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An approach to annual water balance for small mountainous catchments with wide spatial distributions of rainfall and snow water equivalent

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Abstract

One of the major problems in understanding the hydrological cycle in high mountainous regions with much snow is evaluating the spatial and temporal distribution of precipitation. This study evaluates the water balance in two small neighbouring catchments, namely Honryu (HN) and Shozawa (SH) in the Takaragawa Forest Watershed Experiment Station, Japan, by analysing records of precipitation in rainfall seasons and snow water equivalent (SWE) in snowfall seasons. Because both records were satisfactorily correlated with the filtered elevation, the distributed values of precipitation and SWE for each grid point of the digital elevation map were evaluated based on this correlation in each of the seasons. Total precipitation during the snowfall seasons was estimated from the SWE records using a relation of the mean catchment SWE to the cumulative precipitation monitored continuously at the observation station. An annual water balance in the two catchments averaged over several years was calculated from the total mean catchment precipitation in rainfall and snowfall seasons. The catchment evapotranspiration derived from the water balance was compared with Hamon's potential evapotranspiration. Although a large difference between the two catchments in evapotranspiration was calculated from the water balance, several probable causes could be suggested for this difference, and it is concluded that the annual water balance could be estimated with an acceptable accuracy.

1. Introduction

Many hydrological findings have recently been obtained from field studies carried out in small catchments (e.g. Anderson and Burt, 1990). However, these studies were typically conducted on hillsides where precipitation is considered to be uniform. Hydrologists can focus their interests on detailed hydrological processes in such a small experimental

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catchment as if it were a kind of 'natural' lysimeter applied for estimating evaporation from a small forested area (Calder, 1976), because their studies are based on the verification of the intrinsic water balance. In contrast, knowledge of hydrological processes is limited in catchments in high mountainous regions. Although suitable hydrographs from these larger catchments can be modelled using sophisticated approaches (Abbot et al., 1986; Wigmosta et al., 1994), significant problems still exist. A major difficulty involves the estimation of spatially distributed precipitation. Even an annual water balance usually cannot be determined, although it is an important verification for both precipitation and runoff records. Thus, we believe that the water balance of mountainous catchments should be assessed before detailed hydrological processes are studied.

It is far more difficult to quantify an annual water balance in a mountainous catchment with an abundant snow cover because the distribution of snowfall is wider and more variable than that of rainfall. Many studies have estimated the distributions of snow water equivalent (SWE) based on snow survey data (Kattlemann et al., 1985; Koike et al., 1986; Elder and Dozier, 1990; Ohta, 1994). However, most of them did not achieve an evaluation of annual water balance.

Compared with fluctuations of precipitation during a short-duration storm, however, the distribution of a long-term precipitation tends to have a clear pattern, owing to time averaging (Foster, 1949), and can be estimated from the records at a relatively small number of observation points. It seems practical to estimate annual precipitation averaged over several years and to check an averaged annual water balance to achieve an initial understanding of the characteristics of water balance of a mountainous catchment.

The spatial distribution of a long-term precipitation has been analysed based on the correlation with elevation (Sugawara et al., 1984; Koike et al., 1986; Ohta, 1994), although this correlation was sometimes found to be poor (Blumer and Lang, 1993). Tani (1994) estimated the spatial distributions of rainfall and SWE in small catchments in a mountainous region of Japan by analysing the correlation of their point observation records to elevation. In this analysis, he used a spatial filtering procedure for a digital map to improve the correlation. However, he produced the distribution of rainfall estimated from only one summer observation and that of SWE from a snow survey at only one time. Based on these results, this paper attempts to estimate the distribution of annual precipitation averaged over several years and an approach to annual water balances in those catchments.

2. Study site

Data for this study were collected in small catchments of the Takaragawa Forest Watershed Experiment Station of Forestry and Forest Products Research Institute (FFPRI) (Fig. 1). The site is located in a mountainous region, at the headwaters of the Tone River, which flows to the Tokyo metropolitan area. The study area consists of two neighboring catchments, 'Honryu' (HN) of 19.06 km^2, and 'Shozawa' (SH) of 1.18 km^2. The elevation of HN ranges from 780 to 1945 m above sea-level (a.s.l.) (Mt. Asahi), with an eastern aspect, whereas the elevation of SH ranges from 800 to 1370 m with a southern aspect. The geology mainly comprises granite rocks and a Tertiary layer called the Mikasa

