

Characteristics of Snowmelt and Runoff in a Mountain Basin in Tohoku District, Japan

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Abstract: Winter snow cover occurs widely in Japan at comparatively low elevation and latitude. The water stored as snow is very important for irrigation of the rice crop during its early growth. We investigated the snowpack and snowmelt quantitatively on a daily and hourly basis during the snowmelt seasons from 2001 to 2008 in the Akazawa basin, Tohoku district. We used a snowmelt model to simulate the flux of snowmelt and rainwater in the snowpack. Among the years, total snowmelt varied between 190 to 540 mm from the beginning of Mar. to the late of Apr. We characterized the snowmelt runoff by making an index from the parameters of a storage routing model. We used the optimized parameters of the storage routing model to examine the output differences between the case in which the snowmelt flux at the snow surface was input, and the case in which the snowmelt flux from the base of the snowpack was input. The results clearly indicated that a snowpack influences the characteristics of snowmelt runoff.

Keywords: Snowpack Outflow, Snowmelt Runoff, SNTHERM, Storage Routing Model

1. Introduction

In high-latitude or alpine regions, many studies have investigated various hydrological processes related to snow, such as the snowmelt heat budget and its infiltration through the snowpack, and the spatial distribution of snow cover. Some examples are given as follows.

Williams and Tarboton [1] developed a snowmelt model that they verified with data from Smithfield Dry Canyon (United States), an alpine site above 3000 m. *Colbeck* [2] investigated the percolation of water in the snowpack using observations of the Seward glacier firn in the St. Elias mountains on the Alaska–Canada border. *Erickson, Williams and Winstral* [3] estimated the spatial distribution of snow cover in the Green Lakes Valley watershed, Colorado, at 3500 to 4000 m of elevation. *Lundquist and Dettinger* [4] studied the relationship between the heterogeneity of snow coverage, the scale of a basin, and the timing of runoff in a basin at comparatively low latitude (about 37°30') but at elevations of 2000 to 2700 m in Sequoia National Park, California.

Winter snow cover is found widely and the flow in many streams at relatively low elevation are seasonally dominated by snowmelt in Japan, despite its low latitude of temperate zone location. Even though snow affects many human

activities in Japan, observations of snowmelt and runoff from snowmelt are not widely reported. Moreover, no organization in Japan collects detailed snow cover information as the National Weather Service (NWS) now does in the United States, even though having information on the present snow cover in Japan, before worldwide shortage of water occurs, is important [5, 6]. For this study, we selected the Akazawa River basin in the Iwate University forest in the Tohoku district. Meteorological elements were observed at the Shimotakizawa meteorological observation station, and stream discharge was measured at the lower end of the basin.

First, we applied an energy balance model to the snowmelt period over 8 years and showed that in general it could reproduce snow depth. Second, since the snowmelt period in this area coincides with the rainy season, we separated the water arriving at the ground surface according to its origin (snowmelt or rain), to clarify differences in the timing and quantity of snowmelt, and runoff from snowmelt among the years.

The feature of above-mentioned researches individually treats the snowmelt, infiltration, and the river runoff, and little has been reported the research that integrates these processes. From this background, third, using the daily variation in snowmelt (excluding the contribution of rainfall) during the 8

years, we characterized the snowmelt runoff by optimizing the parameters of a storage routing model, a type of lumped runoff model.

2. Materials and Methods

In the recent snow study, tendency to the lateral observation are remarkable. For example, the distribution of large area of SWE has been presumed by using remote sensing technique [e.g.7]. The forecast of the snowfall was tried with mobile polarimetric X-band radar [8]. This study belongs to the class that aim at basic findings and information intended for experimental basin.

2.1. Site Description

This research focused on the melt period during the 8 years from 2001 to 2008 in the Akazawa basin (Fig. 1), Iwate University Omyojin forest. The Akazawa River (Shizukuishi, Iwate Prefecture) is a branch of the Shizukuishi River, and the Akazawa basin is on the east side of the Ou Mountain Range (the longest in Japan at about 450 km), which traverses the Tohoku district from north to south. The area of the basin is 8.8 km², and its elevation range is from about 230 to 670 m. The meteorological observation station at Shimotakizawa is at 310 m elevation, 1.5 km southwest of the stream's outlet.

Most of the basin (97%) is forested, and the remaining area is occupied by a forest road, grassy areas, and a lumberyard. This cool-temperate forest comprises both natural forest (72%) and artificial plantations of mainly Japan cedar or

Japanese red pine. The bedrock in the basin consists of Neogene tuffs and shale, with some rhyolite and dacite. The dominant soil (93%) in the basin is Brown Forest soil, and podzolic soils cover the remaining area.

