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Characteristics of summer water balance in eastern Siberian tundra watershed

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Abstract: Hydrological observations were carried out in a tundra watershed near Tiksi, eastern Siberia during summer in 1997, 1998 and 1999, and characteristics of the summer water balance are discussed. Due to strong wind in winter and low height vegetation, snowdrift is a common feature in the tundra area, and remain as snow patches until the middle of summer, supplying meltwater to rivers. In the summer water balance analysis, we introduced the ratio a of the snow patch area to the whole watershed area as a tuning parameter for taking into account the meltwater from the snow patches. The obtained results are as follows: The contribution of meltwater from snow patches to discharge continued in summer after the main snowmelt period. Rainfall and snowmelt amount changed substantially from year to year. The major output component was discharge. The contribution of evapotranspiration to the output varied year to year. The storage change was small compared with the other components.

1. Introduction

It is important to understand energy and water transport in the Arctic in order to determine the role of the Arctic on global atmospheric water cycles and in turn, global weather processes. In the Arctic, permafrost is expanded and it affects the water and energy circulation. The thawing layer during summer is called the active layer, and water moves only in this layer because the permafrost layer is impermeable, preventing from infiltration of rainfall and snowmelt water into the deep ground. Therefore, water cycle and soil storage amount depend on the thickness and soil properties of the active layer (Kane, 1997). Since the storage capacity of the active layer is small when the active layer is not so deep, rain and snowmelt water run off quickly to rivers. Thus, permafrost affects the water circulation. Snowdrift is a common feature in the tundra region due to strong wind and low vegetation such as mosses, sedges and lichens. It is usually formed leeward of hills and ridges, and in depressions such as streambeds (Liston and Matthew, 1998). Since precipitation is small in the Arctic, meltwater from snowdrifts, which remains as snow patches until the middle of summer, is one of the main sources of summer runoff. Therefore we must consider the amount of meltwater from the snow patches as an input to the watershed in order to estimate the water balance of the whole summer.

The objective of this study is to examine the summer water balance of a tundra watershed where snow patches exist. In order to explain the water balance, the all components of the water balance must be obtained, however, it is very difficult to discuss them in detail because the estimation errors are hardly eliminated. Therefore, we wish to focus attention on the contribution of the snowmelt water from the snow patches to the water balance during the summer.

The observation was carried out as part of the GAME-Siberia project. There have been few reports made on hydrological studies in this region of the Siberian Arctic. In the analysis, we introduce a method using a ratio, a, of snow patch area to watershed area as a tuning parameter for the water balance, and the contribution of snowmelt amount to the summer water balance is considered. In the reports of the water balance in the Arctic, many process studies, such as evaporation and discharge have been done, however, there are few reports on water balance between input and output (Woo, 1986). Characteristics of seasonal and yearly change of water balance will also be discussed.

2. Observations

An experimental watershed, 5.5 km^2 in area, is located near Tiksi, eastern Siberia (71°40′N, 128°50′E). It is near the mouth of the Lena River and 5 km from the Laptev Sea coast (Fig. 1). The altitude of the watershed ranges from 40 to 300 m a.s.l. Permafrost completely underlies this region, its thickness reaches over 500 m (Fartyshev, 1993). The active layer thickness varies from 20 to 70 cm. The surface conditions in the watershed are classified into three types: snow patch; wet land with mosses including sphagnum, sedges and dry tundra with lichens. The wet tundra is distributed on flat plains and the lower parts of slopes, where the slopes are gentle. The dry land is distributed on ridges and upper parts of slopes. The leeward sides of ridges and hills, where snowdrift are formed, is usually dry because vegetation other than lichens cannot live and no soils are developed.

There were two main observation stations (1: hydrological station, 2: meteorological station in Fig. 1). The measured items at these stations are shown in Table 1. At the hydrological station, which was located at the outlet of the watershed, the water level of the stream was continuously measured by a pressure transducer at 30 minute intervals. The continuous discharge was calculated from the relationship between the water level and discharge. Several times in a season, the discharge was calculated by manually measuring the flow rate and cross-sectional area of the stream and relating it to the water level at the time, in order to obtain the water level-discharge relationship.

At the meteorological station (2 in Fig. 1), a meteorological mast and a radiation measuring system were installed on wet land at the end of August 1997. Wind speed, air temperature, relative humidity, air pressure, soil temperature, net radiation and precipitation were measured at the station. A digital data logger (LS3300PtV, Hakusan Corp., Japan) was used to record the data, 10 minute average. Soil water content was measured by TDR sensor at 1 hour intervals. Because we could not obtain air temperature, relative humidity and windspeed data before August, 1997, the daily mean was created from the data at the Russian Hydro-Meteorological Station in Tiksi,

