Rate of floodplain reworking in response to increasing storm-induced floods, Squamish River, south-western British Columbia, Canada

Gregory D. Bauch* and Edward J. Hickin
Department of Geography, Simon Fraser University, Burnaby, Canada

ABSTRACT: Relations among hydroclimatic and channel planform changes on Squamish River are presented for the period 1956–2007. Squamish River basin occupies 3600 km² of mountainous terrain in south-western British Columbia, about 50 km north of Vancouver. The magnitude, volume and duration of extreme floods (Q ≥ 1500 m³/s) exhibit respective temporal increases of 50, 450 and 300%. The increase in extreme floods is attributed to the intensification of late-season (August–December) Pacific storms that have produced increases in precipitation amounts, intensity and duration of respectively 340, 200 and 200% over the same period. Changes in floodplain-surface area calculated from the geographic information system (GIS) differencing of sequential large-scale aerial photographs indicate that the rate of geomorphic change in Squamish River has accelerated during the 1980s to the mid-1990s. Among four study reaches of varying planform, erosional, depositional and cumulative changes in floodplain surface-area have rapidly increased. Channel-change activity after 1980 has increased by a factor of two to six compared with the period prior to 1980. Erosion is currently outpacing deposition in the majority of study reaches. Although channel geometry generally exhibits no uniform pattern of response to the increase in extreme floods, the meandering reaches have straightened over the duration of the study period. The increase in the magnitude and duration of the annual flood appears to be the principal cause of this recent acceleration of channel change. Copyright © 2010 John Wiley & Sons, Ltd.

KEYWORDS: channel change; climate change; flood regime

Introduction

The morphology of river channels is determined in part by the character of their channel-forming flows that in turn are driven by hydroclimatic forcing. Furthermore, it is generally accepted that a stationary hydroclimatic regime is the necessary basis of equilibrium channel morphology.

But what if the hydroclimatic regime is changing? Will such changes be reflected in the morphodynamics of rivers in the region?

In this context, one of the most compelling recent environmental concerns is the character and significance of global climate change on a decadal to century timescale. It is generally agreed that, in many parts of the world, climate is exhibiting change although the reasons for it remain hotly debated.

In western Canada there is evidence that warming over the last half century has occurred (Danard and Murty, 1988; Gan, 1992; Gullet and Skinner, 1992; Koshida and Avis, 1998; Whitfield and Cannon, 2000) and that other related climate change is underway (Moore et al., 2002; Cunderlik and Ouarda, 2009). Physical reasoning suggests that, because there has been a general increase in thermal energy available in the ocean-atmosphere (Cunderlik and Ouarda, 2009), the intensity of local atmospheric circulation (especially as it is expressed in storm frequency, duration and intensity) may be increasing, trends supported by climate modeling for British Columbia (Whitfield et al., 2002; Whitfield et al., 2003). Indeed, the amount of precipitation being delivered to the Canadian west coast is increasing and the temporal pattern of its delivery is changing (Danard and Murty, 1988; Gan, 1992).

Increasing precipitation implies related changes in river hydrology (Zhang et al., 2001) and therefore potential changes in the intensity of geomorphic activity in the rivers of the region. This climate/hydrology/geomorphology linkage in general has been the subject of recent technical discussions and conference themes in geomorphology (Pike et al., 2008a, 2008b) but there remain very few quantitative field studies to assess the nature and significance of this line of deductive reasoning about environmental dynamics.

The purpose of this study is to describe, for a large coastal drainage basin in British Columbia near Vancouver, for the last half-century (a) the temporal pattern of the storm climate, (b) shifts in the formative discharge of a major river (Squamish) in the region, and (c) the related changes in river channel morphology exhibited by Squamish River. This study is in part an extension and updating of an earlier record of Squamish River environmental change to 1984 documented by Sichingabula...
The Field Area

Squamish River basin was chosen as the test area for this study because it is a large (3200 km²) representative drainage area in the Coast Mountains of south-western British Columbia. Although it is very accessible, located just 40 km north of Vancouver (Figure 1), it remains in a relatively pristine state. The river and its basin have been the subjects of numerous scientific studies over the previous several decades so that the environmental data-base for the area is extensive.

