PALEOHYDROLOGY OF THE BELLA COOLA RIVER BASIN: AN ASSESSMENT OF ENVIRONMENTAL RECONSTRUCTION

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ABSTRACT

Recent geomorphic and hydrologic environments of a mid-latitude alpine basin are investigated under the integrative theme of paleohydrology. The aims of this research are: 1) to characterize the response of selected biological and geophysical elements to recent climatic change; 2) to determine the resolution and length of paleoenvironmental records in the study area; and 3) to ascertain the significance of observed and inferred environmental change over the Little Ice Age interval.

Bella Coola River drains 5050 km^2 of glacierized mountains along the central coast of British Columbia. Biological elements examined on a basin-wide scale included: tree-growth in temperature and moisture-stressed environments, damage to trees in glacial and fluvial settings, pollen variations in a variety of sedimentary deposits and soil development. Geophysical elements include primarily glacio-lacustrine and floodplain sediments, glacier deposits and river channel morphology. A retrospective strategy was adopted by testing initially for the nature of relationships between synoptic climate, basin hydrology and element response during the period of instrument record (1900 AD to present). Inferences about pre-instrument environments were then made using the proxy data.

Events of several types are characteristically mixed in a response record. Variations in Douglas and subalpine fir growth, glacio-lacustrine sedimentation rates, glacier fluctuations and shifting of the Bella Coola River reflect a combination of persistent and episodically extreme behavior. Glaciers appear to respond by advancing or retreating after departures in winter precipitation persistent for several years. Extreme events, particularly high-magnitude autumn floods, are not exclusively linked to a particular set of mean climatic departures. This makes inferences from proxy data such as floodplain deposits and flood-damaged

vegetation difficult. Periods of increased flood frequency are supposed to relate to an increase in floodplain sedimentation.

Except in very favorable circumstances, paleoenvironmental methods do not have the resolution promised. Climatic information recoverable from tree-ring data and glacio-lacustrine sediments is of considerably lower than annual resolution. Statistically based climate models using proxy data as independent variables produce low levels of explained variance. Proxy data sources in the basin were largely restricted to the last 300 to 400 years or Little Ice Age interval.

Most glaciers in the basin reached Little Ice Age maxima in the middle of the 19th century in response to below average temperatures and above average precipitation between approximately 1800 and 1855 AD. Tree-ring data and equilibrium line altitudes on glaciers indicate that precipitation was on average 25 to 30% greater than the 1951-1980 mean. Inferred below average temperatures in the early 18th century probably signaled the beginning of the Little Ice Age along the central coast; however, there was not a major response in glaciers until persistent positive departures in precipitation occurred. Recession of glaciers from Little Ice Age maxima was slowed by cooler and wetter conditions between 1885 and 1900 AD. The persistence of warmer and drier conditions in the first half of the 20th century was exceptional in comparison with inferred climate of the last 330 years. Major floods in 1805/06, 1826, 1885 and 1896 correspond to intervals of increased precipitation.

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Chapter I
INTRODUCTION

It has long been recognized that certain biological and geophysical components of the environment respond in a predictable manner to external forces. Remaining evidence of these components allows inferences to be made about the distribution and magnitude of forces in the past. Confident identification of past environmental settings relies upon some knowledge of the attendant physical processes, the nature of system responses and how the system might be affected when certain boundary conditions are altered. An integrated approach towards understanding environmental change prior to the period of instrument records utilizes evidence from several biogeophysical subsystems in order to assess the consistency of results and to evaluate errors involved in interpretations and related inferences. The purpose of this thesis is to assess the utility of several methodologies in the analysis, interpretation and reconstruction of recent hydrological and geomorphological processes. The test field area is in a mid-latitude, glacierized alpine basin of west-central British Columbia.

(1.1) Systems Framework

In this work the interactions amongst atmospheric, hydrological, biological and earth surface processes are analyzed around the integrative theme of paleohydrology. Leopold and Miller (1954) first introduced this term to describe "the interaction of climate, vegetation, stream regimen and runoff obtained under climates different from that of the present". Schumm (1965) suggested that the term be "restricted to that portion of the hydrological cycle that involves the movement of water [and sediment] over the surface of the earth . . .". Brown (1982) has proposed that the term be even more narrowly defined as "the applications of concepts, methods and models derived from hydrogeomorphological studies to fluvial processes, sediments and forms of the past". This latter definition stems from the frequent emphasis on fluvial processes but is considered here as unnecessarily restrictive. The common theme in paleohydrological studies is an attempt to provide a more comprehensive understanding of environmental settings over longer time scales and with increasing temporal resolution.