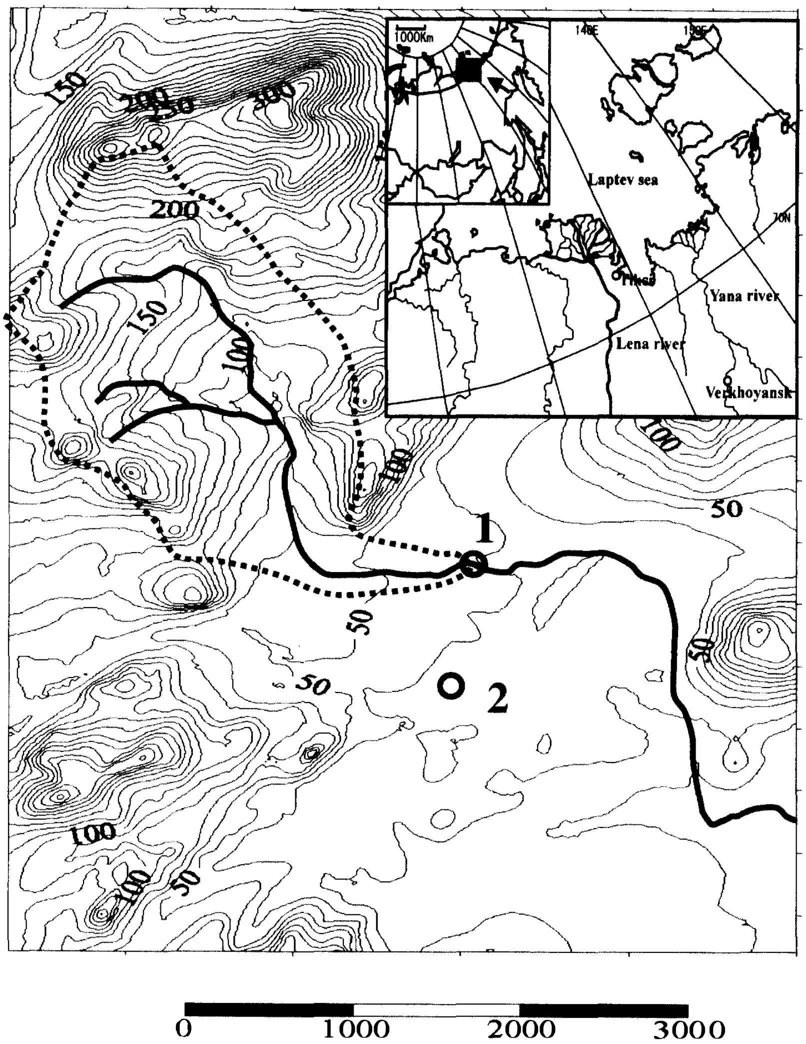


Fig. 1. Observation site. 1: hydrological station, 2: meteorological station, broken line: boundary of watershed. Area of experimental watershed was 5.5 km^2.

by comparing the mast data (MS) with the data at Russian Hydro-Meteorological Station (HM) in June, July and August, 1998, as follows:

$$\begin{aligned} T_{MS} &= 0.99\, T_{HM}, \qquad (R^2 = 0.97) \\ e_{MS} &= 0.68\, e_{HM}, \qquad (R^2 = 0.91) \\ u_{MS} &= 0.92\, u_{HM}, \qquad (R^2 = 0.89) \end{aligned} \tag{1}$$

where T is the daily mean air temperature, e the daily mean water vapor pressure and u the daily mean wind speed. Because the Russian station was 5 km away to the east and located near the seacoast, e and u were smaller at our station than at the Russian station. On dry land near the meteorological mast, turbulent fluxes were measured in July 1999 for 8 days using an eddy covariance system with a 3-dimensional sonic anemometer and a krypton hygrometer. Fluctuations of wind speed, temperature and humidity were sampled at 20 Hz, and the 10 minute averages of the covariance were recorded by a data logger (CR-10, Campbell Scientific Inc., USA).

Table 1. List of observed items and instruments.

	Measurements	Auto/Manual	Period	Interval	Equipments
Hydrological Station	Water level	Auto	June to September	30min.	Pressure transducer (C50, CTI Science System, Japan)
	Discharge	Manual	June to August	Several days	Electromagnetic current meter
Meteorological Station	Air Temperarure	Auto	After September, 1997	10min. Average	1, 2, 4, 10m height (HMP-35, Vaisala)
	Relative humidity	Auto	After September, 1997	10min. Average	1, 2, 4, 10m height (HMP-35, Vaisala)
	Windspeed	Auto	After September, 1997	10min. Average	2, 4, 10m height (AC860, Makino, Japan)
	Net radiometer	Auto	After September, 1997	10min. Average	1.5m height (Q7, REBS)
	Upward Shortwave radiation	Auto	After September, 1997	10min. Average	1.5m height (MS-801F, EKO, Japan)
	Downward longwave radiation	Auto	After September, 1997	10min. Average	1.5m height (MS-201F, EKO, Japan)
	Upward longwave radiation	Auto	After September, 1997	10min. Average	1.5m height (MS-201F, EKO, Japan)
	Air Pressure	Auto	After September, 1997	10min. Average	(PTB100B, Vaisala)
	Soil Temperature	Auto	After September, 1997	10min. Average	-0.01, -0.05, -0.10, -0.20, -0.30, -0.48m depth, Pt100
	Volumetric soil water content	Auto	After September, 1997	1hour	-0.05, -0.15, -0.30m depth, TDR (TRIME-MUX6, IMKO)
Dry land near Met. station	3D-Sonic anemometer	Auto	29 July to 5 August, 1999	10min. Average	20Hz sampling (Campbell Scientific Inc.)
	Hygrometer	Auto	29 July to 5 August, 1999	10min. Average	20Hz sampring, Krypton hygrometer (KH-20, Campbell Scientific Inc.)
Snow patches	Snow ablation rate	Manual	June to July at 4 to 6 points	about 10days	Snow stakes method
	Snow density	Manual	June	Several times	Snow sampler (100cc)

In order to estimate snowmelt amount, snow ablation and the surface snow density were measured from the end of May to July in 1997 and 98. The snowmelt amount was obtained as a product of the snow ablation and surface snow density. Snow ablation measurements using snow stakes were carried out at several points on snow patches in the watershed, at about 10 day intervals. The surface snow density was measured on a snowdrift near the meteorological station in June in both years. The average snow density was about 500 kgm^{-3}, and we set it to 500 kgm^{-3}.

3. Method of analysis

3.1. *Water balance equation*

In general, the water balance equation in a watershed is as follows:

$$P=Q+E+dS, \qquad (2)$$
$$dS_{\mathrm{water}}=dS_{\mathrm{snow}}+dS_{\mathrm{soil}},$$

where P is the precipitation, Q the discharge, E the evapotranspiration and dS_{water} the storage change of water in a watershed, including both the snow storage change dS_{snow} and the storage change in soil, dS_{soil}. In this study, we have dealt with only the liquid phase of water balance in the watershed. Therefore, the meltwater from snow patches are considered as a liquid water input to the watershed, and negative snow storage change corresponds to snowmelt amount when there is no snowfall event (eq. (2)).

$$P+M=Q+E+dS,$$
$$dS=dS_{\mathrm{soil}}, \qquad (2)'$$
$$M=-dS_{\mathrm{snow}},$$

where dS is the storage change of liquid phase, and M the snowmelt amount.

