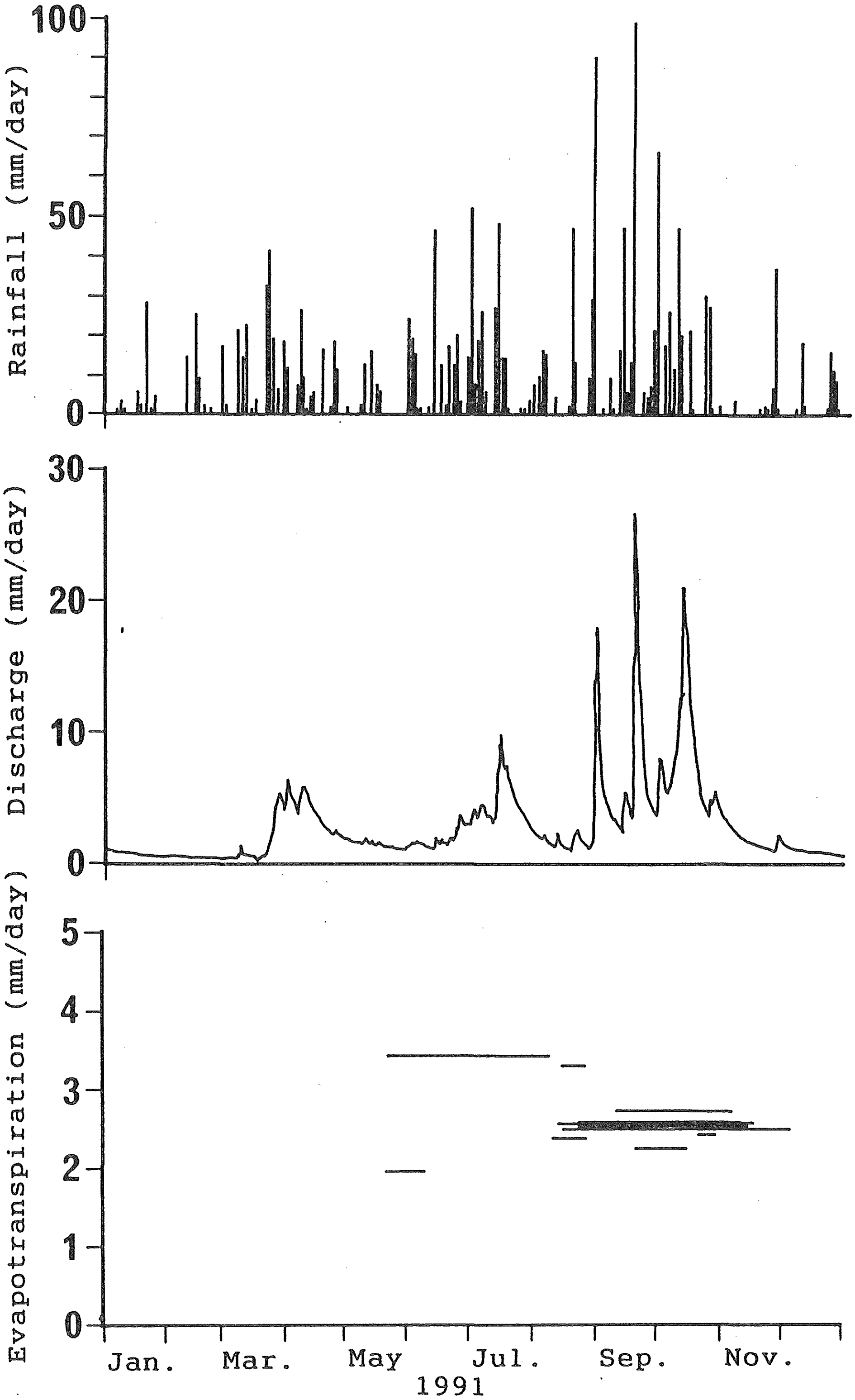


Fig. 21-c Change of daily evapotranspiration rate in 1991

small amount of rainfall in rainy season in June, the evapotranspiration rate showed relatively high value of 4.5 mm/day. The relationship between the minimum discharge for the water budget period and the evapotranspiration rate in each period is plotted in Fig. 22. The figure indicates that the highest evapotranspiration rate was observed when the minimum discharge was the lowest. This suggests that the control of evapotranspiration by the shortage of soil moisture might not occur in the Kawakami basin. This is consistent with the fact that the temporal variation of soil moisture content is small. This appears to be caused by the high water retentivity of B horizon of the soil mantle.

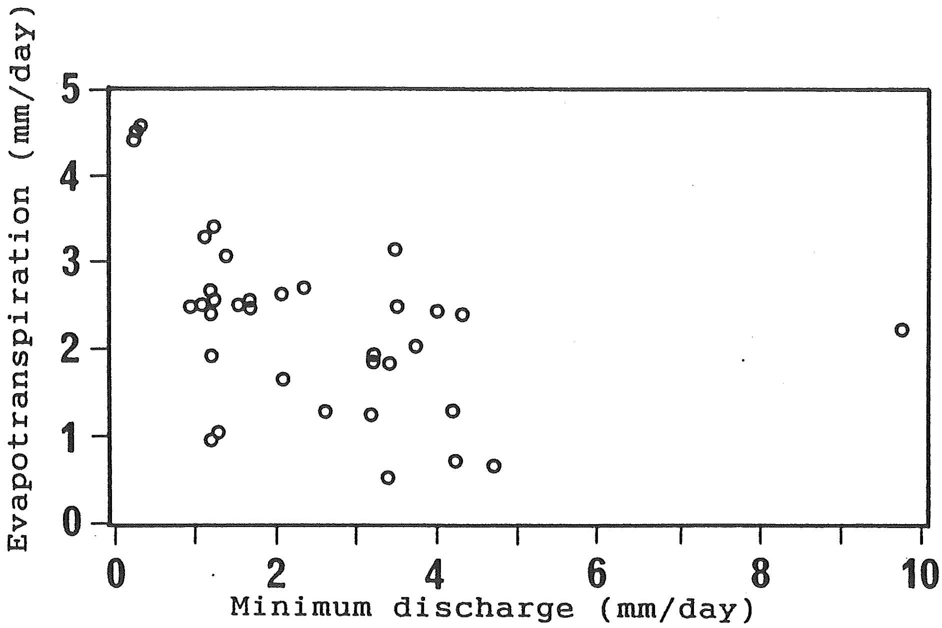


Fig. 22 Relationship between the minimum discharge and evapotranspiration.

Chapter V

Spatial and temporal variation of hydraulic head of soil water

5.1 Spatial and temporal variation of hydraulic head of soil water during and after a storm event

In this section the temporal change of hydraulic head during and after a storm event at some points of the slope will be described and the temporal change of soil water flux will be estimated to understand two dimensional soil water movement in the slope in the short-time period. The emphasis will be on the change of direction and magnitude of soil water flux. As mentioned in chapter 3, intensive measurements of pressure head at upper, middle and lower points of the slope were performed in the end of rainy season and during a rainless period in July and August, 1989. The hydrograph, hyetograph and the change in pressure head during the storm events between 7th and 21th of July are shown in Fig. 23. The total amount of rainfall was 158.5 mm and the highest rainfall intensity of 17 mm/hr was observed at 9:00 on 13th during this period. Also, the hydrograph and the change in pressure head during the rainless period from 5th through 17th, August are shown in Fig. 24. The 63 mm rainfall occurred on 5th and 6th, followed by the rainless period for one week. The data mentioned below were observed within these periods.

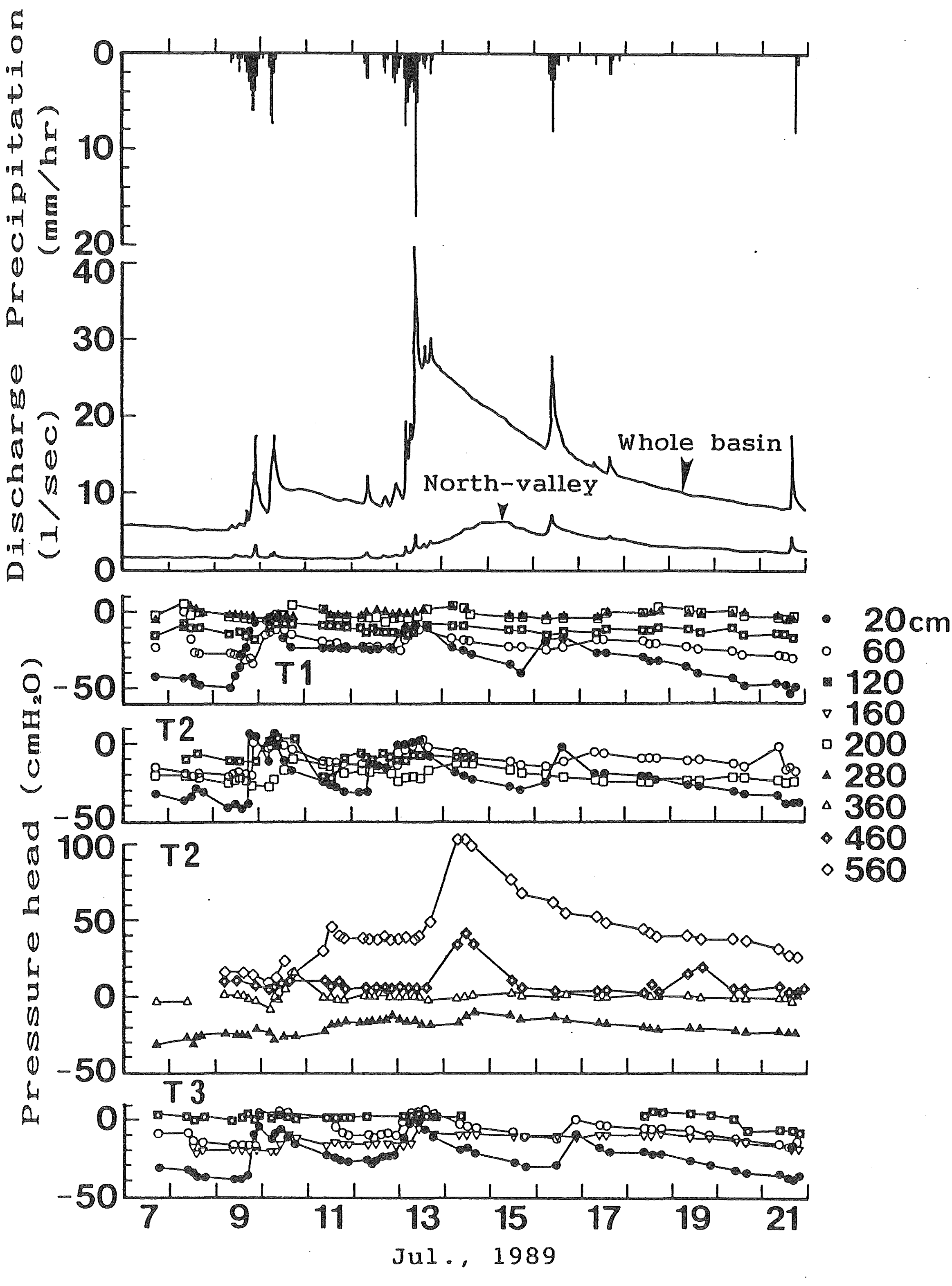
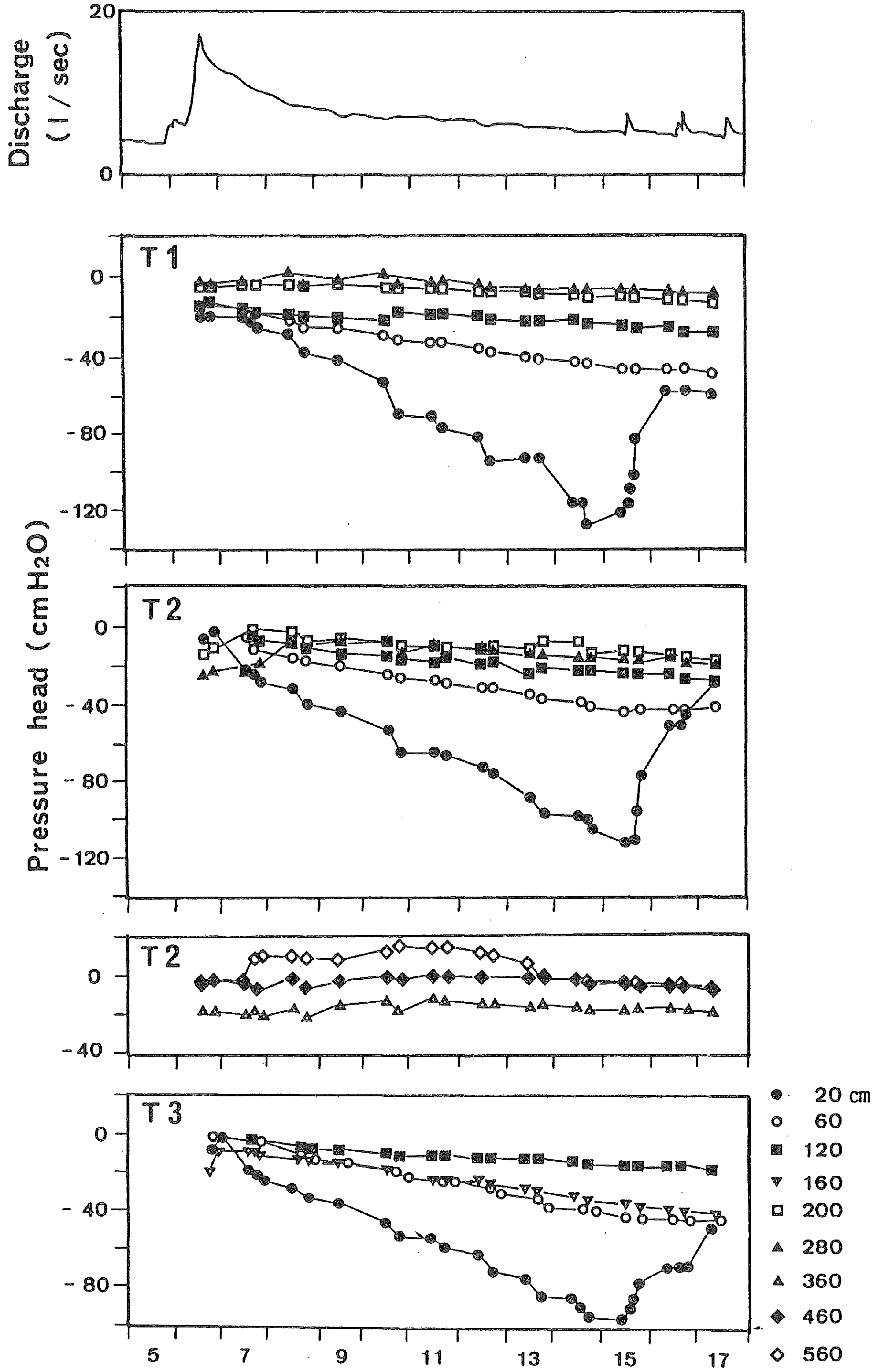


Fig. 23 Hydrograph, hyetograph and change of pressure head between 7th and 21th of July, 1989.



Aug., 1989

Fig. 24 Hydrograph and change of pressure head between 5th and 17th of August, 1989.

The hydraulic head profiles at T1, T2 and T3 during the storm are illustrated in Fig. 25. The point symbols of T1, T2 and T3 are corresponding to the lower, middle and upper points of the site C slope respectively. The datum for the elevation head is the deepest depth where the porous cup of the tensiometer was installed at each point. It is clear from the figure that the hydraulic gradients of the profiles were around 1.0 during the storm which includes the event of 17 mm/hr rainfall intensity at all points of the slope. The pressure head below the 0.6 m depth remained over -20 cmH_2O so that the soil moisture below 0.6 m was in the saturated capillary zone. However the rising of water table up to the soil surface was not observed at every point of the slope and the overland flow did not occur even at the lower-slope. It is clear that the soil water movement dose not change during the storm events especially in the deeper layer of the soil mantle.