It is convenient to view the environment as a system comprised of several morphogenetic components linked by a series of functional relations and affected by a set of boundary conditions. Figure 1.1 illustrates four morphogenetic subsystems of interest in this study: glacial, glaciolacustrine, fluvial and mass wasting. A set of functional relations between the atmospheric environment and the earth surface environment can be defined. These relations are formulated on the basis of exchanges of energy and mass and are influenced by boundary conditions such as topographic and geologic controls, neotectonics and former glaciation. Energy from the atmospheric and hydrological components of figure 1.1 drives the system resulting in the redistribution of mass and the formation of new sedimentary deposits. Inferences about former hydrological settings, then, are made in the reverse direction from that of the functional relations.

The transfer of information between the hydroclimatic environment and any one of the four morphogenetic subsystems identified in figure 1.1 is subject to a variety of temporal and spatial filtering effects. Deposits within any of the subsystems integrate the effect of short-term hydroclimatic processes, or events, hence have a characteristic low resolution of the true amplitude and frequency of climatic and hydrologic variations. For example, maximum advances of ice are frequently marked by

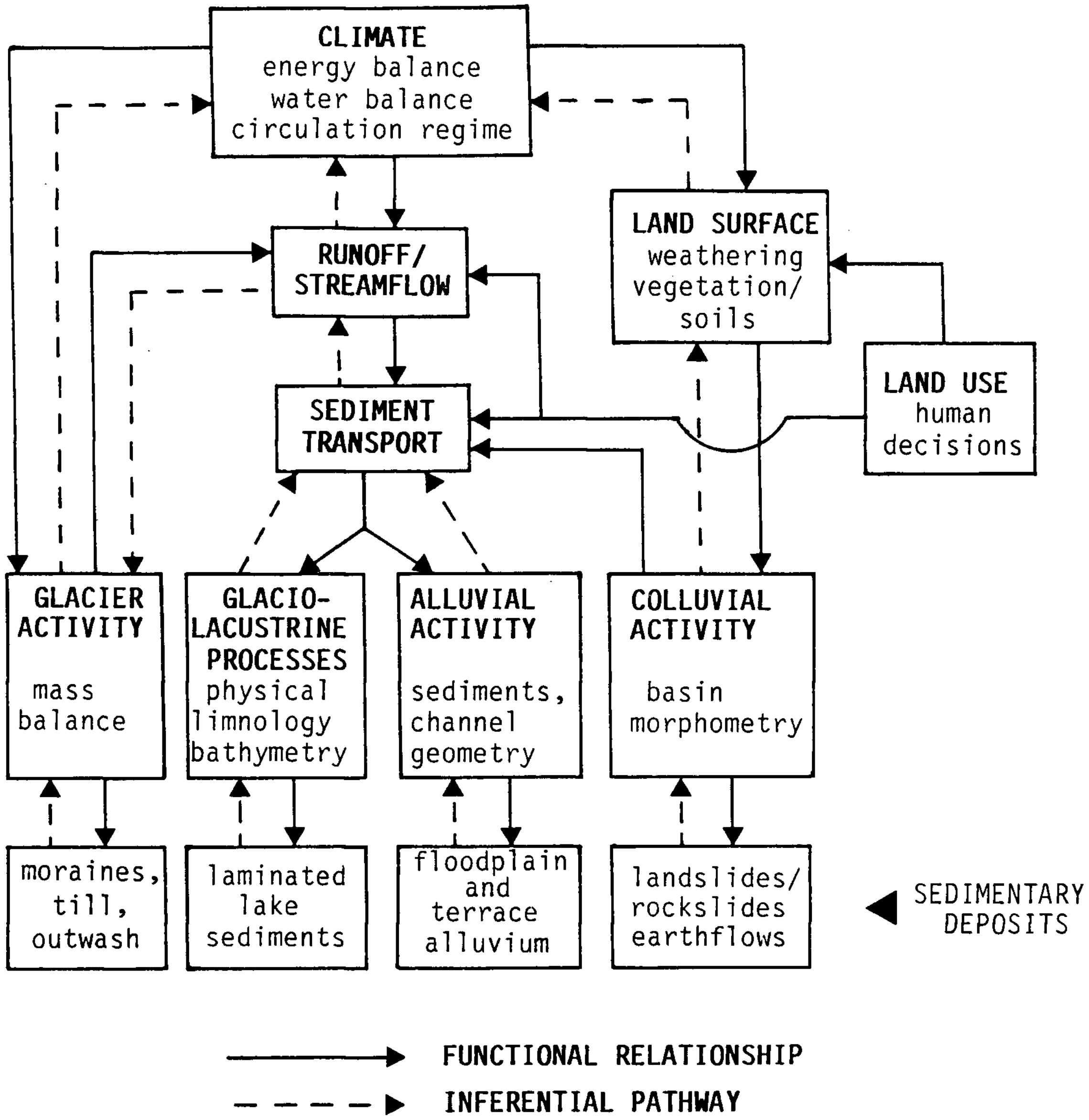


Figure 1.1 Functional and inferential pathways connecting four morphogenetic subsystems of interest in this project. Not shown are boundary conditions which include geological and neotectonic factors as well as past glacial activity (after Starkel and Thornes, 1981). Other subsystems not considered include eolian, marine and glaciomarine environments.

