DELAYING SEASONAL SNOWMELT WITH AVALANCHE ACTIVITY: SOME RESULTS FROM THE CASCADE MOUNTAINS, SOUTHERN BRITISH COLUMBIA

by

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ABSTRACT

Winter-time avalanche activity has the potential to delay a portion of annual snowmelt runoff from high-elevation basins, if enough avalanche transported snow is deposited in sites where it is protected from subsequent ablation. In the Cascade Mountains of southern British Columbia, streamflow after the main snowmelt period in three years from a small (5 km^2) avalanche prone basin is substantially greater than from an adjacent forested, avalanche-free basin. The total excess of discharge over an eleven-week period after the complete disappearance of undisturbed snow cover totals 80,000 m^3 km^{-2} (80 mm) in 1992 and 1993, and 130,000 m^3 km^{-2} (127 mm) over a thirteen-week period in 1994. Despite much lighter than normal avalanche activity in all three winters, the delayed melting of avalanche snow accounts for 18 to 30% of this excess and 12 to 21% of the total streamflow from the avalanche prone basin over these summer periods. Other factors which might play a role in producing these differences in basin yield are also discussed.

INTRODUCTION

The ablation of large avalanche snow deposits during the warm season can be substantially delayed on avalanche paths with a suitable topographic configuration (Iveronova, 1966; Sosedov and Seversky, 1966; Zalikhanov, 1975; de Scally, 1992; 1993; de Scally and Gardner, 1990). Specifically, avalanche snow which has dropped a relatively small vertical distance from the starting zone, and which is protected from ablation in deeply confined track and runout zones, has the potential to linger for weeks after the undisturbed snow cover has disappeared from the highest elevations. Yet, except for some estimates for a relatively large river in the Punjab Himalaya (de Scally, 1992; 1993), there has been no attempt to quantify the effect of such delayed ablation of avalanche snow on streamflows from mountain basins. This effect would be expected to be greatest in high elevation basins where avalanche slopes constitute a significant percentage of the basin area. The purpose of this paper is to report on the contribution that delayed melting of avalanche snow makes to runoff after the main snowmelt period in one such basin in the Cascade Mountains of southern British Columbia, Canada.

STUDY AREA AND METHODS

This ongoing study is being carried out in Manning Provincial Park, located astride the broad crest of the Cascade Mountains approximately 170 km east of Vancouver, British Columbia (Fig. 1). The park's environment is characterized by a sharp transition from wet coastal conditions on the west side to much drier rainshadow conditions on the east side. The average snowpack water equivalent on 25 April (the date of the average annual maximum) is 842 mm at the "Blackwall Peak" snow pillow (Fig. 2), situated at an elevation of 1935 m a.s.l. six kilometres north of the study area. At the "Lightning Lake" snow course situated at the mouth of one of the study basins (elevation 1220 m a.s.l.; S in Fig. 1) the 1 April snowpack water equivalent averages 319 mm (Fig. 2). These stations should be approximately representative of high and low elevations respectively within the study basins. The annual hydrologic regime of the region is typical of nonglacierized, seasonally snow covered mountains, as illustrated by the Similkameen River headwaters inside Manning Park (Fig. 3).

A paired basin methodology is used to study the effect of avalanche snow transport on subsequent snowmelt runoff, since the data requirements for a water balance approach such as employed by Kattelmann and Elder (1991) could not be met. The Frosty Creek basin (5.0 km^2), ranging in elevation from 1220 to 2423 m a.s.l., was chosen as representative of high-elevation basins intensely affected by avalanche activity (Figs. 1 and 4). An immediately adjacent, forested "Control" basin (11.8 km^2), ranging in elevation from 1160 to 2300 m a.s.l., was chosen as representative of the montane-subalpine headwaters of the Similkameen River (Fig. 1). The differences in forest cover and extent of avalanche slopes between the two study basins are shown in Table 1. The forest cover consists mainly of Engelmann spruce and alpine fir with some lodgepole pine. The topographic characteristics of the basins are summarized in Fig. 5.