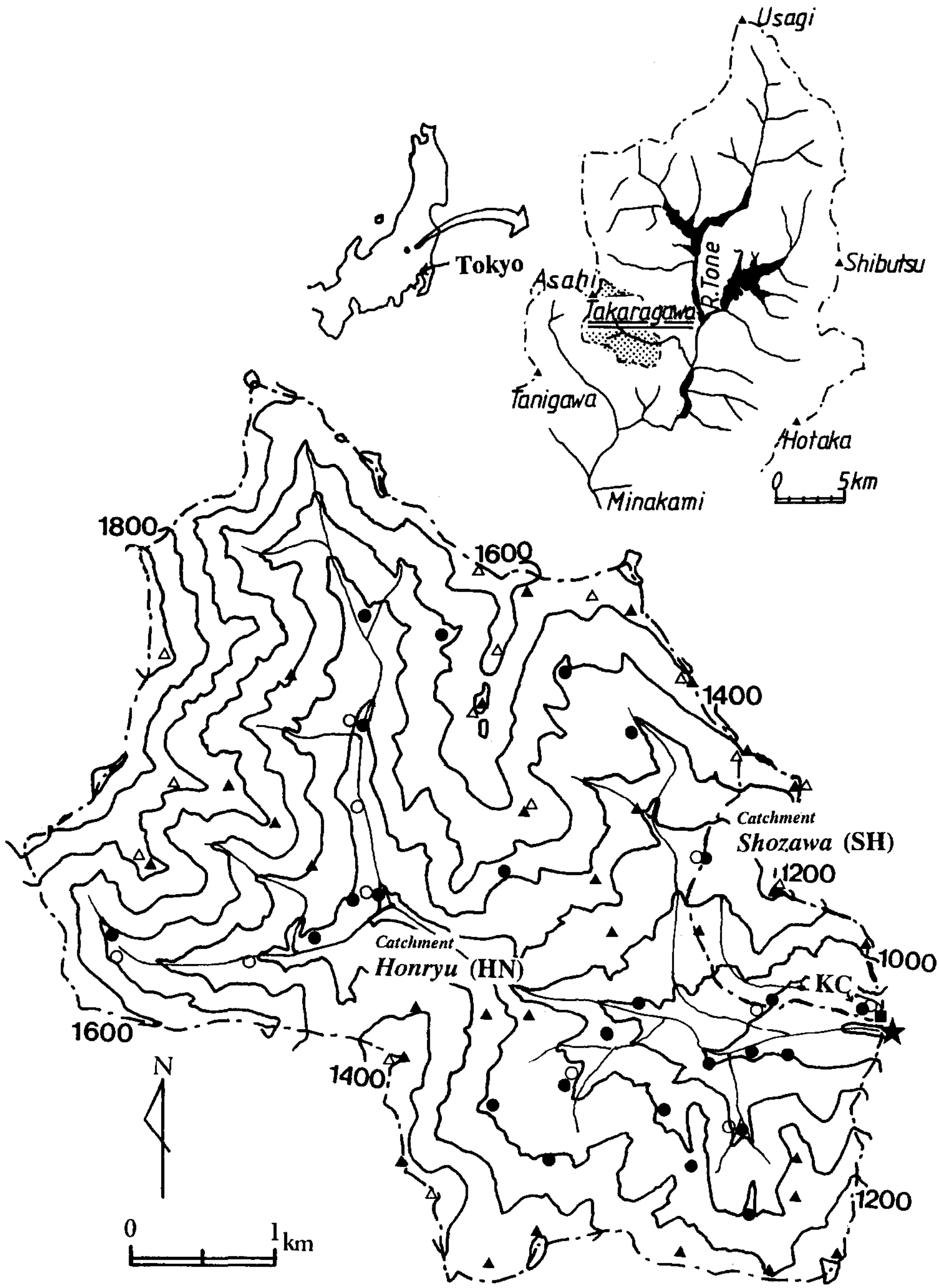

Fig. 1. Observation design at the Takaragawa Forest Watershed Experiment Station. ○, Rain gauge near a stream valley; △, rain gauge near a ridge line; ●, snow survey near a stream valley; ▲, snow survey near a ridge line; KC, observation field 'Kichi'; ★, stream gauge of HN; ■, stream gauge of SH.

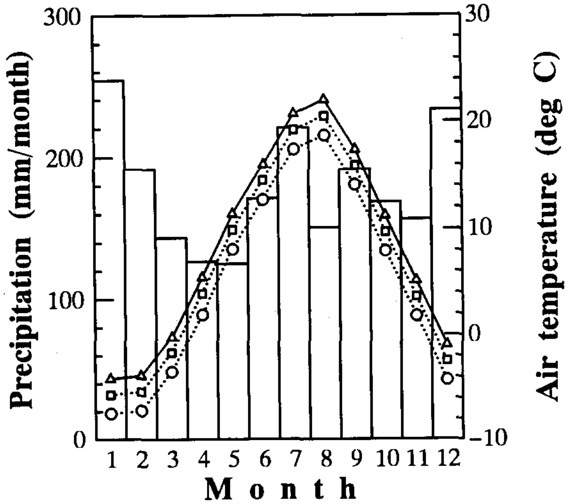

Fig. 2. Monthly values of precipitation and air temperature. △: Observed at KC; ○, estimated for the mean elevation of HN catchment; □: estimated for the mean elevation of SH catchment.

Layer. Deep seepage which may sometimes influence the water balance is regarded as negligibly small, as rocks derived from the geology cover steep slopes all over the mountainous catchments.

Most of the study area, including HN and SH catchments, was covered with natural forest of cool temperate deciduous trees. The main species were *Fagus crenata*, *Quercus serrata* and *Thujopsis dolabrata*. The upper 31% of the HN catchment was located above the treeline (1500 m a.s.l.) (Yamada, 1943). No vegetation harvesting occurred in the HN catchment during our study period, whereas 50% of the forest volume was removed from 1948 to 1952 in the SH catchment (Government Forest Experiment Station, 1961).

Stream gauges for catchments HN and SH are located just upstream of the junction of the two streams. Meteorological monitoring was conducted at an observation station near the runoff gauges called 'Kichi' (KC) (36°51'N, 139°01'E; 816 m a.s.l.) from 1937 to 1956 (Government Forest Experiment Station, 1961). Mean annual precipitation and air temperature were 2134 mm and 8.3°C. Monthly mean values of precipitation and air temperature which were estimated from KC by a lapse rate of 0.006°C m^{-1} at the mean elevations of the two catchments (1372 m for HN and 1057 m for SH) are given in Fig. 2. Because the air temperature for transition from rain to snow generally ranges from 1.5 to 2°C (Ohta, 1989), almost all precipitation from December to March occurred as snow throughout the catchments. Thus, we define a rainfall season from April to November and a snowfall season from December to March. The water year is defined as a period from 1 December to 30 November.

Fog precipitation caught by tree canopies (Geiger et al., 1995) may arise in forest in high elevation areas, which are located at a slightly lower elevation than the tree line in catchment HN. This influence should be evaluated if the quantity of usual precipitation is small. As large amounts of rain frequently fall in the catchment, however, the influence of this factor is assumed to be negligible for our precipitation estimation.

3. Data description and strategy for estimation

3.1. Background and data period

First, the history of hydrological study in the Takaragawa Forest Experiment Station is summarized to clarify the meaning of the present paper as well as the selection of data period for the analysis. Hydrological monitoring there was established in 1937. The purpose of this study was to understand the influences of forest cutting on stream runoff, considering the importance of this mountainous region in flood control and water resources for the metropolitan area. The well-laid plan at the beginning covered various kinds of field investigations including rainfall observations and snow surveys over both study catchments HN and SH (Yamada, 1943). The observation points gradually increased till 1953, to obtain more accurate catchment precipitation. However, mainly studies in which evaluations of catchment precipitation were not critical were applied to the data from Takaragawa. In fact, although a forest treatment was carried out for the SH catchment from 1948 to 1952 as mentioned above, only runoff increase in summer seasons owing to the treatment was investigated by means of statistical analyses comparing runoff before and after the treatment using the 'single catchment method' (Nagami et al., 1964; Nakano, 1971). Analyses of hydrological responses requiring catchment precipitation were put off for a long time, although the study plan at the beginning in the 1930s looked widely at climatic, geomorphological and geological effects on runoff as well as the forest influences (Yamada, 1943). After the most intense investigations for estimating catchment precipitation were carried out in 1953, investigations with similar intensity continued till 1955. Only precipitation observations at several points and runoff monitorings for HN and SH have been continued since then. Though studies on the effects of forest change on runoff were conducted through statistical analyses (Yoshino and Kikuya, 1984, 1985; Shimizu et al., 1992, 1994), few studies focused on the hydrological responses.