Snowdrift remains as snow patches in the watershed in summer, and the meltwater runoff from it contributes to the discharge. In order to determine the average amount of snowmelt in the watershed, the area of snow patches is required. We defined a ratio of snow patch area, a, as the ratio of snow patch area to the whole watershed area, and tried to obtain the average snowmelt amount of the watershed. Introducing the ratio, a, average snowmelt amount in the watershed can be expressed as follows:

$$M=am, \qquad (3)$$

where M is the basin average snowmelt amount, and m the snowmelt amount in snow patches.

To estimate areal evapotranspiration and storage change in the watershed is not easy, when compared it with other components. Evapotranspiration depends strongly on surface condition, such as snow patch, wet land or dry land. According to Sand and Bruland (1999), on a snow surface, evaporation and condensation rates balance each other. We assumed that no net sublimation or evaporation from the snow surface. We further assumed that the snow patch exists only on dry land, as stated in Section 2, and the ratio, f_{d}, of evapotranspiration on dry land, E_{d}, to evapotranspiration on wet land, E_{w}, is constant. Thus, evapotranspiration of the watershed becomes as follows:

$$E=S_{\mathrm{w}}E_{\mathrm{w}}+(S_{\mathrm{d}}-a)E_{\mathrm{d}}=S_{\mathrm{w}}E_{\mathrm{w}}+f_{\mathrm{d}}(S_{\mathrm{d}}-a)E_{\mathrm{w}}, \qquad (4)$$

where S_{w} and S_{d} are the ratios of wet land area and dry land area to the watershed area, respectively.

The estimation of storage change is also difficult due to the heterogeneous property of soils in the watershed. The active layer thickness and the storage capacity are small, therefore, the storage change and its estimation error should be small when compared with the other water balance components. The some methods of estimating the storage change are considered. In this analysis, we adopted the method using hydrograph as

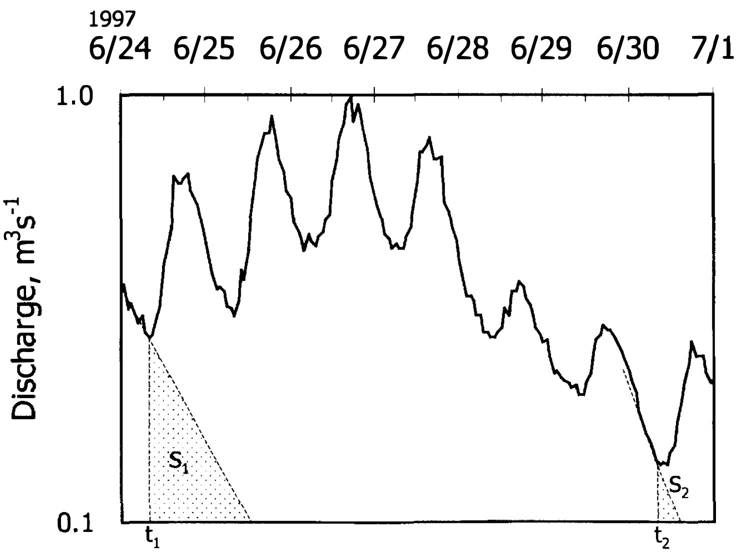


Fig. 2. Estimation of storage change, dS. Each triangle area means storage in watershed at time t. Constant recession coefficient was $\alpha_1 = 9.1 \times 10^6\ s^{-1}$ at t_1 (8:00 June 24) and $\alpha_2 = 11.9 \times 10^6\ s^{-1}$ at t_2 (9:00 June 30). dS was -3.4 mm.

follows. The concept of the storage change estimation is shown in Fig. 2. When there is no input to the watershed, discharge $Q(t)$ at time t decreases exponentially with a constant recession coefficient. According to, after $t=0$, total discharge S can be calculated by the following equation.

$$Q(t) = Q(0)\exp(-\alpha t),$$
$$S = \int_0^\infty Q(t)dt = \frac{Q(0)}{\alpha}. \tag{5}$$

Considering that total discharge S is equal to storage amount in the watershed, storage change dS from t_1 to t_2 can be expressed as follows:

$$dS = S_1 - S_2 = Q(t_2)/\alpha_2 - Q(t_1)/\alpha_1, \tag{6}$$

where S_1 and S_2 are the storage amounts at time t_1 and t_2, and α_1 and α_2 are the recession coefficients at t_1 and t_2, respectively. $Q(t)$ and α at times t_1 and t_2, were determined using the $\log Q$-t graph. A linear recession line was drawn on a decreasing part of discharge. $Q(t)$ is the discharge at the intersection with the discharge and recession line, and α was gradient of recession line. The storage change dS obtained by the above method is the storage which can be discharged later, that is, the 'dischargeable' storage. It is not the total storage of the watershed.

The ratio 'a' is expressed by merging eq. (2)$'$, eq. (3) and eq. (4) as follows.

$$a = Q + dS + (S_w + f_g S_d)E_w - P/m + f_g E_w. \tag{7}$$

In this study, all components and parameters except the ratio a were determined from field data, and the ratio was calculated using eq. (7). In the following section, estimations of the components of water balance and the parameters are discussed.