terminal moraines. However, fluctuations in the direction or rate of ice movement may be a response to a combination of hydroclimatic changes which occur over time scales much shorter than the interval required for formation of the deposit. In contrast, landslides, flood deposits or subaqueous slope failures may represent 'key events' without reference to a particular set of departures in long-term mean conditions of hydroclimate. These two situations, which generally limit the inferences that can be made, are commonly associated with primary morphological sources of information and therefore complicate the exercise.

A precursor to successful application of the model in figure 1.1 to past environments is the interpretation of various morphological components such as vegetation, soils, surface forms and sedimentary sequences (Gregory, 1983a). Environmental reconstructions then, can only be made after absolute and relative dating techniques are accurately applied to sequences of sediment and preserved landforms which have clearly defined depositional environments. In this context, a true paleohydrological model is rarely achieved.

(1.2) Operational Approaches

Brown (1982) suggested that three types of paleoenvironmental modeling can be undertaken: process, empirical and inferential model development. A modified version of this scheme discussed here includes three general categories: (1) process or physical models, (2) empirical models and (3) inductive models.

Process Models

The most difficult of all three categories is the process model where the functional relationships between various components of the system are quantified and grouped to form a general physical model which operates according to physical and geochemical laws. The model is constructed and calibrated under modern conditions and then applied to past environments. The approach is purely deductive and requires quantitative estimates of input parameters. General circulation models are of this type.

Clarke *et al.* (1984) utilize a process-model, verified using modern outburst floods, to estimate former discharges of ice-dammed glacial Lake Missoula in Washington state. Ice-flow and heat transfer relations are used as governing equations to predict the size of the subglacial discharge tunnel. Various parameters such as lake volume, water temperature and ice-thickness are required input variables. Model similitude and assumptions regarding ice tunnel geometry and flow resistance remain problematic. Power and Young (1979), among others, have used an integrated watershed model to predict future and former runoff from glacier-covered catchments. This type of model is calibrated against observed flow variations and adjusted accordingly.

The major assumptions underlying process-models are: (1) all dominant processes have been identified and contributing factors parameterized, or, in the case of simulations, unknown but important factors are controlled; (2) governing relationships are properly calibrated for the expected range and variance of input data; and (3) the temporal and spatial resolution of the input data correspond with the calibration data. Violation of the first or second assumption results in an unreliable model whereas violation of the third assumption yields unreliable results.

Empirical Models

Though the deductive approach used in process modeling is also the basis for this second group of models, the underlying physical processes are not quantified. A particular set of processes (rainfall, stream flow, glacier flow) is related to responses in biological or geophysical attributes of the environment and then relationships are developed, usually using statistical techniques. The relations are then inverted to make predictions of former process based on response chronologies preserved in the environment. These include the quantitative analysis of vegetation growth patterns, variability in glaciolacustrine sedimentation, valley or river channel morphology, paleochannels and fossil fluvial sediments. Nominal, interval or ratio scale data can be used.

Mathewes (1973) and Mathewes and Rouse (1975) have attempted qualitative reconstruction of Holocene paleoclimates in the lower Fraser Valley of British Columbia using paleoecological evidence preserved in lake cores. Quantitative estimates of temperature and precipitation for the same area and period were then made using calibration techniques in which the spatial distribution of modern vegetation is calibrated against contemporary climates (Mathewes and Heusser, 1981). These relations are then used to formulate transfer functions in order to estimate former hydroclimatic conditions based on the inferred paleoecology. Two problems complicate the "transfer" of these relations derived from spatial associations. The first is related to the fact that spatial variations in climate can be greater than those experienced at a particular site through time. The second is that factors other than the assumed dependent variable (in this case climate) may explain a significant proportion of the temporal variance and are not included in the development of the transfer function (Imbrie and Webb, 1983).

The major underlying assumption in the retrospective application of empirical models is that an equilibrium has been achieved between a specific attribute and a dominant process for <u>both</u> the calibration and application intervals. Certain biological or geophysical attributes may adjust very slowly to imposed change and thus may not reflect higher-frequency perturbations in the system. In contrast, adjustments may be rapid but sensitive only to forces beyond an exceedence threshold and thus are representative of short-term conditions only.

Linear least-squares statistical techniques are commonly used in the formulation of empirical models. The application of these relationships for environmental reconstruction is based on the assumption that the explained portion of the total variance reflects the primary response characteristic of the system and that the residual variance is random "noise". This constraint can seriously affect the significance of interpretations from a linear model if the system is also known to respond significantly to extreme departures. These less frequent, higher magnitude events become ′key′ elements which are lost in the assumption of linearity.

