Soil water behaviors and their effects on river runoffs: The Saromabetsu River basin, Hokkaido

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Abstract

The soil water behaviors during rainfall or snowmelt events were explored by monitoring the soil moisture on the slope of forest and grassland in the Saromabetsu river basin, Hokkaido, in the summer season of 2008 and the snowmelt season of 2009. The river basin is covered mostly by forest (75.2% in area) and farmland (21.5% in area; mainly grassland, wheat field and corn field). The soil layer, common to the slope of forest and grassland, consisted of porous organic layer (A layer) less than 0.3 m thick and organic/inorganic (B layer) about 10 m or more thick. The B layer is silty clay with low hydraulic conductivity of 10^{-4} cm/s order. The 4-channel profilers (8cm, 18cm, 28cm and 38cm in soil depth) for measurements of volumetric water content (cm^3/cm^3) indicated that, common to rainfall and snowmelt events, the 8-28 cm layer at forest stores infiltrated water for a few days after events, and then returns to the previous moisture level by gradual drainage. At grassland, the drainage after events was slight except for 8 cm depth. At a rainfall of more than 20 mm/h, the saturated zone appeared at 38 cm depth in forest and at depths of more than 8 cm in grassland. Hence, in the events, the percolation and subsurface flow are judged to be active. The percolation at 38 cm depth of forest and grassland was calculated for rainfall and snowmelt events. Meanwhile, runoff analyses by the tank model revealed that surface and intermediate runoffs occupy more 70% of rainfall or snowmelt runoffs in the Saromabetsu River. As a result, there existed the linear relationship between the total percolation and surface and intermediate runoffs for the runoff events. Hence, it is suggested that the surface and intermediate runoffs from runoff analyses occur as the percolation and subsequent subsurface flow below the soil surface layer.

Key words: soil moisture, rainfall runoff, snowmelt runoff, percolation, tank model, surface runoff, intermediate runoff

I. Introduction

Soil moisture is widely recognized as a key parameter in environmental processes, including meteorology, hydrology, agriculture and climate change. From a hydrologic viewpoint, soil moisture controls the partition of rainwater into percolation,

storage and drainage, and therefore has an important effect on the runoff behavior of catchments (Aubert *et al.*, 2003). Runoff generation is a highly nonlinear process and threshold processes are observed in many cases (Grayson *et al.*, 1997; Meyles *et al.*, 2003), e.g., saturation excess flow starts when soil storage capacity is filled. The soil type is mainly de-

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pendent on parent material (bedrock), local climate and topography, but its management can affect the properties very profoundly. In the previous studies, many researchers pointed out that the macropore flow and the matrix flow in the vadose zone play the important role on runoff generation and streamflow recession, respectively (Dingman, 2002). However, it is very difficult to predict how such flows are formed in the subsurface layer and affect the runoff generation and depletion, since a number of factors affect soil hydrology, including texture, structure, hydraulic conductivity, pore size distribution, and retention characteristics (Nielsen and Hansen, 1973). There are many studies on relations between observed soil moisture distributions and runoff generation on the basin slope (Calvo-Cases *et al.*, 2003; Penna, 2008) or at the small basin scale of 0.01 to 10 km² order (Harmes and Chanasyk, 1998; Gautam *et al.*, 2000; van Asch *et al.*, 2001; Meyles *et al.*, 2003; Martinez *et al.*, 2008; Penna *et al.*, 2011), but very few investigations on how different the soil moisture variability is between the summer and snowmelt seasons, and on how the observed soil moisture is related to river runoffs modeled on the basin scale of 10² km² or more order. In this study, in a river basin of 10² km² order, it is explored how soil water behaviours on the basin slope of different land use affect river runoffs by monitoring the soil moisture content in the summer and snowmelt seasons and by analyzing the river runoffs with a tank model (Sugawara, 1972).

II. Study area and filed observations

1. Study Area

The Saromabetsu river basin in Hokkaido, Japan (43°49' to 44°04'N and 143°04' to 144°00' E), has the area of 387.0 km² with the maximum elevation of 829 m a.s.l. (above sea level) at Mt. Nikoro (mean, 203.3 m a.s.l.) and the mean basin

slope angle of 9.73° (Fig. 1 and Fig. 2a). Fig. 2a was made up on the base of the 10 m GIS data of Gespatial Information Authority of Japan, Ministry of Land, Infrastructure, Transport and Tourism. The Saromabetsu River is the largest river flowing into a brackish lake, Lake Saroma connected to the Okhotsk Sea. The meteorological records by the Automated Meteorological Data Acquisition System (AMeDAS) at the Saroma town (site S in Fig. 1; 59.0 m a.s.l.) in 1980 – 2007 show the mean annual precipitation of 789 mm (583mm rainfall of May – November and 206mm snowfall of December – April) and the annual mean air temperature of 5.3°C (12.4°C in May – November and -4.6°C in December – April). The rainfall records are conducted also at site E and site H by the Abashiri District Public Works Management Office. The basin geology is Cretaceous to Paleogenic marine sedimentary rocks in the upstream and downstream areas, and Jurassic to Cretaceous basaltic rocks in the middle area (National Institute of Advanced Industrial Science and Technology, Japan; URL <http://www.aist.go.jp/GSJ/bHOK/gguide/Contents/GeoMap/all/HokGAh.html>). The gauging station of the Saromabetsu River is located at the Eidai Bridge in