Generally speaking, to understand the influences of various kinds of catchment properties on runoff, which was the purpose at the beginning of the monitoring, is still a critical hydrological subject, especially in mountainous catchments with much snow cover. Estimating areal evapotranspiration from rugged terrain is becoming important not only for water resources management but also to determine energy and water exchange between land and atmosphere. In consideration of these needs, it is important to utilize data for catchment precipitation in the period when the most intense observations were carried out. Once the catchment precipitation is estimated without contradiction against the water balance, one can move on to analyses of runoff responses. Even though many coarse assumptions are introduced to estimate catchment precipitation from the limited rainfall observations and snow surveys, the estimation may be regarded as an approximation that will ensure a start of the hydrological analyses for specifying the effects of catchment properties on runoff.

3.2. Data description

Spatially distributed precipitation data were measured from storage rain gauges and snow surveys, and the intensity of data collection changed each year. The most intense investigations were conducted in 1953, as mentioned above, when rainfall was measured at 24 points and SWE at 49 points. The number of points was similar or somewhat decreased in other years. Rainfall in high elevation areas in the western part of HN was not measured before 1947, and the number of observation points decreased after 1955. Although the intense snow surveys at about 50 points continued from 1940 to 1958, the density of snow cover was not measured before 1953. Considering these circumstances, the estimation of annual precipitation was calculated from data for 1947–1955. In this estimation process, snow depth data for 1940–1958 were also employed for analysis of the characteristics of SWE distribution.

Precipitation at KC was recorded by a storage type rain gauge with a siphon drainage mechanism, except for the snowfall season. However, daily precipitation measured there throughout a year is mainly used for the analysis in the present paper. Runoff rates at gauging stations for HN and SH were recorded throughout each year, except that runoff data from SH were missing for 1948 and 1950. Runoff rates from HN are monitored by a rectangular channel system with a bridge. The functional relationships of the rates to the water levels were decided based on measurements on the bridge for cross-sectional distributions of water level and runoff velocity in the channel (Takeda, 1950). Runoff rates from SH were monitored by a weir system consisting of seven rectangular weirs. Discharge from the notches of the weirs could be calibrated by a measuring tank.

3.3. Estimation strategy

The data used in our analysis are the total amount of precipitation from early summer to mid-autumn and snow cover information in early spring, at each observation point in our study area. To approach the annual water balance, precipitation amounts in both the entire rainfall and snowfall seasons averaged for the catchment have to be estimated from the observed data. As a digital map is used for the estimation, the procedure can be divided into the following steps: (1) precipitation in an observation duration and SWE at a time of snow survey, for each grid point on the map, are estimated from their point observation data; (2) total precipitation during the entire rainfall or snowfall season for each grid point is estimated from each of them; (3) mean catchment precipitation during each season is calculated by averaging the total for each grid point over the catchment area; (4) annual mean catchment precipitation is given as the sum of the precipitation values in both seasons.

Tani (1994) analysed distributions of rainfall and SWE in the study area in 1953 and determined their values for every grid point (100 m × 100 m) on a digital map. This method is employed for Step (1). For Step (2), a simple extrapolation is applied to calculate precipitation during the entire seasons. Because the records for the snowfall season were based not on precipitation but on SWE or snow depth, additional calculations are necessary to estimate precipitation in the snowfall season within Step (2).

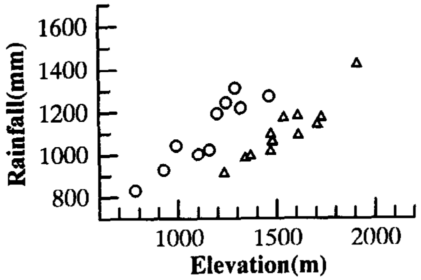

Fig. 3. Relationship of precipitation in the rainfall season of 1953 to the unfiltered elevation. ○, Near a stream valley; △, near a ridge line.

4. Estimation processes

4.1. Estimation of precipitation in rainfall season

Tani (1994) investigated precipitation distribution in the study area (Takaragawa) in the summer season of 1953, and found that observed precipitation roughly increased with elevation but that the precipitation near ridge lines was less than near stream valleys when elevations were similar to each other (Fig. 3). He attempted to improve a correlation between precipitation and elevation based on the concept that the precipitation increase would be dominated by an orographic scale which is larger than the scale of the rugged terrain. A spatial filtering procedure (Nogami, 1985) was applied to a digital map with a unit length of 100 m which included the study area. The following filtering technique developed for summit level analysis was employed for the procedure:

1. Consider a grid point (i,j) on a digital map and a square network the corners of which are grid points $(i-k, j-k)$, $(i+k, j-k)$, $(i+k, j+k)$ and $(i-k, j+k)$. The network consists of $(2k-1) \times (2k-1)$ grid points, the centre of which is grid point (i,j). The sides of the network are $(2k-1) \times 100$ m.
2. Replace the elevation at grid point (i,j) by the maximum elevation in the network. The summit of the network at a scale of $(2k-1) \times 100$ m is obtained from the replacement.
3. Apply the replacement throughout the digital map. Elevations for the summit level

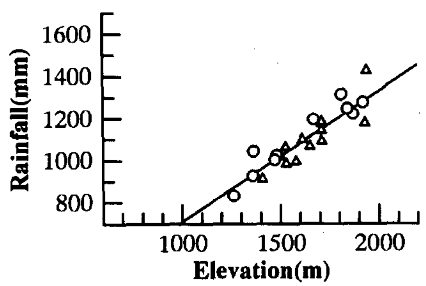

Fig. 4. The same as Fig. 3, but showing the relationship to the filtered elevation when the scale is 2300 m. The line indicates the regression.

cover the whole area and a filtered map is produced. The scale of filtering is characterized by (2k–1) × 100 m, increasing with the value of k.