3.2. *Estimations of water balance components and parameters*

In this analysis, we used rainfall amount P measured by a tipping bucket rain gauge at the meteorological station and discharge Q observed at the hydrological station. Snowmelt amount, m, was estimated by the degree-day method in the watershed. The relationship is as follows:

$$m = 5.9\,T + 10.2, \tag{8}$$
$$m = 5.9\,T, \tag{9}$$

where m is the accumulated snowmelt amount (mm), and T is the accumulated daily mean air temperature (°C). Equation (8) was the relationship in 1997 and eq. (9) was in 1998. These formulas were obtained using the relationship between the snowmelt amount by the snow stake method and the accumulated air temperature (Fig. 3). In 1999, there was little snowpack when we started the observation. Therefore, we used the same relationship obtained in 1998 for the case of 1999 because the observation period of snowmelt amount in 1998 was longer than in 1997 and the relationship of 1998 was considered better.

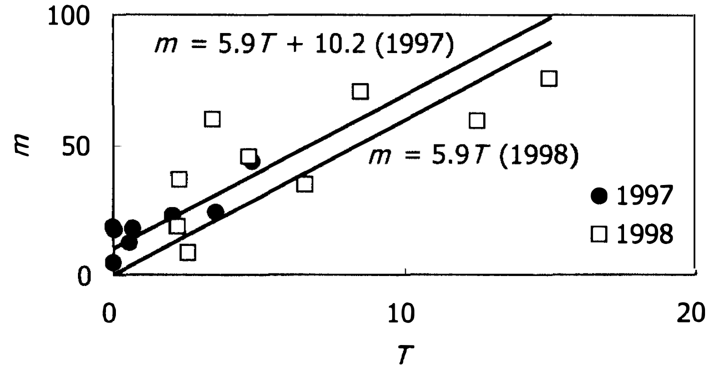


Fig. 3. Estimation of snowmelt amount m. Relationship between accumulated daily mean air temperature, T (°C) and accumulated snowmelt amount, m (mm) in 1997 and 1998.

Evapotranspiration on wet land was estimated from potential evapotranspiration using the Penman method. The relationship between potential and actual evapotranspiration is as follows:

$$E_w = f_p \cdot E_p, \tag{10}$$

where E_w and E_p are the actual evapotranspiration on wet land (mm) and the potential evapotranspiration (mm). The parameter f_p was determined from the relationships between 5 day accumulated E_p and E_w (Fig. 4a). E_p was estimated using the Penman method. The equation used in the Penman method includes net radiation and surface soil heat fluxes. The net radiation was measured by net radiometer. The surface soil heat flux was calculated as the sum of the heat flux at the lower boundary of the surface soil layer, which was measured by a heat flux plate at depth 0.05 m, and the heat storage change in the surface layer between the surface and depth 0.05 m. The heat storage change was estimated from the soil density, soil temperature and the changeable specific

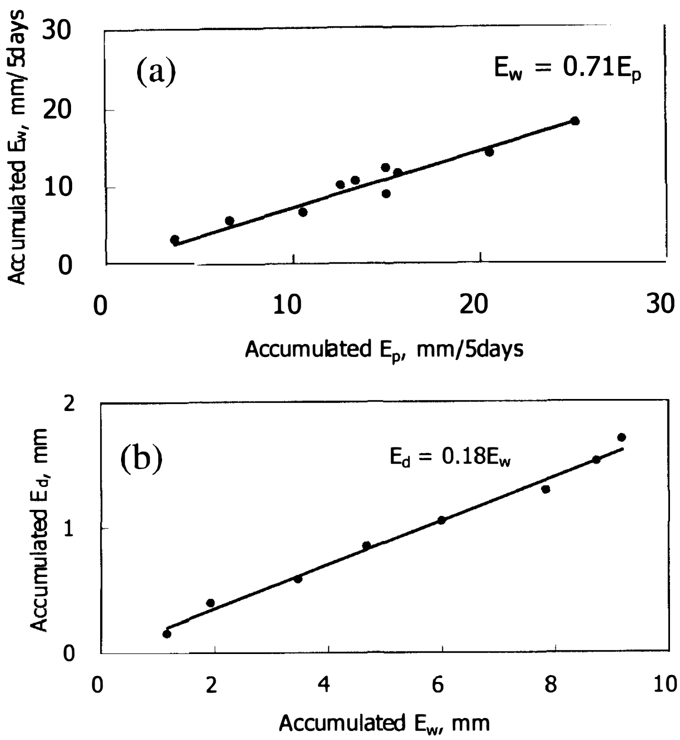


Fig. 4. Estimate of parameter for evapotranspiration. (a) Relationship between E_p and E_w accumulated in 5 days. (b) Relationship between accumulated E_w and E_d.

heat capacity with water content in the layer, which was measured by a TDR sensor. E_w was calculated by the Bowen ratio method using the gradient of air temperature and water vapor pressure at the heights of 1 and 4 m from June to August 1999. In this study, f_p was given as a constant value, 0.71. In order to obtain watershed-mean evapotranspiration, we must know the parameters, S_w, S_d and f_d used in eq. (4). According to Sato *et al.* (2001), S_w and S_d were 0.64 and 0.36 in the snow free period, using a result from the analysis of video camera pictures taken on board of a helicopter. The parameter f_d was obtained from the relationship between accumulated daily evapotranspiration on dry land and wet land from July 29 to August 5 in 1999 (Fig. 4b). Evapotranspiration on dry land, E_d, was obtained by the eddy covariance method using a sonic anemometer and krypton hygrometer, and f_d was 0.18. The storage change was obtained from eq. (6).

In this analysis, we calculated the short term water balance in order to obtain the seasonal change of the ratio of snow patch area. The length of each term of the water balance depends on the estimation of storage change, because the recession line of the hydrograph cannot be drawn when the hydrograph is affected by precipitation. In the snow-free period, an imbalance component X was added to the outputs in eq. (2) because there is no tuning parameter '*a*' for the snow-free period.