Squamish River rises in the Tantalus Ranges in the north-west and drops approximately 2700 m to its sea-level terminus in the fjord-head of Howe Sound. It is a high-energy multi-planform river composed of braided, structurally confined and single-channel meandering reaches. Inter-planform transitions are in places abrupt and in others gradual. Most of the 150 km-long coastal river is flanked by steep mountain slopes composed primarily of granitic rocks with some gneiss and schist (Woodsworth, 1977) formed during the Late Cretaceous (Matthews, 1958). Much of the eastern side of the valley is formed in intermediate volcanics associated with a series of now dormant volcanoes. About 640 km² (about 20%) of the drainage area is glacier covered (Paige and Hickin, 2000). Although Squamish River is free of man-made sediment input and channel geometry. Minor (~3 km) dyking of Squamish River is present in the estuary.

Sediment sources of Squamish River are moraine veneers situated on the steep mountain slopes (Brooks and Hickin, 1991) as well as vesicular volcanics forming Mount Cayley and Mount Garibaldi on the eastern side of the basin. Mass wasting events such as debris avalanches and debris flows originating from weathered vesicular volcanic andesitic lavas and tuff also produce significant but temporarily localized sediment supply inputs. Hickin (1989) estimated from Squamish delta progradation rate that 1.29 × 10⁶ m³ of Squamish River sediment is deposited in Howe Sound in an average year.

Wet mild winters and dry warmer summers characterize the maritime climate of Squamish River Valley. Monthly average temperature ranges from 0 to 18°C. As a result, annual snowfall only accounts for approximately 11% of the total precipitation at lower elevations (46 m above sea level (a.s.l.); Environment Canada climate stations, 2007). The bulk of snowfall accumulates between December and February and usually melts by the end of July.

The majority of the precipitation captured by Squamish River basin is generated by cyclonic frontal systems that typify the fall and winter months (August–December). They develop over the North Pacific Ocean and track eastward (Loukas et al., 2002) or south-southwest (Loukas and Quick, 1993). The north/south orientation of Squamish basin makes it an efficient interceptor of precipitation events. Rainfall intensities recorded for fall events commonly exceed 40 mm/day with larger storms producing rainfall intensities greater than 100 mm/day. Approximately 75% of all precipitation is delivered during October–March.

Vegetation of Squamish River basin reflects the location within the Western Hemlock bioclimatic zone. The abundant moisture of the Pacific Coast supports dense growth of conifers and deciduous trees and most of the basin below the treeline is forested.

The annual discharge regime of Squamish River is melt-water driven with precipitation events producing localized spikes in the annual hydrograph, especially during the fall season (Figure 2). At the village of Brackendale the average annual discharge of Squamish River is 240 m³/s and flow typically ranges from an average-monthly low of 100 m³/s from (December–February) to a high of 500 m³/s in August. Intense and prolonged storm events can produce floods with a mean daily discharge 20 times greater than that of average seasonal baseflow, often exceeding bankfull stage (1216 m³/s).

The character of channel changes was documented for four, non-braided study reaches (Figure 1) first outlined by Sichingabula (1986). We use Sichingabula’s original naming and designation protocol for each reach: Squamish-Mamquam Reach (A); the Lower Meandering Reach (B); Upper Meandering Reach (C); and Squamish-Ashlu Reach (D). Our channel-change observations, from 1984 to 2007, are a continuation of Sichingabula’s original study period (1947–1984). Thus, the morphologic record of Squamish River now spans 60 years.

Data and Analytical Methods

Hydroclimatic data

Data consisting of river discharge, precipitation and temperature records, were used to evaluate possible shifts in hydroclimate. These records (mean daily discharge, daily precipitation and mean daily temperature) were assembled from the archives of Water Survey of Canada gauging stations and Environment Canada climate stations. From these compiled data sets, continuous daily records for all hydroclimatic variables were produced for the years 1956–2006. Gaps within records were filled by interpolating daily values using linear regression equations derived from relationships between the main (primary) and secondary stations. Regression estimates between stations require at least two years of daily data recorded for the same period of time by each station and coverage of a secondary station must overlap the gap in the primary record. Variance (R²) values for these relations indicate strong statistical associations between the primary and secondary data sets for all hydroclimatic variables (R² > 0.92). Regressed daily values represent approximately 15% and 45% of the total daily data for the hydrometric and climatic records respectively. The locations of the primary gauging and climate stations are shown in Figure 1.