A much better correlation between precipitation and the filtered elevation based on a square network of k = 12 (a scale of 2300 m) was found in Fig. 4 compared with the correlation using unfiltered elevation in Fig. 3. The correlation with filtered elevation is described by

$$P_r/P_{KCr} = 0.000618(H + 147) \quad (1)$$

where P_r and P_{KCr} are cumulative precipitation amounts at each grid point and at KC (mm), during the observation period in the rainfall season, and H is the filtered elevation (m) when k = 12. The precipitation at each grid point on the digital map could then be calculated by Eq. (1). Tani (1994) finally obtained the catchment means of precipitation for HN and SH as 1122 mm and 915 mm, respectively, in the observation period in summer 1953, when precipitation at KC was 833 mm.

Precipitation at each grid point during the observation period for each year from 1947 to 1955 is calculated by the same method. Before applying the above method to the data for years other than 1953, however, it may be better to confirm that 1953 was not an abnormal year from a meteorological point of view. The precipitation totals in rainfall and snowfall seasons at KC were 1575 mm and 640 mm in 1953, whereas the average values for the 9 years from 1947 to 1955 were 1419 mm and 716 mm. Therefore, the method developed for the 1953 data can be reasonably applied to those for other years from 1947 to 1955 to estimate normal values of catchment precipitation. Because Eq. (1) is a linear function, the 9 year average relationship between P_r/P_{KCr} and H is easily calculated, and is expressed as

$$P_r/P_{KCr} = 0.000742(H + 80) \quad (2)$$

The distribution of this ratio gives an average distribution of precipitation in a rainfall

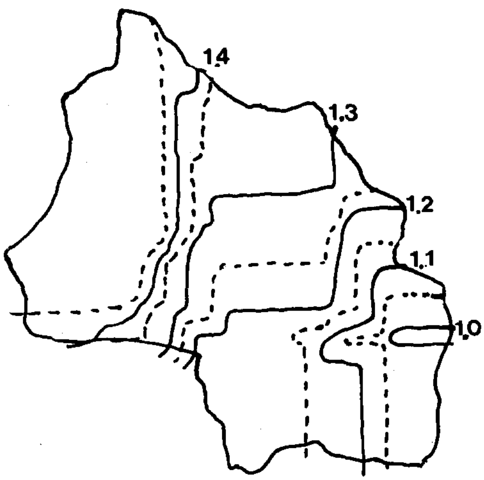

Fig. 5. Contour lines for the ratio of precipitation at each grid point to precipitation at KC in the rainfall season averaged over 9 years from 1947 to 1955.

season. Contour lines for the ratio of precipitation at each grid point to precipitation at KC are shown in Fig. 5.

Because the observation period for rainfall was limited to a portion of the entire rainfall season (e.g. from 30 June to 3 October 1953), the distribution of rainfall outside this measured period must be extrapolated from the shorter-term observations. We assume that the coefficients obtained in Eq. (2) can be applied to the rainfall distribution outside the measured period. The mean catchment precipitation outside the measured period can be obtained by averaging the ratio (P_r/P_{KCr}) at each grid point in Fig. 5 over the catchment area and multiplying this average by the cumulative precipitation outside the measured period that was continuously recorded at KC. The total of these two precipitation amounts in the observed period and outside this period gives the mean catchment precipitation for the rainfall season from April to November. Using this method, the 9 year average precipitation in catchment HN is 1831 mm compared with 1419 mm at KC. Because runoff from SH had missing data in 1948 and 1950, the average precipitation for the 7 years of complete runoff records is 1456 mm compared with 1383 mm of precipitation at KC.

4.2. Distribution of SWE at the time of snow survey

Snow density was measured in surveys from 1953 to 1958, but only snow depths were measured in 1952 and earlier. Thus, it is necessary to analyse the records of density to calculate SWE values from snow depth data before 1953. The seasonal distribution of snow density is given in Fig. 6, based on the 15 snow surveys conducted in the study area from 1953 to 1958. As has been widely noted (Sato and Kurashima, 1988; Braun, 1991), the density tends to increase with time. The correlation for the study area shown in Fig. 6 can be described as

$$Dn = 0.00263T + 0.193 \tag{3}$$

where Dn is the snow density (g cm^{-3}) and T is the number of days since 1 January. The duration of the density increase described by Eq. (3) applies from early February

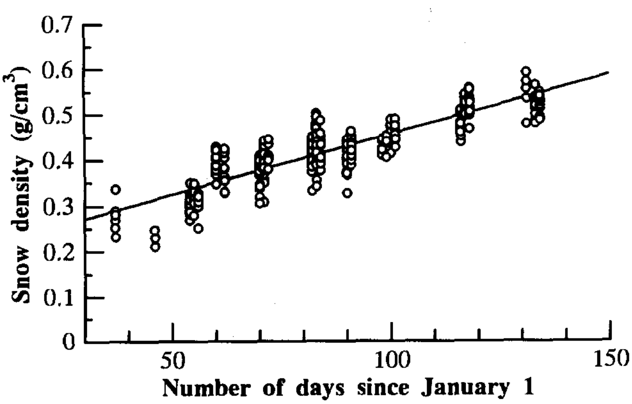

Fig. 6. Records of snow density against the days of snow survey. The day was expressed as the date since 1 January.

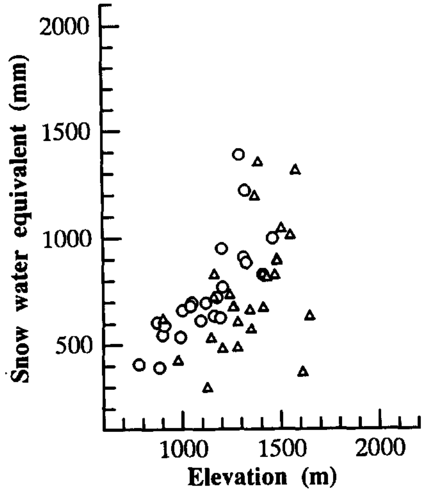

Fig. 7. Relationship of SWE surveyed on 23 and 24 March 1953 to the unfiltered elevation. ○, Near a stream valley; △, near a ridge line.

to mid-May. Because snow surveys were conducted during this period, each SWE value before 1953 was simply calculated as the product of snow depth and *Dn*.