Fig. 1. Location of the Saromabetsu river basin, Hokkaido, and observation sites (site E, river stage and rainfall; site S, AMeDAS; site H, rainfall).

the middle reach (site E in Fig. 1), and indicates the discharge of $3.6 \text{ m}^3/\text{s}$ averaged in June – November. The basin area upstream of the gauging station is 259.3 km^2 . The land cover of the upper river basin

Fig. 2. (a) Digital elevation model with main five mountains and (b) land cover of the Saromabetsu river basin.

in 2006 is composed of 75.2% forest, 21.5% farmland (60% grassland and 40% field), 1.3% urban area, 1.1% water surface and 0.9% wasteland (Fig. 2b: National and Regional Planning Bureau, Ministry of Land, Infrastructure, Transport and Tourism, Japan; URL <http://nlftp.mlit.go.jp/ksj/jpgis/datalist/KsjTmplt-L03-b.html>). The forest is mixed by the needleleaf and broadleaf trees (e.g., *Larix kaempferi*, *Abies sachalinensis*, *Betula platyphylla* var. *japonica*). The urban area is sporadically located along the river channel, but mostly at the Saroma town just upstream of site E. In this study, the soil water behaviors in the upper Saromabetsu river basin is specified by relating runoff analyses at site E to the soil moisture measurements on the slope upstream of site E (Fig. 1).

The soil moisture was measured at site A in forest and site B in grassland in the slope area upstream of site E (Fig. 1 and Fig. 3). The land cover in the slope area consists of forest, grassland, corn field and bareland (white) with a deserted house (black). Ditches 60 to 80 cm deep below the ground surface are excavated in or around the grassland and field to easily drain snowmelt water or rain-water (Fig. 3b). The slope faces the river channel in the downslope direction (Fig. 3a). Thus, part of

Fig. 3. Location of (a) slope area and (b) sites (site A at forest and site B at grassland) for soil moisture measurements. The infiltration capacity at the ground surface and the hydraulic conductivity of the soil were also obtained at site A, site B and site C (field) (Table 1).

meltwater or rainwater infiltrated into soils in the slope area could directly flow into the Saromabetsu River. Site A and site B are located at the basal parts of the forest slope and the grassland slope, respectively. The ditch at near site A could then work as a barrier obstructing the input of shallow subsurface water in forest into the grassland. The slope angle is 5.0° at site A (84.5m a.s.l.) in forest, 7.1° at site B (68.7 m a.s.l.) in grassland and 3.5° at site C (59.3 m a.s.l.) in field (Fig. 3b). The well drilling on the flood plain at 500 m north of the slope area furnishes the sedimentary information on silty deposits about 35 m thick above the bedrock with the bottom mineral layer 5 m thick (personal communication with a dairyman, Mr. K. Yano). The flood plain is developed up to around site S, 7.6 km upstream of site E, and upstream of around site S, the farmland slope or relatively steep forest slope exists as part of the riparian zone (Fig. 1 and Fig. 2).

2. Field Observations

The soil moisture in the slope area was monitored by 2 soil moisture profilers (4-channel probe; Delta-T Devices Ltd., UK, type PR2; accuracy, $\pm 0.002 \text{ cm}^3/\text{cm}^3$) in June 2008 – May 2009. One soil moisture profiler was set at site A on the slope base of forest, and the other at site B in grassland. In June 2008, access tubes of the profilers were inserted carefully into holes in soil layers of sites A and B, and then the profile probes were set into the access tubes. The soil moisture by the profilers was monitored at the same holes in June 2008 – May 2009 by keeping dry the interior of the access tubes. The measurement intervals are 20 minutes. The profilers measure soil moisture under condition that the square root of the permittivity, ϵ , for an electromagnetic field is proportional to the volumetric soil moisture content, θ (cm^3/cm^3), i.e., $\sqrt{\epsilon} \propto \theta$. The voltage from the probe, which is proportional to $\sqrt{\epsilon}$, was stored in a data logger,

and, using an empirical formula, changed into θ . The profiler can measure the soil moisture within about 100 mm in the radial direction from the probe at each depth of 8 cm, 18 cm, 28 cm and 38 cm below the ground surface. The saturated hydraulic conductivity and volumetric moisture content of the soils at site A, site B and site C were measured in laboratory. The infiltration capacity at the ground surface in the slope area was *in situ* measured by a double-ring infiltrometer (Type IV-40063, Turf-Tec, Inc., USA) at site A, site B and site C in July 2008. The seven trials at each site gave us the mean and observational scatter. The density of covered snow was obtained at site A and site B and site S in March and April, in order to get the degree-day factors in the degree-day approach for the calculation of snowmelt amount (Rango and Martinec, 1995).