The hydraulic head profiles at each point during the rainless period are illustrated in Fig. 26. All over the slope a zero flux plane was formed at 0.6 m depth under the dry condition from 11th through 14th in August and below this depth the hydraulic head profiles did not change much as compared with those during the storm events (Fig. 25). The pressure head below the 0.6 m depth kept the value of -20 to -40 cmH_2O, and the degree of saturation was quite high. It was clear from the results that the hydraulic head above 0.6 m changed considerably in response to the storms or rainless conditions. On the other hand, below the 0.6 m

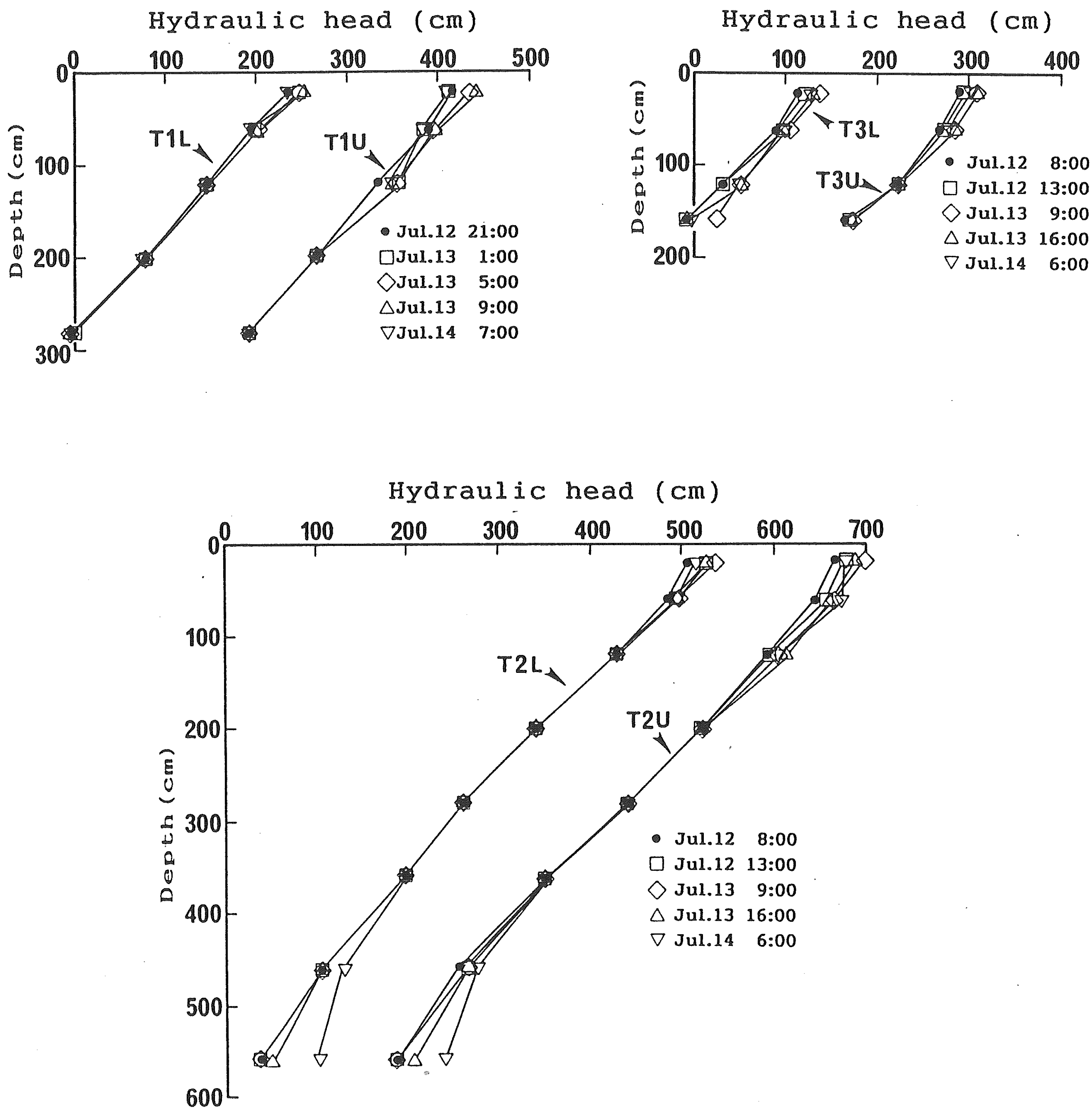


Fig. 25 Hydraulic head profiles during a storm event occurred in July, 1989.

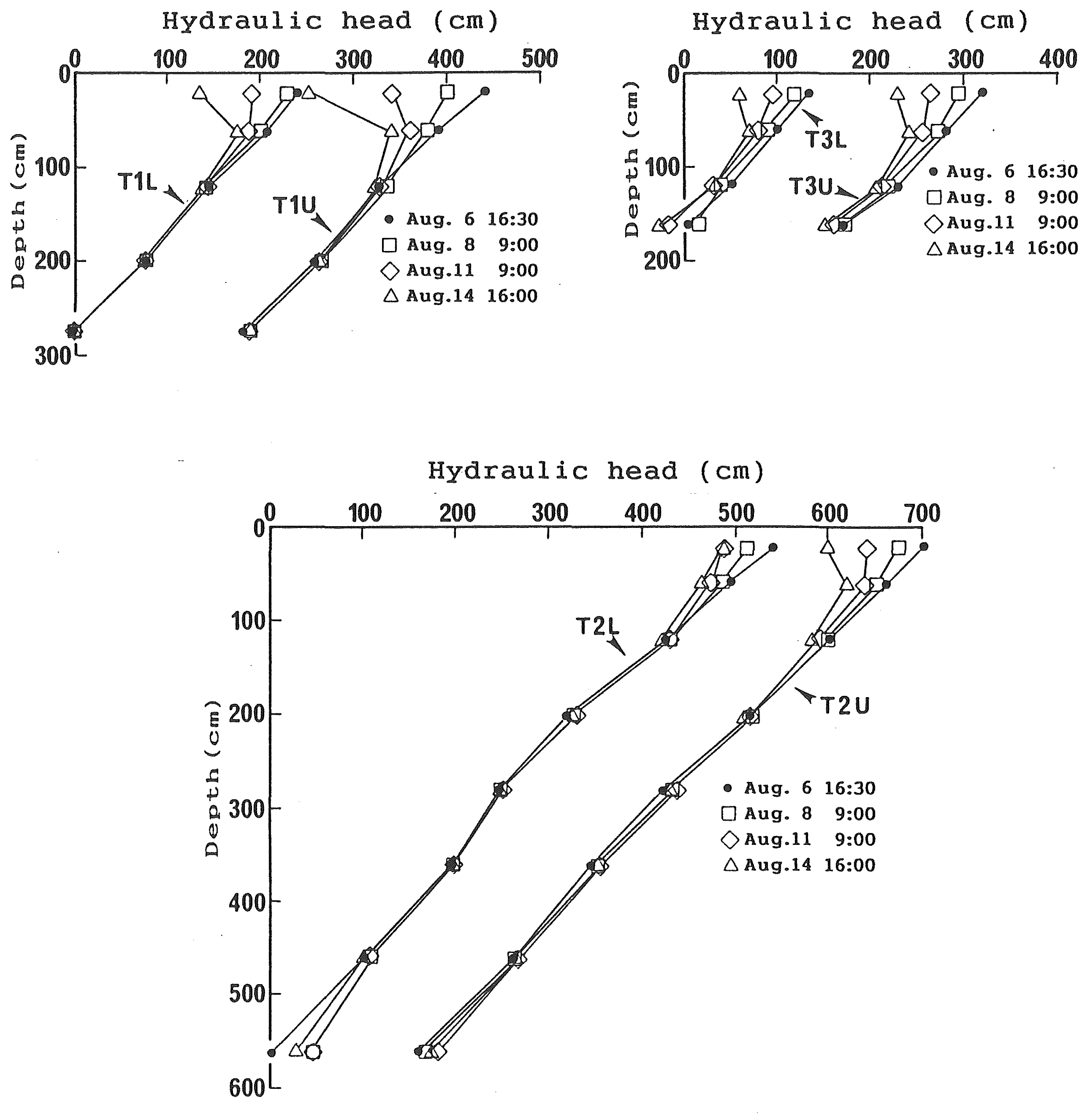


Fig. 26 Hydraulic head profiles during a rainless period in August, 1989.

the pressure head values did not change and the hydraulic gradient remained the value of about 1.0 during and after the storm events.

For the estimation of soil water flux in the slope, the two dimensional axis are defined as shown in Fig. 27. The soil water flux of x and z components is described as follows (Harr, 1977; Ohta and Takahashi, 1986; Tsujimura, 1993):

$$q_x = K \left(\frac{\partial \Psi}{\partial x} + \sin \omega \right) \tag{5.1}$$

$$q_z = K \left(\frac{\partial \Psi}{\partial z} + 1 \right) \tag{5.2}$$

where q_x is the soil water flux of x component, q_z is the flux of z component, ω is the gradient of the slope and K is the hydraulic conductivity. The hydraulic conductivity is estimated by Campbell (1974) as follows:

$$K = K_s \cdot S^{2b+3} \tag{5.3}$$

where K_s is saturated hydraulic conductivity, b is the gradient of soil water retention curve plotted in logarithmic scale and S is the degree of saturation. The actual soil water flux is represented as a vector compound form of the fluxes of x and z components as follows:

$$q_r = [(q_x + q_z \cdot \sin \omega)^2 + (q_z \cdot \cos \omega)^2]^{0.5} \tag{5.4}$$

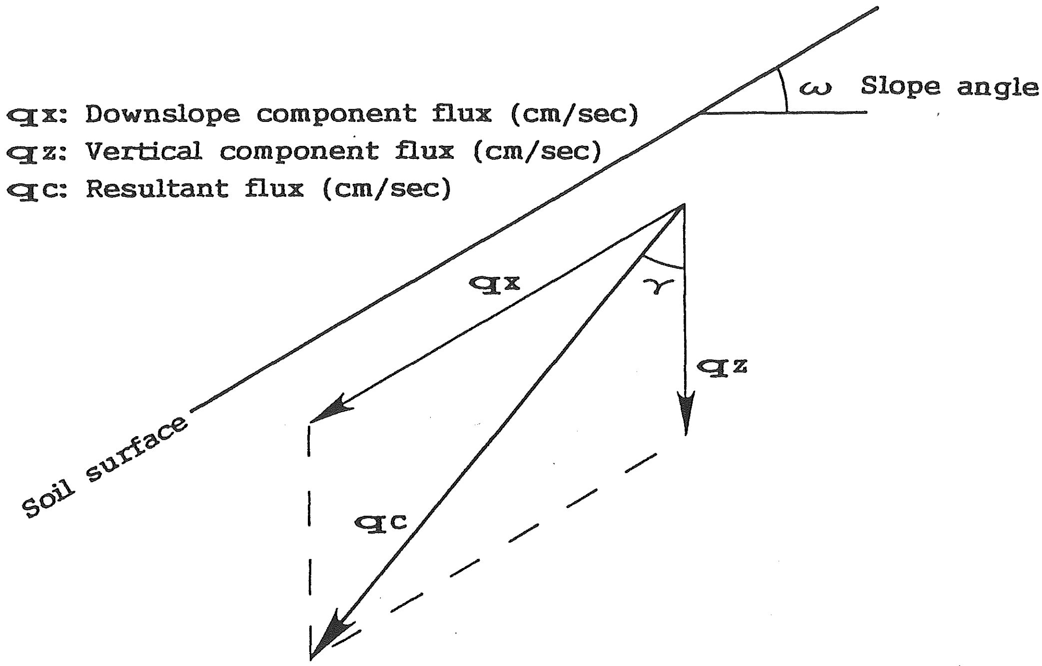


Fig. 27 Definition of the axis for calculation of soil water flux.

where q_r is the resultant soil water flux. The angle of flux from the vertical direction (γ) is given by,

$$\gamma = sin^{-1} \left(q_x \cdot \frac{cos\ \omega}{q_r} \right) \tag{5.5}$$

By applying the equations from (5.1) through (5.5) outlined above, the soil water flux in the slope was estimated during the storm events and the rainless period.

The temporal change of the magnitude and the angle of soil water flux during the storm events is shown in Fig. 28. The soil water flux from 0.2 to 0.6 m depth increased from order of magnitude of 10^{-5} to 10^{-3} cm/sec in response to the on set of rainfall event, and immediately decreased with ending of the storm. On the other hand, the flux under the depth of 0.6 m always remained at the steady value of the order of magnitude of 10^{-5} cm/sec. The angle of soil water flux of 0.2 to 0.6 m depth became small in response to the rainfall events. When the rainfall infiltrates the soil surface, the hydraulic gradient of the shallow soil water becomes large and the vertical component of soil water flux increases. In consequence the angle of soil water flux tends to be relatively small. The angle of flux under the 0.6 m depth kept the value of about 20° throughout the storm event. As a result, the soil water flux including the magnitude and the angle above 0.6 m changed considerably in response to rainfall event. Contrary to the case, under the

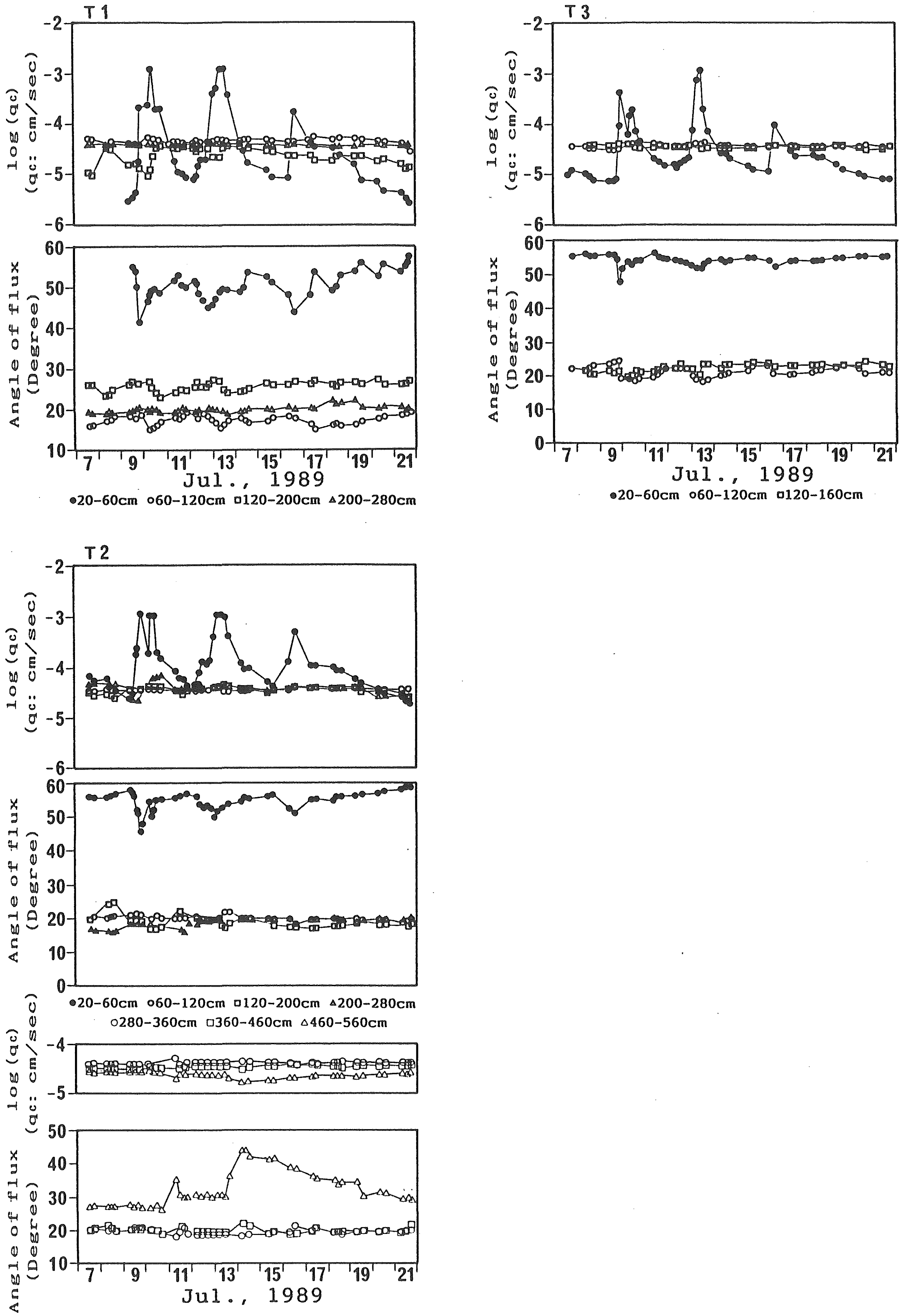


Fig. 28 Temporal change of soil water flux during a storm event occurred in July, 1989.

0.6 m the soil water flux was relatively stable and did not change during the storm events.

During the rainless period, the temporal changes in soil water flux from 5th through 16th in August are shown in Fig. 29. The soil water flux from 0.2 to 0.6 m depth decreased to a value of the order of magnitude of 10^{-6} cm/sec under the driest condition in the period, and the direction of the flux became upward. This clearly shows that the zero flux plane was formed around the depth of 0.6 m. The flux from 0.6 to 1.2 m depth also decreased a little, though the direction of flux did not change much. On the other hand under, below 1.2 m the soil water flux remained steady value of the order of 10^{-5} cm/sec which is the same value as that during the storms. The trend of soil water movement was common at every point of the slope, and the considerable difference of soil water movement among the three positions on the slope have not been observed.