The primary gauging station (Squamish River near Brackendale) is the only hydrometric station on Squamish River. It has a continuous record since 1955 and was used by Sichingabula (1986) and Hickin and Sichingabula (1988) when they documented Squamish River channel change for the period 1947–1984. The Brackendale gauging station represents flow derived from approximately 72% of the Squamish River catchment (the remaining 28% is added downstream of this station).

The primary source of climate data is Upper Squamish climate station (Figure 1). Data have been recorded here since May 1979. Precipitation data prior to May 1979 were extrapolated from the nearby Clowhom Falls climate station which has been recording precipitation data since 1932. The remaining temperature data were extrapolated from the Britannia Beach
Near Furry Creek climate station which has been collecting rainfall data semi-continuously since 1913. The secondary stations are located within a 35 km radius of the primary station.

To determine if shifts in any of the hydrologic variables of interest have occurred within Squamish River basin over the 50-year study period, time-series analyses of the continuous discharge, precipitation and temperature records were performed.
conducted over a variety of timescales (annual, seasonal and specific extreme-flood events).

The relationship between storm characteristics and extreme-flood events receives the most attention of all the timescales because floods produce rapid channel change on Squamish River (Sichingabula, 1986). Here a mean daily discharge equal to or exceeding 1500 m³/s (approximately 1.5 times morphological bankfull discharge) is arbitrarily chosen to represent an extreme flood. Floods of this magnitude are geomorphically important in that they are capable of drastically reworking both in-channel and floodplain materials.

Fourteen floods qualify as extreme but three of these were disqualified from the analysis. In one case, a 1700 m³/s flood on June 27, 1968 occurred outside the study months. Because it was caused by rapid snow melt or rainfall on snow, it is not part of the hydrologic population being considered. In the other case, the remaining two disqualified flood events, the 1620 m³/s extreme flood of September 26, 1957 and the 1560 m³/s flood of October 11, 1967 exhibit hyetographs that display no definitive onset or terminus. Therefore, defining the storm precipitation and storm intensity inducing the extreme floods is not possible. The remaining 11 extreme flood events used for this study are shown in Table I.

Two attributes, in addition to the maximum daily discharge of each event, are determined for each extreme flood: flow duration and flood volume. Flood duration is defined as the time taken, in days, for mean daily discharge to return to that flow magnitude recorded on the day prior to the initial, steep rise in the flood hydrograph (Figure 3). Flood volume is defined as the total flow of an individual flood calculated by summing the mean daily discharge for the flood event and multiplying by the number of seconds in a day.

Similarly, three attributes of storms that produced the extreme floods were also determined: storm duration, event rainfall, and rainfall intensity (maximum-daily and average-daily). Storm duration is defined as the period of time, in days, from the storm onset to terminus, a transition described by the storm hyetograph. The onset of a storm is characterized by an initial steep rise from zero or near-zero daily precipitation and the terminus of a storm is marked by the next successive day in which no precipitation is recorded. Event rainfall is defined as the total amount of rainfall recorded over the storm duration and is an indicator of the magnitude of a storm. Two rainfall intensities of different timescales are calculated for each storm: maximum single-day intensity (maximum daily intensity) and average rainfall intensity for a given storm. Rainfall intensity for both timescales is based on daily data.

Regression analysis is the primary analytical method employed to determine the strength of various relationships among storm and flood characteristics and among individual hydroclimatic characteristics with respect to time.

Channel planform change data

In order to identify and compare changes in channel form of Squamish River over the full study period (1947–2007), the measurement protocols of Sichingabula (1986) for channel change were adopted in this study, although instrumentation has changed between the two studies. In the 1986 study photographic images of channel changes were processed optically (using a C-240-D (ST) Photo Ace Process Camera and a Bausch and Lomb Zoom Transferscope) and areas were measured manually using a compensating polar planimeter. Sichingabula’s analysis was based on British Columbia Government photography for 1947, 1951, 1952, 1958, 1960, 1964, 1969, 1976, 1977, 1980, 1982 and Pacific Survey Corporation photographs for 1978 and 1980 and Simon Fraser University aerial photography for 1984. Scales of these photographs varied from 1 : 68 000 to 1 : 10 000.