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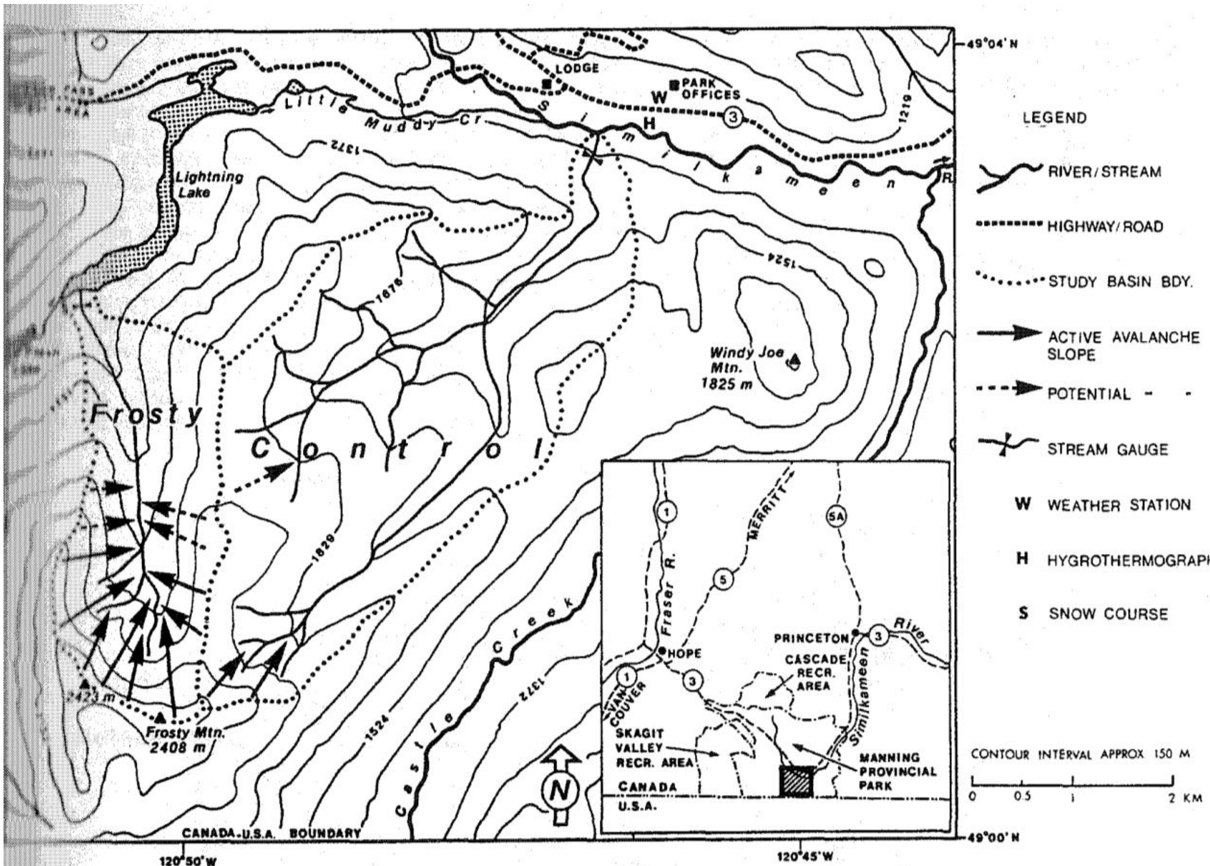

Figure 1. Location of study area.

Table 1

Forest cover and avalanche slope data for Frosty and Control basins.

	Percentage of basin area	
	Frosty	Control
Mature coniferous forest	40	74
Sparse krummholz-type or avalanche-disturbed forest	14	17
Active avalanche slope	43	6
Potential avalanche slope[1]	13	1

[1] Vegetative and other evidence indicates an absence of avalanche activity for several years, but such activity is possible during severe avalanche conditions.

Streamflow has been continuously recorded from early April to October at the mouth of both study basins (Fig. 1) for three years (1992-94). Eight-hourly stage measurements, made using float-type water level recorders housed inside stilling wells, are converted to discharge using exponential or logarithmic rating equations for each stream reach (Frosty R^2 = 0.91-0.99; Control R^2 = 0.98). The minimum stage resolution that could be obtained from the charts represents between two and four percent of discharge in the creeks over the full range of flows measured during the study periods. The eight-hourly instantaneous discharges are used to calculate streamflow totals in Frosty and Control Creeks, and after averaging to obtain mean daily values, are also used for graphing instantaneous unit-area discharges. In order to compare streamflows from the study basins to the timing of regional snowmelt runoff, unit-area discharge from the upper Similkameen River basin (407 km^2) is included on these graphs.