At the Shizukuishi Automated Meteorological Data Acquisition System (AMeDAS) observatory of the Japan Meteorological Agency, which is about 4 km east of the basin, the average annual precipitation was 1600 mm, and the annual mean air temperature was 9.6°C.

2.2. Data

Because the data were collected in a university forest, where traffic is restricted, there were few artificial influences on the observations. The meteorological measurements were carried out at Shimotakizawa. The observed parameters, the sensors used, and their accuracies are summarized in Table 1. Air temperature, relative humidity and wind velocity were observed 2 m above the ground. From 2001 through 2005, downward and upward shortwave radiation were measured with the MR-22 sensor and downward longwave radiation with the MS-200 sensor (Eko Instruments, Tokyo, Japan). Beginning in 2006, the EKO MR-50 sensor was used to measure net radiation of the four radiation components. Precipitation was measured with a tipping bucket rain gauge (Ikeda, Keiki, Japan) with a precision of 0.5 mm equipped with an overflow container filled with antifreeze (ethylene glycol). Snow depth was measured on a calibrated staff inserted in the snowpack at 9: 00 LT every day.

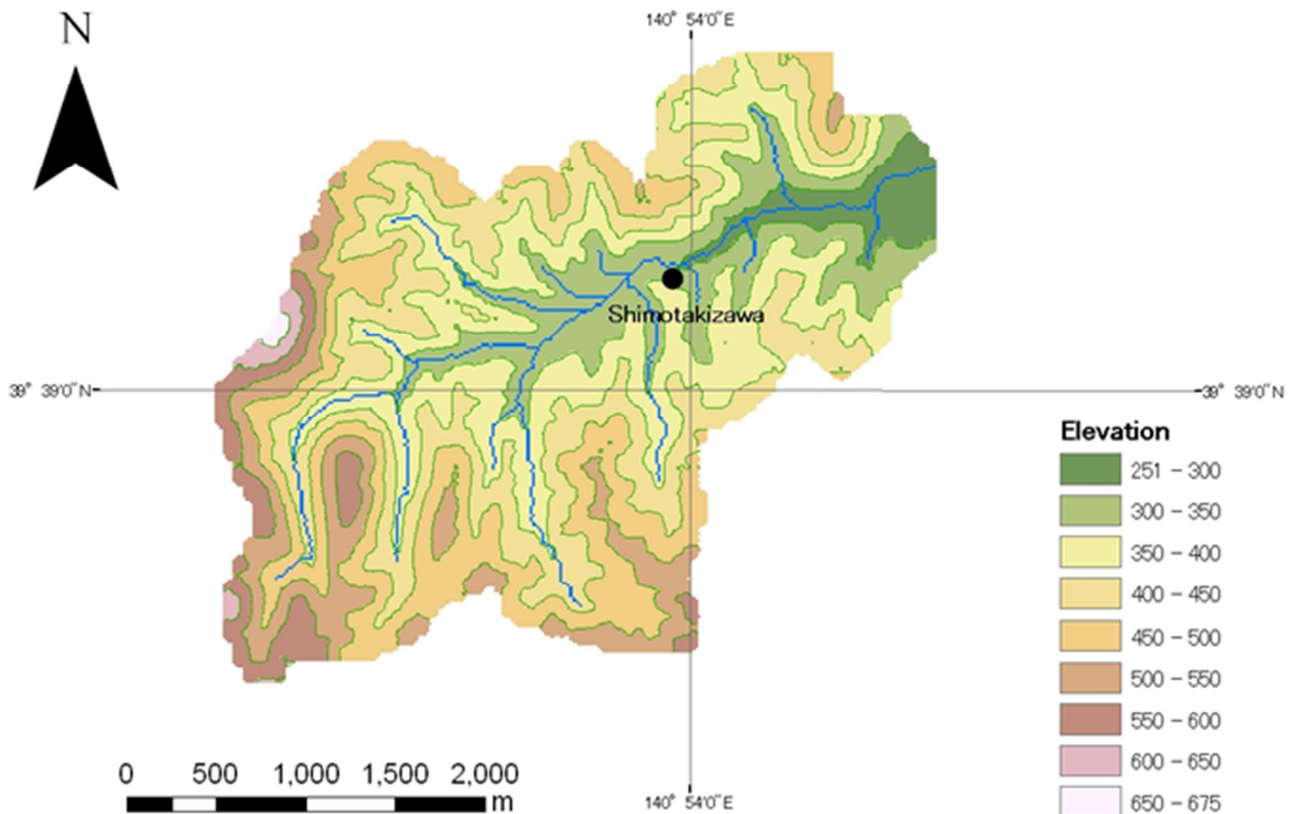


Figure 1. The Akazawa basin and the location of the Shimotakizawa meteorological observation station.