In conclusion, it has been observed that the soil water flux between the depth of 0.2 m and 0.6 m changed considerably in response to the storm events and the soil water flux above 1.2 m decreased under the effect of evapotranspiration during the rainless periods. Under the depth of 1.2 m the soil water movement was relatively stable and the soil water flux kept steady value during the storm events and the rainless period. Also observed were that the vertical downward component of the soil water flux was much larger than the downslope component of the flux in percolation processes. Contrary to that, during the rainless

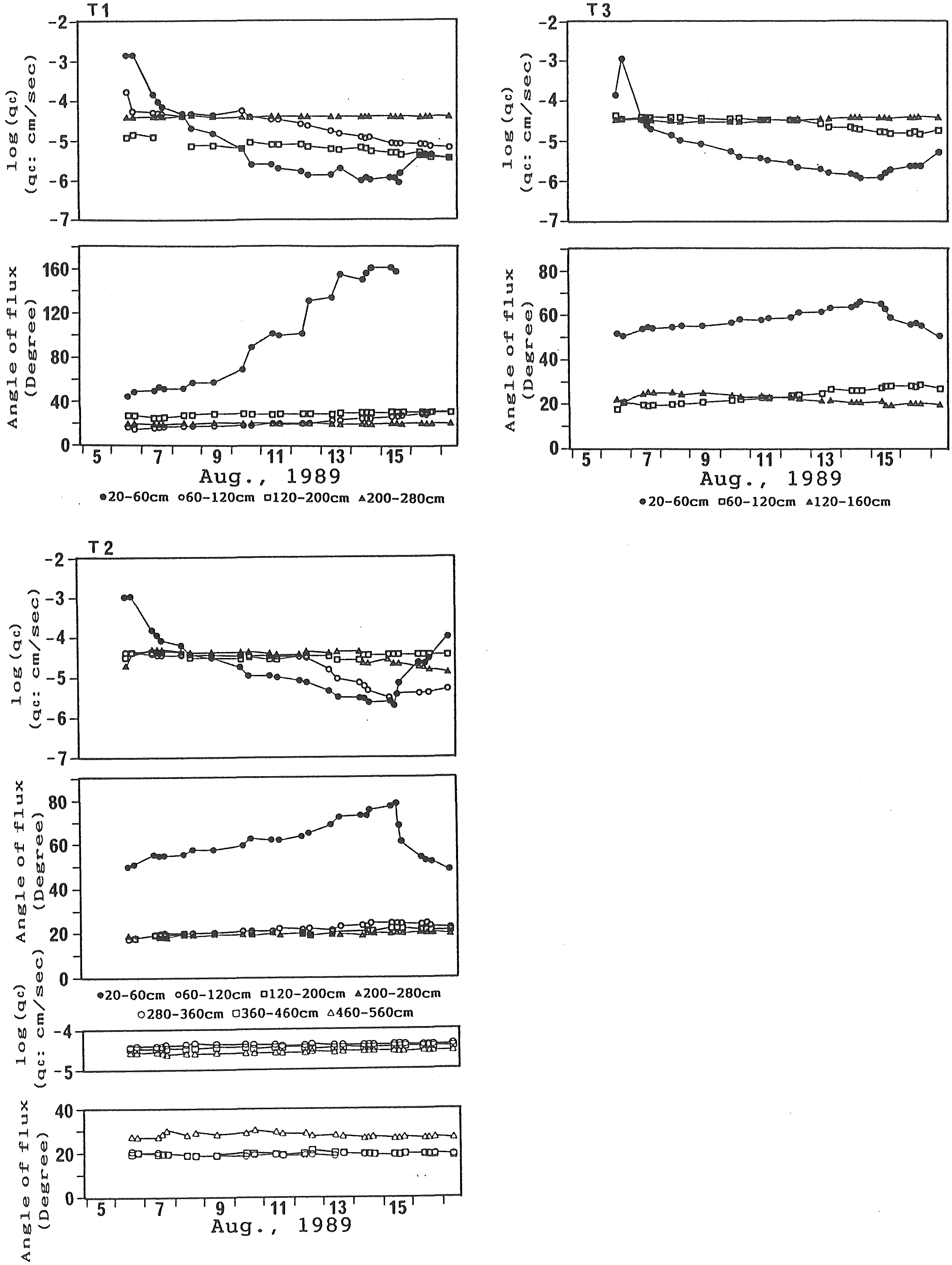


Fig. 29 Temporal change of soil water flux during a rainless period in August, 1989.

periods the downslope flow became large component between 0.2 m and 0.6 m depth, even though the magnitude of the flux was considerably small. These characteristics were observed at every point of the slope.

5-2. Soil water movement estimated from the change of hydraulic head in the long-time period

The temporal change in hydraulic head at 0.1, 0.3, 0.5, 0.7, 1.0 and 1.5 m depth (the datum for calculation of elevation head is 1.5 m depth) at sites A and C from August through December, 1991 is shown in Fig. 30. The hydraulic heads in the figure are the daily average values calculated from the data measured continuously at 30 minutes interval. It can be pointed out that the hydraulic head values over the 1.0 m depth changed considerably throughout the whole period, while those below 1.0 m did not show the large variation and maintained stable values except for the period shortly after the rainfall at site C.

The isopleth diagrams of change of daily average hydraulic head in a monthly period are shown in Fig. 31 to understand the soil water movement in more detail. The period of each diagram corresponds to the interval of water sampling of rainfall, throughfall, soil water, groundwater and discharge water, which is about one month. In the figure the depth of divergent zero flux plane (D-ZFP) and convergent zero flux plane (C-ZFP) is also represented. The D-ZFP is located at a depth where the upward and downward

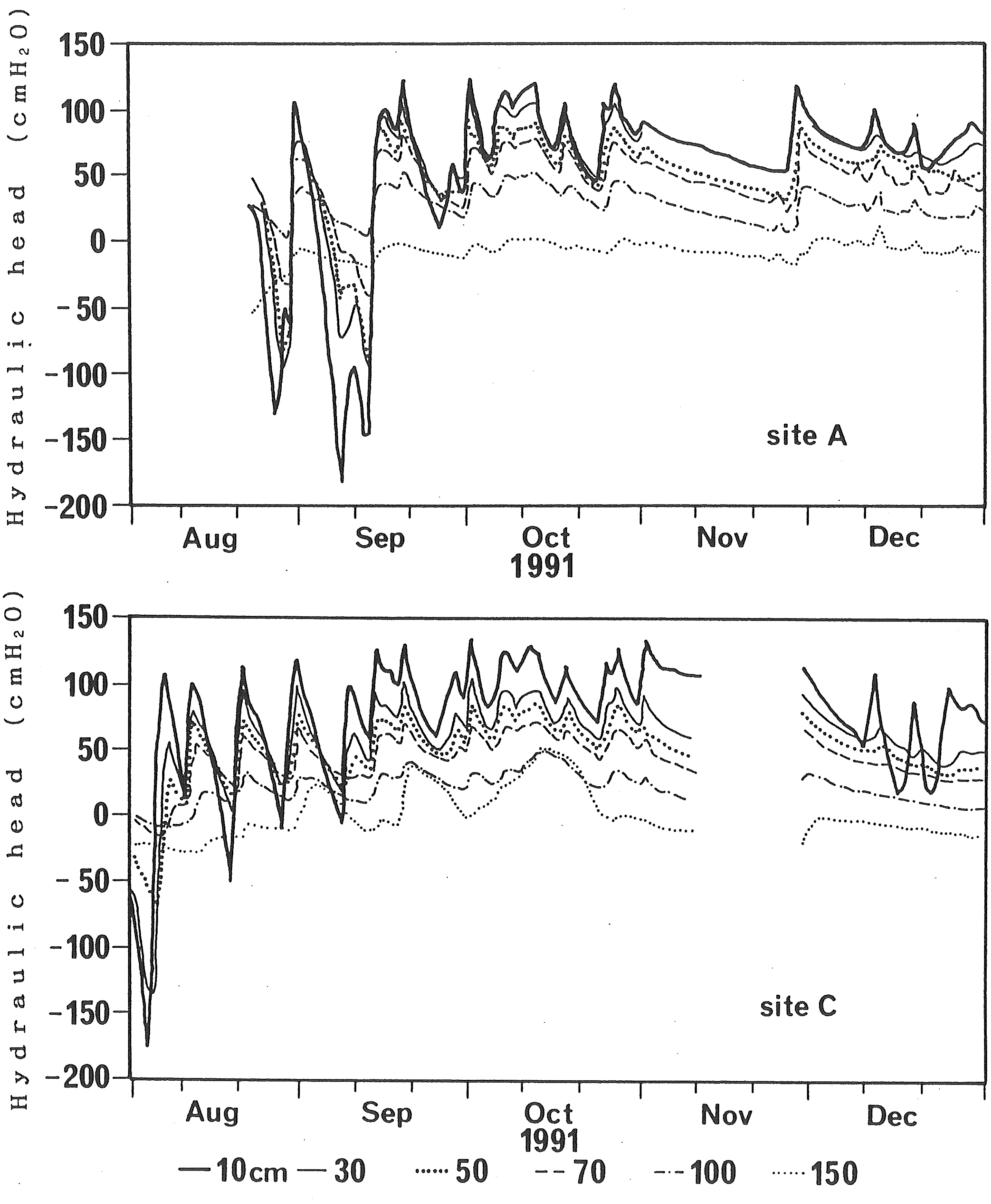


Fig. 30 Temporal change of hydraulic head between August and December, 1991. Datum: 1.5 m depth

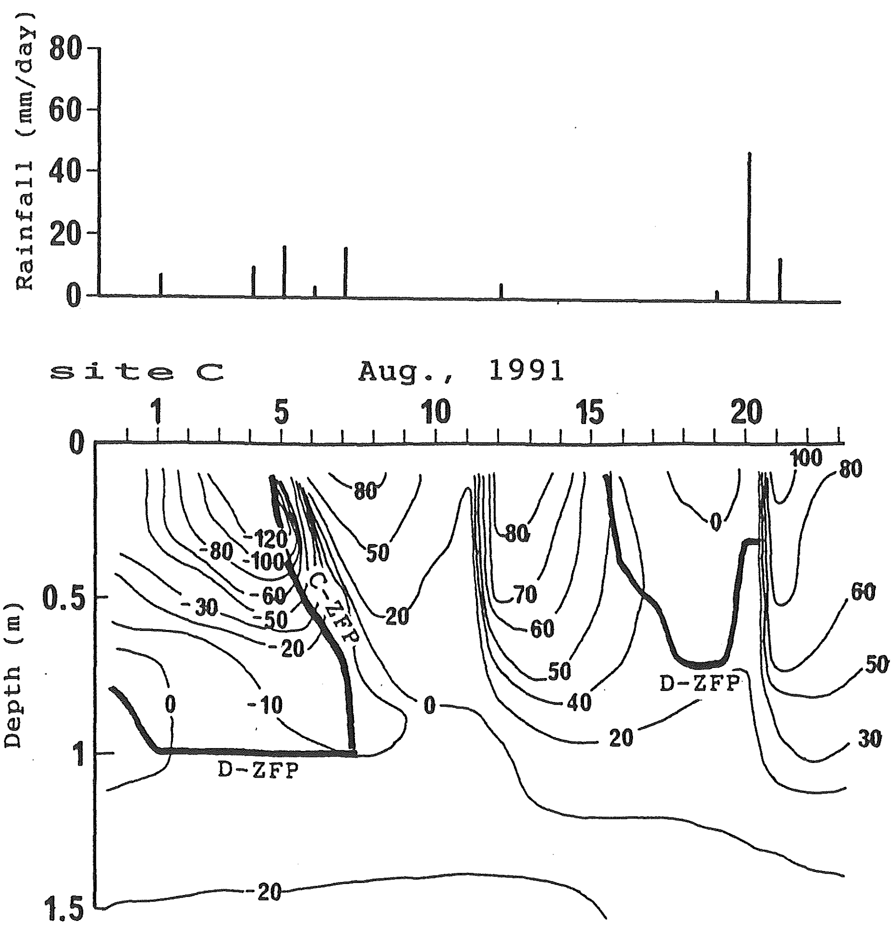


Fig. 31 Isopleth diagram of daily average hydraulic head (cmH_2O, Datum: 1.5 m depth).
D-ZFP: Depth of divergent zero flux plane
C-ZFP: Depth of convergent zero flux plane

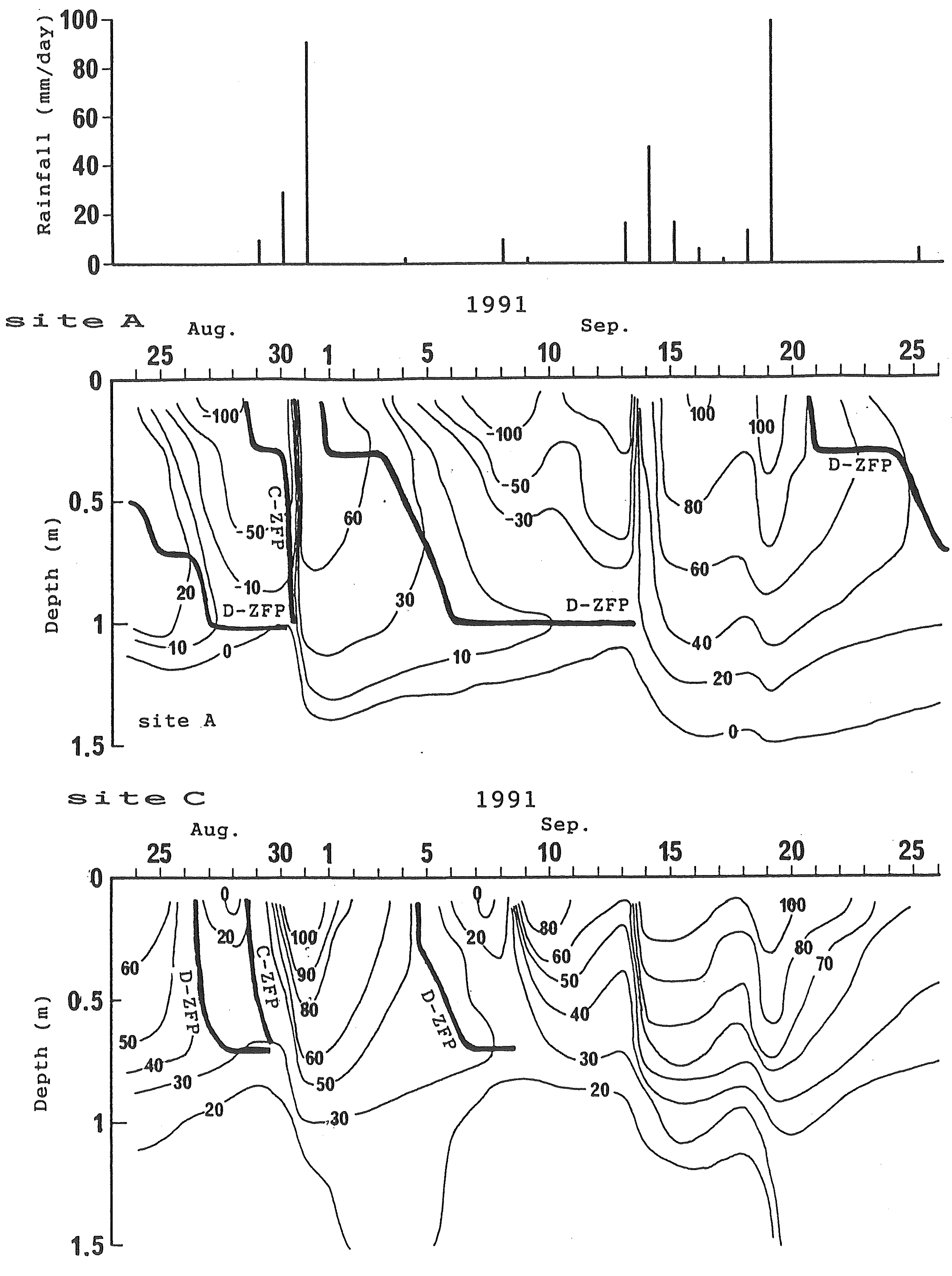


Fig. 31 Continued (cmH_2O, Datum: 1.5 m depth).