For the present study provincial aerial photographs, in the form of contact prints were obtained through the British Columbia Crown Registry and Geographic Base (1990, 1991, 1994, 1996), and positive-film aerial photographs (large-scale 1984, 2007) were used as a basis for measuring channel form at various points in time. The scale of aerial photographs ranges from 1 : 10 000 to 1 : 40 000. These photographs were scanned at high resolution (1200dpi) and imported into the geospatial program ER Mapper 7.1 where the images were geocoded/georeferenced. A minimum of seven and average of 10 ground control points were selected for each photograph. The spatial coordinates of the ground control points were obtained through Arc Mapper 7.1 using CamMap DMTI Street-files data, a Universal Transverse Mercator (UTM) projection referenced to the NAD83 datum.

Outlines of the channel form for the rectified images were manually drawn in ER Mapper 7.1. These outlines are referred to as annotations and consist of two features: boundaries of the channel and boundaries of vegetated islands/bars.

Channel boundaries are defined as the boundary between permanent vegetation (the line of riparian grasses and trees) and unvegetated channel sands/bars or the river itself. This definition obviates any problem of defining channel boundaries from year to year when discharge, and therefore river stage and width of the water surface, are not constant. Within a given reach, for the photographic record, discharge fluctuations range from 100 to 364 m³/s, with no discharge exceeding bankfull.

Islands are defined as any accumulation of sediment supporting sub-areal vegetation, separated from a channel boundary by a distinct filament of flow. Woody-debris dams that initiate the deposition of alluvial material are also recognized as islands. It should be noted that changes to unvegetated gravel bars are not incorporated into the areal measurements. These features are constantly reworked, as indicated by the lack of bar-top vegetation, which makes them difficult to characterize. Although this is a limitation of the measurement protocol, bars are generally of low relief so they appear to constitute a small proportion of the total change in sediment storage in a reach. Therefore it is appropriate to classify unvegetated surfaces as products of active bed load transport rather than stable features.

It is important to re-state that floodplain changes reported here are two-dimensional. Bed incision and aggradation are certain to simultaneously accompany lateral changes to the floodplain. However, they are ignored here for two reasons.
First, to determine volumetric changes requires some method of defining and measuring the average depth of the mobile bed-material layer as well as adjusting the areal extent of channel change to account for variation in water level (Ham and Church, 2000). Second, Sichingabula’s (1986) channel change measurements were present in two-dimensions. In order to make useful comparisons between those and our findings a two-dimensional approach is adopted here.

Errors associated with the annotations arise from the inaccuracy of the source data (which varies from ±7 to ±15 m depending on proximity to major urban centers; CanMap DMTI Spatial Streetfiles specification) used in the geocoding process, from rectification errors [root mean square error (RMSE) ~5 m; also see Nicoll and Hickin, 2010], and from the operational error associated with manually geocoding the images (±1-5 m; Ham and Church, 2000). In combination these absolute position errors are estimated as ±9 m for the more suburban reaches (A and D) and ±16 m for the more rural reaches (B and C). These absolute errors propagate in quadrature through the image differencing process yielding an uncertainty in displacement measurement of ±12 m (Reaches A and D) ±23 m (Reaches B and C). The actual uncertainty in the relative measures presented here, however, is likely to be much less because the differenced errors are almost certainly partly compensating. In any case, only significant bank shifts are described here and these range from the minimum detection level (12–23 m) to 414 m over the photograph period.

Uncertainty in the area measurements is difficult to determine because of the unknown error compensation and because the minimum detection level depends on the shape of the depositional/erosional unit on the photograph. Compact geometries such as a square or circle are detectable if the area is greater than about 500 m² but irregular shapes pose a more complex problem. For example a long elongated deposit may represent a significant area but if the width of the unit is less than the displacement error the area will not be measurable. In the present study all unit areas mapped are associated with boundary displacements that significantly exceed the displacement error (12–23 m).