Two pressure gauges for air pressure and water pressure plus air pressure (HOBO water level loggers; a range of 0 - 207kPa and accuracy of $\pm 0.10 \text{ kPa}$ for pressure and a range of -20 - 50°C and accuracy of $\pm 0.5^\circ\text{C}$ for temperature) were set at site E, in order to record the pressures at 30 min intervals for a period of January 2008 to May 2009. The water pressure was converted into water depth (m) by dividing the gravitational acceleration ($= 9.8 \text{ m/s}^2$) and the water density (kg/m^3) at water temperature recorded in the data logger underwater. A rating curve was established on the basis of frequent discharge measurements and simultaneous water depth records. As a result, the discharge includes the 10 – 15% observation error.

III. Observational results and discussion

1. Soil moisture variability

Fig. 4 shows hourly variations of volumetric moisture content (cm^3/cm^3) at (a) site A (forest) and (b) at site B (grassland), and (c) discharge and (d) rainfall at site E in the summer season

Fig. 4. Temporal variations of volumetric moisture content at (a) forest and (b) grassland, (c) river discharge at site E, and (d) rainfall at site S in the summer season of 2008.

of 1 June – 7 November 2008. At forest (Fig. 4a), the temporal variability of moisture content, being similar at all the soil depths, occurred at rainfall of more than 10 mm/h, and the drainage after rainfalls is to some degree faster at 8 cm and 18 cm than at 28 cm and 38 cm. The soils at depths of 8 cm, 18 cm, 28 cm and 38 cm exhibited water saturation at 0.40, 0.44, 0.43 and 0.43 cm^3/cm^3 , respectively. All of the water storages and drainages thus occurred under unsaturated condition. The similar variability at all the depths probably reflects non disturbance

in structure, and the gradual downward increase of the moisture content as the whole indicates the downward decrease of pore size or the downward fineness in texture.

At grassland (Fig. 4b), the moisture content varied in different ways at 0 cm to 38 cm depth; at 8 cm depth, water content increased greatly at the rainfall of 4 mm/h on 11 July. Thereafter, following the continual rainfalls, increased gradually until 4 September, and then decreased greatly under no rainfalls. At 18 cm, the moisture content similarly

increased in response to the rainfall of 4 mm/h on 11 July, but, thereafter, maintained the similar moisture level of 0.53 to 0.58 cm^3/cm^3 . At 28 cm and 38 cm depths, the moisture content increased in response to the small rainfall of 17 July and the large rainfall of 23 July, respectively, and, thereafter, maintained the similar moisture level of 0.63 – 0.65 cm^3/cm^3 and 0.45 – 0.48 cm^3/cm^3 , respectively. The grassland soils at 8 cm, 18 cm, 28 cm and 38 cm depth showed water saturation at 0.60, 0.65, 0.68 and 0.63 cm^3/cm^3 , respectively. The soils at 18 and 28 cm depth thus tend to become nearly saturated and almost saturated by the small rainfalls, respectively, while they at 8 and 38 cm depth get into saturation only by large rainfalls of more than 20 mm/h. In a period of 22 June to 10 July with few rainfalls, the moisture content consistently decreased especially at depths of 8 cm and 18 cm. Part of the decrease is probably due to the evaporation of soil water.

On the whole, the sensitivity to rainfalls is higher at grassland than at forest. This probably results from the high effective rainfall or the small interception of rainwater at grassland. The moisture content at grassland is more than 0.3 cm^3/cm^3 at all the depths, whereas that at forest is at any time more than 0.3 cm^3/cm^3 only at a depth of 38 cm. This means that the A layer, including the root zone, at forest is relatively porous and thick.

At the large rainfalls producing river discharge of more than 20 m^3/s , the forest soil stores infiltrated water similarly at all the depths, and, after the rainfalls, releases the stored water smoothly (Fig. 5a). At grassland, the drainage after the rainfalls occurs significantly only at 8 cm depth (Fig. 5b). The vertical measurement of soil moisture at wheat field in the slope area by Yagisawa (2008) indicated that the rainwater storage and drainage occur markedly only at 8 cm depth. In Fig. 4, it is seen that the soil moisture at 8 cm depth of grassland is sensitive even to rainfalls of less than 10 mm/h producing

Fig. 5. Temporal variations of volumetric moisture content at (a) forest and (b) grassland, (c) river discharge at site E, and (d) rainfall at site S for 21 – 25 July 2008.

river discharge of less than 10 m^3/s . Meanwhile, the forest soils, showing a definite response to rainfalls of more than 15 mm/h generating discharge of more than 20 m^3/s , tend to produce the rainwater storage and drainage at all the depths (especially, at 8 cm, 18 cm and 28 cm depth) (Fig. 5a). Thus, the large river runoffs likely occur in relation to the water storage and drainage in forest and the 0 – 10cm layer of farmland. On daytime with no rainfalls, the moisture content increased by 0.012 to 0.014 cm^3/cm^3 at depths of less than 38 cm at grassland (Fig. 5b). This results from the rise of soil water, probably due to the water absorption by grass

roots, following the transpiration. On daytime of 24 July, when the water rise was largest at grassland, the forest soil also increases the moisture content slightly. Little or no water rise at forest (Fig 5a) is probably caused by the little or non transpiration, which is produced during the evaporation from wet leaves between the rainfalls (Fig. 4d).