4. Results and discussion

4.1. Meteorological and hydrological conditions

Daily mean air temperature is shown in Fig. 5. Air temperature rose above 0°C in June. After the snowmelt runoff started at the beginning of June, the snow pack became patchy, and air temperature rose quickly and substantially. In July, air temperature reached nearly 20°C; in September, it returned to 0°C. Mean air temperature from June 1 to September 15 was 5.9°C in 1997, 5.8°C in 1998 and 8.5°C in 1999. The mean air temperature in 1999 was higher than in other years because snowpack disappeared earlier in early June and air temperature in June was higher than in 1997 and 1998.

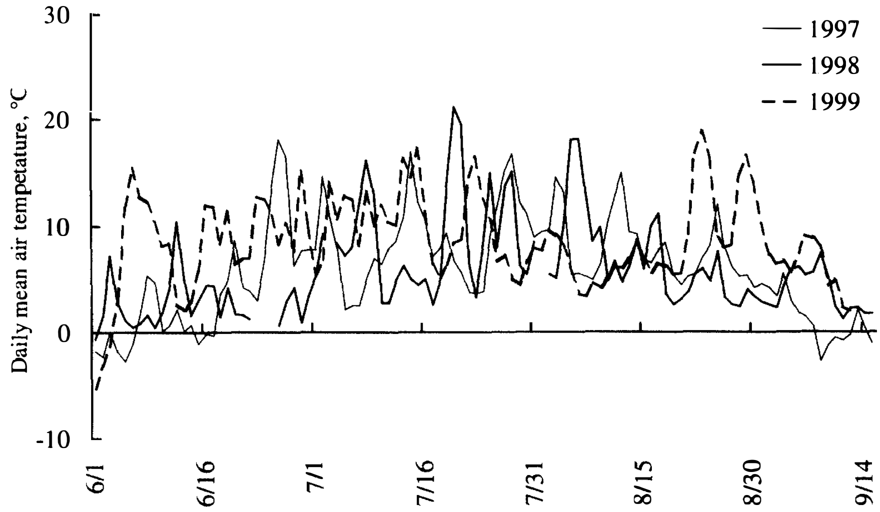


Fig. 5. Daily mean air temperature during summer in 1997, 1998 and 1999.

Precipitation and discharge are shown in Fig. 6 for the period from 1 June to 15 September in 1997, 1998 and 1999. The precipitation in 1997 was larger than in other years. There were remarkable rainfall events in 1997, reaching sometimes above 25 mm day^{-1} (Fig. 6). The precipitation in June and August 1999 was large, but no rainfall continued for 23 days from 12 July to 3 August. The discharge to the stream began a week after the beginning of the snowmelt period in all three years. The discharge in summer varied considerably with rainfall events. The rainfall runoff was characterized by a quick rise and a quick recession. Church (1974) reported the same characteristics for the runoff in Canadian tundra. When precipitation was large and continuous in August 1997 and 1999, base flow of discharge became large (broken line in Fig. 6) and it is kept large even after rainfall event. The precipitation in these months was large.

4.2. Ratio of snow patch area to watershed area

The ratio of snow patch area, a, is shown in Fig. 7. The analyzed period was 82 days from 19 June to 8 September in 1997, 78 days from 18 June to 3 September in 1998 and 76 days from 23 June to 6 September in 1999. The period when snow and ice covered stream beds was excluded from this analysis, because the discharge measure-

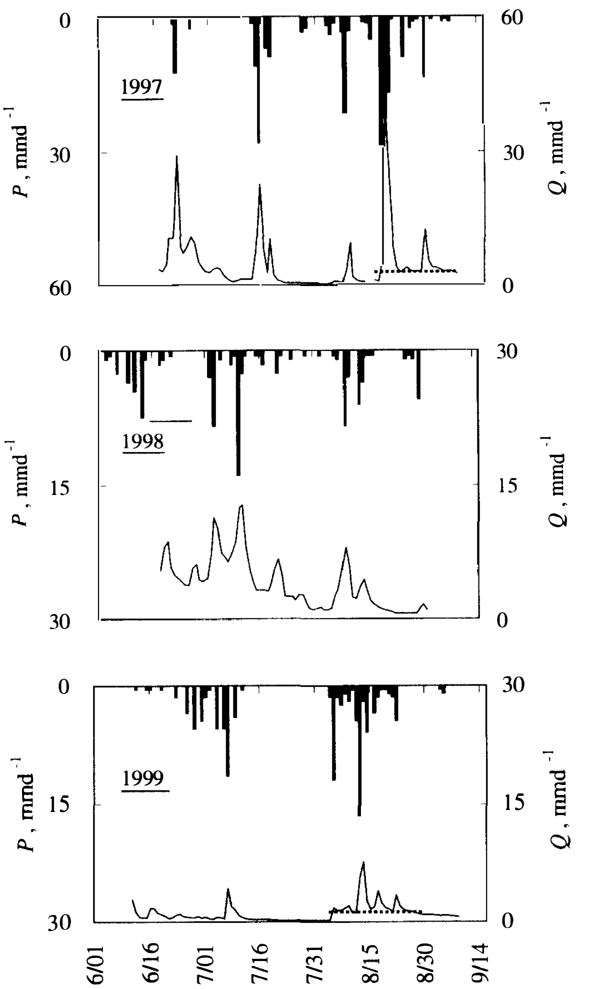


Fig. 6. Precipitation (P) and discharge (Q), from 1 June to 15 September in 1997, 1998 and 1999. Broken line indicates large base flow.