A crude and conservative estimate of maximum error and corresponding minimum-area detection level can be obtained as the product of the displacement error and the length of the perimeter bounding each area of deposition/erosion. In all cases the calculated areas reported here exceed this minimum detection area by almost three times on average and often by an order of magnitude. The median size of the primary erosion/deposition units mapped from photographs is 5250 m² and areas range up to 160 000 m².

The measurement errors reported by Sichingabula (1986) for the mapping based on the aerial photographs prior to 1984 are similar to those noted here (~25 m displacement error and a 1250 m² minimum area detection limit).

The measurement protocol adopted here is concerned only with planform changes evident on sequential aerial photographs. No account has been taken of independent vertical changes in the channel boundary. Although field surveys undertaken at other times suggest that Squamish River is neither aggrading nor degrading, short-term scour and fill occurs on a seasonal timescale (Paige and Hickin, 2000). Although the geographic information system (GIS) differencing reported here provides a useful relative measure of river activity, its utility as a basis for determining the sediment budget is more constrained.

### Results

#### Hydrologic change

Most of the discharge, precipitation and temperature data (1956–2007) at decadal, annual and seasonal timescales are stationary in the mean for the last half of the century (Figure 4). But these means obscure significant non-stationary elements of the record.

For example, the annual precipitation record exhibits phases in which precipitation increases with respect to time.

### Table 1. Extreme August–December floods (maximum discharge ≥ 1500 m³/s) on Squamish River, 1956–2005

<table>
<thead>
<tr>
<th>Extreme flood date</th>
<th>Maximum daily flood discharge (m³/s)</th>
<th>Flood volume (km³)</th>
<th>Flood duration (days)</th>
<th>Storm precipitation (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>September 6, 1957</td>
<td>2230</td>
<td>0.28</td>
<td>5</td>
<td>62.2</td>
</tr>
<tr>
<td>October 12, 1958</td>
<td>1580</td>
<td>0.22</td>
<td>6</td>
<td>136.6</td>
</tr>
<tr>
<td>October 31, 1967</td>
<td>1610</td>
<td>0.19</td>
<td>5</td>
<td>94.6</td>
</tr>
<tr>
<td>October 29, 1968</td>
<td>1570</td>
<td>0.11</td>
<td>4</td>
<td>97.2</td>
</tr>
<tr>
<td>November 4, 1975</td>
<td>1800</td>
<td>0.43</td>
<td>9</td>
<td>150.9</td>
</tr>
<tr>
<td>December 27, 1980</td>
<td>2020</td>
<td>0.41</td>
<td>7</td>
<td>203.2</td>
</tr>
<tr>
<td>November 1, 1981</td>
<td>2110</td>
<td>0.35</td>
<td>6</td>
<td>226.2</td>
</tr>
<tr>
<td>October 8–9, 1984</td>
<td>2150</td>
<td>0.68</td>
<td>13</td>
<td>305.6</td>
</tr>
<tr>
<td>November 1–12, 1990</td>
<td>1720</td>
<td>0.5</td>
<td>13</td>
<td>285.6</td>
</tr>
<tr>
<td>August 30, 1991</td>
<td>2120</td>
<td>0.4</td>
<td>8</td>
<td>259.2</td>
</tr>
<tr>
<td>October 17–19, 2003</td>
<td>2630</td>
<td>0.86</td>
<td>15</td>
<td>329.6</td>
</tr>
</tbody>
</table>

![Image](image-url)
The most compelling temporal hydroclimatic shift, however, is the intensification of fall storms generating extreme floods over the period of record (Figures 5 and 6). Storms generating extreme floods exhibit a three-fold increase in precipitation over the period of record and a doubling of the associated rainfall intensity and storm duration (Figure 5). The effect of this intensification of storms on the size of extreme floods over the same period is clearly evident in Figure 6. Extreme flood discharge, duration and flood volume have each respectively increased over time by about 50, 300 and 450%.

Channel change

The amount and intensity of channel planform change is quantitatively expressed as two-dimensional surface-area changes and as maximum bank-retreat rates.

Floodplain deposition (FD), floodplain erosion (FE), total change in floodplain surface-area \( \Delta A_1 = |FD| + |FE| \) and net change in floodplain surface-area \( \Delta A_2 = FD - FE \) constitute the variables used to assess channel change. The four variables are presented as cumulative-mass plots organized by reach in Figures 7–10.
The shape of the cumulative mass-plot curves indicates the nature of change over time. A linear plot implies a constant rate of change while concave and convex plots imply acceleration or deceleration in the respective rates of change.