Table 1 shows infiltration capacity (mm/h) at the ground surfaces of site A (forest), site B (grassland) and site C (field), and saturated hydraulic conductivity (cm/s) at 35 cm and 75 cm depth of the three sites. The similar hydraulic conductivity at forest and grassland reflects the similar pore size or texture at depths of 35 cm or more. The infiltration capacity much larger than the rainfalls in Fig. 4d means that any Hortonian overland flows do not occur at the three sites. Meanwhile, the hydraulic conductivity at depths of 35 cm and 75cm is relatively low at the order of 10^{-4} cm/s. This suggests that, because of the small percolation, the saturated zone is produced at depths of 38 cm or more.

Fig. 6 shows hourly variations of volumetric moisture content (cm^3/cm^3) at (a) site A (forest) and (b) at site B (grassland), and (c) discharge, (d) precipitation, snow depth and air temperature at site E in the snowmelt season of 4 March – 22 May 2009. The soil moisture measurement at forest ceased on 15 April, due to the battery trouble. The snowmelt river runoffs in March appeared sporadically by snowmelt (and rainfall) under positive air temperature, but the relatively large runoffs after 3 April occurred by continuous positive air temperature (Fig. 6c and Fig. 6d). The largest snowmelt river runoffs

occurred on 10 April after the snow at site S disappeared. As a characteristic of the moisture records in the snowmelt season, the moisture content varied daily at all the depths of forest and at 8 cm depth of grassland. This indicates that the percolation in the surface layer of forest and just below the ground surface of grassland is sensitive to the snowmelt amount at the snow surface. After 3 April, showing positive air temperature on almost all the days, the moisture content varied daily also at all the depths of grassland. Especially, the moisture content at 38 cm depth varies greatly until 11 April, corresponding to the great daily variation of river discharge. The soil moisture variability at 38 cm depth of grassland also corresponds to the river runoffs before 3 April. This suggests that, at the abrupt large snowmelt, in addition to the percolation in the soil layer of forest, the percolation at depths of more than 38 cm of grassland plays the important role on the runoffs.

On the whole, the temporal patterns of soil moisture are similar to those in the summer season (Fig. 4a and Fig. 4b); the meltwater (sometimes, plus rainwater) storage and drainage occur markedly at 8 to 28 cm depths of forest and at 8 cm depth of grassland. Especially, the moisture variability at 18 cm and 28 cm depth of forest corresponds to the temporal variability of discharge including the peak discharges of 10 and 14 April. This suggests that the main snowmelt runoffs are produced in close relation to the meltwater storage and drainage between 18 cm and 28 cm depths at forest.

The large peak discharge of 1 May was produced by the melt of snow newly accumulated on

Table 1. Infiltration capacity and saturated hydraulic conductivity, K_s , at 35 cm and 75 cm depth of forest (site A), grassland (site B) and field (site C).

	Infiltration capacity (mm/hr)	K_s at 35cm (cm/s)	K_s at 75cm (cm/s)
Forest	$3,240 \pm 280$	1.7×10^{-4}	1.3×10^{-4}
Grassland	270 ± 65	5.3×10^{-4}	3.9×10^{-4}
Field	576 ± 92	4.3×10^{-4}	3.9×10^{-4}

Fig. 6. Temporal variations of volumetric moisture content at (a) forest and (b) grassland, (c) river discharge at site E, and (d) snow depth, air temperature and precipitation at site S in the snowmelt season of 2009.

26 – 27 April (Fig. 6c and Fig. 6d). Corresponding to the runoff, the moisture content at grassland varied greatly at depths of 8 cm, 18 cm and 38 cm, and, at the maximum moisture content of 1320h – 1520h, 30 April, the soil at depths of 8 – 38 cm was simultaneously saturated. Then, the saturation overland flow, contributing to part of the runoff,

possibly occurred. At 0500h – 1600h of 10 April, the air temperature increased up to 20 $^{\circ}\text{C}$ at maximum. Then, the moisture content at 8 cm depth of grassland abruptly increased from 0.29 to 0.44 cm^3/cm^3 , and thereafter, maintained the similar moisture level (0.44 – 0.52 cm^3/cm^3) until the morning of 30 April. The low drainage, compared with that at 8 cm

depth of forest, could thus provide the occurrence of such an overland flow.

As shown by Fig. 4b and Fig. 6b, at 38 cm depth of grassland, there appeared the very high response (rapid water storage up to the near saturation and rapid drainage to the previous moisture level) to rainfall and snowmelt events. This may occur, also when there is some space between the access tube of the profiler's probe and the surrounding soil layer. However, the effect of such space during the moisture measurements is considered to be low, if the linear relationship is established between the percolation amount and simulated runoffs calculated for the events.