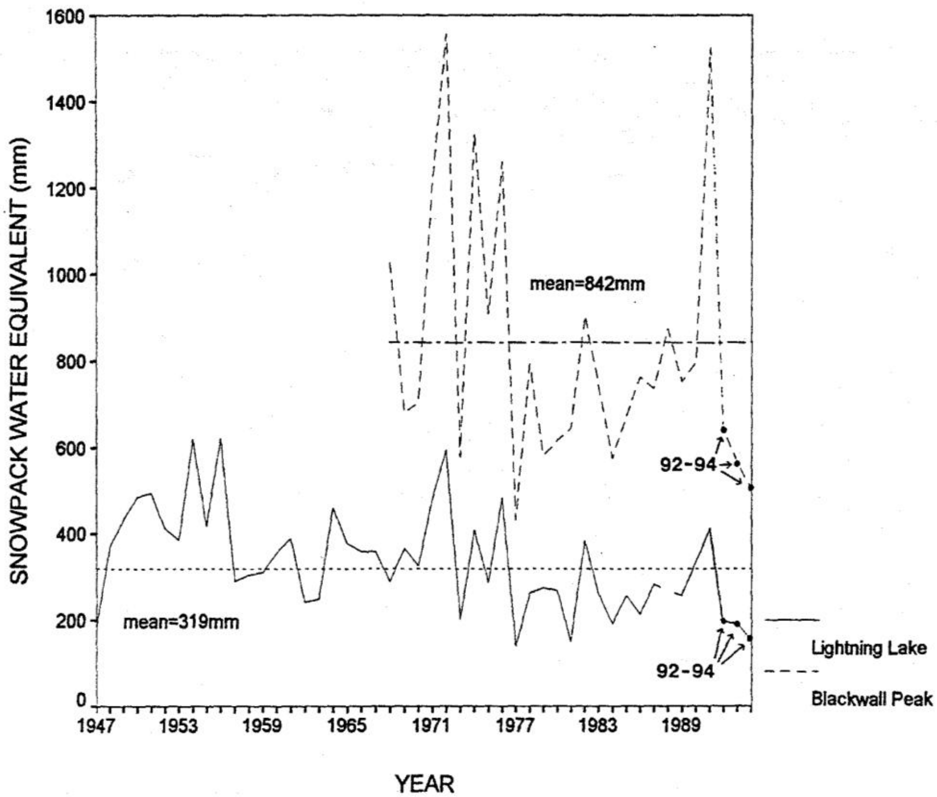

Figure 2. Snowpack water equivalent on 25 April at "Blackwall Peak" (no. 490042DE) snow pillow and 1 April at "Lightning Lake" (no. 3D02) snow course. (Data from B.C. Ministry of Environment, Lands and Parks.)

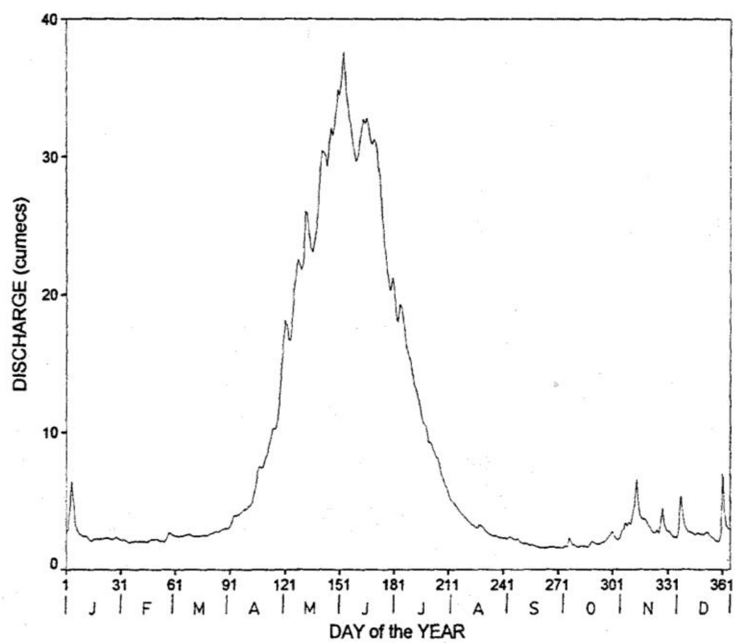

Figure 3. Mean daily discharge, Similkameen River above Goodfellow Creek (no. 08NL070), Manning Park, 1974-93. (Data from Water Survey of Canada).

The volume of avalanche transported snow in the Frosty basin has been surveyed each spring for comparison with the above streamflow totals. The surveys were carried out as close as practicable to the date of disappearance of the undisturbed snow cover in both study basins (Table 2). This was done to ensure that the effect of delayed ablation of avalanche snow could be isolated from differences in the pattern of normal snowmelt between the two basins, and because of considerable difficulties of access into the upper Frosty basin (Fig. 4). Most avalanche deposits were surveyed by hip-chaining and probing in order to derive cross-sections (or on smaller deposits an average depth) and surface area, from which the snow volume could be calculated. The volume of inaccessible deposits could only be derived by measuring their approximate surface area from photographs and estimating the average depth (usually only one or two metres). For conversion of the deposit volumes to water equivalents, measured or estimated densities of spring-time avalanche snow are employed (Table 2). The latter are averages based on figures from a variety of sources (Schaerer, 1967; 1988; Martinec and de Quervain, 1975; Zalikhanov, 1975; de Scally and Gardner, 1990; McClung and Schaerer, 1993). The range of error arising from using such average densities should not exceed eight percent, and is therefore smaller than errors in some of the surveyed snow volumes.

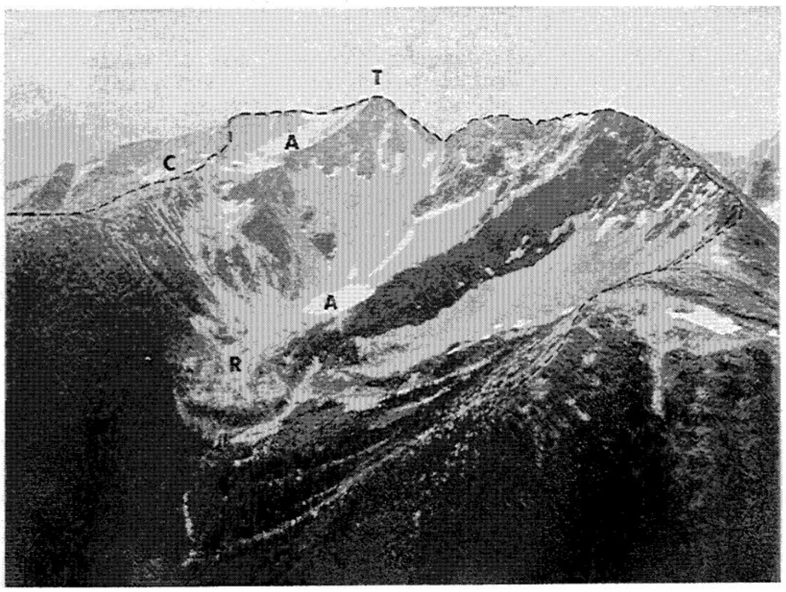

Figure 4. Upper Frosty basin viewed from the north-northwest on 22 July 1994. T: "true" summit of Frosty Mountain. A: largest avalanche deposits surveyed in all three years. R: rock glacier. C: upper part of Control basin.