From the late-1950s to 1970 the rates of erosive, depositional and total channel change in all reaches remain moderate and relatively constant. This period was followed by extremely low rates of channel change extending into the early 1980s. However, from approximately 1982 until 1996 the four non-braided study reaches exhibit a period of heightened geomorphic activity. From 1996 to 2007 the rate of geomorphic change appears to decrease, but this may be a statistical and methodological artifact related to the long photographic interval rather than to a real decline in geomorphic activity.

The significant concave-upward profile of the FE, FD and $\Delta A_1$ cumulative-mass plots of all reaches indicate that the rate of erosion, deposition and total channel change have accelerated in the last 50-years. The magnitude of total change (bottom left graphs in Figures 7–10) are approximately equal.
among Reaches A, C and D, while Reach B has only experienced approximately half the change exhibited by the more active reaches.

The greatest average annual rate of increase in FD, FE and $\Delta A_1$ between individual measurement periods, can be attributed to the passage of extreme floods (1980–1982; 1982–October 1984; October 1984–November 1984; 1990–1991). In general, study periods that closely bracket extreme floods are predominantly characterized by erosive changes. More specifically, in all study reaches the total surface-area change recorded in the 1990–1991 period is approximately two times greater than any other prior average rate of change recorded within their respective reaches.

Although the increase in ratios of floodplain surface-area erosion and deposition are approximately equal, the absolute rates of increase vary considerably. The cumulative-net change in floodplain-surface area (lower-right, Figures 7–10) indicates that, with the exception of Reach B, floodplain erosion always (Reaches C and D), or recently (Reach A since 1980) has out-paced deposition. In a few cases erosion may appear to be the dominant process because the time interval between aerial photographs is too short for vegetation to establish (October 1984–November 1984) so that the measurement protocol does not recognize the deposition.

If cumulative erosion and deposition mass-plots are compared for each reach, the erosion plots are noticeably 'smoother'. The more step-wise nature of the deposition curves is difficult to explain. Perhaps enhanced deposition occurs after an extreme flood. A large flood event could mobilize a large amount of sediment that eventually could be colonized by vegetation, thereby temporarily increasing the rate of deposition before achieving a more steady state. However, this notion is contradicted by the fact that large and small amounts of deposition are recorded by periods that proceed extreme floods in 1984–1990 and 1980–1982, respectively.

Maximum and average bank-retreat rates for eight freely-migrating bends were determined from 1984 to 2007 in Reaches B, C and D. Sichingabula (1986) did not record maximum bank-retreat rates in his earlier inventory of Squamish River channel change and therefore long-term relation between bank retreat and the increasing trend in extreme flood magnitude cannot be determined from the existing data base.

Although simultaneous erosion and deposition in banks may occur within a bend, bank retreat is the dominant process associated with extreme flooding on Squamish River. Thus bank erosion is viewed as the more relevant bank-specific measure for comparing the degree of channel change within a bend.
Like the total areal channel change ($\Delta A_1$), bank-retreat rate within all of the study reaches appears to be significantly affected by extreme floods (Figure 11). Rates for the 1990–1991 period are the greatest on record at each location except for Bend 6 which may have been influenced by a hard point (bedrock outcrop) upstream of the bend. Unlike the trends displayed by the cumulative mass-plots (Figures 7–10), however, there is no temporal or spatial trend in the migration rates at each location. Rather, retreat rates generally are closely clustered within each location (with the exception of Bend 8).

Discussion and Conclusions

Hydrologic change

The most significant findings of the hydrologic analysis are the linear increases in, and the correlations among, the intensity, magnitude and duration of extreme storms and of peak flows over the period of record. Physical reasoning suggests that the climate and flood series are causally linked.

Although an analysis of the fundamental cause of this shift to a more energetic climatic environment over the Squamish basin is beyond the scope of our paper, others have provided some useful insights. Within the Pacific Northwest geographical region local climatic variables have been linked to the Pacific Decadal Oscillation (PDO) index (Jakob et al., 2003; Cunderlik and Burn, 2004; Stahl et al., 2005). Within Squamish River basin mean annual temperature appears to mimic the ‘cool’ and ‘warm’ phases of the PDO cycle although similar relations are not exhibited by mean annual discharge and mean annual precipitation.