Table 1. Instruments used for observations to obtain model input and verification data.

Input or verification parameter	Mfr. and model (accuracy)	
	2003–2005	2006–2008
Downward and upward shortwave radiation	EKO MR-22 (3%)	—
Downward longwave radiation	EKO MS-200 (5%)	—
Net radiation (of 4 components)	EKO MR-50 (nonlinearity 3%)	—
Air temperature Relative humidity	Vaisala CS500 (0.4 K 3%)	—
Wind speed	Yokogawa A-702 (cup type) (5%)	—
Precipitation	IKEDA RT-5 (3%)	—
Water level in the stream	KONA KADEC-MIZU (1 mm)	—



Figure 2. The lysimeter collector; of dimensions 1 m × 1 m.

+The snowpack was excavated and a plastic barrier was installed around the collector just before the melt season.

Lysimeters can be used to observe water generation from snowmelt. We used a lysimeter with a 1 m × 1 m collector. We enclosed the snow above the lysimeter collector to prevent lateral flow as follows. In late February each year, just before the melt season, we excavated the snowpack surrounding the lysimeter, wrapped plastic sheeting around the snowpack on the lysimeter collector, and then backfilled the excavated area with the removed snow (Fig. 2). Then, as the snowpack melted, we cut off plastic sheeting protruding above the snow surface. Meltwater and rainwater that passed through the snowpack to the lysimeter collector were conducted to a tipping bucket with a capacity of 200 cm³ and measured with a precision of 0.2 mm. However, such an attempt did not always succeed, it was judged that the reliability of data was low in some application years. Therefore, this data was not used except the small amount of adjustment of an initial value of SWE.

A timber bridge was built across the stream at the lower end of the basin for observing streamflow. The water level, measured with a pressure-type water-level gauge at the bridge, was used to calculate the flow volume.

2.3. Application of SNTHERM

In this research, we studied the hourly generation of meltwater and its movement through the snowpack. For

example, the lysimeter measures only the quantity of water released at the base of the snowpack, water originating as rain cannot be distinguished from snowmelt. SNTHERM [9] is detailed one-dimensional model-which handles snow-cover as multilayer. The model can reproduce processes of accumulation, compaction, generation of the liquid water, water flow etc. that occur within a snowpack. The model generally require air temperature, relative humidity, wind velocity, downward and upward shortwave radiation, and downward longwave radiation. The model has been applied to snowpacks in various regions and environments. The utility of SNTHERM was shown by other models' comparisons [10], and it was shown that the model was useful at mid-Scandinavian winters conditions [11]. SNTHERM was applied at Greenland ice sheet in summer, and simulated SWE and snow height agreed with actual measured values [12]. Furthermore, the model was used for an effective snow melting promotion in the ski area [13].

We therefore applied the SNTHERM model, making certain following assumptions and selecting parameters by referring to past studies conducted near our study site [14, 15].

The energy equation I_{top} for the top layer of snow is given by

$$I_{top} = I_s \downarrow (1 - \alpha_{top}) + I_{ir} \downarrow - I_{ir} \uparrow + I_{sen} + I_{lat} + I_{conv} \quad (1)$$

where I_s (Wm⁻²) is downward shortwave radiation, α_{top} is albedo of the snow surface, $I_{ir} \downarrow$ (Wm⁻²) is downward longwave radiation, $I_{ir} \uparrow$ (Wm⁻²) is upward longwave radiation, I_{sen} (Wm⁻²) is sensible heat flux, and I_{lat} (Wm⁻²) is latent heat flux. Then, turbulent heat fluxes I_{sen} and I_{lat} can then be expressed as follows:

$$I_{sen} = c_{air} \rho_{air} C_H u (T_{air} - T^n)$$

$$I_{lat} = L_{vi} \rho_{air} C_E u (q_{air} - q^n)$$

where ρ_{air} (kg m⁻³) is air density, c_{air} (J kg⁻¹ K⁻¹) is specific heat of air at constant pressure, u (m s⁻¹) is wind velocity, L_{vi} (J kg⁻¹) is latent heat of sublimation for ice, T_{air} (K) is air temperature, T^n (K) is the temperature of snow surface layer n (snow and soil layers are numbered from bottom to top), q_{air} is the specific humidity of air, and q^n is saturation specific humidity at T^n . Though bulk transfer coefficients depend on the atmospheric stability and the distance of the sensors (for air temperature, relative humidity, wind speed) from the snow surface, the bulk transfer coefficients of sensible heat C_H and latent heat C_E are each fixed at 0.002.