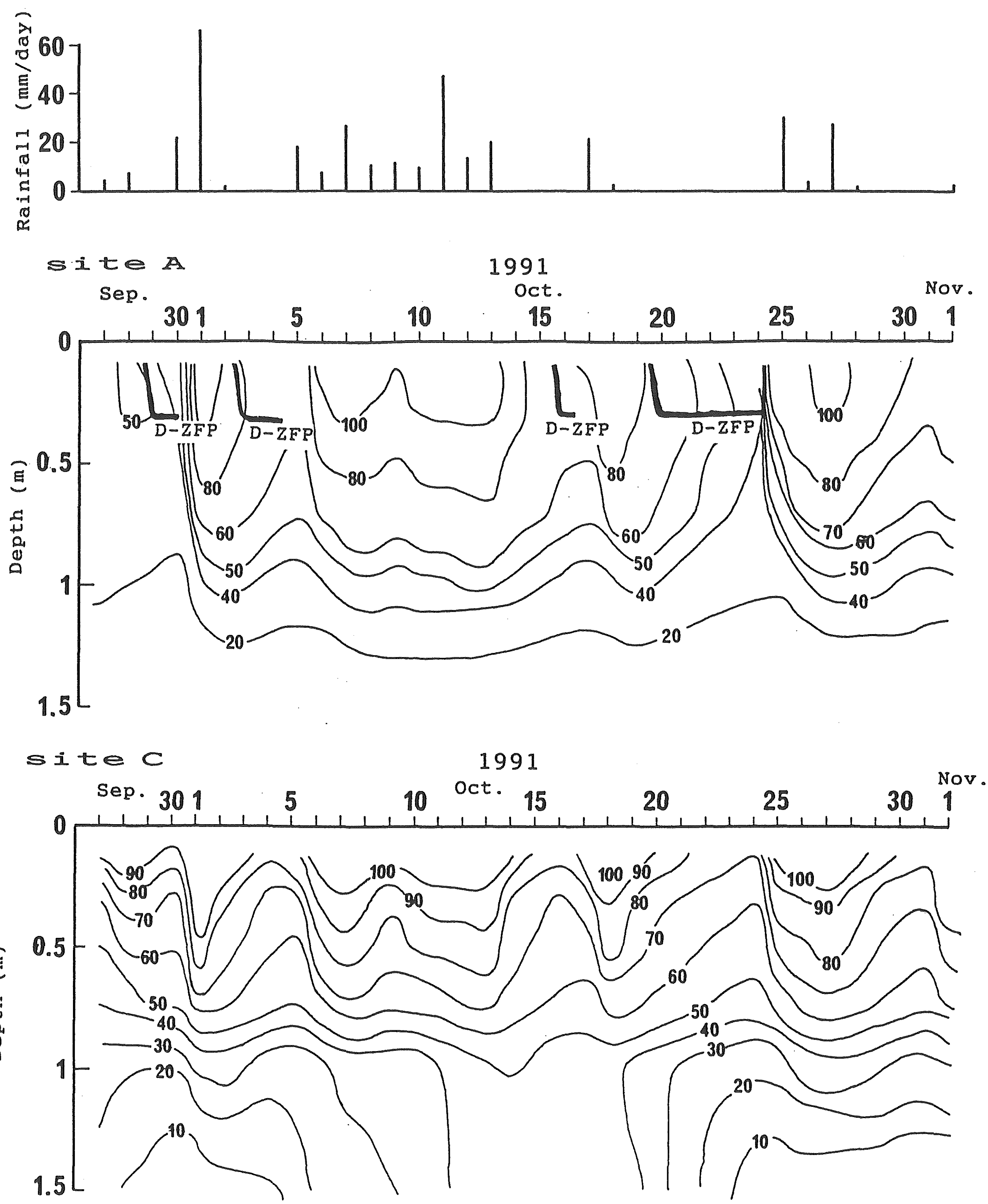


Fig. 31 Continued (cmH_2O, Datum: 1.5 m depth).

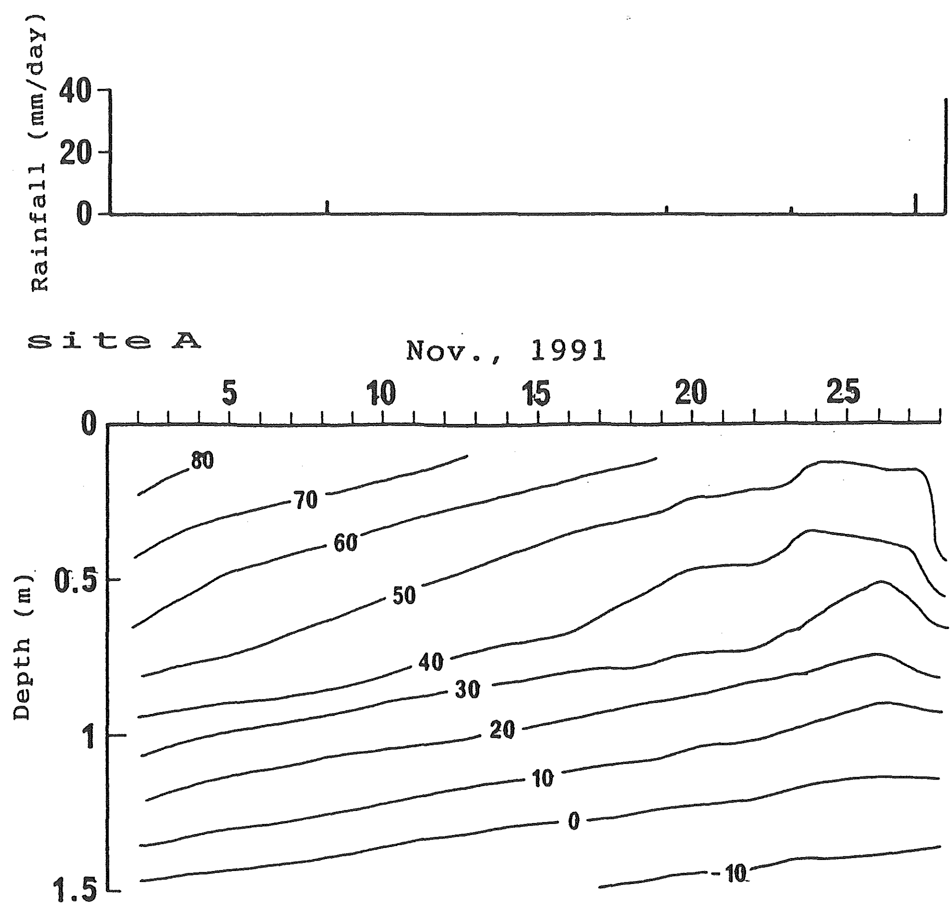


Fig. 31 Continued (cmH_2O, Datum: 1.5 m depth).

soil water fluxes diverge. In contrast, the C-ZFP corresponds to a depth where the upward and downward soil water fluxes converge. (Wellings and Bell, 1980). The C-ZFP is formed generally when the D-ZFP disappears in response to the rainfall event.

Between 31th, July and 23th, August at site C, the hydraulic head value changed considerably above 1 m depth, while below 1 m depth it did not vary much. The D-ZFP was formed during the rainless periods. When the rainfall event occurred, the C-ZFP fell down from the soil surface to the depth of D-ZFP. At last the D-ZFP and C-ZFP disappeared.

Between 24th, August and 26th, September at sites A and C, the same feature was observed. In site A the depth of D-ZFP was deeper and the period over which the D-ZFP existed was longer than that observed at site C. This appears to suggest the difference of behavior of soil water movement in response to the evapotranspiration between the two sites.

Between 27th, September and 1st, November at sites A and C, the variation of hydraulic head became relatively small as compared with the case in August and September. The D-ZFP was sometimes formed at 0.3 m depth in site A.

Between 2nd and 28th, November at site A, the variation of hydraulic head became very small. The zero flux plane was not observed.

The temporal change in hydraulic head profiles when the D-ZFP disappeared at sites A and C is shown in Fig. 32. It is clear from this figure that the depth of C-ZFP fell down gradually as the wetting front penetrated downward; then the

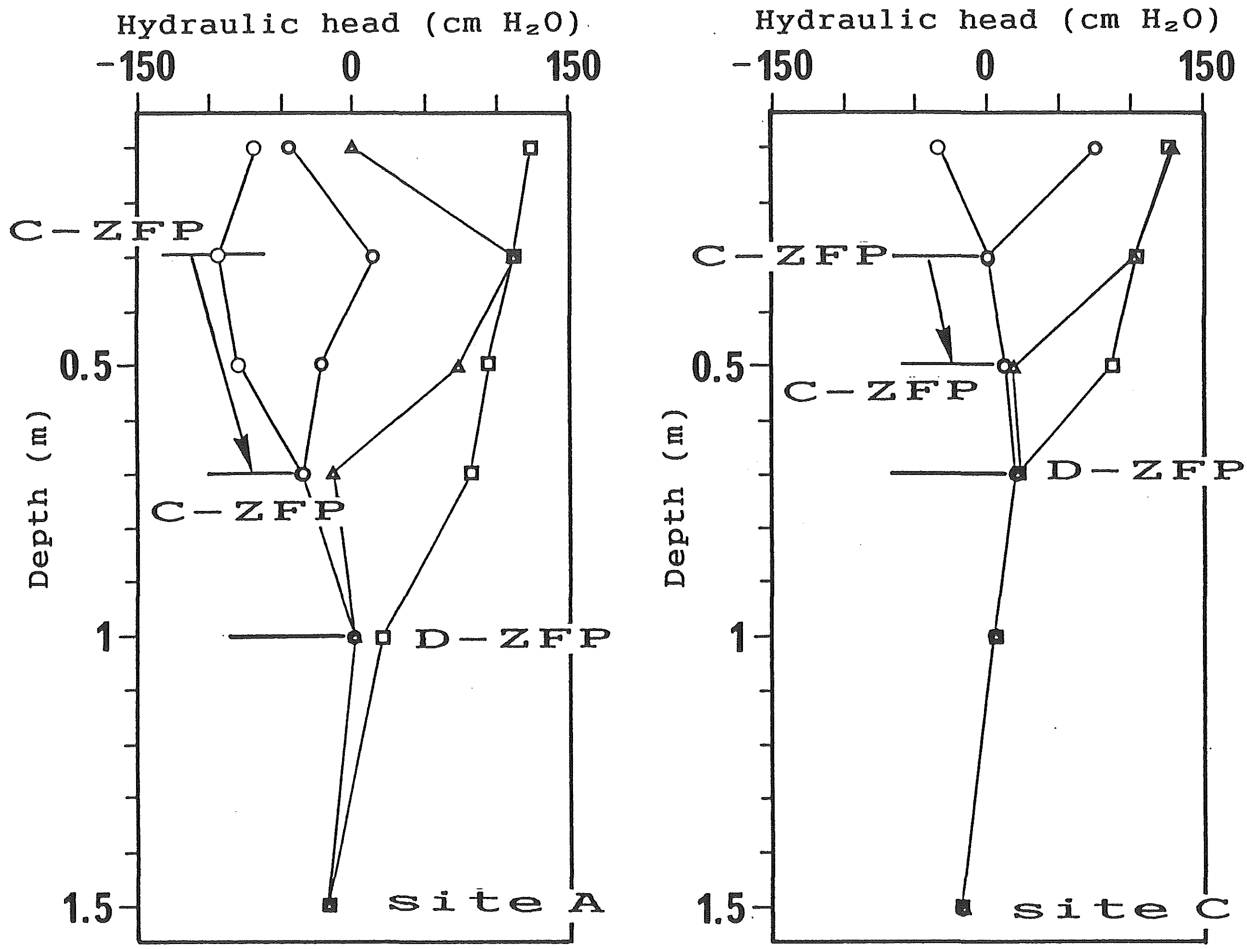


Fig. 32 Temporal variation of hydraulic head profiles when the D-ZFP disappeared.
site A: 14th Sep., 1991
○6:00, ●7:00, ▲8:00, □12:00
site C: 20th to 21th Aug., 1991
○12:00, ●16:00, ▲20:00 of 20th
□0:00 of 21th

D-ZFP disappeared at last and the hydraulic gradient became around unity at all depths. The behavior of soil water movement described above is schematically illustrated in Fig. 33.

In this section the soil water movement is discussed based on the temporal change of hydraulic head profiles observed continuously for about 4 months. The behavior of soil water movement above 1 m depth is significantly different from that below 1 m depth. The hydraulic head profile changes considerably in response to the rainfall events and the effect of evapotranspiration above 1 m depth. Also, the formation and disappearance of the divergent or convergent zero flux plane are observed frequently. The maximum depth where the D-ZFP is formed is about 1 m. In contrast, the hydraulic head profile is quite stable below 1 m depth. This stable nature of soil water movement is not inconsistent with the characteristic of soil moisture contents as shown in section 4-3. This is caused mainly by the physical properties of the soil mantle, i.e. that the variation of soil moisture with change of pressure head is quite small. The maximum depth of main root of the larch is about 1 m as determined by the observation of fallen trees. This suggests that the depth where the effect of transpiration extends is around 1 m. Therefore transpiration process also appears to have an affect on the behavior of soil water movement, and the boundary depth of 1 m seems to be determined by both effect of physical properties of soil and the transpiration.

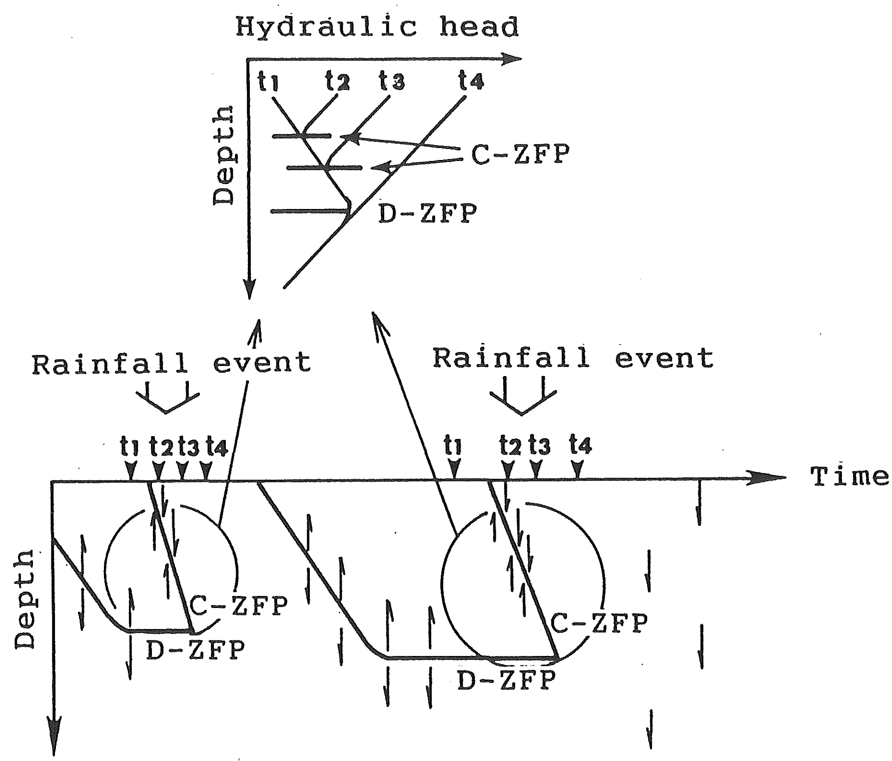


Fig. 33 Schematic diagram for the behavior of soil water movement. (The arrows show direction of soil water flux)
D-ZFP: Depth of divergent zero flux plane
C-ZFP: Depth of convergent zero flux plane

Chapter VI

The change of stable isotopic ratio of water with hydrological cycle in the basin

6-1. Stable isotopic compositions of rainfall, throughfall, soil water, groundwater and discharge water

In this section, the temporal change of isotopic ratios of rainfall and discharge water will be described at first. Then the stable isotopic compositions of rainfall, throughfall, soil water, groundwater and discharge water will be described to clarify the isotopic characteristics of water in the Kawakami basin.

The temporal variations of δD and $\delta^{18}O$ values of the monthly rainfall and discharge water in 1990 and 1991 are shown in Fig. 34. The δD value of rainfall ranged from -60.25 to -113.34 per mil and the $\delta^{18}O$ value from -9.51 to -15.60 per mil. The the weighted mean values of δD and $\delta^{18}O$ were -88.32 and -12.84 per mil respectively between 1990 and 1991. The δ values of discharge water did not change throughout two years and the mean values of δD and $\delta^{18}O$ were -80.82 and -11.88 per mil respectively.

The relationship between δD and $\delta^{18}O$ values of all water samples in 1991 is plotted in Fig. 35 as a δ-diagram, and regression analysis yielded from the data of rainfall is the following equation,

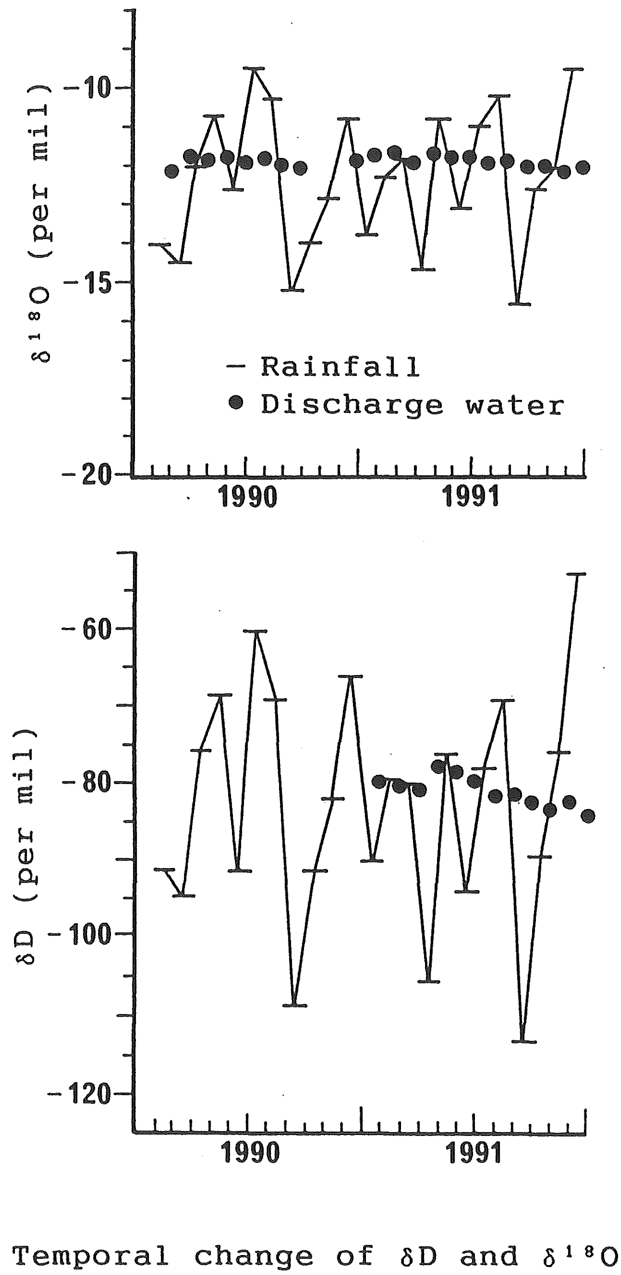


Fig. 34 Temporal change of δD and $\delta^{18}O$ of monthly rainfall and discharge water.