2. Calculation of the percolation amount

Preparation

The soil moisture observations indicated that the percolation in the soil surface layer of forest and at 38 cm depth of grassland may be related to the rainfall and snowmelt river runoffs. The amount of percolation is here calculated to know what volume of soil water moves in each of rainfall and snowmelt events. The soil layer at 0 to 38 cm depths is divided into four layers ($i = 1, 2, 3, 4$) of 0 – 8 cm, 8 – 18 cm, 18 – 28 cm and 28 – 38 cm depth, respectively. First, the total moisture content, W_{it} (mm), of the i layer at a time t was calculated as follows:

$$W_{it} = \begin{cases} \theta_{i,t} \times 80 & (i = 1) \\ \frac{(\theta_{i,t} + \theta_{i-1,t})}{2} \times 100 & (i = 2, 3, 4) \end{cases} \quad (1)$$

where $\theta_{i,t}$ is the volumetric moisture content (cm^3/cm^3) at depths of 8cm, 18cm, 28cm and 38 cm ($i = 1, 2, 3$ and 4, respectively). As shown by the first equation of Equation (1), the volumetric moisture content within the ($i=1$) layer was assumed to be constant at $\theta_{1,t}$. Using the water content, $W_{i,t}$, and assuming the moisture content to be horizontally uniform, the equation of continuity is given as

$$\frac{\partial W}{\partial t} = - \left(\frac{\partial q_z}{\partial z} \right) \quad (2)$$

where q_z is the downward flow rate (mm/s). Here z is positive downward from the ground surface. Applying Equation (2), the percolation amount, q_{i,t_1-t_2} (mm), at a depth of i at between t_1 and t_2 is approximately obtained by the following equations (Nakano *et al.*, 1995):

$$q_{i,t_1-t_2} = q_{0,t_1-t_2} - \sum_{m=1}^i (W_{m,t_2} - W_{m,t_1}) \quad (3)$$

Here, the effective rainfall or snowmelt amount, $P_{t_1-t_2}$ (mm), at the ground surface at between t_1 and t_2 was given to q_{0,t_1-t_2} as shown by equation (4):

$$q_{0,t_1-t_2} = P_{t_1-t_2} \quad (4)$$

The infiltration, q_{0,t_1-t_2} , at the ground surface of site A and site B was assumed to be equal to rainfall and/or snowmelt amount, although it is probably overestimated by the interception of rainwater and snow, especially in forest. The snowmelt amount at site A and site B was obtained by applying the data of air temperature and snow depth at site S to the degree-day method and by obtaining air temperature at the altitudes of the two sites from the lapse rate. Using the air temperature data at the meteorological stations in or around the Saromabetsu river basin, the lapse rate was obtained at $0.3^\circ\text{C}/100\text{m}$. Then, the degree-day factor was obtained by measuring the snow density *in situ* at forest and grassland and site S, and by comparing the snow depth records at site S with the positive degree days. As a result, the snow density was $380 \text{ kg}/\text{m}^3$ on average, and the degree-day factors were determined at $4.2 \text{ mm}/^\circ\text{C} \cdot \text{day}$ at forest and at $5.1 \text{ mm}/^\circ\text{C} \cdot \text{day}$ at grassland and site S. The snow-covered conditions at sites A and B were judged by knowing a temporal change of snow-covered area in the river basin. The snow cover change was acquired by the digital

elevation map (Fig. 2a), the lapse rate and the degree-day factor. The spatial snow cover distribution obtained was ascertained by the MODIS/Terra Snow Cover 8-day L3 Global 500m Grid, version 5 (URL: <http://nsidc.org/data/mod10a2v5.html>) (Hall *et al.*, 2006).

Calculated results

Fig. 7 shows time series of the hourly percolation, q_{4,t_1-t_2} (mm/h), at 38 cm depth of (a) site A (forest) and (b) site B (grassland), and (c) the rainfalls at site S in a period of the rainfall events of 2008. As a result, the percolation at grassland is smaller than that at forest except for the first large rainfall of 23 July, but occurs a longer time by the earlier

response to the rainfalls.

Fig. 8 shows time series of the hourly percolation at 38 cm depth of (a) site A and (b) site B, and (c) air temperature and rainfall at site S, and the rainfall plus the calculated snowmelt amount at sites A and B in a period of the snowmelt events of 2009. After 24 March, the percolation at forest was consistently larger than at grassland. This is probably because the meltwater infiltration occurred also on the forest slopes of the altitude higher than at site A. After 1500h, 4 April, the meltwater infiltration at sites A and B was depleted because the snow almost vanished there. Discharge of more than $10 \text{ m}^3/\text{s}$, beginning at 1500h, 3 April, continued until 1100h, 26 April, including the two peak river

Fig. 7. Time series of the hourly percolation at 38 cm depth of (a) site A and (b) site B, and (c) rainfalls at site S in the summer season of 2008.

Fig. 8. Time series of the hourly percolation at 38 cm depth of at (a) site A and (b) site B, and (c) air temperature and rainfall at site S and snowmelt amount plus rainfall at site A and site B in the snowmelt season of 2009.

discharges at 1900h, 10 April and at 2100h, 13 April to 0000h, 14 April (Fig. 6). Such large river runoffs appear to have been produced by the relatively large percolation at forest.