Snow density increases over time as a result of metamorphism and viscous compaction, but because viscous compaction is considered the dominant process during the snowmelt period, the influence of metamorphism was disregarded. The snowfall density was assumed to be 80 kg m⁻³ and the liquid water content

of the snowfall was assumed to be 0.

Because bottom snowmelt commonly occurs in almost all of Japan, at this study site, the temperature at the bottom of the snowpack and the soil layer in contact with the snowpack were assumed to be 273.15 K, reflecting the actual condition. The snowpack was divided according to snow depth, and the divided snow layers were set to 5 cm each. The density of all layers was set to the average value as measured by a tubular-type snow sampler (5.0 cm in diameter). The snow was assumed to be isothermal, and the temperature of all snow layers was set to 273.15 K.

Snowfall occurs frequently in the Akazawa basin even when the air temperature is above 0°C. Therefore, for the purpose of this research, precipitation was judged to be snow when the snow depth increased or when no water reached the lysimeter collector, regardless of the air temperature.

2.4. Extraction of the Snowmelt Runoff Characteristics with a Runoff Model

Meltwater generated near the snow surface contributes to runoff through various processes. *Colbeck* studied the gravitational flow of water through the snowpack as a snowmelt runoff process [16-18] and investigated infiltration of snowmelt through a snowpack on a slope, and through a heterogeneous snowpack with intermingled ice layers [19, 20]. Snowmelt runoff on slopes has also been modeled by varying parameters such as slope angle, slope length, and vegetative cover [21, 22]. *Akan* [23] modeled meltwater infiltration on a slope of similar scale by taking into account the snow cover heat budget.

Basin level studies have examined how snowpack thickness affects the movement and storage of snowmelt in the snowpack. *Woo and Slaymaker* [24] showed that a remarkable time lag existed between the melting of snow and increased stream flow from snowmelt runoff. *Braun and Slaymaker* [25] examined the daily variation in snowmelt runoff at different spatial scales in basins of less than 23 km². *Lundquist and Dettinger* [4] showed that the characteristics of snowmelt runoff depend on the basin size (<30 km², 30–200 km², and >200 km²).

In this study, we characterized snowmelt runoff by using a lumped runoff model. We applied a storage routing model [e.g.26], which has mainly been used for flood runoff analysis in Japan:

$$\frac{dQ}{dt} = \frac{1}{Kp} [f r(t - T_l) - Q]$$

where t (h) is the time, r (mm h⁻¹) is the inflow rate into the basin (rain or snowmelt), Q (mm³) is direct runoff, T_l (h) is the time lag between rain or snowmelt and runoff i.e. duration of storage, f is the runoff coefficient, and K and P are model constants.

Numerical solutions for Q can be calculated by the Runge-Kutta method. By determining the values of parameters K , P , T_l , and f the runoff can be

characterized. We used an iterative gradient descent method [e.g.27] for finding a local minimum of a function of several variables J ;

$$J = \frac{1}{n} \sum_{i=1}^n \frac{(Q_{oi} - Q_{ci})^2}{Q_{oi}}$$

where n is total number of target data and Q_{oi} and Q_{ci} are the observed and calculated runoff amounts during the i -th hour, respectively.