Inductive Models

This category is based primarily on the inductive approach utilizing site-specific evidence for drawing inferences about former environments. Sedimentary sequences (lithofacies) and the remains or by-products of biological activity are used to reconstruct a series of possible environmental conditions. Ordinal and nominal scale data are used frequently and mostly qualitative results are achieved. Inferences usually relate to a specific set of environmental conditions such as spatial and temporal variations in runoff, or the chronology and distribution of glacier ice movements.

Starkel (1984) has used various alluvial sequences to infer the character of climate transitions over the Holocene Epoch in central Europe. Ryder and Thomson (1986) have used several sources of evidence from glacial deposits throughout southwestern British Columbia to infer the timing and magnitude of glacial activity in post-Pleistocene time. The application of several dating techniques is crucial for establishing chronostratigraphic control both locally and regionally. Preservation of datable materials and inaccuracies associated with radiocarbon dating become major limitations of the exercise.

Commonly, particularly for alluvial sediments, the stratigraphic record represents extreme or 'key' events only (*e.g.* flood deposits). Erosional hiatuses or other unconformities may complicate interpretations if a substantial proportion of the intervening sequence is missing or otherwise disturbed. In this way a substantially different record of environmental change may evolve using these data sources. Recognition of these differences is important for successful model development.

In summary, empirical and inductive approaches, despite their problems, are commonly employed to draw inferences regarding past hydrological and climatological conditions in the reverse direction from that of the usual functional relations. The process-based approach yields estimates of these functional relations and, then the boundary conditions and initial state are changed to approximate former environments. While initially more attractive, this latter type of model is the most difficult to develop and still frequently suffers from linear parameterizations which do not easily accommodate conditions beyond the calibration interval. The three types of models yield different characteristics of the past environment, and therefore should be viewed as complementary approaches to environmental reconstructions.

(1.3) Objectives and Research Strategy

The specific study objectives are:

(1) to identify and relate the response characteristics of selected geophysical and biological attributes of the study basin to recent fluctuations in hydroclimate;

(2) to determine the resolution and length of record available in these attributes and evaluate several methodologies for inferring the character of past hydroclimates; and

(3) to assess the within-basin and regional (southwestern British Columbia) significance of measured and inferred environmental change in the study area.

The primary emphasis is on problems associated with methods used in the reconstruction of past environments. In particular, consideration is given to sources of error and the strength of inferences drawn from indirect and proxy data which may exist in quantitative, semi-quantitative or qualitative form. Empirical and inductive methods only are considered because of difficulties in applying and linking existing process-based models. The study considers the utility of the model presented in figure 1.1 as a framework for drawing inferences regarding recent paleoenvironments in a mid-latitude alpine basin.

In order to assess critically the real resolution of responses in the system and to provide a basis for calibration of paleoenvironmental data, a retrospective strategy has been adopted. This is operationalized by considering first the relationship between hydrologic variability and geophysical responses over the period of recent and detailed instrument records (post-1945). The relationships are then tested using less detailed but long-term instrument records as confirmatory data (1900-1945). This strategy was adopted as a means of assessing the strength of inferences from biogeophysical data in the pre-instrument period. Inferences about environmental conditions prior to 1900 are then made from the evidence of biogeophysical attributes alone.

Initially, it was anticipated that this methodology would allow for a resolution of at least annual, or perhaps even seasonal, biogeophysical responses over the testing period. However, in view of possible persistent trends in biological and geophysical attributes, it was reasonable to expect a resolution only on the order of years to decades. The period after 1650 AD defines the study interval. This was found to be consonant with preserved evidence. Details are given in the following sections.

(1.4) Data Sources

Biological, and geophysical data can be grouped and classified on the basis of temporal continuity, temporal resolution and interval of application (Bradley, 1985). Some provide only a discontinuous record of changes and may represent only extreme events (*e.g.* flood deposits), while others may appear to give a continuous record (season-to-season, year-to-year) of changing conditions (*e.g.* tree-rings). Certain sources are applicable to time scales of millennia or longer and others may yield data concerning only recent (decennial, centennial) processes.