IV. Simulations of rainfall and snowmelt river runoffs

1. Preparation

The rainfall and snowmelt runoffs of the Saromabetsu River are simulated by the tank model

(Sugawara, 1972). The adoption of a lumped model such as the tank model is due to unsatisfactory results by many distributed models for snowmelt river runoffs (Hollander *et al.*, 2009). The tank model (or the linear reservoir model or the HVB-ETH model) is also applicable to simulations of glacier-melt runoffs (Braun and Renner, 1992; Hannah and Gurnell, 2001; Kido *et al.*, 2007) and to those of river runoffs including the rainfall and glacier-melt runoffs (Wada *et al.*, 2007). The three sequential tanks were used to separate the runoffs into three flow components,

surface runoff, intermediate runoff and baseflow (Fig. 9).

The actual evapotranspiration from the river basin upstream of site E as output from the top tank was calculated as follows: First, the potential evapotranspiration, PE (cm/day), upstream of site E was determined by the following Penman equation (Kayane, 1980).

$$PE = \frac{[8.64 \times 10^6 \cdot (Q^* - Q_G) / \rho \lambda] \Delta + \gamma E_a}{\Delta + \gamma} \quad (5)$$

where λ is the latent heat of vaporization in J/kg, ρ is the water density in kg/m³, $\Delta \equiv de_s/dT$ (e_s and T , the saturation vapor pressure in hPa and air temperature in °C at the evaporating surface, respectively), $\gamma \equiv c_p \cdot P / (0.622 \lambda)$, (c_p , the specific heat for constant pressure in J/kg·K; P , the air pressure in hPa), Q_G is the conductive heat in W/m² to the ground, Q^* is the net radiation in W/m² given by $Q^* = (1 - \alpha)K \downarrow + L^*$ (α , the albedo; $K \downarrow$, the downward shortwave radiation; L^* , the net longwave radiation), and E_a is the drying power of air, given by $E_a = (0.013 + 0.0016u_2)(e_{sa} - e_a)$ in cm/day (u_2 , the wind speed in km/day at 2 m above the ground surface; e_{sa} and e_a , the saturation vapor pressure and the vapor pressure in hPa at the air temperature, T_a , respectively) (Dunne and Leopold, 1978). The

coefficient of 8.64×10^6 in Equation (5) is used to change m/s to cm/day for numerical values of the bracket. Referring to Kayane (1980), the downward shortwave radiation in ly/min, $K \downarrow = (0.18 + 0.55 \cdot n / N) K \downarrow_p$ (n , the daylight time in hr, N , the potential daylight time in hr; $K \downarrow_p$, the potential shortwave radiation) and the net longwave radiation in ly/min, $L^* = -\sigma (T_a + 273.2)^4 \cdot (0.366 - 0.036\sqrt{e_a}) (0.1 + 0.9 \cdot n / N)$ (σ , the Stefan-Boltzman constant, 1.183×10^{-7} cal/cm²·day·K) were applied, and the albedo, $\alpha = 0.2$, was adopted.

The spatial distributions of air temperature corrected in elevation were then applied to the air temperature, T_a . The air pressure, P , was also corrected in elevation by using the air pressure record at site E and the Abashiri Meteorological Observatory (37.6 m a.s.l.) about 42 km east of site E. The other meteorological factors for the calculation of the potential evaporation were observed at site S and the Abashiri Meteorological Observatory. The actual evapotranspiration, AE , was numerically obtained by the short-term water budget method for periods before and after rainfall events. As a result, the evapotranspiration ratio, $\beta (=AE/PE)$, was estimated at 0.80. The total snowmelt amount in the river basin as input to the top tank in the snowmelt season was calculated by applying the spatial distributions of snow cover and air temperature to the degree-day method. Then, referring to Rango and Martinec (1995), the degree-day factors, $f = 4.2$ and 5.1 in mm/°C·day, were applied to calculate the snowmelt amount at site A and site B, respectively. The comparison with the MODIS/Terra Snow Cover indicated that the temporal change of the snow-covered area in the river basin was reasonably built up.

Parameters of the tank model are side outlet coefficients a_{ij} (for the j th side outlet of the i th tank), bottom outlet coefficients, b_i ($i=1, 2$), heights, h_{ij} , of side outlets and initial storages, h_i ($i=1, 2, 3$),

Fig. 9. Schematic diagram of the tank model applied in this study.

in tanks (Fig. 9). The discharge, q (mm/day), and percolation, p (mm/day), from the tanks can then be described by the following equations.