3. Results and Discussion

3.1. Calculated Daily Snowdepth over the 8 Years

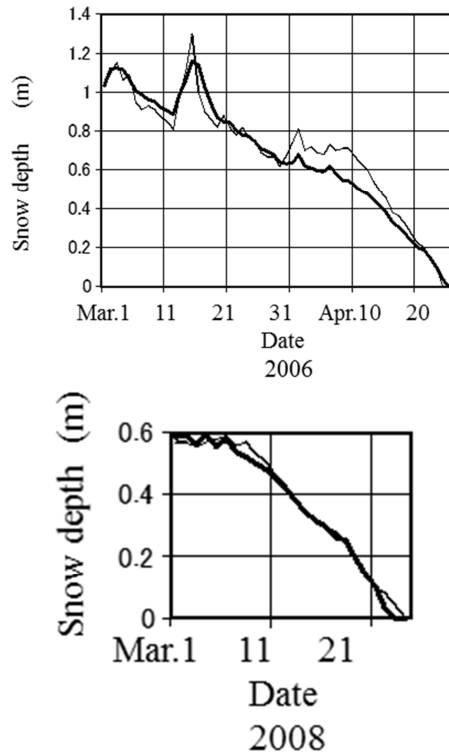
We applied SNTHERM to data collected from 1 March of each year to the day that the snow cover disappeared in the model. The initial snow depth, the snow water equivalent (SWE) of the snowpack, and snow density are in Table 2. In most years, the snowpack SWE was observed several days before 1 March. Therefore, we added to the observed SWE the amount of precipitation that fell, and subtracted from it the outflow from the snowpack measured by the lysimeter, during the days before 1 March to obtain the initial SWE value. The computational time interval of the model was 1 h. The computed snow depth at 9:00 LT was compared with the observed snow depth. Although the ending date of continuous snow cover varied from the end of March (2008) to the end of April (2006) during the study (Fig. 3), it had no particular trend. We calculated the relative errors of the modeled snow depth for each year (Table 2). We judge the results of this application of SNTHERM to be mainly satisfactory.

If we assume that snowmelt and rainfall contribute to the moisture in the snowpack to the same extent that they likewise contribute to the outflow of water from the snowpack, then it is possible to distinguish the origin of snowpack outflow as either rain or snowmelt (Fig. 4a, b). Because in Japan, the rainy season and the snow melt period overlap, more rainfall is observed during snow melt season in Japan than where rain seldom falls during the snow melt period [4]. We also determined the central day of the snow melting period by multiplying the number of days from 1 March by the daily snowpack outflow originating from snowmelt until the snow cover was gone, and then dividing that product by the integrated value of the snowpack outflow (Fig.4c, arrow). The central day ranged from the middle of March (in 2008) to early April (in 2005) during the study period. The cumulative outflow from snowmelt was greater than 500 mm during two of the eight years (2005 and 2006), but was less than 200 mm in 2008 (Fig. 4b). The average value over the eight years was 370 mm, which is equivalent to 23% of the annual precipitation of this area. Though the occurrence of rain obscured the pattern, the runoff rate leveled off at around 0.8 about 10 days after the snowpack outflow ended (Fig. 4c). The runoff rate indicated that in this basin, the measurements made at Shimotakizawa were fairly representative.

Table 2. Initial snow depth, snow water equivalent (SWE) in the snowpack, and snow density for each year.

Year	2001	2002	2003	2004	2005	2006	2007	2008
Snow depth (cm)	75	61	80	57	130	103	48	57
SWE (mm)	280	230	230	228	410	420	180	180
Snow density (kg/m ³)	380	390	280	400	310	410	370	310

+Snow density is the mean density of all snow layers measured with a tubular-type sampler.

**Figure 3.** Observed (thin line) and computed (thick line) snow depth for selected year.**Table 3.** Mean relative errors of calculated snow depth at 9: 00 LT.

Year	
2001	0.33
2002	0.12
2003	0.06
2004	0.11
2005	0.11
2006	0.01
2007	0.14
2008	0.03

Table 4. Optimized parameters f , K , p , and T_l in equation (4) for r set to the flux near the surface of the snowpack.

Year	Date	Total snowmelt flux (mm)	f	K	p	T_l	$J \times 10^{-3}$	Calculated snow depth (m)
2001	8-9 Apr.	25.4	0.311	16.1	0.727	2.04	0.7	0.31
2001	9-10 Apr.	19.4	0.148	12.9	1.11	1.72	0.8	0.26
2003	25-26 Mar.	14.2	0.595	16.0	0.708	4.57	0.3	0.72
2003	7-8 Apr.	24.2	0.448	14.1	0.784	2.66	2.0	0.31
2004	28-29 Mar.	26.9	0.227	11.8	0.668	1.27	0.4	0.21
2004	29-30 Mar.	22.2	0.432	13.7	0.752	3.28	1.0	0.15
2005	23-24 Mar.	13.3	0.637	14.7	0.801	4.53	2.0	1.0
2007	8-9 Apr.	17.8	0.695	19.8	0.736	1.62	0.4	0.21
2008	11-12 Mar.	9.6	0.475	10.7	1.00	3.38	0.1	0.45
Average			0.441	14.4	0.810	2.79	0.9	