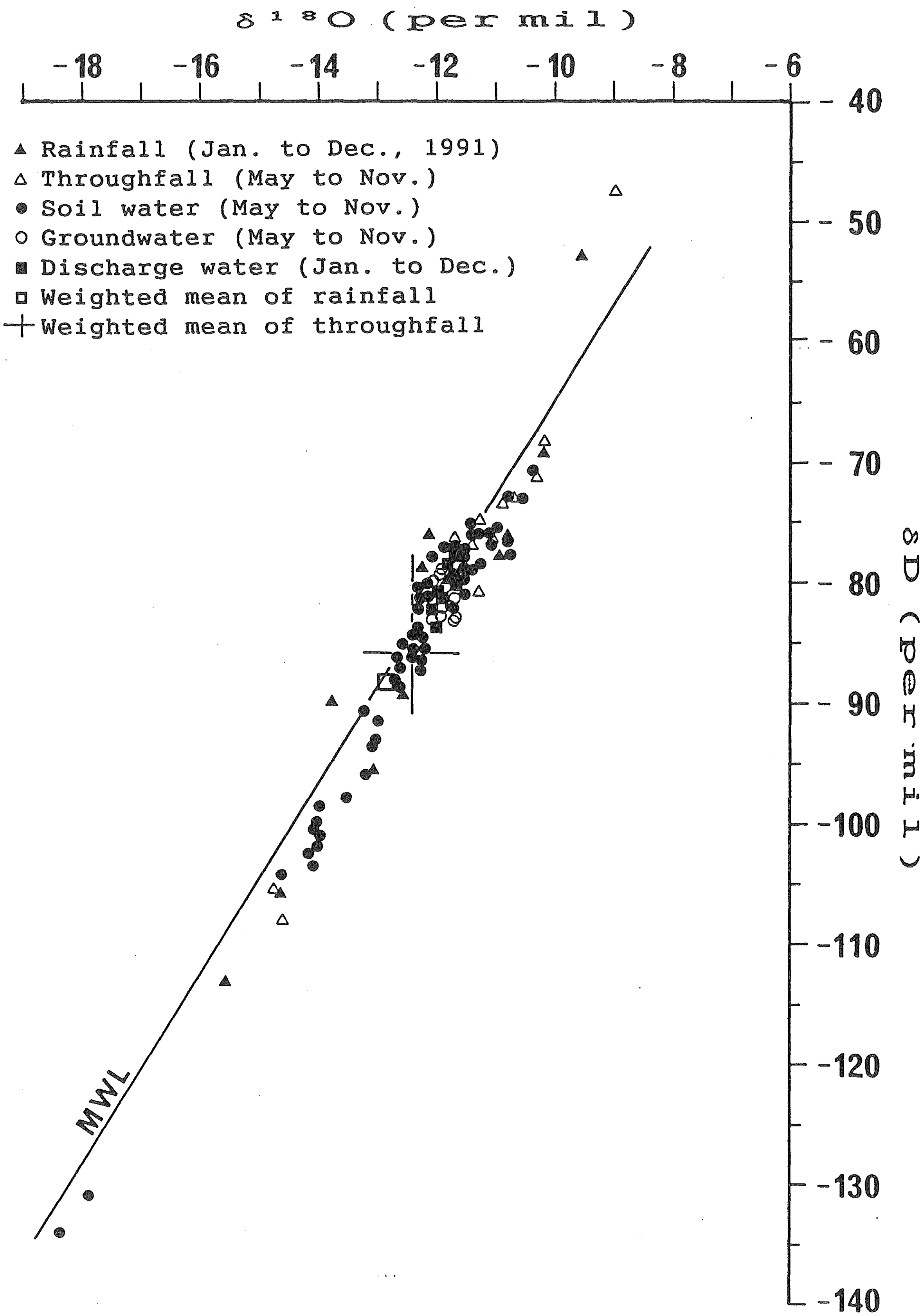


Fig. 35 δ-diagram of all water samples in the Kawakami basin in 1991.
MWL: Meteoric Water Line

$$\delta D = 8.048\delta^{18}O + 16.393 \quad (r^2 = 0.9104) \tag{6.1}$$

which is indicated by the line of MWL in Fig. 35. The fact that all points in the diagram distribute in the vicinity of this line means that the water in the Kawakami basin is formed by a Rayleigh process under the equilibrium condition (Mizutani, 1986; Itadera, 1993). It is obvious from this figure that the data of groundwater, discharge water and part of soil water converge on a value which is a little higher than the value of weighted mean of throughfall. This suggests that enrichment of water may have occurred through the process from infiltration to groundwater recharge. This will be discussed in the following.

The δ-diagram of rainfall in 1990 and 1991 is shown in Fig. 36. The figure clearly shows that the data could be categorized into two groups, i.e. the data of warm season (from June through September, 1990; April through October, 1991) and those of cold season (January, February, November and December, 1990; from February through April, October through December, 1991). The correlation of each group is also presented in the figure. The gradient of all the correlation lines show around 8, while the y-intercept, which is called d-excess, shows the seasonal variation ranging from 10.53 to 31.84 per mil. In winter the d-excess value tends to have higher one, while that of summer shows lower value. This characteristic have been pointed out by some researchers (e.g. Sanjo, 1991; Yamamoto *et al.*, 1993). Sanjo (1991) showed that the δ values of rainfall reflected

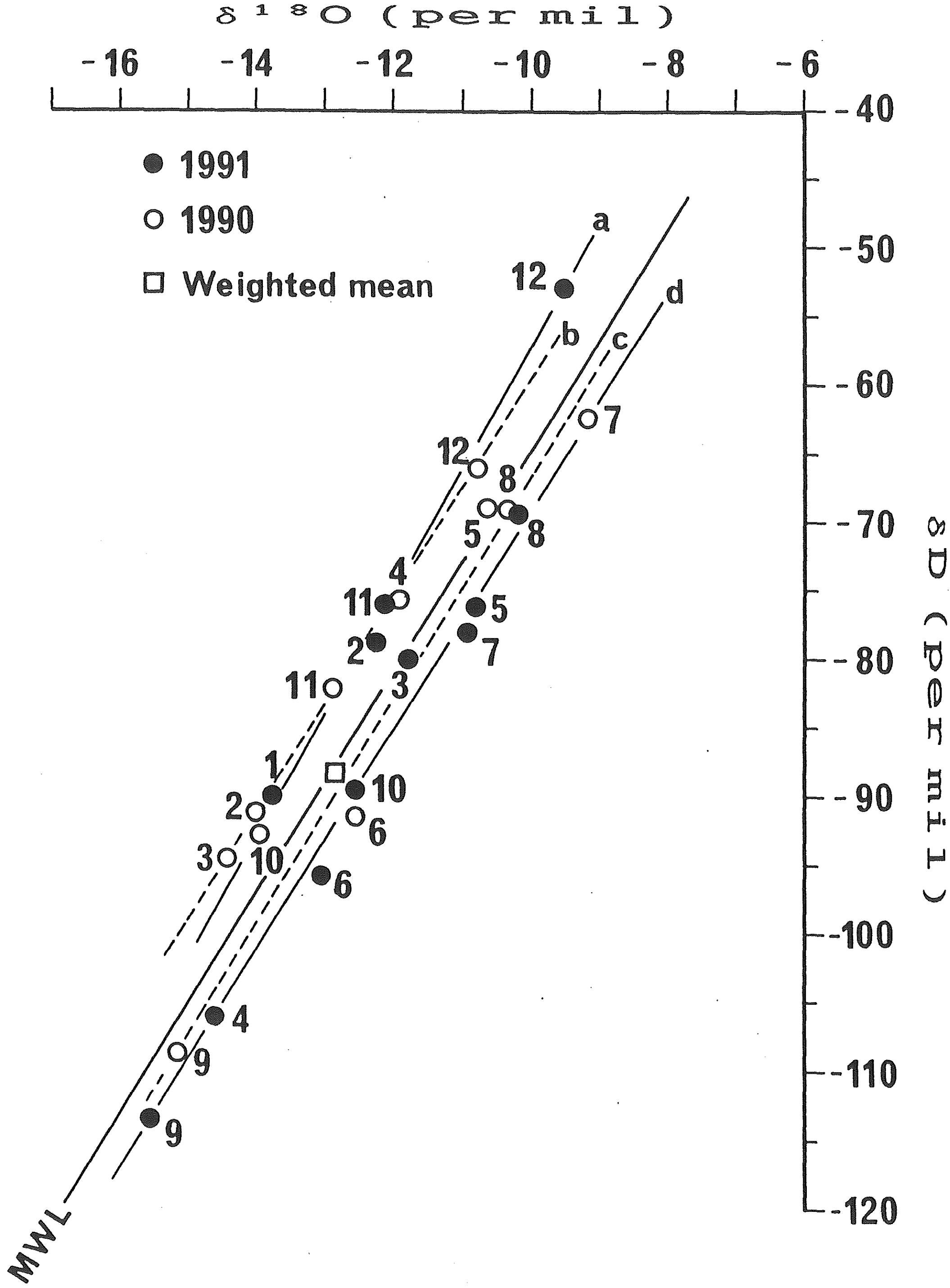


Fig. 36 δ-diagram for monthly rainfall in 1990 and 1991.
MWL: Meteoric Water Line
a: $\delta D = 8.92\delta^{18}O + 31.84$
(Jan., Feb., Nov. and Dec., 1991)
b: $\delta D = 7.88\delta^{18}O + 19.05$
(Feb. to Apr. and Oct. to Dec., 1990)
c: $\delta D = 8.18\delta^{18}O + 14.65$
(Jun. to Sep., 1990)
d: $\delta D = 7.99\delta^{18}O + 10.53$
(Apr. to Oct., 1991)

the δ values of air mass which is a source of the rainfall. Because the Kawakami basin exists in the mountainous region at higher elevation in central part of Japanese Islands, the seasonal variation of isotopic compositions of rainfall appears to be much clearer than that of previous studies.

The relationship between the $\delta^{18}O$ value of rainfall and that of throughfall is shown in Fig. 37. The weighted mean value of $\delta^{18}O$ of rainfall is -12.78 per mil, and that of throughfall at sites A and C is -12.42 and -12.39 per mil, respectively. Therefore the water enrichment of 0.4 per mil must have occurred with the interception loss of water by canopy.

The δ-diagram for soil water is shown in Fig. 38. The majority of the data below 1 m depth converges on a value which is a little higher than the weighted mean of throughfall. Most part of the data above 1 m depth also converges on that value. All data for groundwater converge on a value which is a little higher than the weighted mean of throughfall (Fig. 39). Mean isotopic compositions of rainfall, throughfall, soil water and groundwater are shown in Fig. 40. The mean value of shallow (0.1 to 0.7 m depth) soil water agrees with that of throughfall. This is because the isotopic ratio of shallow soil water may reflect that of throughfall which fell immediately before the water sampling. The mean $\delta^{18}O$ value of deep (1.0 to 2.0 m depth) soil water and groundwater is respectively 0.2 per mil and 0.4 per mil higher than the weighted mean of throughfall. This suggests that the enrichment of water occurs between

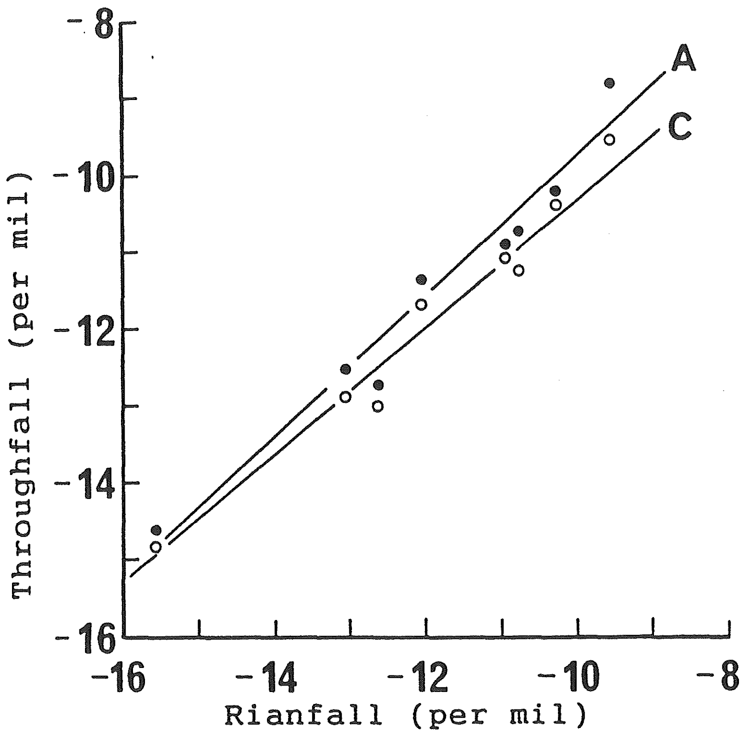


Fig. 37 Relationship of $\delta^{18}O$ between rainfall and throughfall.
site A: $y=0.903x-0.762$ ($r^2=0.957$)
site C: $y=0.815x-2.211$ ($r^2=0.965$)