The primary data sources for this study can be divided into four groups of attributes: biological, geophysical, historical and archaeological, which are based on the most useful indicators of environmental change in mid-latitude, glacierized alpine basins (figure 1.2). Biological sources include trees, flowering vegetation and lichen. Tree ring-width time series possess a resolution of seasonal (late-

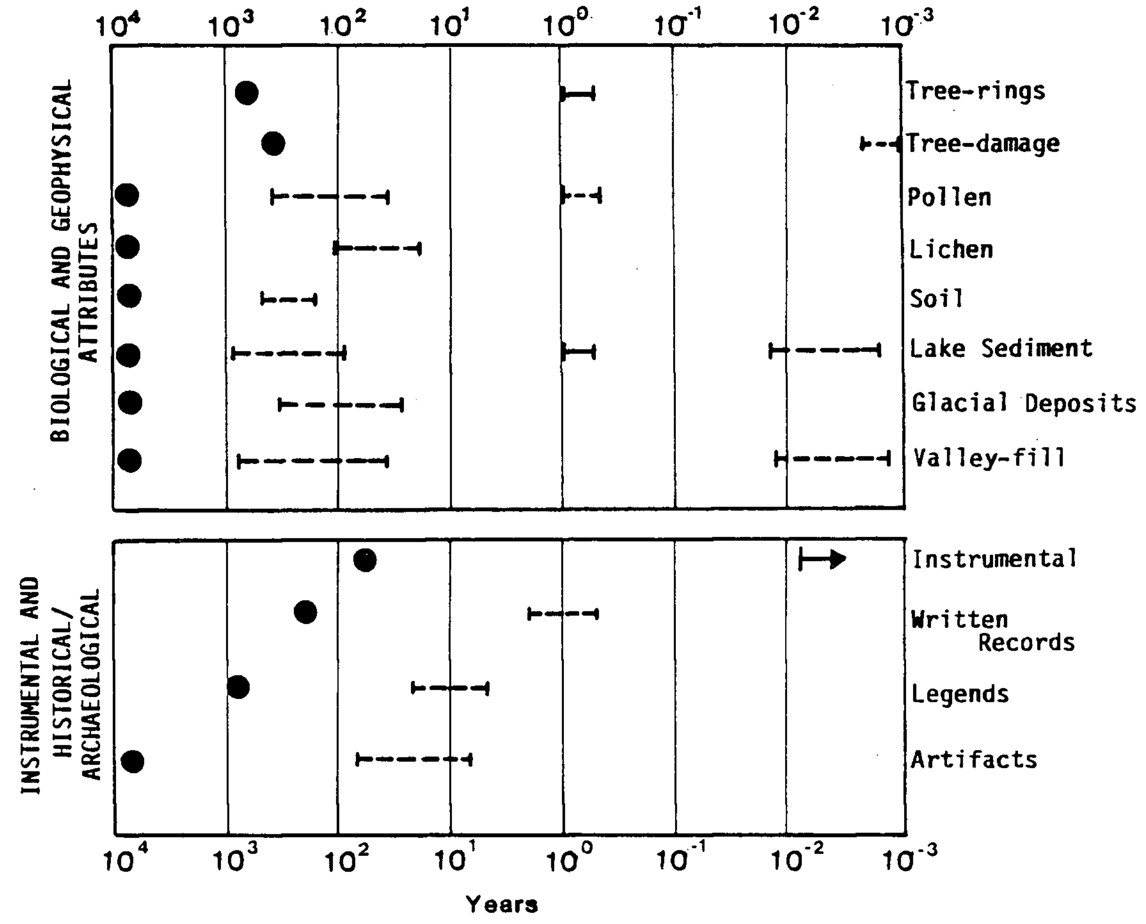


Figure 1.2 Biological, geophysical and historical/archaeological sources used for environmental reconstruction in this study. The minimum sampling interval (attribute resolution) in years is given by a solid horizontal line for continuous variables and a dashed line for non-continuous variables. Some attributes have more than one sampling interval. Solid circles indicate the period open for study in years. Resolution of pollen depends on the character of the host sediments.

wood/early-wood) to annual (total ring-width) growth variations and are continuous through time. Damage to trees by discrete events, such as floods, or more continuous stress, like insect damage, provide some indication of low-frequency, high-magnitude events but generally these records are discontinuous. The oldest trees available for sampling are usually less than 10^3 years.

Pollen and spores are produced seasonally, and if preserved in sediments also characterized by a high temporal resolution, may yield a record of changing pollen contributions and thus ecological variability. Mostly, pollen is useful for identifying changes over long-term intervals (10^4 years) with a resolution of 50 to 500 years (figure 1.2). Lichen and soil development can be used to establish relative ages for different surfaces as old as 10^4 years usually with a resolution of 10 to 500 years. In some instances it may be possible to produce continuous form/time curves with an absolute temporal reference but they are difficult to achieve, particularly with soils.

Lake deposits form during an influx of sediment from the surrounding terrain into a nearly closed system and may represent several time scales and degrees of continuity depending on sedimentation controls. Low-frequency, high-magnitude deposits may be the result of isolated runoff events which occur over a period of hours or days. These sediments are distributed discontinuously throughout the sequence. Alternatively, autogenous events may occur, such as density underflows, produced by the buildup of deltaic deposits which then exceed stability limits. The event is the consequence of continuous sedimentation, but does not mark an extrinsic event. However, in a long record the changing frequency of failures may be significant. If certain runoff regimes prevail, a continuous record of seasonal or annual sedimentation might be preserved (*e.g.* annually-laminated sediments or varves). In shallow water away from primary sediment sources, sedimentation rates may be low and represent long periods of accumulation (10^2- 10^3 years).