3.2. Characterization of Stream Runoff from Snowmelt

To examine the characteristics of the hourly snowmelt runoff, we selected data that met two requirements: first, no precipitation included in the streamflow, and second, the daily variation resulting from snowmelt could be identified in the streamflow. Nine cases met these requirements. We used a runoff model as a tool to examine the infiltration of snowmelt in the snowpack. The existence of a snowpack affects the parameter values used in the runoff model. In equation (4), r (inflow rate) can be set to one of two values: the melt flux near the surface of the snowpack (surface case), or the melt flux at the base of the snowpack (bottom case). In the surface case, r is the flux, that is, at about 5 cm below the surface, and in the bottom case, r is the flux (i.e., snowpack outflow) from the base of the snowpack.

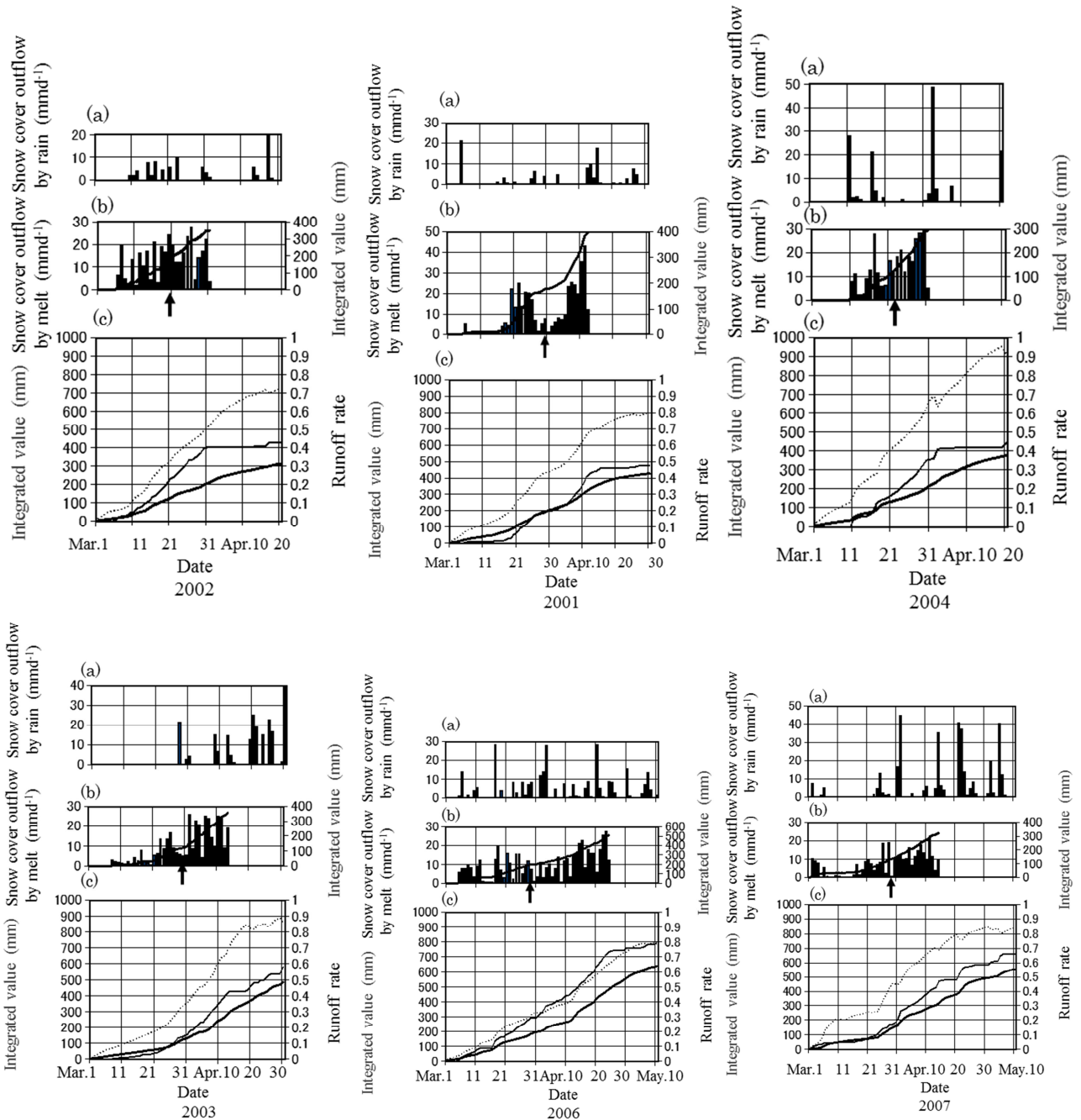
We optimized the model parameters (Tables 4 and 5) for each r of the nine selected cases. The parameters were optimized for the runoff from the time when a snowmelt flux was first simulated to the time when the hydrograph began to rise after the runoff peak.