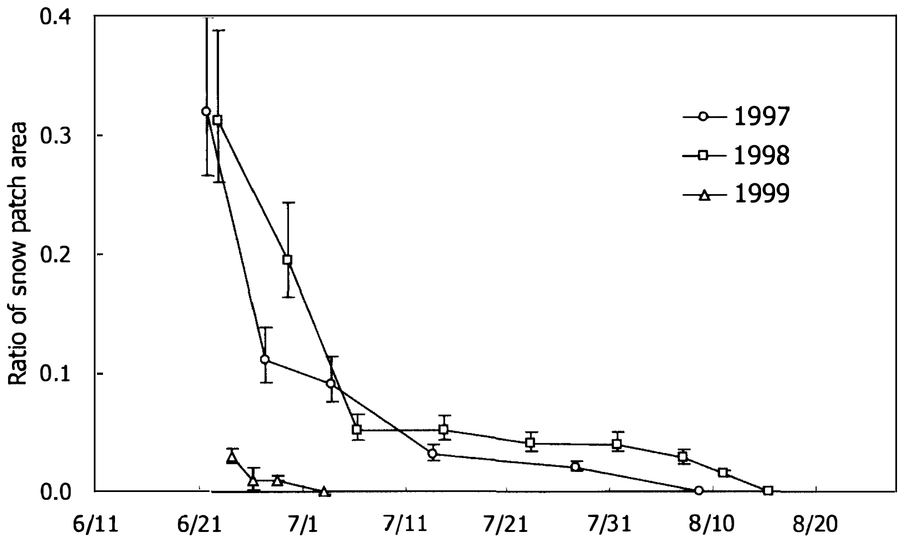


Fig. 7. Seasonal change of ratio of snow patch area 'a' in 1997, 1998 and 1999. Error bars are the ranges when snowmelt amount has ±20% errors.

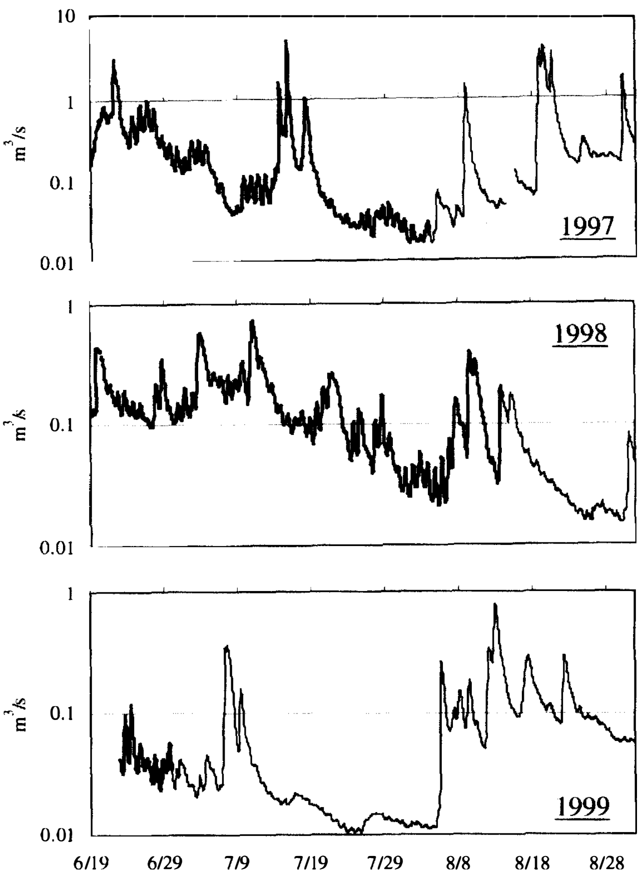


Fig. 8. Discharge from June to August, 1997, 1998 and 1999. Thick line indicates the period when the ratio, a, is positive.

ment was not accurate. From the end of June to the beginning of July in 1997 and 1998, the parameter decreased rapidly, and then it decreased gradually. It became nearly 0 at the end of July or early August. Because the storage as snow before the snowmelt was small in 1999, the parameter was smaller and became 0 earlier in this year than in the other 2 years. When 'a' became 0, diurnal fluctuation of the discharge became small (Fig. 8). Because the daily change of the discharge was unlikely occurred by evapotranspiration, which was true for forested watershed in temperate region, and it depended on the snowmelt from the snow patches, the change of the ratio a indicated the contribution of the snowmelt from snow patches. We have set the value of 'a' to 0 after the periods when the ratio, a, was nearly 0 or negative, that is, 4 August in 1997, 13 August in 1998 and 30 June in 1999 to the end of the observation. Since 'a' is calculated using eq. (7), it contains all errors in the observed and estimated values of the water balance components. This problem will be discussed in the following section.

4.3. Estimation errors

Input amount and output amount of water balance to the watershed are defined as follows:

$$I = P + M,$$
$$O = E + Q + dS, \qquad (11)$$
$$X = I - O,$$

where I and O is the input amount and output amount, respectively, and X the imbalance. In the snow melting period, the output balances with the input, since the ratio 'a' acts as a tuning parameter in the water balance equation. However, during the snow-free period with $M = 0$, the output may not balance the input because the tuning parameter is excluded from the equation. In order to evaluate the estimation error of water balance components, the relationship between the input and the output in the snow-free period is shown in Fig. 9. In this figure, group A includes the events of large continuous precipitation in August and September 1997 and 1999, group B includes for small rainfall event from 10 July to 5 August in 1999, and group C includes the rest. The output of group C is mostly balanced with the input. For group B, the precipitation was small (14 mm in 27 days) with no rainfall for 23 days (Fig. 6). In this period the evapotranspiration was 36 mm and the discharge was 10 mm; therefore, the storage change must be -32 mm to be balanced. However, $+0.1$ mm was obtained for the storage change by the hydrograph analysis (Fig. 2). As mentioned in Section 3.1, the storage calculated here is the dischargeable storage, not the total storage. The evapotranspiration in this dry period with quite small precipitation was as same as the evapotranspiration in the same period in 1997 and 1998 (1–1.5 mm per day in all years, the line with triangle on the right figures in Fig. 10). Therefore, there must be an additional storage change, which supplies the water to the evapotranspiration. When precipitation occurs at a certain interval, recession analysis of the hydrograph for obtaining the storage change could be adequate, however, when precipitation is small and infrequent, and soil is dryer, recession analysis of the hydrograph would not give an adequate storage change estimate. We treat the imbalance as it is and express it as X. In group A, all components of the water balance were larger when compared with group