Distribution of SWE in the study area was analysed by Tani (1994), based on the data of the survey conducted on 11 and 12 March 1953. Contour lines for SWE on the digital map were drawn according to this analysis. The same procedure of spatial filtering that was used for the precipitation distribution in a rainfall season (described in Section 4.1) was also applied to SWE. A *k* value of 12, which gives a filtering scale of 2300 m, was suitable for both the precipitation and the SWE. Fig. 7 and Fig. 8 display the observed

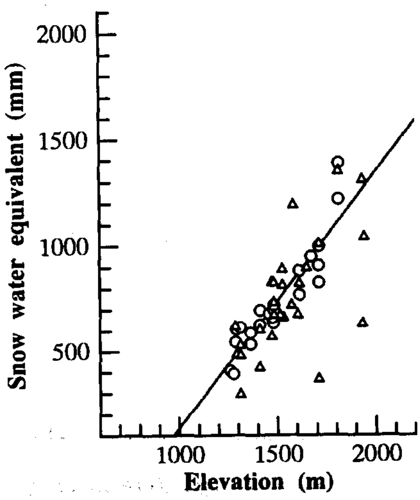

Fig. 8. The same as Fig. 7, but showing the relationship to filtered elevation when the scale is 2300 m. The line indicates the regression.

SWE values against the unfiltered and filtered elevations, respectively. SWE (Sw, in mm) was much more highly correlated with the filtered elevation (H, in m) and is expressed as

$$Sw = 1.202(H - 889) \tag{4}$$

The mean SWE values for catchments HN and SH averaged over their areas were 937 mm and 534 mm, respectively, in 1953 (Tani, 1994). The mean catchment SWE at the time of snow survey for each year from 1947 to 1955 can be obtained through the same procedure. This provides the following relationship for Sw and H at every grid point on the digital map averaged over the 9 years:

$$Sw = 1.109(H - 760) \tag{5}$$

The 9 year averaged means of SWE were calculated as 1009 mm for HN and 637 mm for SH.

4.3. Estimation of precipitation in snowfall season

As our measurements of SWE in the study area do not span the entire snowfall season, we must develop procedures to extrapolate these results to the entire season. The approach

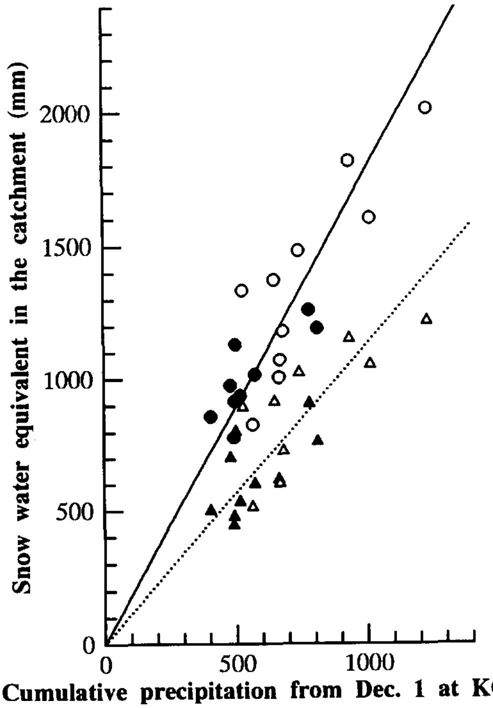

Fig. 9. SWE values averaged over each catchment area of HN and SH, against the cumulative precipitation at KC integrated from 1 December to the last day of the snow survey. ●, For HN from 1947 to 1955; ○, for HN of other years; ▲, for SH from 1947 to 1955; △, for SH of other years. Lines are drawn so that each of their slopes is equal to the ratio of the mean catchment SWE to the cumulative precipitation at KC. Continuous line, HN; dotted line, SH.

we attempt to use here is to extrapolate precipitation to the entire snowfall season based on relationships between SWE records by snow survey and the precipitation data monitored continuously at KC.

First, we should check relations of mean catchment SWE to cumulative precipitation at KC integrated from 1 December (the first day of the snowfall season) to the last day of the survey. A snow survey conducted before the snowmelt season (from mid-February to mid-March) in each year was used for the analysis. A mean catchment SWE was averaged in a catchment by using a functional relation between surveyed SWE and filtered elevation in each year such as Eq. (4). These values for each of HN and SH obtained from the snow surveys from 1940 to 1958 are plotted against cumulative precipitation at KC in Fig. 9. Good correlations are obtained for each catchment. Data for the 9 years from 1947 to 1955 used in our estimation process are marked with filled symbols (and others with open symbols), and the lines are drawn so that each of the slopes is equal to the ratio of mean catchment SWE to the cumulative precipitation at KC, based on the 9 year records. The ratio for each catchment is expressed as

$$\alpha_{\mathrm{HN}} = Sw_{\mathrm{HN}}/P_{\mathrm{KC}} = 1.795 \tag{6}$$

$$\alpha_{\mathrm{SH}} = SW_{\mathrm{SH}}/P_{\mathrm{KC}} = 1.133 \tag{7}$$

where α_{HN} and α_{SH} are the ratios of the catchment mean SWE for HN ($\mathrm{Sw}_{\mathrm{HN}}$) and for SH ($\mathrm{Sw}_{\mathrm{SH}}$) to the cumulative precipitation at KC (P_{KC}). These relationships demonstrate that mean catchment SWE was roughly in proportion to the cumulative precipitation at KC integrated from 1 December to the last day of the snow survey.

Because we need the mean catchment precipitation instead of SWE, the water balance of snow cover should be examined to translate SWE into precipitation. The water balance at any point in a catchment is

$$\frac{\mathrm{d}Sw}{\mathrm{d}t} = p_{\mathrm{r}} + p_{\mathrm{s}} - e - q_{\mathrm{b}} - q_{\mathrm{f}} \tag{8}$$

where p_{r} is the rainfall rate, p_{s} is the snowfall rate, e is the evaporation rate from the surface of snow cover, q_{b} is the discharge rate from the melting snowpack at the ground surface, and q_{f} is the discharge rate produced when liquid water storage within the snow cover, originating from either snowmelt on and near the snow surface or rainfall, exceeds the retention capacity of snow cover. It is reasonable to assume $p_{\mathrm{r}} = 0$ and $q_{\mathrm{f}} = 0$ at the time of snow survey and earlier because the survey was conducted before the snowmelt started in early spring. Discharge from the melting snowpack at the ground surface (q_{b}) is assumed to be negligible at present, although it will be discussed in Section 4.4. As for the evaporation, e, Nakai et al. (1993) recently demonstrated that the evaporation rate from the intercepted snow cover on the canopy of an evergreen conifer was 5.6 times that from bare land. However, as most of our study area was covered with deciduous trees and a treeline existed in the upper portion of catchment HN, the area where evaporation from the intercepted snow occurred was small. Thus, only evaporation from the snow surface is considered here. Kojima et al. (1985) measured the evaporation from the snow cover at an observation station from December to March and evaluated the total amount as only 16 mm. Accordingly, it can be assumed that $e = 0$ for our approximation also. Based on these assumptions, an approximate form of Eq. (8) can be simply written as