K is considered to depend on the slope, roughness, soil permeability, etc., of the basin, and we surmise that the response of runoff to storage becomes blunter as K becomes large.

K of the surface case was larger than K of the bottom case, on average. The difference in the time lag T_l between the two r values clearly shows the effect of snow cover, especially when the snow depth is large. In Fig. 5, we show two examples among the nine selected cases when snow depth was comparatively large. We compared the two kinds of presumed snow melt flux and snowmelt runoff with the actual runoff measurements. Reproducibility was almost excellent, and was able to suggest the influence that the snowpack exerted on the outflow.

Table 5. Optimized parameters f , K , p , and T_i in equation (4) for r set to the flux at the base of the snowpack.

Year	Date	Total snowmelt flux (mm)	f	K	p	T_i	$J \times 10^{-3}$
2001	8-9 Apr.	23.9	0.302	15.1	0.737	1.26	0.7
2001	9-10 Apr.	19.3	0.215	11.4	0.871	0.07	0.6
2003	25-26 Mar.	13.1	0.332	11.7	0.828	1.76	0.5
2003	7-8 Apr.	23.8	0.378	12.1	0.821	2.12	3.0
2004	28-29 Mar.	25.9	0.223	11.9	0.682	0.99	0.4
2004	29-30 Mar.	28.2	0.334	12.2	0.735	2.33	0.3
2005	23-24 Mar.	15.6	0.296	9.64	0.917	3.06	0.4
2007	8-9 Apr.	18.0	0.501	15.6	0.722	1.89	0.3
2008	11-12 Mar.	10.0	0.394	10.0	1.08	1.92	1.0
Average			0.331	12.2	0.821	1.71	0.8



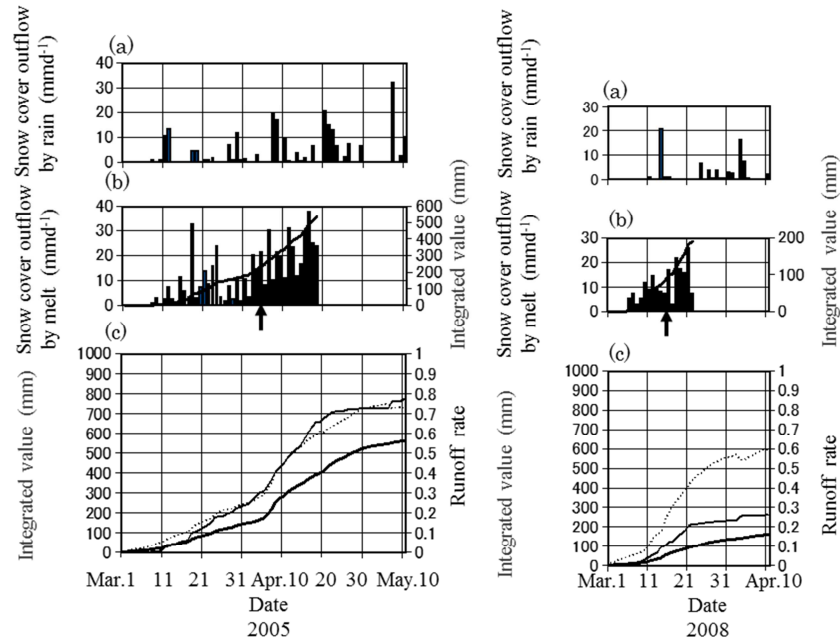


Figure 4. Daily amounts of snowpack outflow separated according to origin, and cumulative snowpack outflow, discharge, and runoff rates.

+(a) Simulated amount of daily snowpack outflow originating as rain. (b) Simulated amount of daily snowpack outflow originating as snowmelt, its cumulative value, and the central snow melt day, reckoned from March 1 (bold arrow). (c) Cumulative value from March 1 of observed stream discharge (thick line), snowpack outflow (thin line), and the runoff rate (dotted line) (i.e., the ratio of cumulative discharge to cumulative input). However, the initial value of the observed precipitation is SWE on 1 March.