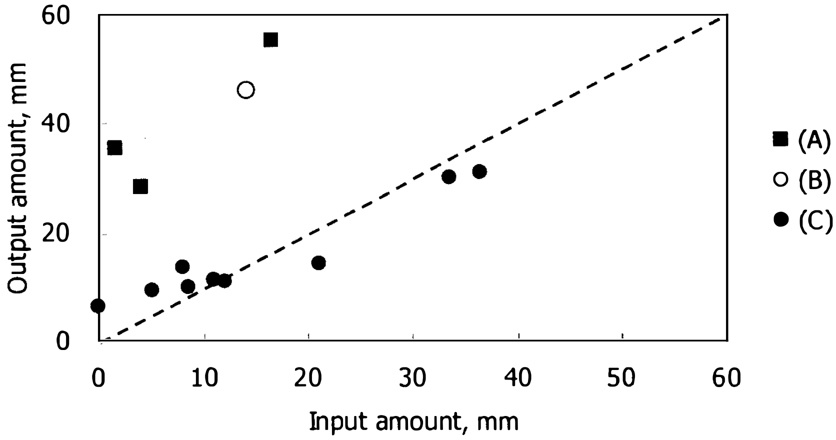


Fig. 9. Relationship between input amount and output amount in snow-free period. (A) for large and continuous rainfall events in August 1997 and 1999. (B) for quite small precipitation from 10 July to 6 August 1999. (C) the others.

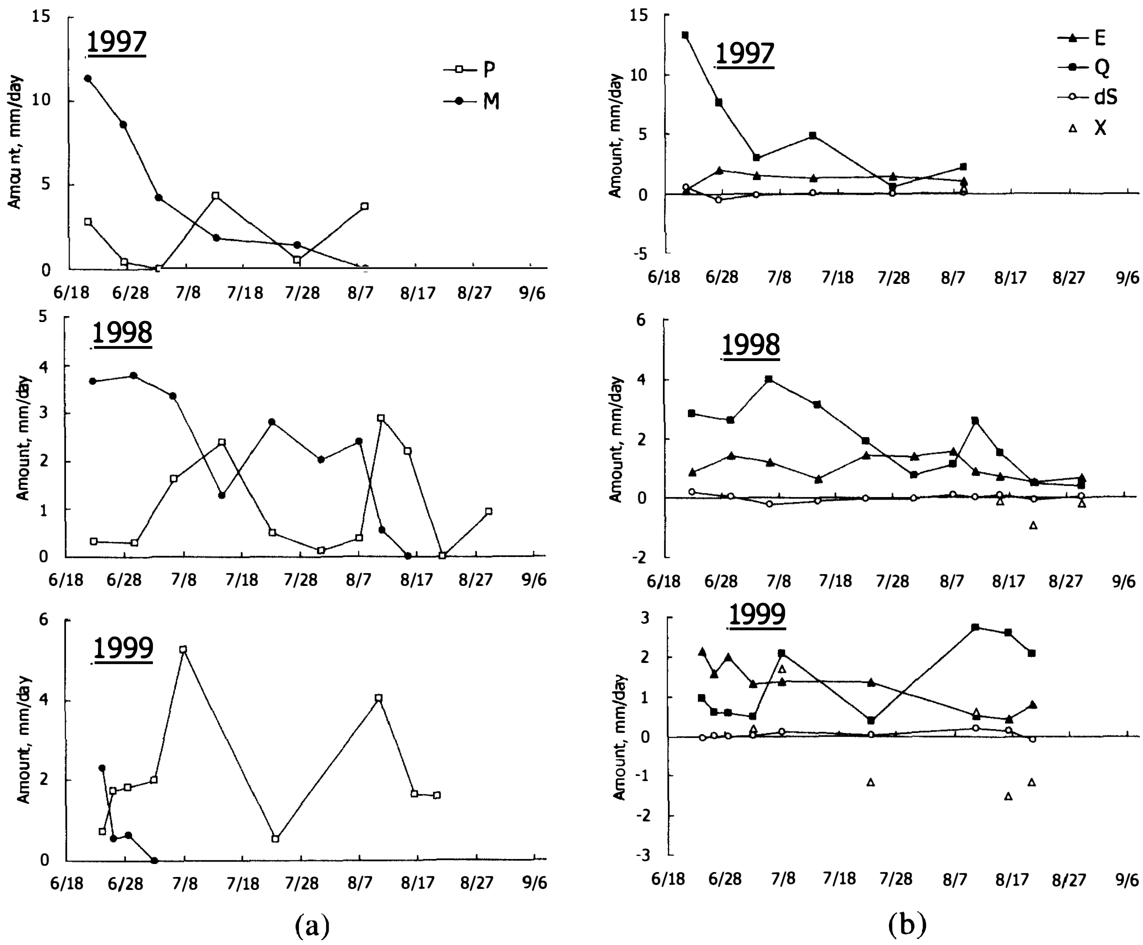


Fig. 10. Seasonal variation of water balance components during summers of 1997, 1998 and 1999. (a) the components of input, (b) the components of output. The evapotranspiration E in this figure means (1 − a)E. The plots indicate the mean value (mm day⁻¹) in each period of short term water balance.

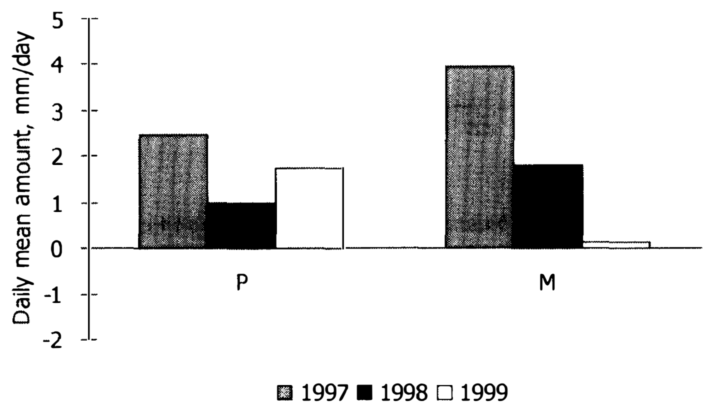
B. The reason for the large imbalance could be sought in the large errors due to the large values of the components. Since we could not specify the reason, we excluded this event from further analysis.