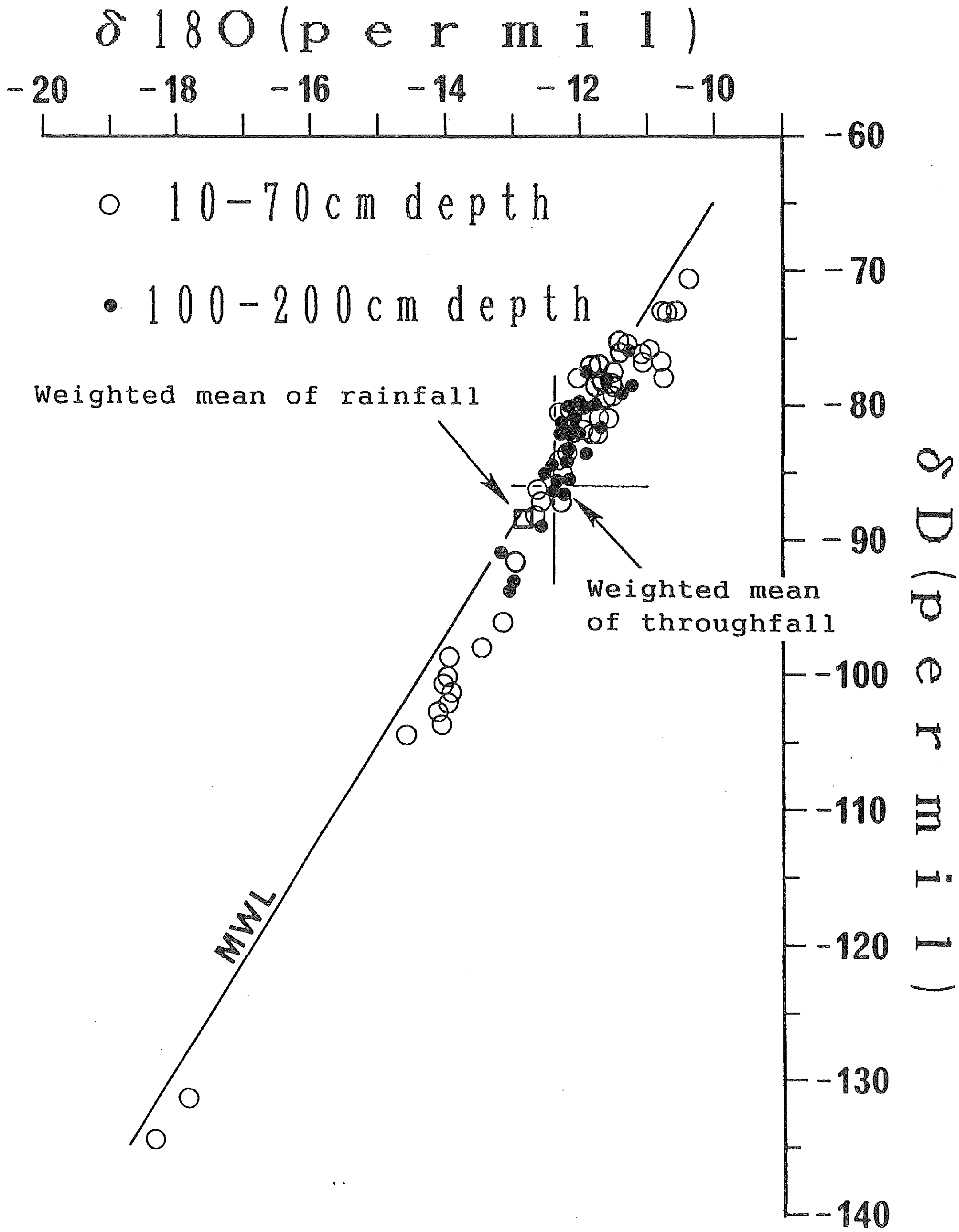


Fig. 38 δ-diagram for soil water.
MWL: Meteoric Water Line

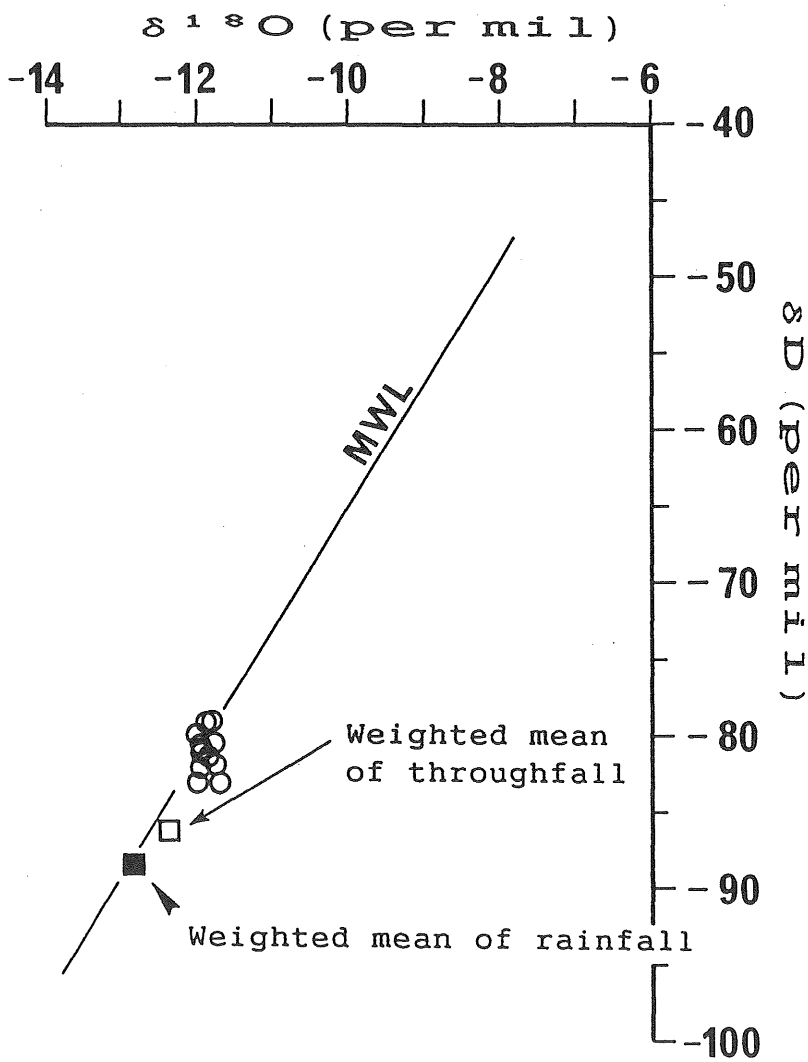


Fig. 39 δ-diagram for groundwater.
MWL: Meteoric Water Line

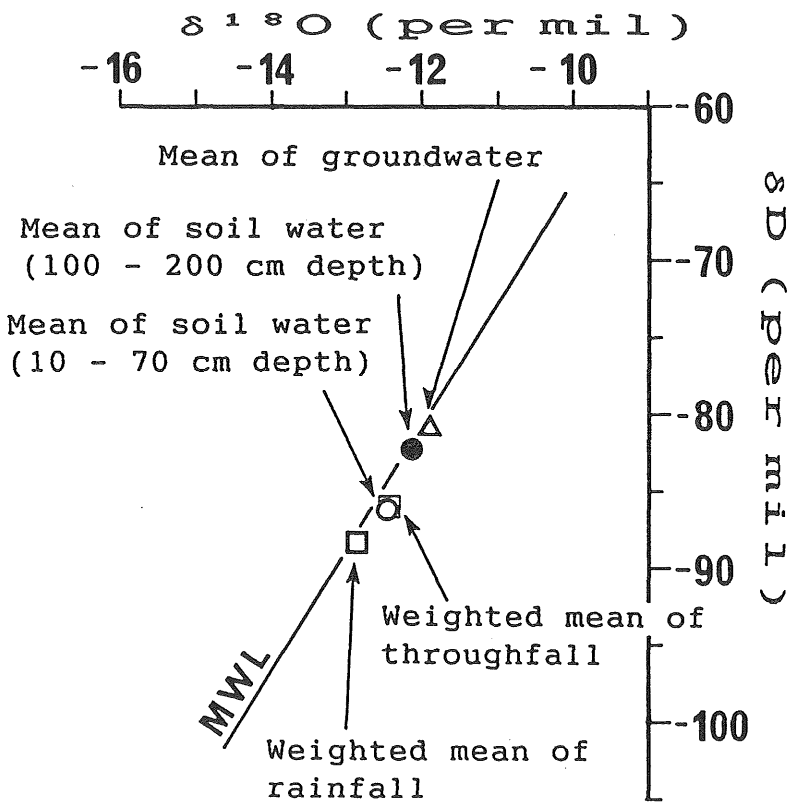


Fig. 40 Comparison of mean δ values of rainfall, throughfall, soil water and groundwater. MWL: Meteoric Water Line

the soil surface and water table. As Mizutani (1986) mentioned, it is generally accepted that the δD or $\delta^{18}O$ value of shallow groundwater in a region corresponds to the annual weighted mean of δ value of rainfall unique in that region (Dansgaard, 1964), unless groundwater with a different δ value discharges into the region. Nakai (1986) confirmed this by comparing δ value of rainfall and that of groundwater in Nobi plain, Japan. However, Sanjo (1991) pointed out that the effect of enrichment by evapotranspiration should have been considered to account for this characteristic in the forested region.

The average δ values of rainfall, throughfall, soil water, groundwater and discharge water are compared in Table 6. The differences of δ values among them are significant, because the accuracy of analysis of δD and $\delta^{18}O$ is ±1.0 and ±0.1 per mil, respectively. Therefore it can be concluded that the water enrichment of 0.2 per mil in $\delta^{18}O$ occurs from the soil surface to the depth of 2.0 m, and enrichment of 0.5 per mil of water occurs through the process from infiltration to groundwater recharge. It seems reasonable that the enrichment of water may be explained by evaporation from the forest floor. The difference of δ values between the deep soil water and the groundwater appears to be caused by the difference of residence times of both water. The residence time for the soil water from 0.5 to 2.0 m depth was calculated to be 4 months, and that for groundwater was 19 to 20 years by using tritium as a tracer by Matsutani *et al*. (1993). Therefore the average evaporation rate from the

Table 6 Mean isotopic ratios of rainfall, throughfall, soil water, groundwater, seepage water and discharge water.

Site	Type	Depth (m)	δD Mean (per mil)	δD S.D. (per mil)	$\delta^{18}O$ Mean (per mil)	$\delta^{18}O$ S.D. (per mil)
	Rainfall		-88.32*	15.47	-12.84*	1.72
A	Throughfall		-86.45*	13.43	-12.42*	1.42
	Soil water	0.1	-85.72	21.66	-12.51	2.62
		0.2	-84.67	9.91	-12.11	1.21
		0.3	-84.89	11.84	-12.29	1.44
		0.5	-88.10	10.47	-12.56	1.16
		0.7	-81.88	3.36	-12.10	0.20
		1.0	-82.22	3.25	-12.08	0.26
		1.5	-83.58	1.63	-12.26	0.18
		2.0	-80.76	3.37	-12.02	0.36
	Groundwater		-81.03	1.26	-11.84	0.07
C	Throughfall		-85.43*	10.55	-12.39*	1.39
	Soil water	0.1	-87.11	21.57	-12.39	2.63
		0.2	-87.58	11.14	-12.50	1.25
		0.3	-86.05	8.84	-12.36	1.14
		0.5	-89.02	9.91	-12.92	1.04
		0.7	-85.96	8.75	-12.55	0.95
		1.1	-81.04	2.54	-11.97	0.30
		1.5	-82.66	3.31	-12.14	0.33
		2.0	-86.95	6.37	-12.52	0.65
	Groundwater		-81.17	1.65	-11.94	0.04
	Seepage water		-80.83	2.05	-11.93	0.09
	Discharge water		-80.82	2.12	-11.88	0.14

*: Weighted mean
S.D.: Standard deviation

forest floor for a few months can be estimated from the difference of δ values between throughfall and deep soil water. Furthermore the difference of δ values between the throughfall and the groundwater may be used for calculation of the mean evaporation rate for about 20 years.

Under the equilibrium condition, the ratio of δ value of liquid water and that of vapor is given by,

$$\alpha = \frac{R_l}{R_v} \tag{6.2}$$

where α is the fractionation factor, R is the isotopic ratio, and suffix l and v represent liquid water and vapor, respectively. When the raindrops are formed successively from the vapor in the atmosphere, the relationship between the isotopic ratio of initial vapor, R_{vo} and that of vapor in a certain time, R_{vt} is given by the following equation,

$$\frac{R_{vt}}{R_{vo}} = f_v^{\alpha-1} \tag{6.3}$$

where f_v is the fraction of the remaining vapor at the time. Furthermore the isotopic ratios of the water before and after the enrichment by evaporation are described as follows (Hoefs, 1980; Tanaka *et al.*, 1992):

$$\frac{R_l^*}{R_l^{in}} = f_l^{(1/\alpha-1)} \tag{6.4}$$

where R_l^* is the isotopic ratio of remaining water, R_l^{in} is the isotopic ratio of the water before the enrichment begins and f_1 is the fraction of the remaining water. Using the δ-notation, eqn.(6.4) can be rewritten as follows:

$$\frac{\delta^{*} + 10^{3}}{\delta^{in} + 10^{3}} = f_1^{(1/\alpha - 1)} \qquad (6.5)$$

In general, the approximation is mathematically possible as follows:

$$ln\ (\ 1 + x\ ) \simeq x \quad (\ |x|<<1\ ) \qquad (6.6)$$

Then the eqn. (6.5) can be reduced finally as follows:

$$\delta^{*} - \delta^{in} = 10^{3}\ (\ \frac{1}{\alpha} - 1\ )\ ln\ f_1 \qquad (6.7)$$

In eqn. (6.7), provided the fractionation factor α, the average isotopic ratios of the infiltrating water and the remaining water is known, an estimation of water loss by evaporation is possible. The α for ^{18}O is given by the following equation (Majzoube, 1971),

$$ln\ \alpha = \frac{1.137}{T^{2}} \times 10^{3} - \frac{0.4156}{T} - 2.0667 \times 10^{-3} \qquad (6.8)$$

where T is the temperature. Given the annual average temperature 6.0 °C of the Kawakami basin for T, the α value of 1.011 is derived by eqn. (6.8). Then provided the weighted mean of throughfall -12.4 per mil as $\delta^{in}{}^{18}O$ and the average value of groundwater -11.9 per mil as $\delta^{*}{}^{18}O$, and 1.011 as α for eqn. (6.7), one can get the value of 0.95 for f. In consequence 5 % of the throughfall might be lost by the evaporation from the forest floor. The annual rainfall interception ratio by the tree crown is estimated to be 15 % in the basin (chapter 4). Therefore the evaporation from the forest floor is evaluated to be about 4 % of the annual rainfall. Table 7 shows some data of evaporation from the forest floor estimated by a few researchers. The value of this basin estimated from the isotopic ratios is within the range of the ratio listed in Table 7, although the method for measurement of the actual evaporation from the forest floor in the mountainous basins has not been completed. This means that it is appropriate to assume that the difference of the isotopic ratios between the throughfall and the groundwater is due to the evaporation from the forest floor. Furthermore the present procedure described above appears to provide an excellent method to estimate the mean evaporation rate from the forest floor for the long-time period.

6-2. Characteristics of stable isotopic ratio profiles of soil water

In this section the monthly variation of the vertical profile of stable isotopic ratio of soil water and its characteristics will be discussed.

Table 7 The ratio of evaporation from the forest floor against evapotranspiration.

References	Ratio(%)	Vegetation
Stälfelt (1963)	20	Young pine
Rutter (1966)	8	Red pine
Baumgarntner (1967)	10	Young pine
Tajchman (1972)	5	Young pine
Hattori (1983)	8.9*	Hinoki

*: The value is the ratio against annual precipitation.
(From Hattori, 1991)

The $\delta^{18}O$ values of throughfall, stemflow, soil water at each depth, groundwater and seepage water are demonstrated in Fig. 41 at sites A and C from May through November, 1991 together with the average $\delta^{18}O$ values of rainfall and discharge water. The maximum, minimum and mean values of those components are shown in Fig. 42.

As mentioned in chapter 4 the throughfall is a major component of an input to the forest floor. It is obvious from Fig. 41 that the δ values of shallow soil water above 0.5 m depth at each month reflect those of throughfall in that month. This means that the δ value profiles of soil water above 0.5 m depth may be formed by the displacement flow of soil water like a piston flow. On the contrary, below the depth of 0.7 m at site A and 1.1 m at site C, the $\delta^{18}O$ of each month shows the similar values. This feature is more obvious in Fig. 42. The $\delta^{18}O$ between 0.1 m and 0.5 m depth is considerably variable ranging from -11 to -18 per mil. In contrast, below the depth of 0.7 m $\delta^{18}O$ value keeps the stable value of about -12 per mil at site A. At site C there is a significant variation of $\delta^{18}O$ above 1.1 m and below 1.1 m the δ value dose not change much. This suggests that an isotopic homogenization of soil water may occur between the soil surface and 1 m depth. The main cause of homogenization of solute concentration have been thought to be the effect of diffusion (e.g. Zimmermann *et al.*, 1966). If one assume that the effect of diffusion is major cause of the isotopic homogenization of soil water, the range of variation of monthly δ value should decrease gradually with

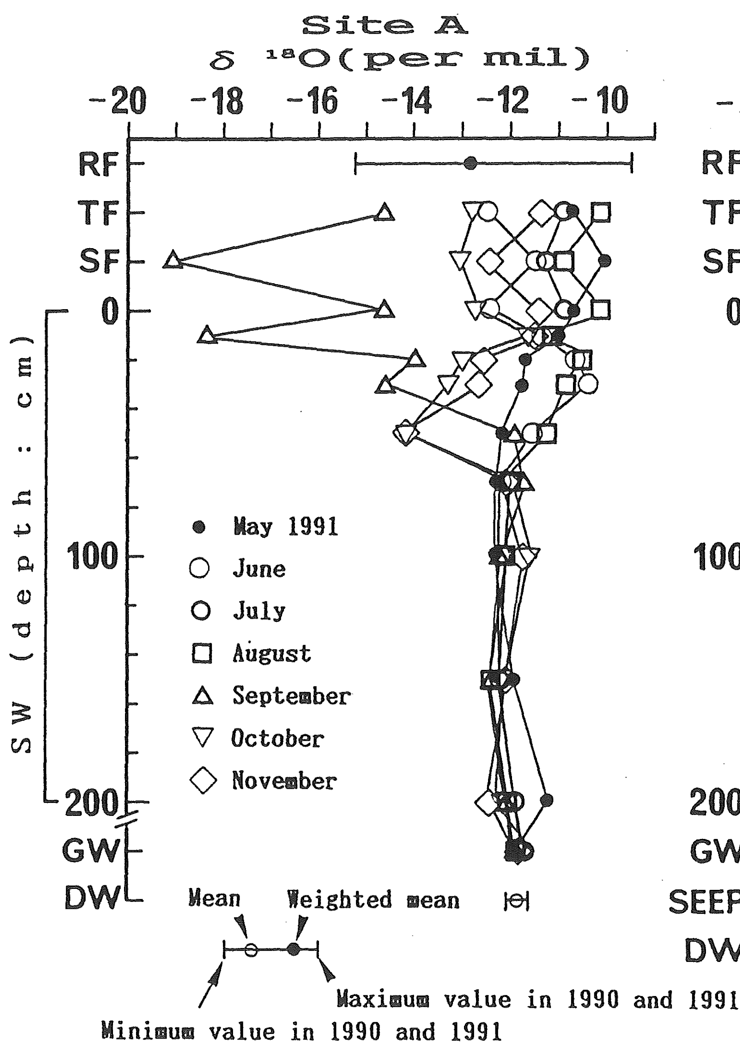


Fig. 41 $\delta^{18}O$ values of throughfall, stemflow, soil water, groundwater and seepage water.