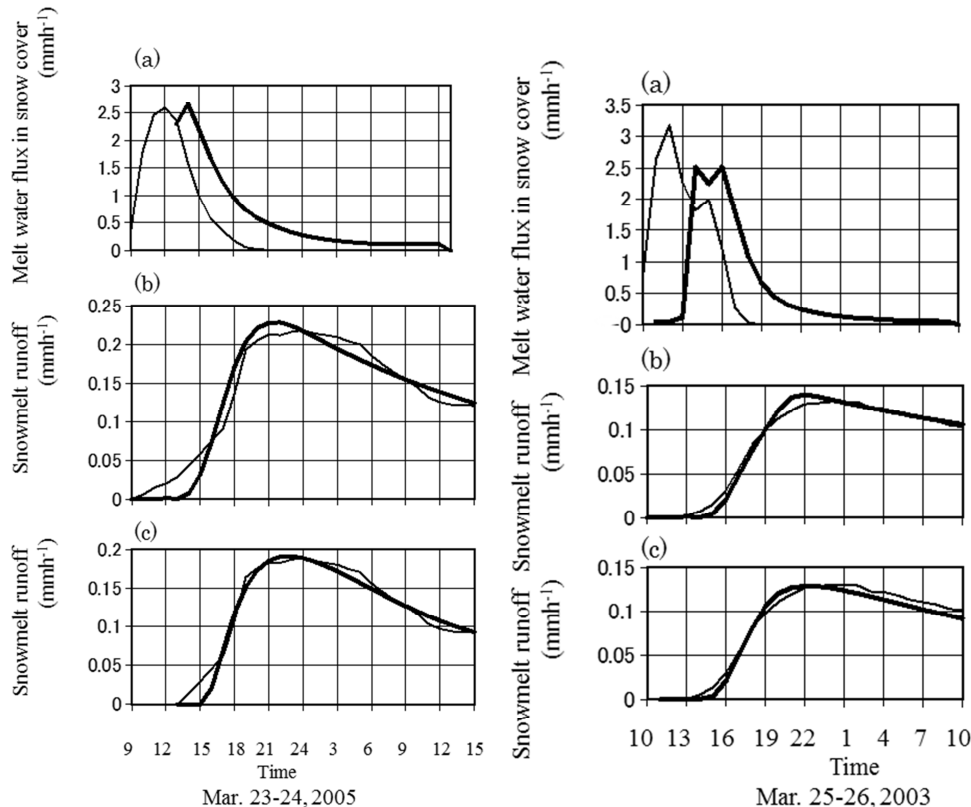


Figure 5. Hourly changes in the snowmelt flux within the snowpack and the snowmelt runoff.

+(a) Snowmelt flux near the snowpack surface (thin line) and at the base of the snowpack (i.e., snowpack outflow) (thick line). (b) Estimated snowmelt runoff (thick line) for r set to the snowmelt flux near the snow surface and observed snowmelt runoff (thin line). (c) Estimated snowmelt runoff (thick line) for r set to the snowpack outflow and the observed snowmelt runoff (thin line).

4. Conclusions

The aim of this research was to gather information about the snowmelt in recent years in a snow cover zone at comparatively low latitude and low elevation in Japan. We collected meteorological and streamflow data during the melt period over the eight years from 2001 to 2008 in Akazawa basin in an Iwate University forest. We then characterized the snowmelt process in this basin on both a daily and hourly basis.

We applied the SNTHERM model, which we modified to take into account actual conditions at the study site, and evaluated the results by comparing the estimated snow depth with the observed data.

Although the results of evaluation varied among the years, we obtained generally good results. Since in Japan the snowmelt period overlaps with the rainy season, we used SNTHERM to divide the snowpack outflow into outflow that originated in snowmelt and outflow that originated as rain, and then determined the distribution of the amount of snowmelt over time, and the central day of the melt period in each year. The average snowmelt during the 8 years was equivalent to 23% of the annual precipitation of this area. We observed no consistent trend in time, although the end of the snow cover season ranged from the end of March to the beginning of April.

We applied SNTHERM and a storage routing model to examine snowmelt and stream runoff on an hourly basis to characterize the snowmelt runoff. The snowmelt flux near the snowpack surface and at the base of the snowpack computed by SNTHERM were input to the storage routing model, and the parameters of the storage routing model were optimized for each value. In the case where the snowmelt flux near the snowpack surface was input, the optimized values of K and T_i (parameters of the storage routing model) were larger than when the flux at the base of the snowpack (i.e., outflow from the snowpack) was input. This clearly indicated that a snowpack influences the characteristics of snowmelt runoff.

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