RESULTS

Snow accumulation was substantially below average in all three winters during this study (Fig. 2), with correspondingly small amounts of avalanche activity in the Frosty basin compared to activity observed in previous years. In the summer of 1992 and 1993 the undisturbed snow cover had disappeared shortly before the avalanche deposit surveys, while in 1994 it is estimated to have disappeared about 18 days after the surveys. A substantial proportion of the total avalanche snow volume in each year (Table 2) is stored in the two large deposits marked A in Fig. 4. The volume of avalanche snow in the Control basin each year was insufficient to have any measurable effect on streamflows during the snowmelt period.

Fig. 6 shows the unit-area discharge of Frosty and Control Creeks and Similkameen River in 1992, 1993 and 1994. Data for Frosty Creek from 5 to 18 May (day 126 to 139) 1992 are missing due to a malfunctioning stage recorder, and gaps exist in the preliminary 1994 record for Similkameen River. However, it is clear that the discharge of Control Creek follows that of Similkameen River quite closely in all years. The exceptions occur from early to mid-May when the Control basin is still largely snow covered while extensive valley-bottom areas of the Similkameen basin are already free of snow, allowing the former to generate larger peaks in snowmelt runoff. Prior to late May in all years, peaks in snowmelt runoff in Frosty Creek are generally smaller than in Control Creek. However, from late May until September or October, the discharge of Frosty Creek is continuously higher compared to Control Creek and Similkameen River.

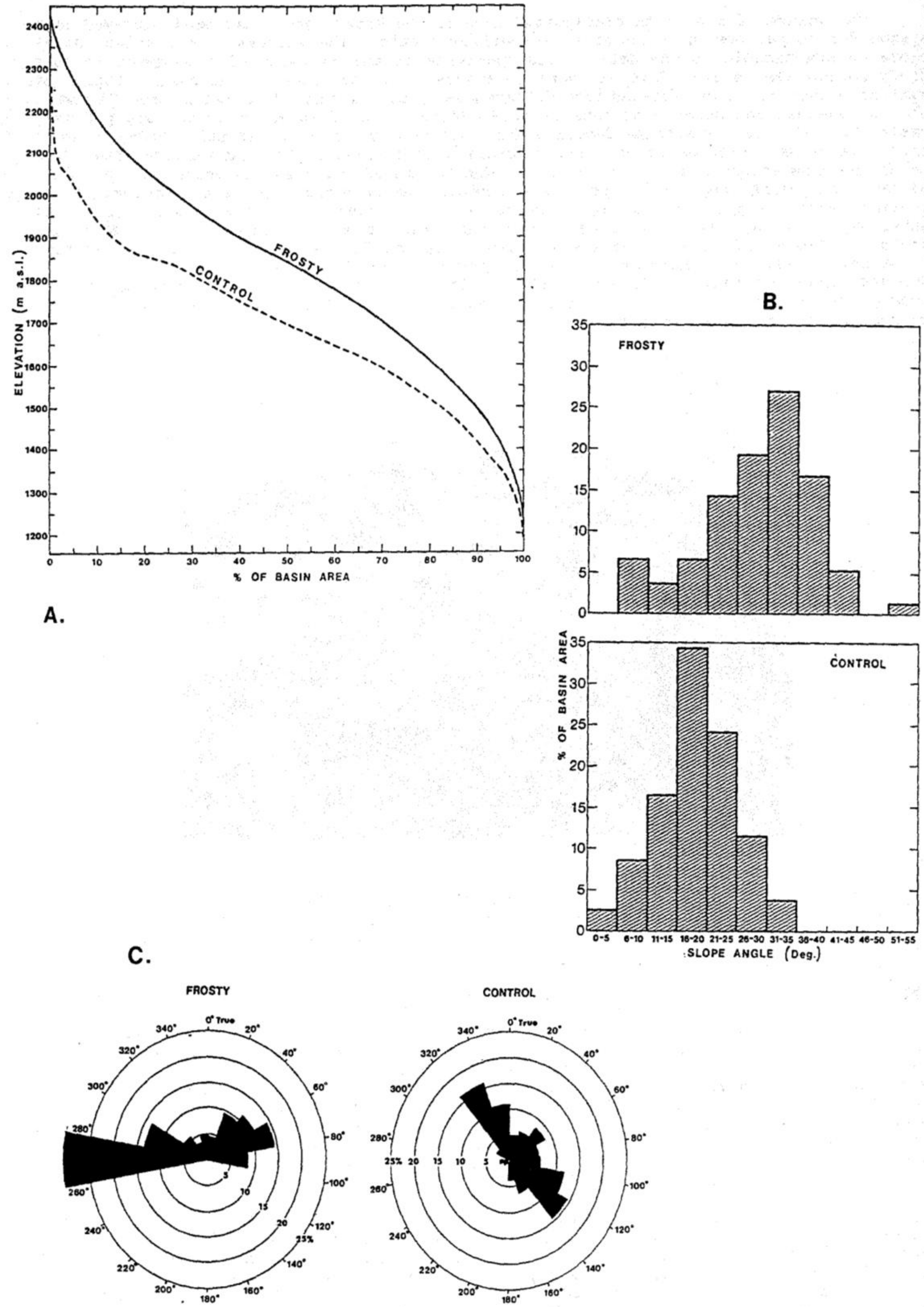

Figure 5. Topometric characteristics of Frosty and Control basins. A: hypsometry. B: slope aspect determined from 250 m by 250 m map grids. C: slope gradient calculated between highest and lowest contours in 250 m by 250 m map grids.