$$q_{ij}(n) = \begin{cases} a_{i,j} \cdot (h_i(n) - h_{i,j}) & \text{when } h_i(n) > h_{i,j} \quad (i=1, j=1, 2) \\ 0 & \text{when } h_i(n) \leq h_{i,j} \quad (i=1, j=1, 2) \end{cases} \quad (6)$$

$$q_i(n) = a_i \cdot h_i(n) \quad (i=2, 3) \quad (7)$$

$$p_i(n) = b_i \cdot h_i(n) \quad (i=1, 2) \quad (8)$$

$$h_i(n) = \begin{cases} h_i(n) - q_i(n) \cdot \Delta t - p_i(n) \cdot \Delta t + P(n) \cdot \Delta t - E(n) \cdot \Delta t & (i=1) \\ h_i(n) - q_i(n) \cdot \Delta t - p_i(n) \cdot \Delta t + p_{i-1}(n) \cdot \Delta t & (i=2, 3) \end{cases} \quad (9)$$

where n is the number of day in a simulated period, Δt is the time step (=1 day), a_i is the discharge rate (1/day), b_i is the percolation rate (1/day), $h_{i,j}$ and h_i are the water levels or storage (mm), p_i is the percolation (mm/day), $P(n)$ is the snowmelt amount and/or rainfall (mm/day), and E is the actual evapotranspiration (mm/day). The total runoff from the drainage basin is given by the sum of discharges from the three tanks as $Q(n) = \sum_{i=1}^3 q(i, n)$.

2. Simulated results

Fig. 10 shows (a) a comparison between simulated discharge and observed one at site E, (b) three flow components separated into surface runoff, intermediate runoff and baseflow, and (c) calculated actual evapotranspiration, and (d) diurnal rainfalls at site S in the summer season of 2008. The parameters applied to the simulation is shown in Table 2. As a result, the total actual evapotranspiration of 1 June – 7 November 2008 was 277 mm (Fig. 10c), corresponding to 74% of the total rainfall. The simulated result is very reasonable to the observed discharge with $r^2 = 0.934$, $RMSE = 0.902$ m³/s, $NASH = 0.953$ m³/s, where r is the correlation

coefficient, $RMSE$ is the root mean square error and $NASH$ is the Nash–Sutcliffe model efficiency coefficient (Nash and Sutcliffe, 1970). As a result, for the three large runoffs of 23 July – 2 August, 15 – 25 August and 27 August – 14 September, the surface runoff, the intermediate runoff and the baseflow occupied 51.1%, 46.5% and 54.4% (mean, 50.7%), 30.9%, 22.7% and 28.2% (mean, 27.3%), and 18.0%, 30.8% and 17.4% (mean, 22.0%), respectively (Fig. 10b). Thus, a quick flow such as the surface flow prevails in the runoff events. Actually, the surface runoff may correspond to the percolation in the surface soil layer, since the infiltration capacity at forest and farmland is much larger than the rainfalls (Table 1).

Fig. 11 shows a comparison between simulated and observed discharges and the three flow components in the snowmelt season of 2009. The simulation was also reasonable to the observed discharge with $r^2 = 0.973$, $RMSE = 0.961$ m³/s, $NASH = 0.974$ m³/s. Over the period of 6 March to 17 May, the surface runoff, the intermediate runoff and baseflow occupied 34.3%, 43.3% and 22.4%, respectively. Thus, the contribution of the intermediate runoff is relatively large, compared with that in the summer season. This results from relatively large a_2 and b_1 values in Table 2, which actually reflects the occurrence of the relatively deep percolation in the snowmelt season, as shown by the increase of soil moisture content at 38 cm depth (Fig. 6a and Fig. 6b).

3. Relations between simulated runoffs and percolation

In order to examine how the calculated percolation affects the river runoff, the simulated surface runoff and intermediate runoff were summed for each event in the summer season of 2008 and the snowmelt season of 2009, and then compared with the total percolation, which was calculated for each event period by using the time series of percolation

Fig. 10. (a) Comparison between simulated and observed discharges, (b) three flow components separated by the tank model, (c) calculated actual evapotranspiration, and (d) daily rainfalls in the summer season of 2008.

Table 2. Parameters applied to runoff simulations.

Parameters	2008 summer	2009 spring
h_{11}	23 mm	36 mm
h_{12}	7 mm	18 mm
a_{11}	0.00848	0
a_{12}	0.00161	0.00839
b_1	0.00484	0.00773
a_2	0.00368	0.00475
b_2	0.00383	0.00173
a_3	0.00105	0.00154

in Fig. 7 and Fig. 8. In the snowmelt season of 2009, the four runoff events of 6 - 10 March, 13 - 17 March, 19 - 28 March and 2 - 17 April were chosen, while, in the summer season of 2008, the three river runoffs of 23 - 29 July, 15 - 25 August and 27 August - 9 September was selected. The summed runoffs were divided by the basin area (259.3 km^2) to be expressed as the runoff height (mm).

Fig. 11. Comparison between simulated and observed discharges and separated three flow components in the snowmelt season of 2009.