The shift in storm characteristics may also be related to synoptic-scale climatic patterns other than the PDO, such as the El Nino Southern Oscillation (ENSO) or Pacific North America (PNA) circulation patterns. For example, Moore and McKendry (1996) found that the intensification of PNA circulation pattern coupled with a deepening Aleutian Low is associated with widespread light snow pack conditions. Similar influences on storm patterns of Coastal British Columbia are likely but unstudied.

Changes in local environmental controls (basin characteristics) do not seem to offer any compelling explanation of the positive shift in the extreme-flood record. Changes in forest cover can alter hydrologic process (see the review in Moore and Wondzell, 2005) although it does not seem to have a significant role in this case. Detailed data on forest harvesting in Squamish River basin are not readily available but from visual inspection of aerial photographs less than 10% of the basin appears to have been harvested. This fact alone, suggests that forest harvesting is not likely to be important in explaining the increase in extreme flood magnitude on Squamish River. In any case, the runoff ratio (event discharge/event precipitation) of each storm/flood event – a derived variable used to approximate forestry influences on...
local hydrology in the absence of monitoring data—supports this conclusion: it is highly variable over time but has remained approximately constant in the average. That is, the amount of discharge produced by a given amount of precipitation for large fall events remains relatively constant across the study period. Similarly, soil moisture variation does not appear to have produced any substantial trend in the flood data set. The Antecedent Moisture Index (AMI) defined as the total precipitation recorded 14 days prior to the onset of a storm—bears no orderly relation to flood magnitude.

The non-stationary flood series for Squamish River has significant implications for estimating flood recurrence interval: the magnitude of the design flood clearly has been increasing over time. Unfortunately, the 50-year hydroclimatic record presented here is not long enough to resolve the fundamental question: Is the decadal increase in annual peak flows part of an ongoing longer-term upward trend or is it part of a decadal oscillation? Observations elsewhere, however, suggest caution in adopting the former assumption. For example, in eastern Australia it has been shown that wet and dry climatic phases are part of the long-term record there and that increases in flood magnitude in some rivers can persist for 50 years or more before the hydroclimatic regime shifts back towards longer-term average conditions (Erskine and Warner, 1988; Warner, 1987, 1997; Vives and Jones, 2005) although this view of climatic oscillation in eastern Australia has not gone unchallenged (see Kirkup et al., 1998; Erskine and Warner, 1998).

Channel change
The evidence suggests that the record of accelerating channel change shown in Figures 5 and 6 is caused by the increasing
size of Squamish River floods generated by the intensifying fall rain storms over time although this trend is not evident for the last two decades because of the lack of data.

Storm-induced extreme floods in Squamish River are often accompanied by debris-flow inputs from several tributary streams. A large debris flow occurred in Turbid Creek, immediately upstream of Braided Reach (Reach E) in 1984. Despite the assertion made by Hickin and Sichingabula (1989) that the majority of debris-flow materials likely were flushed out of Braided Reach and did not cause the major reorganization of the reach at that time, the debris flow may have influenced the total change in floodplain surface area as the materials were transported downstream in the years following this event. Indeed, elsewhere Pelpola and Hickin (2004) suggest that a debris flow is the likely explanation for a period of above-average annual bed-material transport rate in Fitzsimmons Creek to the north of Squamish basin at Whistler BC and this secondary effect could be relevant to the Squamish River case. Paige and Hickin (2000) found that bedload in Squamish River moves in coherent waves or pulses in the downstream direction at an average rate of 15·5 m/day. This process, termed pulse scour and fill, is detectable on sonar surveys. Should the increase in bed-material transport affect the amplitude or downstream celerity of the bed-wave, shoaling or temporary storage of a bed-wave on the streamward edge of a point bar or island head, for example, could provide a stable environment for vegetation to establish. The net result, given the measurement protocol adopted in this study, would be an increase in net areal deposition. Average rates of areal deposition presented in Table II, however, provide no clear evidence that Turbid Creek debris flow left such a depositional signature.