Table 2

Dates of avalanche deposit surveys and measured volumes of avalanche snow, Frosty basin.

Date of surveys	Avalanche snow density ($kg\ m^{-3}$)	Total volume of avalanche snow (m^3 water equivalent)
4 July (Day 186) 1992	616[1]	73,000
24 June (Day 175) 1993	620[2]	120,000
24 June (Day 175) 1994	570[2]	187,000

[1] Average of six measurements taken on the largest deposit (std. error = 16 $kg\ m^{-3}$)

[2] Estimated on the basis of published densities for avalanche snow of similar age and character; see text.

Prior to the dates of the avalanche deposit surveys (Table 2 and Fig. 6), the differences between Frosty Creek discharge and Control Similkameen discharge are largely the result of a later peak in normal snowmelt in the former basin. At this time it is impossible to isolate any role the differential ablation of avalanche snow might play. However, after the surveys the measured volumes of avalanche snow (Table 2) can be compared to streamflows since the surveys were carried out around the time of disappearance of the undisturbed snow cover. Table 3 shows that following the surveys there is still a substantial excess of streamflow from the Frosty basin compared to the Control basin in all years. Over the duration of this excess (11 weeks in 1992 and 1993, 13.5 weeks in 1994), the water equivalent volume of avalanche snow represents 11 to 21% of the total discharge of Frosty Creek and 18 to 30% of the total excess of discharge over Control Creek. In 1994, when it was possible to fix the approximate time period from the disappearance of the undisturbed snow cover to the disappearance of all avalanche snow, the volume of avalanche snow represents 33% of the total discharge of Frosty Creek and 47% of the total excess of discharge over Control Creek. These 1994 figures may be somewhat overestimated, since they neglect the ablation of avalanche snow during the 18-day period between the avalanche deposit surveys and disappearance of the snow cover, and assume that evaporation and sublimation losses from the avalanche snow are negligible. The latter assumption generally is supported by the results of other studies (Sosedov and Seversky, 1966; Lang, 1981; de Scally and Gardner, 1990).

DISCUSSION

The period of peak snowmelt occurs later in the Frosty basin compared to the Control and upper Similkameen basins as a result of the higher elevations and other topographic characteristics of the former (Fig. 5). As an example, ablation on the extensive west-facing slopes of the Frosty basin (Fig. 5b) is reduced by the fact that for a significant portion of the day these slopes are in the shadow of the Frosty Mountain summits. The effect of the extensive forest cover in the Control basin (Table 1), which would help to protect the snow cover from ablation, appears to be outweighed by the topographic characteristics of the Frosty basin. Furthermore, snow accumulation appears to be greater in the Frosty basin, although there are no data available to confirm this observation. The greater accumulation may be the result of regional precipitation variations across the Cascade Mountains, less forest cover in the Frosty basin, and the orientation of the basin relative to prevailing southeasterly to southwesterly winter winds. The presence of cornices each spring suggests that these winds may transport significant quantities of snow into the upper basin. This wind transported snow certainly plays an important role in loading avalanche starting zones below the ridge crests (Fig. 4).

The contributions made by melting avalanche snow to streamflows in Frosty Creek after the snowmelt period (Table 3) follow three winters of light avalanche activity. Even when possible evaporation and sublimation losses are accounted for, the importance of this melting in generating summer runoff may therefore be substantial following winters with severe avalanche cycles. This could especially be true if, despite widespread avalanching, the snowpack at the end of the winter is relatively shallow and little rain falls during the following summer.

The persistence of avalanche deposits in the Frosty basin is somewhat unusual given the unconfined nature of most of the avalanche paths. The runouts of these paths are mostly located on open talus slopes (Fig. 4). Following avalanching of snow to lower elevations in such situations, the increase in ambient air temperature would be expected to outweigh the effects of any concentration of the snow, leading to accelerated ablation (Martinec and de Quervain, 1975). The vertical falls of avalanches in the Frosty basin (maximum on the order of 500 m) do not appear to be sufficiently large for this to occur