Fig. 12. Relations between total percolation and summed runoff height for the runoff events in the summer season (black circles) and snowmelt season (white circles). (a) forest (b) grassland

Fig. 12 shows relations between the runoff height and the total percolation at 38 cm depth of (a) site A (forest) and (b) site B (grassland). There is definitely the linear relationship with $R^2 = 0.945$ ($P < 0.001$) or $R^2 = 0.881$ ($P < 0.002$) between the total percolation and the runoff height at both forest and grassland. At forest, the total percolation is about 2.5 times as large as the runoff heights. The trees, especially the needleleaf trees, tend to hide the ground over the year. Hence, too large the percolation is probably due to the overestimate of the infiltration, $q_{0,t-t_0}$, at the ground surface of site A, since the snow-covered area is relatively small in forest. At grassland (site B), the relationship is relatively close to the one-to-one correspondence. The infiltration given is thus probably reasonable, and the effect of the space between the access tube and the surround soil layer, if any, is judged to be low for the percolation amount. The characteristics of soil moisture variations would be common to grassland and field, as shown by the similar hydraulic conductivity in Table 1. The linear relationship thus suggests that the surface runoff and intermediate runoff extracted by the tank model are produced by the percolation and subsequent subsurface flow at depths of more than 38 cm. According to the runoff separation by $\delta^{18}\text{O}$ and δD (Yamaguchi, 2011), in the Saromabetsu river basin, the intermediate runoff may occur as a runoff from the perched groundwater zone at depths of less than 10 m. Hence, it is supposed that the modeled surface runoff and intermediate runoff corresponds to the shallow macropore flow and the relatively deep subsurface flow through the perched-groundwater zone, respectively.

V. Conclusions and future works

The soil moisture monitoring by the 4CH moisture profilers showed that, common to the

summer and snowmelt seasons, the unsaturated percolation could occur in the porous surface layers at 0 – 28 cm depth of forest and at 0 – 8 cm depth of grassland. The percolation at 38 cm depth of forest and grassland was calculated for the rainfall and snowmelt events. In many cases, the percolation was larger at forest than at grassland, especially during the relatively large runoffs. Meanwhile, the simulations by the tank model were carried out for river discharge in the summer and snowmelt seasons. The simulated results were agreeable to the observed ones with the high correlation. The total percolation at 38 cm depth of forest and grassland during the rainfall and snowmelt events were compared with runoff heights from the surface and intermediate runoffs separated by the simulations. There was definitely the linear relationship between the total percolation and runoff heights. This suggests that the surface and intermediate runoffs obtained by the runoff analysis occur actually as the percolation and subsequent subsurface flow below the soil surface layer.

In this study, it is supposed that the one-point measurements at forest and grassland represents the soil moisture conditions of the soil surface layer all over the river basin. In order to better understand the dynamics of soil water behaviors to produce river runoffs, the information on the distribution and its temporal variability of soil moisture in the B and C soil layers on the basin slope and in the riparian zone is needed. Also, in the snowmelt season, the soil moisture condition could change between the north-facing and the south-facing slopes, since a temporal change of snow cover or snow depth is different between the slopes. Thus, the monitoring at a few points at each of forest and farmland by considering the slope direction is necessary. In addition, the infiltration and snow-covered conditions on the forest floor should be observed in the future.

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この論文に対する「討論」を2012年8月31日まで受け付けます。

土壌水の挙動特性と河川流出への役割： 北海道・佐呂間別川流域

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要 旨

北海道・佐呂間別川（流域面積 259.3 km²）の流域斜面で、2008 年 6 月～2009 年 5 月の期間、土壌水分量をモニタリングし、森林と牧草地における降雨・融雪に伴う土壌水の挙動を調べた。同流域の土地被覆は、主に森林（占有率 75.2%）と農地（同 21.5%）で、測定点での土壌構造は、森林・牧草地共に、透水性のよい有機層（A 層、平均厚さ 0.3m 未満）と透水性の悪い有機/無機層（B 層、平均厚さ約 10m 以上）からなる。用いた土壌水分計は 4 チャンネルのプロファイル土壌水分計で、8 cm, 18 cm, 28 cm, 38cm 深度で 20 分間隔で体積含水率（cm³/cm³）が求められた。結果として、森林斜面では 8-28 cm 深でイベント後の 2～3 日間は貯水状態にあり、その後は速やかに排水され、イベント前の土壌水分レベルに戻ることがわかった。一方、牧草地斜面では、8cm 深でイベント後の排水は速やかだが、さらに深い層では排水はほとんど行われなかった。また、20 mm/h 以上の降雨では、森林斜面の 38cm 深と牧草地斜面の 8 cm 以深で飽和状態となった。以上から、イベント時は透水性のよい層の下で浸透流がさかんであることがわかった。ここでは、得られた土壌水分データを用いて降雨と融雪の各イベントごとの 38 cm 深での降下浸透量を求めた。他方、タンクモデルによって佐呂間別川に対する流出解析を実施し、結果として、降雨・融雪イベント時に応答の早い直接流出（表面流出+中間流出）が 70% 以上を占めることがわかった。この直接流出量と先に求めた降下浸透量を比較すると、全体として明瞭な直線関係が認められた。このことより、流出解析で得られる直接流出は、実際には土壌表層下の浸透流として発生することが示唆された。

キーワード：土壌水分，降雨流出，融雪流出，浸透流，タンクモデル，表面流出，中間流出

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