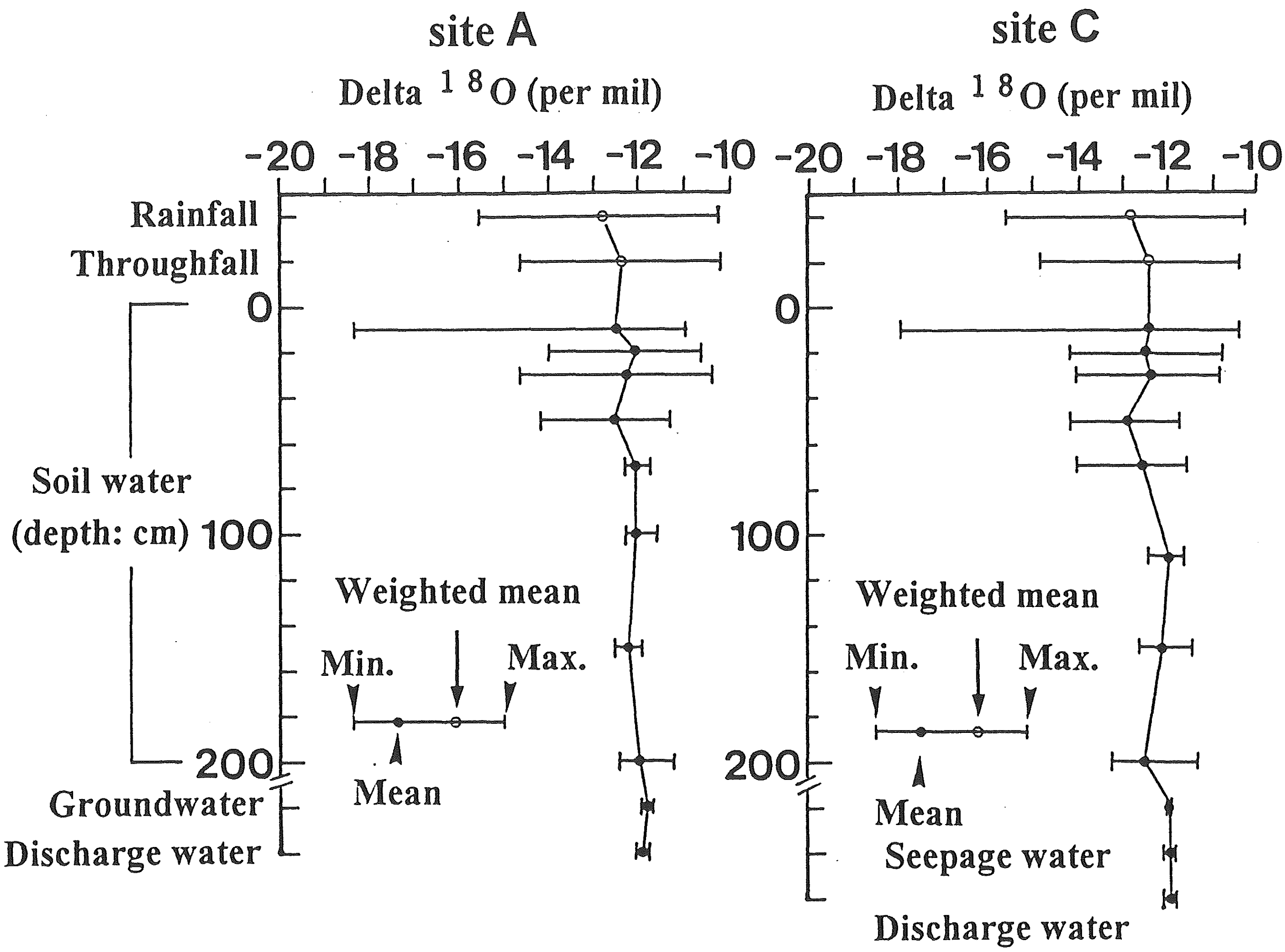


Fig. 42 $\delta^{18}O$ values (maximum, minimum and mean) of rainfall, throughfall, soil water, groundwater, seepage water and discharge water.

depth under steady state condition. However Fig. 42 clearly shows that the range of $\delta^{18}O$ variation decreases discontinuously at a certain depth, which is at around 0.7 m at site A and 1.1 m at site C. Furthermore the range of variation at 2.0 m depth is larger than that of 1.0 m in both sites. To explain these facts another factor in addition to the diffusion should also be considered. It appears to be reasonable to explain these facts by the behavior of soil water movement described in chapter 5.

As mentioned in chapter 5, the hydraulic head profiles of soil water vary dynamically above 1 m in response to the rainfall events or evapotranspiration. When the divergent zero flux plane (D-ZFP) disappears by the rainfall events, the convergent zero flux plane (C-ZFP) is formed and falls gradually down to the depth of the D-ZFP. Mixing of soil water should occur at the depth where the C-ZFP is formed, because the downward and upward soil water fluxes converge at the C-ZFP. The deepest depth of D-ZFP is 1 m. Therefore the depth where the mixing of soil water is caused by this process is limited above 1 m. Furthermore the C-ZFP is formed frequently with the disappearance of the D-ZFP. It should be considered that the mixing of soil water at the C-ZFP causes the isotopic homogenization above the depth of 1 m. In the following sections, the isotopic ratio profiles of shallow soil water determined by the displacement flow model will be discussed at first. Then the displacement flow model with consideration of mixing as mentioned above will be used to explain the isotopic homogenization of soil water.

6-3. The mechanism of isotopic homogenization of soil water

6-3-1. Consideration of isotopic ratio profiles of soil water by a displacement flow model

In the Kawakami basin, a macro pore system such as a pipe network which contributes to the rapid drainage has not been observed. Therefore it is not needed to consider the short-cut percolation system to account for the soil water movement. As mentioned in section 6-2, the isotopic ratio profiles of shallow soil water reflect those of throughfall. It appears to be possible to explain the δ value profiles of soil water by the displacement flow model (Kayane *et al.*, 1980; Shimada, 1988).

In the basin the average value of infiltration capacity is 260 mm/hr (Tsujimura *et al.*, 1991) and Horton overland flow has never been observed during six-years observation. Accordingly, the infiltration rate (I) can be expressed as follow:

$$I = P - E \tag{6.9}$$

where P is the rainfall and E is the evapotranspiration. If the infiltrated water displaces the previously stored soil water without mixing and if the soil water previous stored is pushed down one after another like a percolation of piston flow, the isotopic ratios profile of the displaced soil water during a given month can be found. This is the basic concept of the displacement flow model (Shimada, 1983) and the schematic diagram of the concept is shown in Fig. 43.

The temporal change of soil moisture contents is small as mentioned in section 4-3, so the soil water profile for displacement flow model was determined by averaging the soil moisture contents measured by the neutron moisture meter from May through November in 1991. The evapotranspiration rate for calculation of *I* was evaluated by the short-time period water budget method given in section 4-4 (Table 8), though it was compensated partly by the Hamon method (Hamon, 1961) when the water budget method was not available. The monthly infiltration rates for the displacement flow model are listed in Table 8.

In order to obtain the profile of the $\delta^{18}O$ in soil water at the end of each month by the displacement flow model, the estimated monthly infiltration (*I*) from May through November in 1991 was replaced with the soil moisture content profile and accumulated. The $\delta^{18}O$ of monthly throughfall from May through November in 1991 was used as an infiltration water and the water was replaced (Fig. 44). Finally the vertical $\delta^{18}O$ profile in November was completed by the displacement flow model as shown in Fig. 44. In Fig. 44 the results of model are compared with the observed data. It is obvious from the figure that the displacement flow model can approximate the δ values of shallow soil water between the surface and about 0.5 m

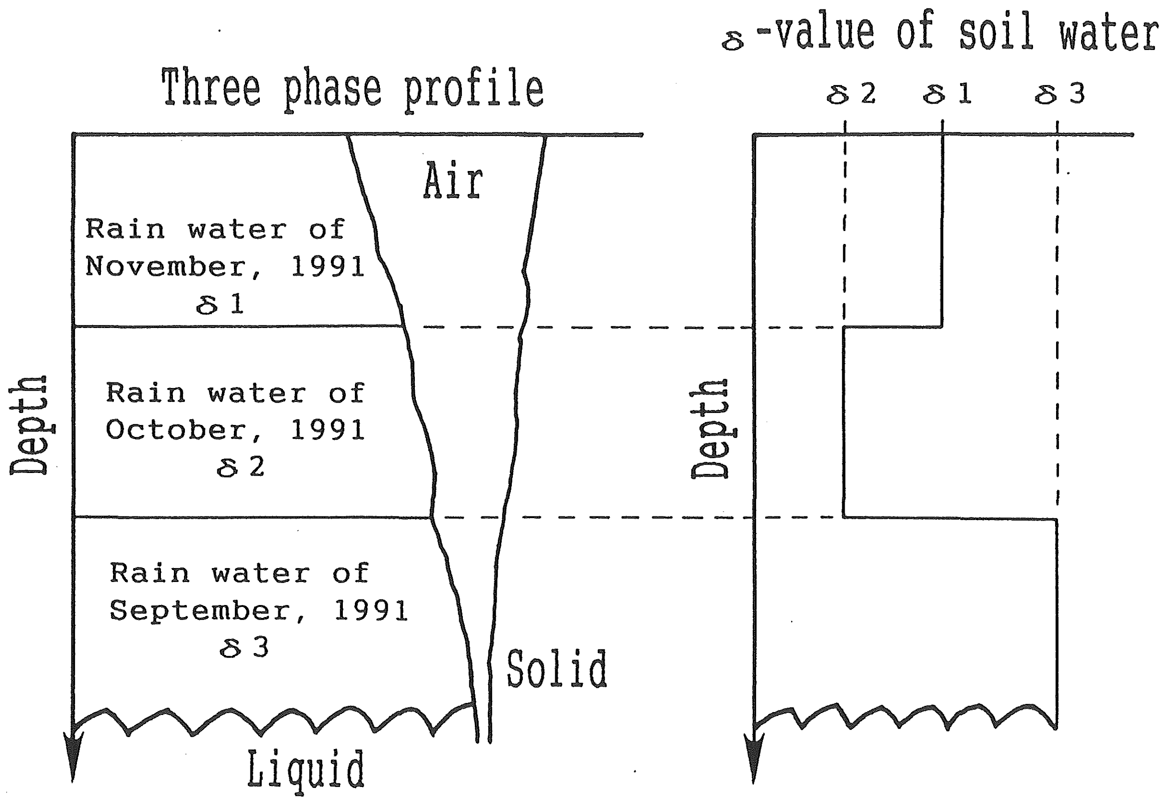


Fig. 43 Schematic diagram of displacement flow model.

Table 8 Infiltration rate for displacement flow model.

Period	Rainfall mm	ET mm	I.R. mm
May, 1 - Jun., 1	69	60.2	9.2
Jun., 2 - Jun., 28	159	89.8	69.2
Jun., 29 - Jul., 29	266	106.0	160.0
Jul., 30 - Aug., 23	119	77.8	41.2
Aug., 24 - Sep., 26	342	86.6	255.4
Sep., 27 - Nov., 1	344	90.0	254.0
Nov., 2 - Nov., 28	50	17.0	33.0

ET: Evapotranspiration estimated by the short-time period water budget method.
I.R.: Infiltration rate = (Rainfall) - (ET)

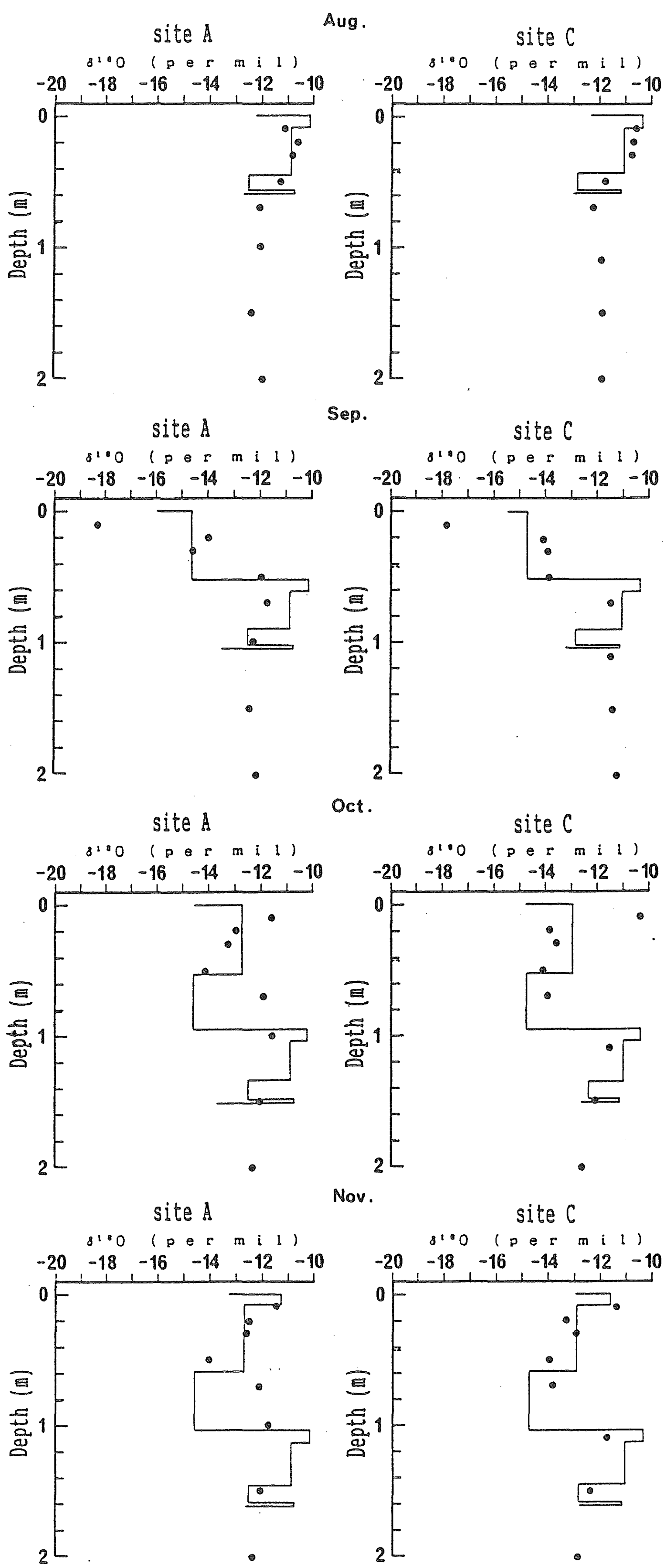


Fig. 44 $\delta^{18}O$ profile of soil water by displacement flow model from August through November, 1991. (solid circles show the observed values) depth. However, below 0.7 m depth the model can not follow the observed isotopic ratios of soil water. Though this model dose not consider the effect of mixing in the percolation, it seems good enough to reproduce the δ values of soil water above the 0.5 m depth. As a next stage, it is necessary that the mechanism of isotopic homogenization of soil water should be explained on the basis of observed data on soil water movement. This subject will be discussed in the following section.

6-3-2. Consideration of isotopic homogenization of soil water by a displacement flow model with mixing

In this section the displacement flow model is modified by considering the effect of soil water mixing occurred at the C-ZFP. In the model without mixing, the monthly infiltration water displaces the previously stored water in a monthly period. Contrary to that, in the displacement flow model with mixing, a month is divided into some periods (Fig. 45). The beginning time of a period corresponds to the beginning of the monthly period or the day when a D-ZFP of the previous period disappears. The ending of the period corresponds to the last date of the monthly period or to the date when the D-ZFP of the present period disappears (Fig. 45). The mixing will occur at the end of the period with C-ZFP in this model. Therefore the mixing dose not occur in the period without the D-ZFP and the C-ZFP. The infiltration rate of each period was determined by dividing the monthly

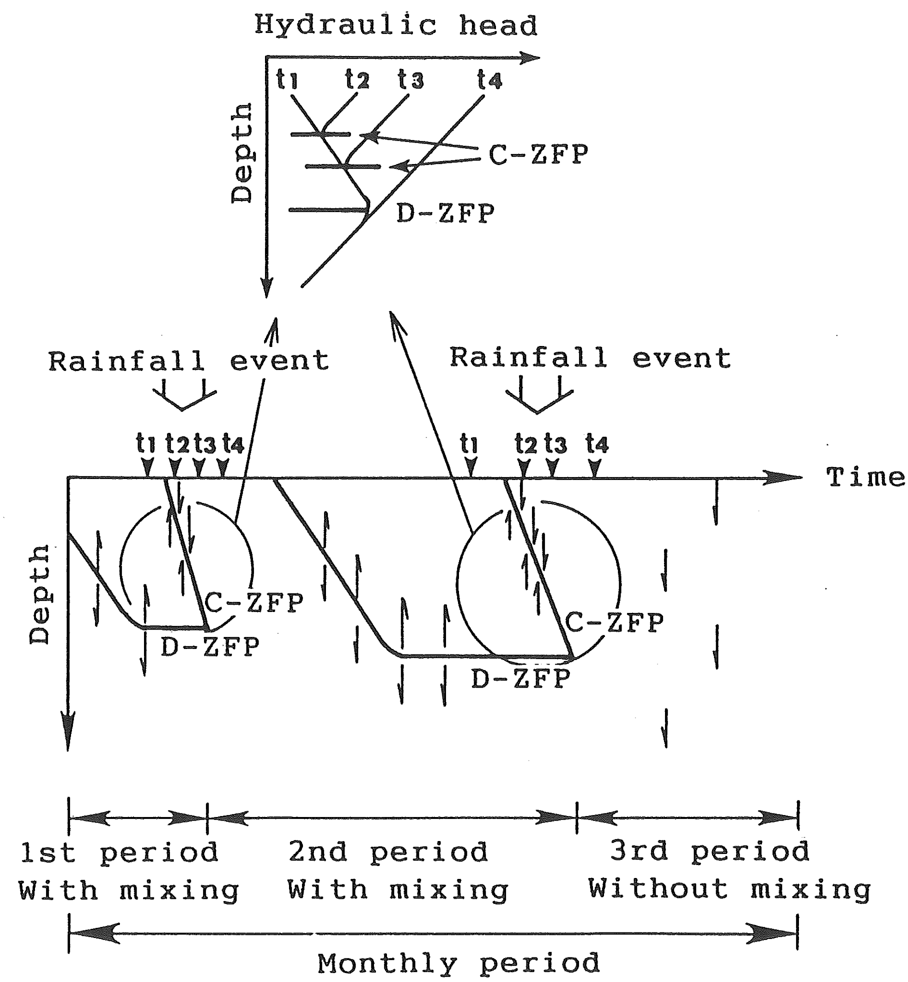


Fig. 45 Schematic diagram of soil water mixing process. (The arrows show the direction of soil water flux)
D-ZFP: Depth of divergent zero flux plane
C-ZFP: Depth of convergent zero flux plane

infiltration rate by days of the period. The δ value of infiltrated water was the same throughout a monthly period.