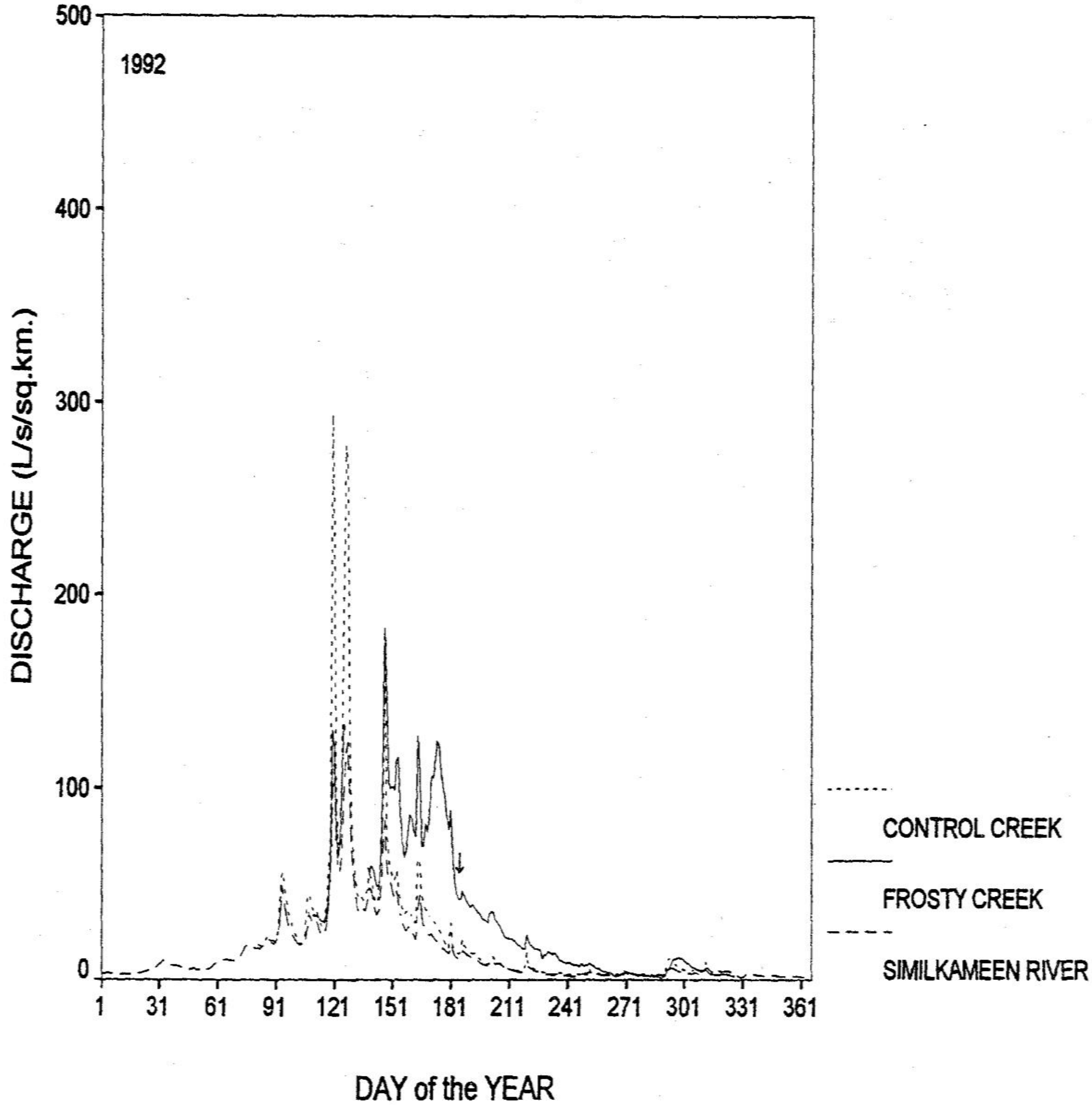

Figure 6. Daily unit-area discharge of Frosty and Control Creeks and Similkameen River, 1992-94. Arrows in 1992 and 1993 graphs indicate dates of avalanche deposit surveys (4 July 1992, 24 June 1993). In 1994 graph, 1 = avalanche deposit surveys (24 June), 2 = disappearance of undisturbed snow cover (12 July), 3 = disappearance of all avalanche snow (20 August). (Similkameen River data from Water Survey of Canada.)