In summary, although other environmental factors may contribute in a minor secondary way to the temporal pattern of channel change, extreme flood magnitude driven by intensifying Pacific storms seems the most likely primary explanation of the acceleration of channel change on Squamish River from 1956 to 2007.

Nevertheless, some anomalous patterns of channel change remain unexplained. For example, bank-erosion rates in some channel bends do not seem to be an orderly response to the flood record. It is somewhat surprising that, despite bracketing the flood of record in 2003, the 1996–2007 period did not register average-annual bank-retreat rates as large as those in the 1990–1991 period. Perhaps significant change did occur but deposition and vegetation establishment in concave bank zones during the period following the flood (2004–2007) concealed or muted any previous erosive changes. Alternatively the pattern of bank retreat may reflect local flow-alignment shifts such as the straightening of the meandering reaches that occurred between 1984 and 2007. This might have reduced the river attack in concave bends, while concurrently shifting the erosional focus within the bend to the convex bank. However, this would only explain the behavior of selective bends in the meandering Reaches B and C.

The behavior of channel change displayed in the cumulative mass-plots (Figures 7–10) is open to several interpretations. For example, the 50-year record of geomorphic change can be divided into two distinct periods: a period of low geomorphic activity (1957–1980) and a period of high geomorphic activity (1980–2007) perhaps mimicking a step-function shift in the hydroclimatic regime (in turn reflecting the segmented record of the PDO). The comparison of average channel change in the earlier relative to the later period is quite striking although

<table>
<thead>
<tr>
<th>Period</th>
<th>Reach A (10^3 m^2/yr)</th>
<th>Reach B (10^3 m^2/yr)</th>
<th>Reach C (10^3 m^2/yr)</th>
<th>Reach D (10^3 m^2/yr)</th>
<th>Reach E (10^3 m^2/yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>November 5, 1984–August 16, 1990</td>
<td>45·8</td>
<td>18·7</td>
<td>39·5</td>
<td>56·4</td>
<td></td>
</tr>
<tr>
<td>August 16, 1990–September 9, 1991</td>
<td>60·6</td>
<td>33·5</td>
<td>97</td>
<td>161</td>
<td>124·81</td>
</tr>
<tr>
<td>September 9, 1991–July 28, 1994</td>
<td>45·9</td>
<td>25·7</td>
<td>79·5</td>
<td>57·7</td>
<td>114·9</td>
</tr>
<tr>
<td>July 28, 1994–September 27, 1996</td>
<td>33·7</td>
<td>23·6</td>
<td>36·1</td>
<td>26·6</td>
<td>79·2</td>
</tr>
<tr>
<td>September 27, 1996–December 10, 2007</td>
<td>32·4</td>
<td>10·1</td>
<td>22·3</td>
<td>28</td>
<td></td>
</tr>
</tbody>
</table>
there is no single event to suggest a dramatic shift in the rate of channel change in about 1980.

Another interpretation, one we favor, is that from 1956 to 2007 geomorphic change has undergone a sensibly progressive acceleration although physical reasoning suggests that it may have been a punctuated acceleration. The channel-change data sets, particularly the cumulative change (bottom left; Figures 7-10) are not inconsistent with the model of change idealized in Figure 12. Here a steep slope represents the stress of an extreme flood (rapid cumulative change) while the subsequent plateau represents a period of system relaxation with declining level of channel change until the next event occurs and the pattern of channel change is repeated. Indeed, the shape of the cumulative erosion and deposition mass-plots hints of this type of behavior (Figures 7-10). That is, in general the long intervals between photographs may have ‘smoothed out’ geomorphic detail embedded in the cumulative mass-plots that only are detectable with yearly data.

Acknowledgments—We are indebted to the reviewers whose thoughtful critique of the original manuscript led to a significantly improved paper. This project, based on the MSc thesis of the lead author (Bauch, 2009), is part of a larger study of the morphodynamics of rivers in western Canada funded by the Natural Sciences and Engineering Research Council of Canada (NSERC) and by Simon Fraser University.

References


Whitfield PH, Reynolds CJ, Cannon AJ. 2002. Modelling streamflows in present and future climates: examples from Georgia Basin,