Glacial sediments are generally deposited discontinuously over time scales commensurate with the response time of the ice to changes in accumulation and ablation (10^1-10^2 years). A record as old as 10^4-10^5 years may be present if sediments are well-preserved. Valley-fill sediments, downstream from the headwater source areas, also represent discontinuous sedimentation over longer time scales but may incorporate sequences deposited during short-duration floods.

In addition to biogeophysical sources, information related to human activity can provide indirect evidence of environmental change (figure 1.2). Data regarding annual or semi-annual changes and extreme short duration events are most common. For western Canada written records prior to the middle and late 1800s are rare. Indian legends and archaeological evidence are also useful, although they are less direct indicators of environmental setting.

(1.5) Organization

This dissertation is composed of eight chapters. The organization is based on a format for assessing the quality of information transfer from the hydroclimatic environment to the earth surface environment in order to reflect the main objectives. Chapter 2 is a summary of the general physical setting of the Bella Coola River basin and characterization of primary sediment sources in the basin. Chapter 3 is a discussion of variations in several hydrological attributes of the recent and long-term instrument period (1900-1983). Chapter 4 documents the relationships between these recent forcing factors and the response of selected geophysical components of the landscape. The discussion is restricted to the post-1945 interval for which there are several types of direct records of environmental change. Chapter 5 is devoted to the interpretation and significance of a high-resolution record of lake sedimentation rates as an independent source of information about environmental change. Chapter 6 is an investigation of the response of growth variations in trees, and an assessment of tree-rings as a source of high-resolution data in environmental reconstruction. Chapter 7 is a discussion of low-resolution data sources and interpretation of hydroclimatic variability during the late Neoglacial interval. The regional significance of the reconstructed records is also considered. A summary of results and conclusions is given in Chapter 8. Data not presented in the main text, plus additional topics and analyses, are given in the appendices.

Chapter II
STUDY AREA

(2.0) Location

Starkel and Thornes (1981) suggest that selection of a drainage basin for paleohydrological analysis should be based on the availability of data regarding landscape morphology, process mechanics and earth surface materials. In addition, long-term instrument records of recent hydrologic change are essential for quantitative reconstructions. In British Columbia there is no basin which fits these criteria well. Bella Coola River basin, a partly glacier-covered catchment located on the central coast, approximately 500 km northwest of Vancouver (figure 2.1), was selected because of the availability of recent hydroclimatic data, comparatively early historical records of environmental conditions and the potential for preservation of paleoenvironmental evidence in upland source areas and lowland floodplains. Accessibility, basin size and the existence of several preliminary studies concerning physical processes and recent environmental change in the basin (Tempest, 1974; Church and Russell, 1977; Hart, 1981a; Church, 1983) were additional factors.

Bella Coola River drains 5050 km^2 of the glacierized Coast Mountains of southwestern British Columbia. Two tributaries join to form Bella Coola River (figure 2.2): the larger Atnarko River (2,430 km^2) originates in gently sloping, uplifted terrain to the east, and Talchako River drains 1300 km^2 of heavily glacierized mountains to the south. Several smaller tributaries enter the 80 km long Bella Coola River as it flows west to enter North Bentinck Arm (figure 2.2), an inland extension of the Burke Channel and Dean Channel fiord system. Approximately 7% of the catchment is covered by cirque, niche and valley glaciers concentrated in rugged mountainous terrain in the southwest. Two major valley glaciers, Talchako