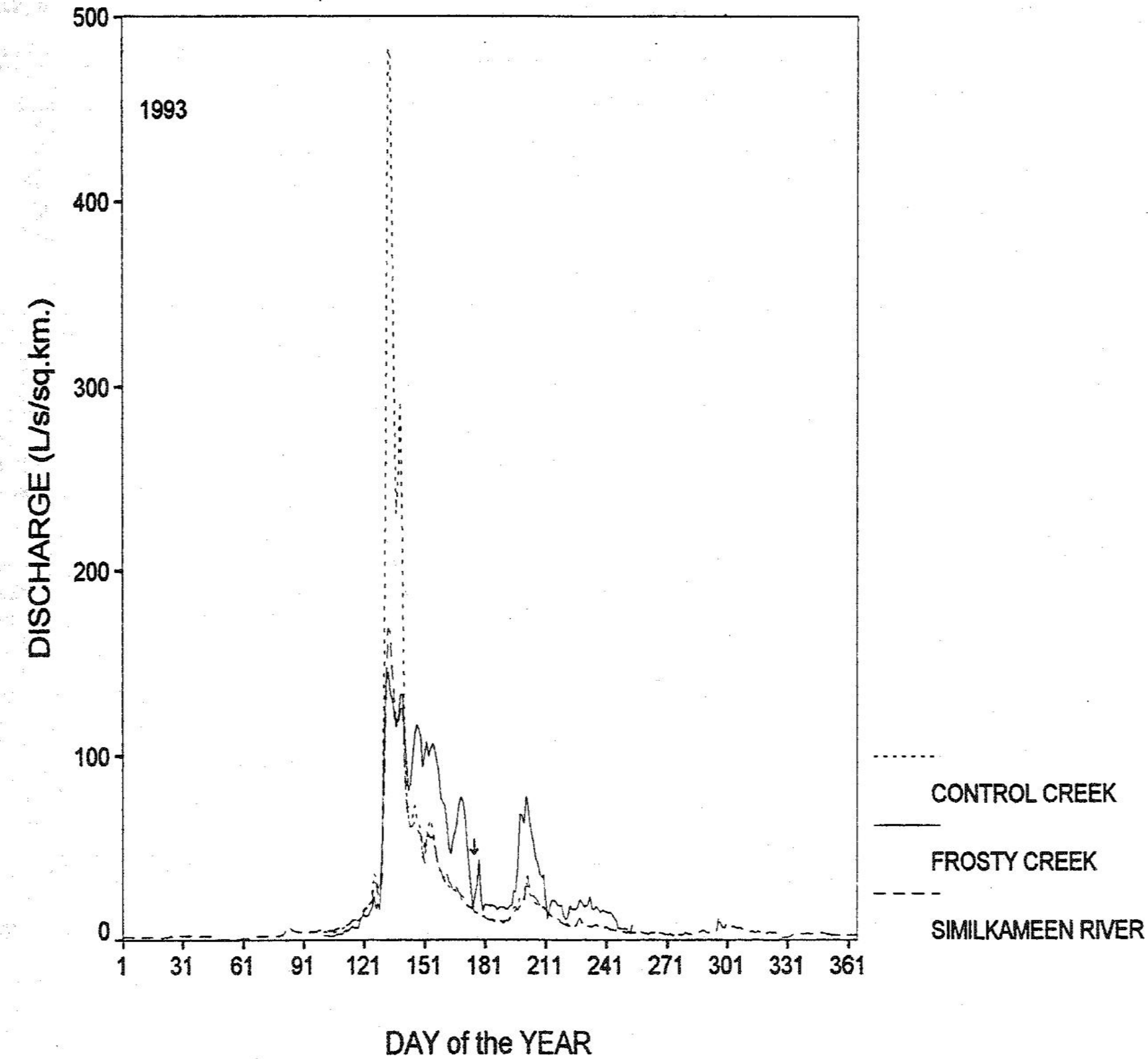

Figure 6. Continued

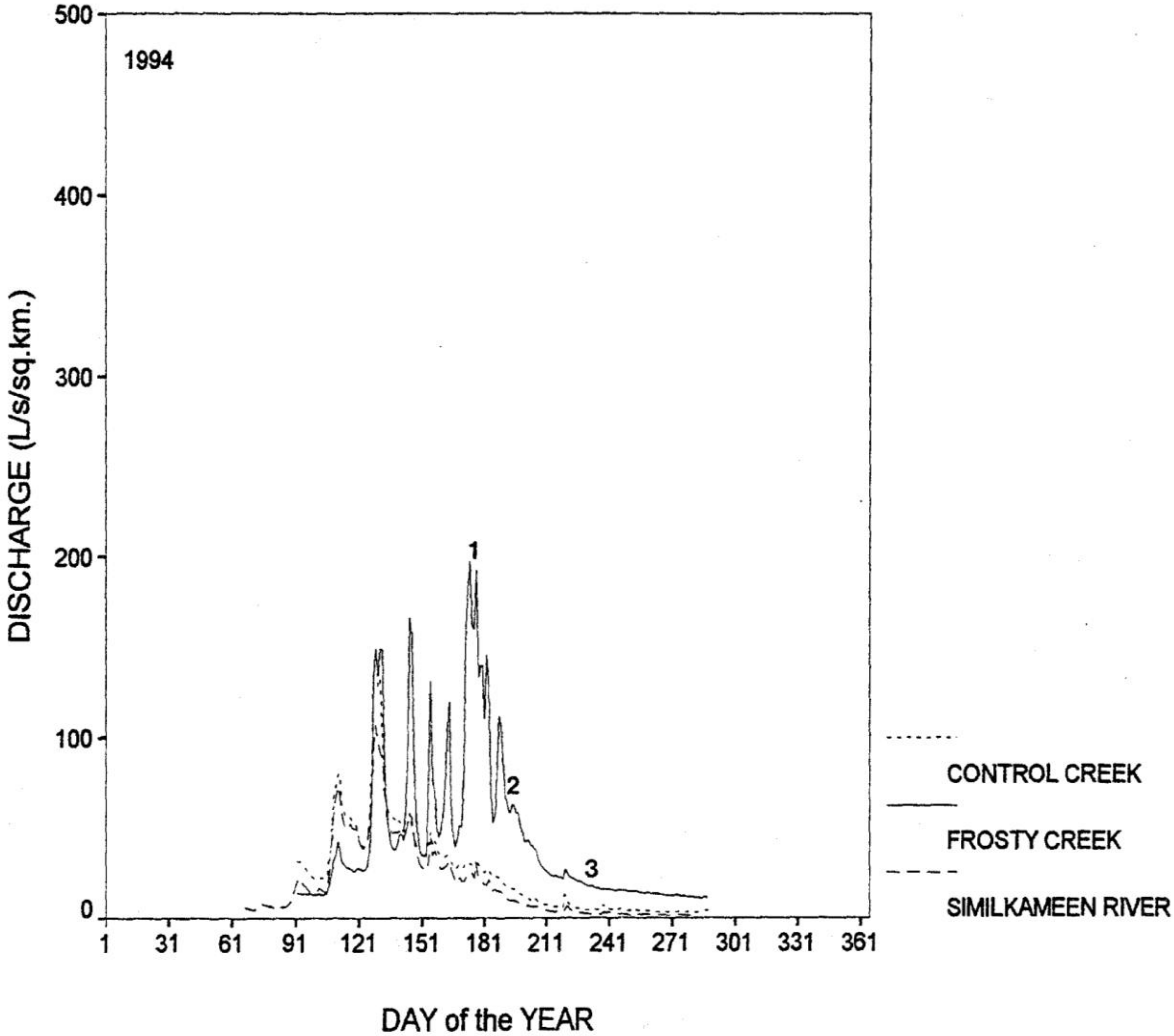

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Table 3

Streamflow totals and differences for Frosty and Control basins following the avalanche deposit surveys. Percentages in brackets are the proportions derived from avalanche snow, assuming all of the snow melts and ends up as streamflow.

Time period after avalanche deposit surveys	Total discharge		Total excess of Frosty Creek discharge over Control Creek discharge	
	Frosty (mm)	Control (mm)	(mm)	(m^3 km^{-2})
04/07/92 → 20/09/92	127 (11%)	46	81 (18%)	80,000
24/06/93 → 08/09/93	161 (15%)	81	80 (30%)	80,000
24/06/94 → 14/10/94[1]	351 (11%)	90	261 (14%)	260,000
12/07/94[2] → 14/10/94[1]	178 (21%[3])	51	127 (29%[3])	130,000
12/07/94[2] → 20/08/94[4]	112 (33%[3])	31	81 (47%[3])	80,000

[1] End of streamflow record; excess in Frosty Creek discharge continues beyond this date (Fig. 6).
[2] Estimated date of disappearance of all undisturbed snow cover.
[3] May be overestimated since the melting of avalanche snow from 24/06/94 to 12/07/94 is ignored.
[4] All avalanche snow gone from Frosty basin.

to any great degree. Moreover, many of the avalanche deposits are partially protected from ablation by being located on steep north- or northeast-facing slopes in the shadow of high cliffs (Figs. 4 and 5b,c). Small remnants of avalanche deposits were observed to be present in high-elevation gullies as late as mid-September (late August in 1994), though not in sufficient quantities to explain the streamflow differences in Fig. 6. An important question that therefore needs addressing, and which this research has so far not answered, is the exact timing of streamflows generated by melting avalanche snow after the main snowmelt period.

Greater transpiration losses from the Control basin appear to be important in explaining the Frosty-Control streamflow differences that are not accounted for by delayed melting of avalanche snow (Table 3). The effect of transpiration on streamflows from similarly forested mountain basins in southern British Columbia is demonstrated by Cheng (1989) and Doyle (1991). Using 1 mm d^{-1} and 3 mm d^{-1} as representative of the range of transpiration rates possible in the study area during summer (Lang, 1981; Doyle, 1991), the difference in total transpiration between the two basins solely as