Snowmelt amount m could also contain some errors. The change of ratio of snow patch area 'a' when m contains 20% estimation error is shown as error bars in Fig. 7. The ratio 'a' fluctuated from −0.05 to +0.08 as a maximum for the largest 'a' calculated. As a result, we can conclude that the ratio was not very sensitive to errors of snowmelt amount estimation.

4.4. Seasonal and yearly variation of water balance components

Seasonal variation of the water balance components in 1997, 1998 and 1999 is shown in Fig. 10. Precipitation was smaller at the end of July, and relatively larger at the beginning of July and the middle of August every year than in other season. The snowmelt amount decreased with decrease of snow patch area in the 3 summer seasons; however, for over a month it affected the water balance in the watershed after a major snowmelt period. The discharge is the major factor in the outputs, but in the period with small inputs such as the end of July, the evapotranspiration was larger than the discharge. The variation of evapotranspiration was small, 1–2 mm per day. This result agrees with the results of the observations in the Alaskan Arctic (*e.g.* Kane *et al.*, 1990). The storage change was smaller than the other components, ranging from -0.6 to $+0.8$ mm/day, and it did not change seasonally.

Summer mean amounts of water balance components for the 3 summer seasons are shown in Fig. 11 and Table 2. The yearly changes of the precipitation, snowmelt amount and discharge were large, whereas the yearly change of the evapotranspiration was small. The ratio of discharge to total input amount ($Q/(P+M)$) was larger than the same ratio of the other output components. $Q/(P+M)$ values during the snowmelt period were 0.77 in 1997, 0.68 in 1998 and 0.27 in 1999; and during the snow-free period, Q/P ($M=0$) values were 0.58 in 1997, 0.75 in 1998, 0.66 in 1999. For the whole period, $Q/(P+M)$ were 0.75 in 1997, 0.59 in 1998 and 0.69 in 1999. These results are consistent with the results in Alaskan and Canadian Arctic, where Q/P, called the flow rate, is typically from 0.7 to 0.8 for rainfall events (Anderson, 1974; Findley, 1969; Kane and Carlson, 1973). Since there are no snow patches in their watershed, the total input is P, whereas ours is $P+M$. The ratio of the evapotranspiration to the total input was 0.24 in 1997, 0.36 in 1998 and 0.64 in 1999, and yearly change of this ratio is large due



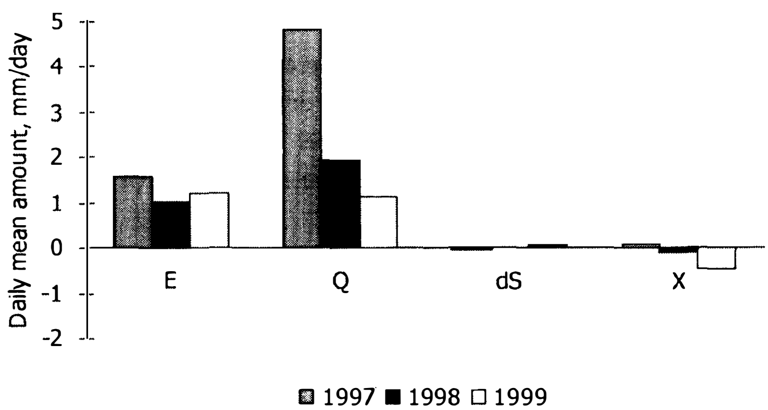


Fig. 11. Summer mean water balance components. The upper figure is for inputs and the lower is for outputs.

Table 2. Summary of water balance during summers of 1997, 1998 and 1999.

Amounts in snowmelt period (mm)

	Period	P	M	E	Q	dS
1997	19 June to 3 August	81	183	63	203	-3
1998	19 June to 13 August	57	139	63	133	-1
1999	23 June to 29 June	11	8	13	5	0

Amounts in snow-free period (mm)

	Period	P	M	E	Q	dS	X
1997	4 August to 12 August	34	-	10	20	1	4
1998	14 August to 2 September	20	-	13	15	0	-9
1999	30 June to 22 August	97	-	59	63	2	-29

Total amounts (mm)

	Period	P	M	E	Q	dS	X
1997	19 June to 12 August	115	183	73	223	-2	4
1998	19 June to 2 September	76	139	77	148	-1	-9
1999	23 June to 22 August	107	8	73	68	2	-29

to large variation in the total input, not in the evapotranspiration. The ratio of meltwater from snow patches to the total input was 0.61 in 1997, 0.65 in 1998 and 0.07 in 1999. Except in 1999, the contribution of snowmelt water to the water balance was large. If the period analyzed was limited to the period when snow patches exist in the watershed, the ratio became larger, 0.69 in 1997, 0.71 in 1998 and 0.42 in 1999.

5. Conclusions

Based on hydrological observations in the eastern Siberian tundra, we have discussed the summer water balance of the tundra watershed. A tuning parameter 'a', which is the ratio of the snow patch area to the watershed area, was introduced in the water balance equation, and seasonal as well as yearly variation of water balance components were discussed. The effect of meltwater from snow patches on the discharge continued in summer. Rainfall and snowmelt amount had a remarkable yearly change. The ratio of snowmelt amount to the sum of rainfall and snowmelt was large, ranging from 7% to 65%. The major output component was discharge; the ratio to the total input was from 59% to 75%. The contribution of evapotranspiration to the input was different from year to year; it changed from 24% to 64%. Storage change contributes little; however, the imbalance was large during the dry season.

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