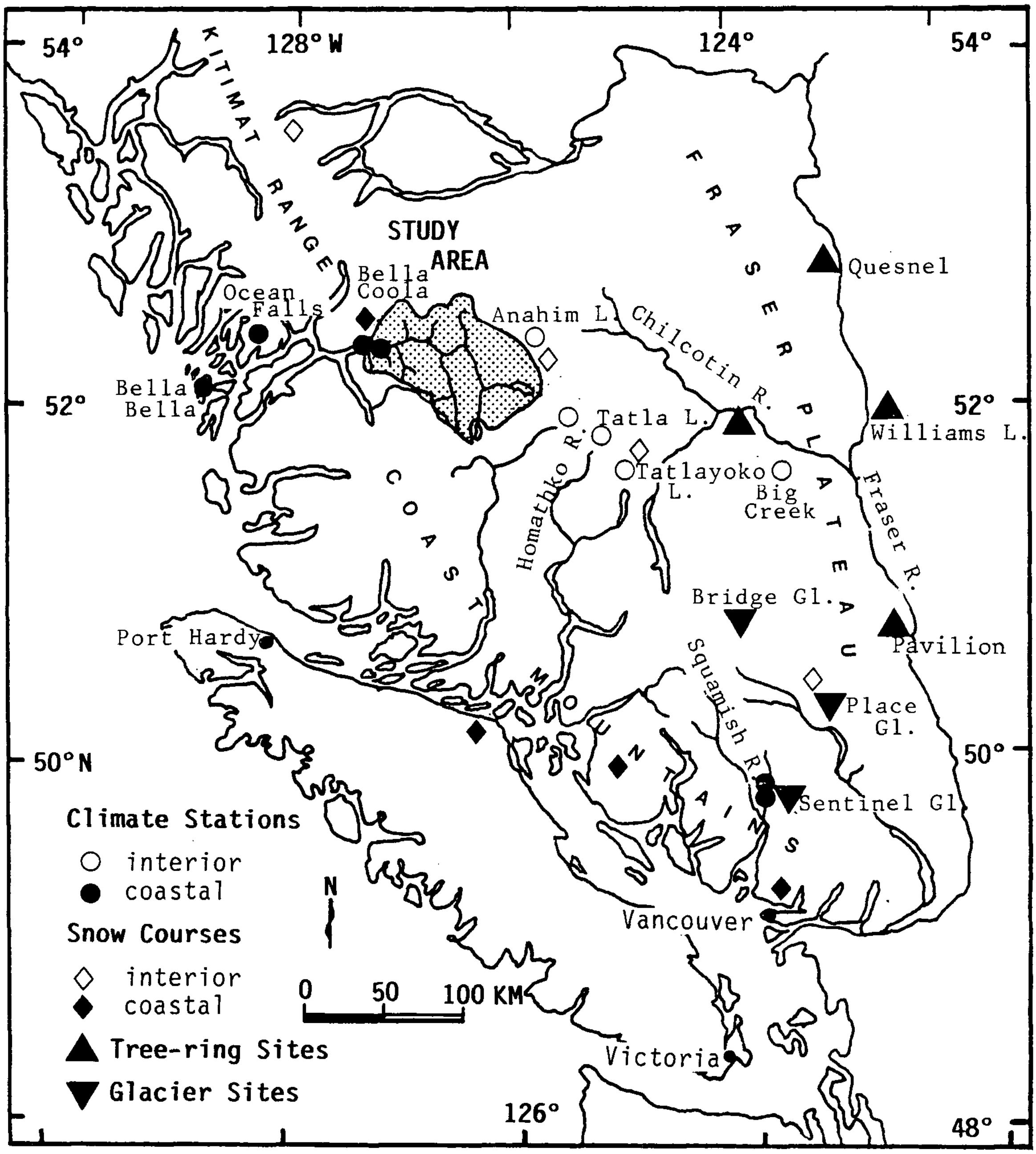


Figure 2.1 General location of study area and hydroclimate stations used in the analysis of climate variations. Axis of the Coast Mountains is an approximate boundary for stations which are referred to in the text as "coastal" and "interior". Also included are locations of glacier and tree-ring data used in the study.