a function of the difference in area of mature forest cover (Table 1) would account for 6800 to 20,000 m^3 d^{-1} of the streamflow differences. The highest transpiration rates are probably achieved late in the main snowmelt period due to high levels of soil moisture and atmospheric energy (Swanson, 1967). However, a stepped pattern indicative of a diurnal transpiration cycle is common in depletion curve portions of the water level record from Control Creek even late in the summer. The same pattern is rarely present in Frosty Creek and then only weakly. A higher runoff coefficient for the Frosty basin may also partly explain the greater streamflows in summer. The best evidence for this occurs during day 194 to 210 in 1993 (Fig. 6), when almost daily rainfall coincides with a substantially higher unit-area discharge in Frosty Creek. An inactive rock glacier in the upper Frosty basin (Fig. 4) can probably be discounted as a source of summer meltwater given its low elevation, and large thermokarst-type hollows on its surface indicative of an absence of internal ice. Even if relict ice did remain inside the feature, the meltwater contribution from it should be negligible.

CONCLUSION

The results of this study demonstrate that, even following winters with light avalanche activity, meltwater from late-lying avalanche deposits can account for up to 20% of the total streamflow from a small avalanche prone basin over a period of 11 to 13 weeks following the normal snowmelt period. More extensive avalanching may therefore have the potential to substantially delay streamflows generated by seasonal snowmelt in such basins. The optimal topographic configurations required for delayed ablation of avalanche deposits are known fairly well: generally north-facing avalanche paths with relatively small vertical falls and steep runout zones, and a tendency to produce concentrated snow deposits with a small surface area-to-volume ratio. Deeply gullied track and runout zones are particularly efficient in the last respect. However, this research has not yet been able to estimate the exact timing of basin streamflows produced by melting avalanche snow. Such estimation is complicated by the highly heterogeneous depths of avalanche deposits, and lack of information about the distribution and size of deposits in larger mountain basins.

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