$$\frac{dSw}{dt} = p_s \tag{9}$$

Using the approximation in Eq. (9), precipitation can be estimated as SWE increase at any point in a catchment before the snowmelt season. Then, the mean catchment SWE in each year plotted in Fig. 9 gives the mean catchment cumulative precipitation integrated from 1 December to the last day of the snow survey, and α given in Eq. (6) or Eq. (7) indicates the averaged ratio of cumulative precipitation between the mean catchment and KC. The approximation in Eq. (9) also shows that the spatial distribution of the cumulative precipitation is the same as the SWE distribution and can be described by a function of filtered elevation such as the right-hand side of Eq. (5). When both sides of Eq. (5) are divided by the cumulative precipitation at KC, which was 562 mm (the 9 year average) and SWE is converted into cumulative precipitation, we can obtain the following equation, which gives an averaged precipitation distribution in the snowfall season as a ratio to the precipitation at KC.

$$P_s/P_{KCs} = 0.00197(H - 760) \tag{10}$$

where P_s and P_{KCs} are cumulative precipitation amounts at each grid point of the digital map and at KC (mm), integrated from 1 December to the last day of the snow survey. The use of eqn (10) produces the isolines of this ratio shown in Fig. 10. The contour lines show a much wider distribution of precipitation during the snowfall season (Fig. 10) compared with the rainfall season (Fig. 5).

Let us assume that the ratio of P_s to P_{KCs} in Eq. (10) is extrapolated to the ratio in the period from the snow survey to the end of the snowfall season (31 March). The mean catchment cumulative precipitation in this period can be calculated by averaging the ratio (P_s/P_{KCs}) at each grid point in Fig. 10 over the catchment area and multiplying this average

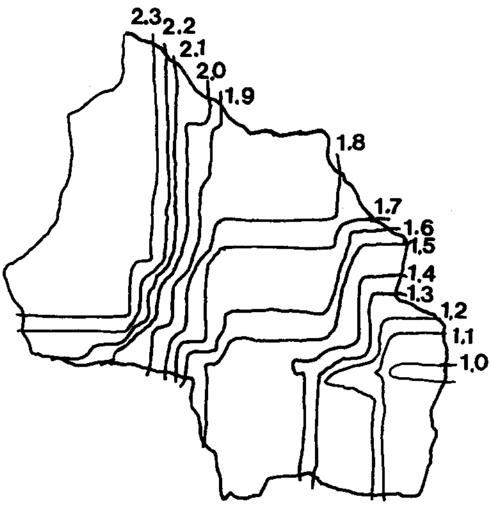

Fig. 10. Contour lines for the ratio of SWE at each grid point to the cumulative precipitation at KC averaged over the 9 years from 1947 to 1955.

by the cumulative precipitation at KC in the period. The total of the cumulative precipitation before the snow survey in each year approximated by the SWE records and the cumulative precipitation after the survey estimated by the above procedure gives the mean catchment precipitation for the entire snowfall season from December to March in each year. Averaging the values from 1947 to 1955, the 9 year average precipitation in catchment HN is 1285 mm compared with 716 mm at KC. Because runoff from SH had missing data in 1948 and 1950, the average precipitation for the 7 years of complete runoff records is 839 mm compared with 732 mm at KC.

4.4. Estimation of snowmelt at the ground surface

The cumulative precipitation in the snowfall season estimated from the surveyed SWE values in the preceding section did not include evaporation loss and discharges from the snow cover. Discharge from the melting snowpack at the ground surface (q_b) may contribute to baseflow rate throughout midwinter. In general, this also depends on rainfall conditions during the previous autumn. However, each of the two factors must cause a different fluctuation in the baseflow rate. The rate responding to the amount of autumn rainfall may fluctuate widely from year to year, whereas the rate originating from the melting snowpack (q_b) may be more constant every year because it is mainly controlled by heat flux from the deep soil. Thus, if we select the minimum runoff rate of each snowfall season, the smallest value within a set of the minimum rates should approximate the mean value of q_b. Fig. 11 shows the minimum daily runoff rates from HN and SH for the snowfall seasons from 1940 to 1958 expressed as the water depth over the unit catchment area. The minimum rate for SH tends to be larger than that for HN. This may be caused by the difference in soil temperature owing to elevation, although other hydrological or geological factors could be involved. The minimum runoff rate from SH begins to increase

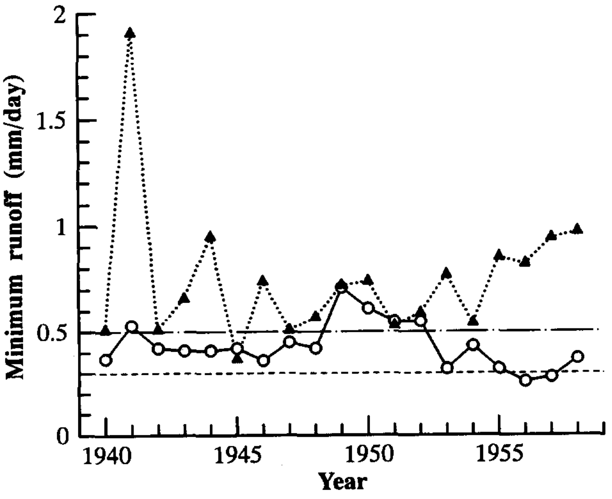

Fig. 11. Minimum daily runoff rates of HN and SH in snowfall seasons from 1940 to 1958. O, HN; ▲, SH; dashed line, 0.3 mm day^{-1}; dashed and dotted line, 0.5 mm day^{-1}.

Table 1
Estimation results of catchment mean precipitation for HN and SH

	P_{rain}	P_{snow}	P_{annual}	Q_{annual}	L_{annual}
HN[a]	1831	1321	3152	2784	368
KC[a]	1419	716	2135		
SH[b]	1456	899	2355	1745	610
KC[b]	1383	732	2215		

P, Precipitation; *Q*, runoff; *L*, loss. Subscript 'rain' indicates rainfall season; subscript 'snow' indicates snowfall season.

[a] Values averaged over the 9 years from 1947 to 1955.