The mixing of δ values of soil water was conducted for every 0.2 m depth interval from the soil surface to the depth of D-ZFP by the weighted mean mixing method after Shimada (1983). The depth interval of 0.2 m was determined based on the depth interval for the measurement of soil moisture contents. As shown in Fig. 46, the soil water from the depth of (n-0.2) m to (n) m includes three periods infiltration waters from m to m+2. The infiltration waters of the periods of m and m+2 are extending over the other zones of n-0.2 and n+0.2, then the estimated infiltration for both periods are expressed as follows:

$$I_m = I_{m,n-0.2} + I_{m,n} \tag{6.10}$$

$$I_{m+2} = I_{m+2,n} + I_{m+2,n+0.2} \tag{6.11}$$

For the total volume of soil water in the zone n (V_n), the following equation is presented,

$$V_n = I_{m,n} + I_{m+1} + I_{m+2,n} \tag{6.12}$$

Therefore the δ value of mixing soil water in the zone n (δ_n) is determined as follows:

$$\delta_n = \frac{1}{V_n} (I_{m,n} \cdot \delta_m + I_{m+1} \cdot \delta_{m+1} + I_{m+2,n} \cdot \delta_{m+2}) \tag{6.13}$$

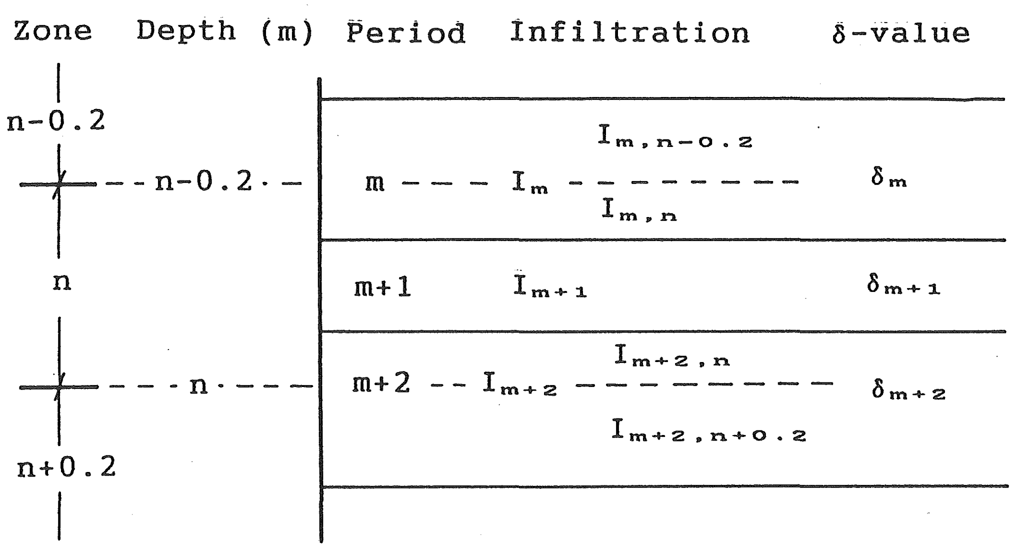


Fig. 46 Schematic diagram for mixing of soil water δ values.

This procedure applied to the soil water from the surface to the depth of D-ZFP produces the profiles of mixed $\delta^{18}O$ values of soil water at the end of a period. Then, the infiltrated water of the next period displaced the stored water having been mixed as described. From May through July, 1991 the displacement flow model without mixing was conducted in same way as in the section 6-3-1, because the volume of infiltrated water was so small that the $\delta^{18}O$ profiles were completed only in the shallow depth before July. After August the displacement flow model with mixing was conducted as mentioned above.

The results of calculation at the end of each month are presented in Fig. 47 together with the observed $\delta^{18}O$ profiles of soil water from August through November, 1991. The estimated profiles of $\delta^{18}O$ appear to be able to reproduce the observed profiles well. The δ value by the model agrees well with the observed data below the depth of 1 m. This means that the isotopic homogenization can be explained by the effect of soil water mixing mentioned above. The estimated profiles of site C are not so smoothed as those of site A. This appears to be caused by the difference of the frequency and the depth of D-ZFP formation between site A and site C. The isopleth of hydraulic head with time at site C (Fig. 31) compared with that of site A in September indicates that the frequency of appearance of D-ZFP in site C is less than that of site A and the depth of D-ZFP at site C is shallower than that of site A. This

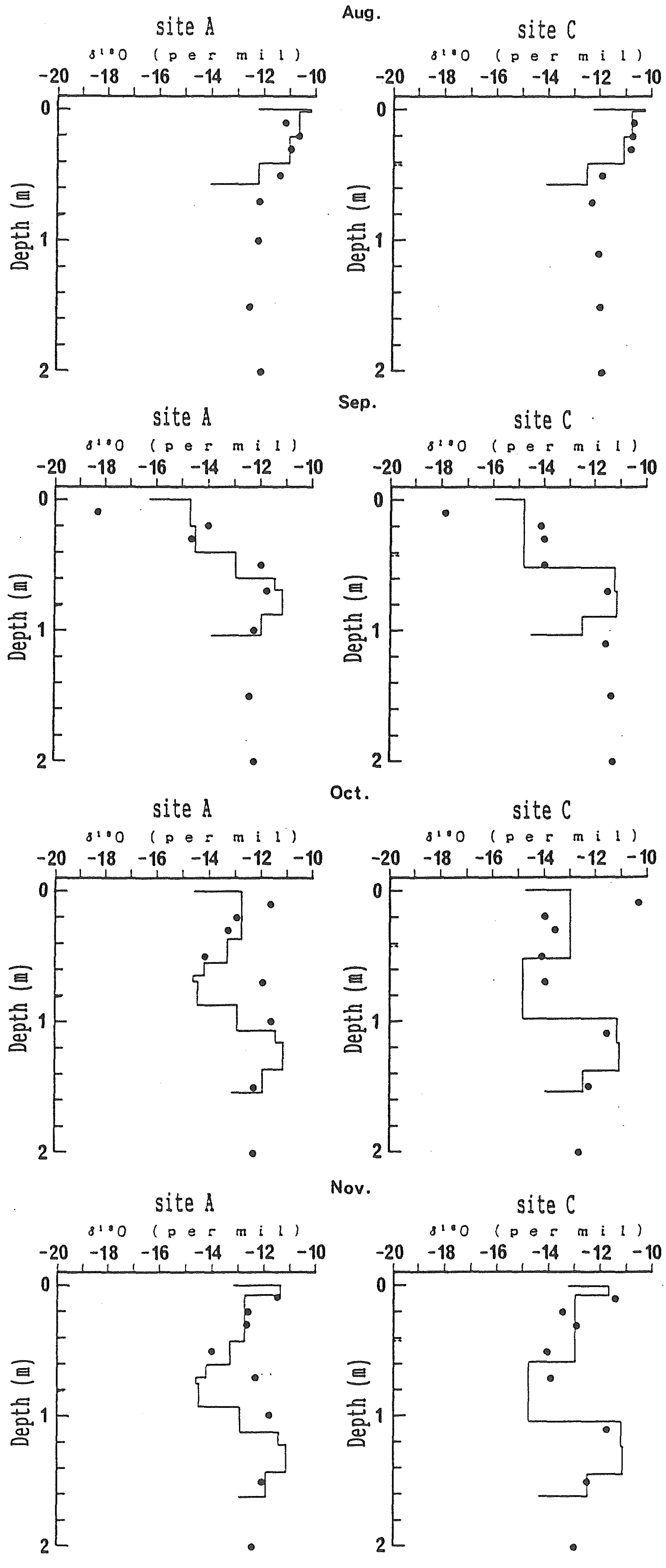


Fig. 47 $\delta^{18}O$ profile of soil water by displacement flow model with mixing from August through November, 1991. (solid circles show the observed values) suggests that the depth which the effect of transpiration extends in site C should be shallower than that of site A. In consequence at site C the mixing of soil water is difficult to occur as compared with site A. Therefore the calculated $\delta^{18}O$ profiles of site C are not so smoothed as in site A.

The estimated profile does not agree well with the observed one between the depth of 0.7 and 1.0 m at sites A and C. This appears to be due to the inappropriate estimation of infiltration rate. Especially, the evapotranspiration rate was probably underestimated. Consequently the infiltration rate appears to be overestimated in September and October, so that the calculated profile may not agree with the observed one from 0.7 m to 1.0 m depth.

In conclusion isotopic homogenization of soil water appears to be caused by the mechanism as summarized below. The infiltrated water displaces the stored soil water like a piston flow at first. During the rainless period the D-ZFP is formed, then the C-ZFP appears with a rainfall event and falls down to the depth of D-ZFP, so that the mixing of soil water occurs around the C-ZFP until the D-ZFP disappears. In consequence below the depth of 1 m the isotopic ratio profiles of soil water keep the stable value. The effect of molecular diffusion is an also important factor for the isotopic homogenization of soil water. This effect appears to be useful for explanation of the gradual isotopic homogenization in a relatively thick soil mantle. In contrast, the mixing of soil water at the C-ZFP can explain the isotopic homogenization in the shallow soil water above the depth of D-ZFP.

Chapter VII

Discussion

The notable characteristic of the Kawakami Experimental Basin is that the B horizon occupies about 70 % of whole soil mantle as mentioned in section 2-2. The soil water characteristic of the B horizon is that the variation of soil moisture with change in pressure head is considerably small. This is common to whole of the basin. On the contrary, in the basin underlain by granite, the soil water characteristic is that the upper boundary of capillary zone is very low (< pF 1.0) and the variation of soil moisture content with the change of pressure head is remarkably large (e.g. Kubota *et al.*, 1987). In contrast in the hilly basins of Tertiary, the soil moisture characteristic is that the soil moisture content varies gradually with change in the pressure head and the upper boundary of capillary zone is not well defined in A and B horizons, and that in C horizon the gradient of the soil moisture characteristic curve is considerably large (e.g. Ohta *et al.*, 1985a). As compared with these basins, it can be pointed out that the Kawakami basin consists of the soil mantle with high water retentivity.

The characteristic of B horizon affects the behavior of soil water movement in the Kawakami basin. The variation of soil water flux in response to the rainfall events is remarkable in the A horizon and the upper layer of the B horizon. On the other hand, the divergent zero flux plane (D-ZFP) falls down to the depth of 1 m, so that the soil water flux above 1 m depth changes considerably during the rainless periods. Furthermore the change of soil moisture contents is notable especially above 1 m depth on every part of the slope (section 4-3). Therefore the soil mantle in the Kawakami basin can be distinguished into two layers with the boundary at 1 m depth from the view point of the behavior of soil water movement. One is an active layer above 1 m and the other is a stable layer below 1 m. In the active layer the soil water movement varies dynamically in response to the rainfall events or the evapotranspiration, in contrast, the soil water movement is not so variable in the stable layer. The existing of the stable layer in which the soil water flux always keeps the stable value may contribute to the diminishment of the runoff peak during the storm events. The boundary depth of 1 m probably is determined by the physical properties of soil mantle and the roots distribution of vegetation.

Furthermore the active layer plays an important role in the isotopic homogenization of soil water as mentioned in section 6-3, since the isotopic homogenization is completed within the active layer. The isotopic homogenization of water in the mountainous basins has been a considerably important subject. It has been suggested by some researchers (e.g. Mizutani, 1986) that the isotope values of groundwater show the stable values as compared with those of rainfall. Kitaoka and Yoshioka (1984) has supposed that the mixing of some groundwater with different residence time occurs in the discharge area, so that the groundwater shows the stable isotope values. McDonnell *et al*. (1991) has mentioned that the subsurface water was mixed with rainfall along with the transfer of the water from the upper to the lower part of the slope. The suggestion by Kitaoka and Yoshioka may be applied to the basin of relatively large scale which includes several groundwater flow systems. The assumption of McDonnell *et al*. appears to be effective in such basin where the soil mantle is relatively thin and the saturated lateral flow occurs frequently on the slope. However an explanation different from these have to be considered for isotopic homogenization in the Kawakami basin, because the area of the basin is relatively small (0.14 km^2) and the lateral flow have not observed in the slope (Tsujimura, 1993).

The previous studies have not paid enough attention to the soil water movement, including mixing process, in the unsaturated zone for the explanation of isotopic homogenization in the headwater basins. The present study has explained this by the mixing process of soil water occurring around the convergent zero flux plane (C-ZFP) which is formed when the D-ZFP disappears. This explanation is more reasonable than that by the molecular diffusion of water alone (Zimmerman *et al*., 1966). Generally in Japan, the deepest depth of D-ZFP being formed appears to be around 1 m in the forested mountainous basins (Ohta, 1992). Furthermore the rainless period dose not persist for a long time in Japan so that the appearance and disappearance of the D-ZFP and the C-ZFP are observed frequently. In consequence the mixing of soil water at the depth of C-ZFP occurs quite often. Therefore it is quite possible that this explanation can be applicable to the other forested mountainous basins in Japan as a likely cause of isotopic homogenization of water.

The mixing process when the groundwater discharges into the stream (Kitaoka and Yoshioka, 1984) and the effect of molecular diffusion (Zimmerman *et al.*, 1966) appears to be also important factors for the isotopic homogenization. However the present study has shown that the isotopic homogenization of water occurs in the shallow soil by the mixing process of soil water. It should be emphasized that the isotopic values of soil water is homogenized considerably before the soil water recharges the groundwater in the forested headwater basins.

Chapter VIII

Conclusions

In the present study the behavior of soil water movement has been investigated by the combination of a hydrometric method and an isotope tracer method in a forested headwater basin, called Kawakami Experimental Basin underlain by volcanic rocks in Neogene and located in Nagano prefecture, central Japan.

The hydrometric observations such as measurement of the temporal change of soil moisture contents and hydraulic head profiles of soil water have shown the dynamic behavior of soil water movement especially in the shallow soil. Furthermore, it became clear that the process of formation of stable isotopic ratio profiles of soil water can be explained by the results of hydrometric observations.

The results of the study are summarized as follows;

1) About 70 % of the soil mantle in the Kawakami basin is occupied by the B horizon, which is characterized by high water retentivity. The B horizon has a physical properties that the soil moisture is relatively high, the upper boundary of the saturated capillary zone is not well defined and the change of soil moisture contents with the variation of pressure head is considerably small.

2) The temporal change of soil moisture contents is notable above 1 m depth and the vicinity of the bedrock surface. On the contrary, the soil moisture is relatively stable except

for those layers. The temporal variation of soil moisture of whole soil mantle is relatively small, with about 5 % by volumetric water content.

3) The soil water flux above 1 m depth shows considerable variation during the storm events and the rainless periods. In contrast, the temporal change of the soil water flux below 1 m depth is much smaller. The vertical downward component of the soil water flux is a major component as compared with the downslope flux component.

4) The behavior of soil water movement is considerably dynamic above 1 m depth, because the appearance and disappearance of divergent zero flux plane and convergent zero flux plane occur frequently by the effect of the rainfall events and the evapotranspiration. The deepest depth where the divergent and convergent zero flux planes are formed is about 1 m, and thus the soil water movement is relatively stable below the 1 m depth.

5) The isotopic composition of deuterium and oxygen-18 of rainfall shows a seasonal variation clearly. On the other hand, the isotopic ratio of groundwater shows a very stable value throughout the seasons. However the average isotopic ratio of groundwater dose not agree with the annual weighted mean value of rainfall. This can be explained by the enrichment of shallow soil water with the evaporation from the forest floor.

6) The monthly isotopic ratios of deuterium and oxygen-18 of soil water above 0.5 m depth show the values corresponding to those of throughfall, and the monthly isotopic ratios of

soil water below 1 m depth converge on the values of groundwater. This trend suggests that the isotopic homogenization of soil water may occur above 1 m depth.

7) The isotopic homogenization of soil water can not be explained solely by the molecular diffusion of water and by the effect of condensation with the evaporation from the forest floor.

8) The isotopic homogenization of soil water can be explained by the dynamic behavior of soil water movement above 1 m depth, that is the soil water displacement flow with consideration of mixing which occurs around the convergent zero flux plane.

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