[b] Values averaged over the 7 years from 1947 to 1955 except for 1948 and 1950 missing runoff data from catchment SH.

after 1953; this is partially attributable to a decrease in evapotranspiration caused by a selected cutting of forest in the catchment (Fig. 11). However, it was believed that the influence of the cutting of deciduous trees on snow cover was small compared with the larger influence of cutting on evapotranspiration and that the cutting had little effect on q_b because it was produced under the snow cover, owing to the ground heat flux. Thus, the average daily rates of q_b are estimated as constant values in the 9 years; these are 0.3 mm day^{-1} for HN and 0.5 mm day^{-1} for SH (Fig. 11). Snowmelt for the entire season would then be 36 mm and 60 mm for HN and SH, respectively. Although these values may have little effect on the spatial distribution shown in Fig. 10, precipitation for the entire snowfall season would be modified to 1321 mm for HN and 899 mm for SH.

5. Estimated annual water balance

The sum of precipitation during the rainfall and snowfall seasons yields annual precipitation of 3152 mm for HN and 2355 mm for SH. Components of the annual water balance (precipitation, runoff and loss) for each catchment are given in Table 1. The annual loss from SH (610 mm) is much larger than that from HN (368 mm). The relationship between annual precipitation and annual runoff yield from 1947 to 1955 is given in Fig. 12. Although some scatter exists and effects of 50% selective cutting during our study duration on losses from SH could not be detected, the data show that for most years losses are about 400 mm for HN and about 600 mm for SH. These data suggest that annual losses for each year are similar to multiyear averages given in Table 1. Assessment of the losses as an estimation of mean catchment evapotranspiration will be discussed in the next section.

6. Discussion on estimated evapotranspiration

Various kinds of measurement and estimation errors may have been included in annual precipitation values estimated for our catchments in Table 1. A large difference in annual losses between HN and SH will require an examination of the physical bases, as it might be caused by large errors included in the estimated precipitation values. It is difficult to find a

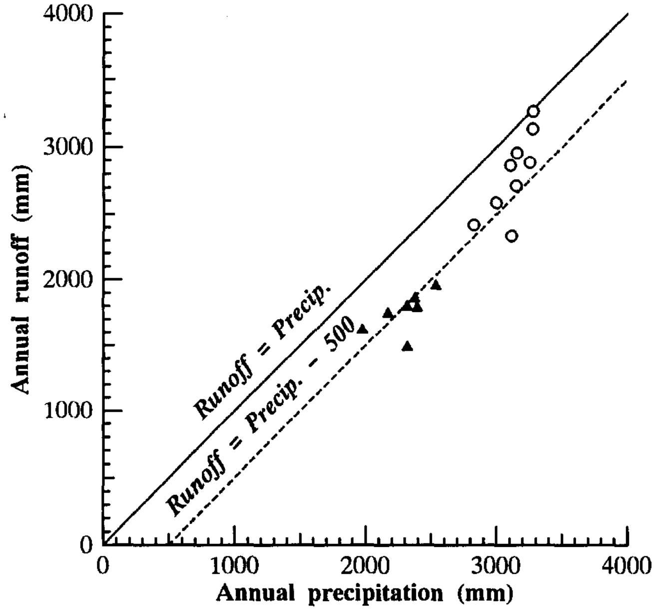

Fig. 12. Relationships of estimated annual precipitation and annual runoff yield in HN and SH catchments. ○, HN; ▲, SH.

suitable method for the assessment of an estimated precipitation value. One of the strategies is to analyse the loss in the annual water balance against the evapotranspiration estimated from the climatological condition. Before the analysis is applied, however, regarding an annual loss as the actual annual evapotranspiration from a catchment generally requires not only an accurate estimation of precipitation but also examination of the fluctuation of catchment water storage in each year and the effect of deep seepage, because the loss depends on errors in each component of the water balance equation. For our estimation, the procedure of averaging in several years (9 years for HN and 7 years for SH) seemed to mask the effect of storage fluctuation on the loss values even though those in each year widely fluctuated according to the effect. The deep seepage can be neglected because rocks cover steep slopes all over our catchment, as described in Section 2. In addition to this, the baseflow rate from each of HN and SH during the snowfall season did not decrease below the runoff rate that must have been produced from the melting snowpack at the ground surface (Section 4.4). This also suggests that the deep seepage was negligibly small for each catchment. Thus, we can assess an estimated annual precipitation value by means of comparing the annual loss with an evapotranspiration obtained from the climatological condition.

Although a large number of estimation methods for evapotranspiration have been developed, many meteorological factors and model parameters are usually required to apply most of these methods. Thus, the simple estimation method of Hamon (1961) for potential evapotranspiration, which needs only air temperature data, is applied for the assessment. This method is defined as

$$e_{hd} = 0.14 Ds^2 Vs \tag{11}$$

Table 2
Catchment evapotranspiration (E_w) based on the water balance and potential evapotranspiration (E_h) calculated by Hamon's method

Catchment	Evapotranspiration		Duration	Years for averaging
HN	E_w	368	Annual	1947–1955
	E_h	428	Apr.–Nov.	1937–1956
	E_h	397	May–Nov.	1937–1956
	E_h	347	Jun.–Nov.	1937–1956
SH	E_w	610	Annual	1947–1955
	E_h	479	Apr.–Nov.	1937–1956
	E_h	447	May–Nov.	1937–1956
	E_h	390	Jun.–Nov.	1937–1956
T1	E_w	651	Jun.–Oct.	1982–1987
	E_h	406	Jun.–Oct.	1937–1956
K1	E_w	461	Jun.–Oct.	1939–1946
	E_h	457	Jun.–Oct.	1938–1956
K2	E_w	494	Jun.–Oct.	1939–1946
	E_h	457	Jun.–Oct.	1938–1956

HN, Honryu in Takaragawa; SH, Shozawa in Takaragawa; T1, Takaragawa 1-gosawa; K1, Kamabuchi 1-gosawa; K2, Kamabuchi 2-gosawa.

where e_{hd} is Hamon's potential evapotranspiration rate (mm day^{-1}), Ds is the possible hours of sunshine in units of 12 h, and Vs is the saturated water vapor density at the daily mean temperature (g m^{-3}). This method has been used as one of the standards of evapotranspiration from a climatological viewpoint in Japan (Ishii, 1987; Rampisela et al., 1990).

Table 2 compares cumulative values of evapotranspiration (E_w) based on the water balance with cumulative values of Hamon's potential evapotranspiration (E_h). In general, an annual loss calculated as the difference between precipitation and runoff is considered as an annual evapotranspiration. However, we should note that losses from our two catchments listed in Table 1 were estimated as the annual totals on the assumption that no evaporation occurred from the snow surface. Therefore, E_h, to be compared with E_w, should be calculated for periods without snow cover. As the area of snow cover gradually decreases in spring, three periods (from April to November, from May to November and from June to November) are used for the comparison. Hamon's potential evapotranspiration is calculated from monthly values of air temperature at the averaged height in each catchment area estimated from that monitored at the observation station using a lapse rate of 0.006°C m^{-1}. In Table 2, E_h was calculated in different years from E_w because it can be obtained only in periods when air temperature data were available. However, this difference is acceptable, as our comparison is made from a climatological viewpoint.

E_w and E_h values for some smaller forested catchments in snowy mountainous regions in Japan are also listed in Table 2. Because each area of these smaller catchments is of hectare scale, the mean catchment precipitation is believed to be the same as that monitored at the observation station at least in summer seasons. Therefore, evapotranspiration from each catchment in summer seasons can be estimated from the water balance with higher accuracy than for our catchments HN and SH. Data from smaller catchments listed in Table 2 are outlined as follows. Takaragawa 1-gosawa (T1, 0.065 km^2) is a sub-catchment of SH, and Shimizu et al. (1994) analysed the effects of a contour-line strip cutting on runoff and evapotranspiration in summer seasons based on hydrological data recorded in this catchment. Evapotranspiration from June to October, when effects of snow were negligible, were estimated using the method of short time period water-budget proposed by Linsley et al. (1958). E_w values before the cutting (1982–1987) are listed in the table. In this estimation, the key assumption is that the difference in water storage can be neglected at two time points when runoff rates are the same. Based on this assumption, the catchment evapotranspiration in a period shorter than a year can be calculated from the water balance equation as the difference between the total precipitation and the total runoff yield. The other data in Table 2 are quoted from the estimation by Suzuki (1985) on evapotranspiration values from Kamabuchi 1-gosawa (K1, 0.031 km^2) and Kamabuchi 2-gosawa (K2, 0.025 km^2) calculated by the same method. Clear cutting was conducted in K2, but the vegetation condition was constant in K1. E_w values before cutting (1939–1946) are listed for K1 and K2 in Table 2.

Comparing evapotranspiration based on the water balance (E_w) with Hamon's potential evapotranspiration (E_h), the following findings have been obtained. E_w from HN lies between E_h from May to November and E_h from June to November, but it is smaller than E_h from April to November. As snow covered most of the catchment area in April and part of it in May, and E_h must be an overestimate of actual evapotranspiration in these months, it is concluded that E_w and E_h are similar to each other for HN. This relationship is also detected for K1 and K2 in Table 2. Considering that the values of E_w and E_h from these smaller catchments are estimated for the summer season from June to October, one might consider the value of E_w from HN to be too small as an annual value. Nevertheless, this value is believed to be valid because evapotranspiration seems to be negligibly small during a long period of snow cover from December to May in the catchment area of HN, as mentioned in Section 4.3. On the other hand, E_w from SH is larger than each of the E_h values in the three periods indicated in Table 2. Looking at evapotranspiration from T1, a sub-catchment of SH, E_w is 250 mm larger than E_h. As the catchment T1 is small enough for an accurate estimation of water balance, a large value of E_w for SH compared with E_h may be plausible. According to the above comparisons, different relationships between E_w and E_h for our catchments HN and SH were not attributed only to observation and estimation errors.

Let us consider possible causes for the different relationships which indicate that E_w is similar to E_h for HN and larger than E_h for SH. First, because of the high elevation near the top of HN, evapotranspiration should decrease owing to frequent fog. Also, evaporation from wet canopies may be dramatically reduced in the 31% area of the catchment above the tree line. The eastern aspect of HN would receive less solar radiation than the southern aspect of SH. In addition to these causes, a heterogeneous snow cover distribution in late spring may produce a large effect on evapotranspiration. Ohta et al. (1994) demonstrated that latent heat flux was transported downward continuously to be used for condensation on the snow surface in May and later in a subalpine region of Japan. The same flux condition must continue in late spring in catchment HN except for low elevation areas. On the other hand, upward latent heat flux must actively arise as transpiration from young leaves in low elevation areas including catchment SH. All these characteristics must increase any difference in evapotranspiration between our two catchments. This difference would be based on hydro-meteorological processes peculiar to each catchment in this mountainous terrain. Further field studies will be required to quantify each of the processes.

7. Conclusion

Records of precipitation in rainfall seasons and those of SWE in snowfall seasons obtained in two small neighbouring catchments in a mountainous region with abundant snow cover were analysed to evaluate the annual water balance. Because both of the records were well correlated with filtered elevation, the distribution of precipitation was estimated based on the correlation in each of the seasons. Relationships of the mean catchment SWE to the cumulative precipitation monitored continuously at the observation station were used to translate SWE records into precipitation. Annual water balances in the two catchments averaged over several years were calculated from the annual total values of mean catchment precipitation and runoff yield. The evapotranspiration defined as the loss, the difference between precipitation and runoff yield, was compared with Hamon's potential evapotranspiration from a climatological view point. Although a large difference in annual evapotranspiration was calculated from the water balance between the two catchments, this result was plausible in consideration of comparisons with water balances for smaller catchments with high accuracy, and several probable causes could be suggested for the difference in evapotranspiration. Large values of annual precipitation in the mountainous catchments, especially more than 3000 mm for HN, could be evident based on the water balances with acceptable accuracy.

From another point of view, our results show a difficulty in estimating the water balance in such a mountainous catchment. Though our data were based on dense observations for the distributions of rainfall and SWE, many assumptions were necessary to estimate annual catchment precipitation. This suggests that the precipitation estimation would not be easy for other catchments without such dense data. In this sense, the precipitation estimated in the present study should be efficiently applied for further analyses on the hydrological characteristics of catchments HN and SH such as runoff response to each storm event and runoff response to snow melt in each year. This conclusion is one role of this paper in the hydrological studies conducted at the Takaragawa Forest Watershed Experiment Station.

On the other hand, some of the procedures used in this paper are not limited to catchments HN and SH but are applicable to other catchments. When one analyses a relationship of rainfall or SWE to elevation, a filtered elevation may account for the relationship better than the rugged elevation of the observation point although the filtering scale is different in each region. This must be helpful for estimating the spatial distribution of precipitation in general.

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