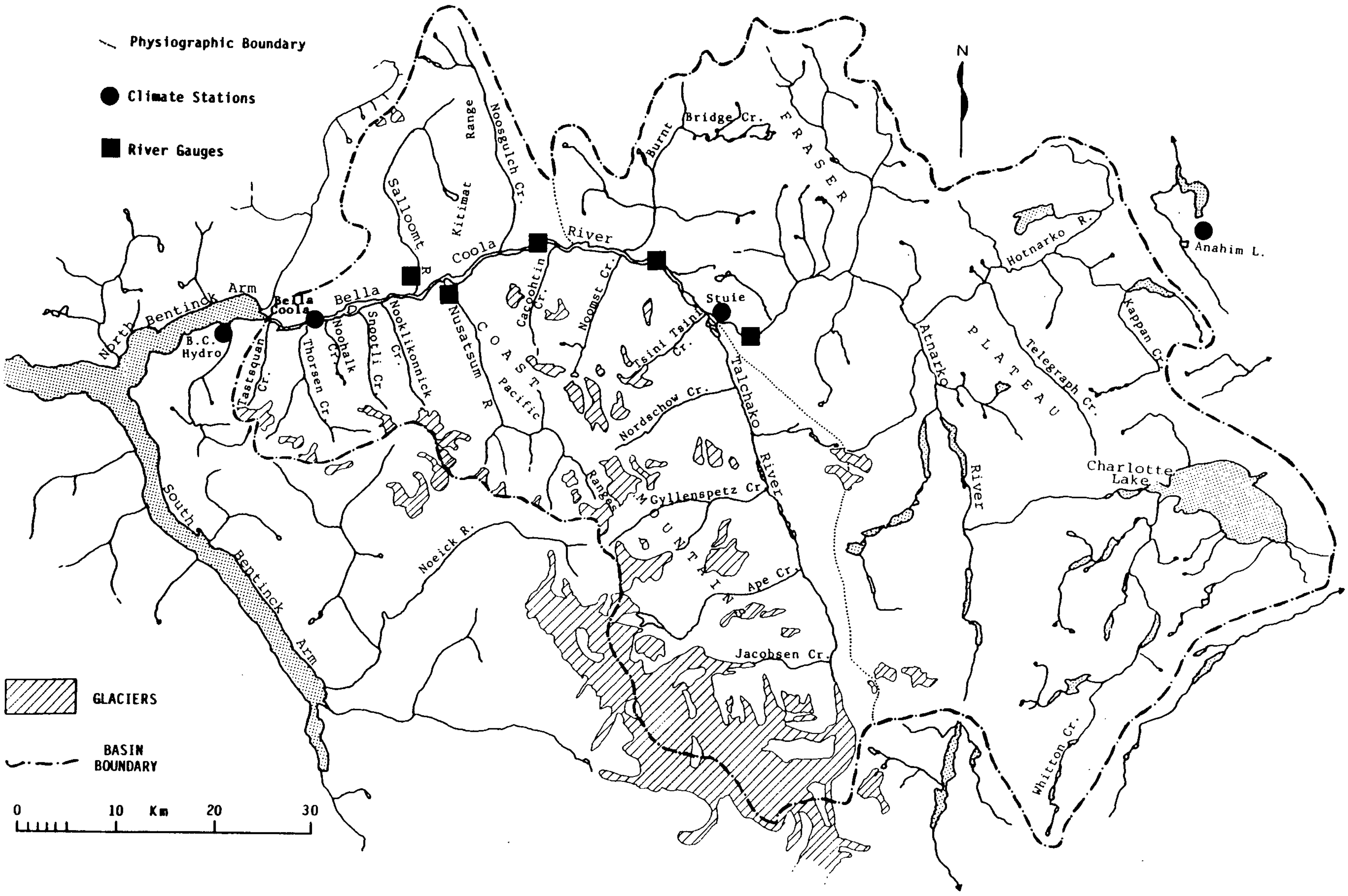


Figure 2.2 Bella Coola River basin including major tributaries and the distribution of upland ice. Also included are the locations of climate and river gauging stations.

and Jacobsen, along with several smaller north-flowing ice masses, emanate from the Monarch Icefield (total area 350 km^2) which straddles the southwest basin perimeter for a distance of 45 km.

(2.1) Physiography

Three different physiographic regions as defined by Tipper (1971a) and more generally by Holland (1964) can be delineated in Bella Coola basin. To the south of Bella Coola River and west of Talchako River are the rugged, deeply dissected Pacific Ranges of the Coast Mountains, extending from sea level to altitudes greater than 3500 m (figure 2.2). The highest peaks bisect Monarch Icefield, some of which are visible only as nunataks, ranging from 2500 to 3500 m. Local relative relief is similar throughout, averaging 1400 to 1600 m. Drainage directions are dominantly eastward or northward resulting in slope aspects which exhibit a strong bimodal distribution of north/south or east/west and a comparatively high mean slope angle of 27 degrees.[1]

Mountains of the Kitimat Range to the north of Bella Coola River and west of Noosgulch Creek (figure 2.2), rarely exceed maximum elevations of 2500 m. Valley bottoms tend to be wider, average relative relief is lower (1000 m) and the mean slope angle (22^0) smaller than those in the Pacific Ranges. Although restricted to headwater regions, landforms typical of the glacial eroded landscape to the south are also found in the Kitimat Range, including U-shaped valleys, cirques, horns, arêtes, hanging tributary valleys and several tarns.

1. Mean slope angle was determined as the average of 300 grid-points sampled systematically from 1:50,000 topographic sheets for each physiographic region. A representative slope distance was defined as ±3 100-foot contours measured perpendicular to the slope (except in low-gradient valley bottoms where ±2 was used). Two sub-samples of 50 grid-points each demonstrated that biases due to grid-spacing and orientation were not significant.

Contrasting with these two regions is the uplifted and less rugged Fraser Plateau which dominates terrain to the east of Talchako River and Burnt Bridge Creek. Gently undulating and hilly topography (mean slope angle = 7^0) characterizes this region in which the average elevation is higher (1300 m) than valley bottom locations to the west. Average relative relief is lower (400 m) than elsewhere in the basin resulting in the formation of hundreds of small lakes along principal drainage lines and on poorly-drained, low-gradient terrain. Charlotte Lake is the largest, covering approximately 65 km^2 (figure 2.2). Drainage from the plateau is concentrated in only a few major streams (Atnarko R., Hotnarko R., Young Cr., Burnt Bridge Cr. and Hunlen Cr. - see figure 2.2) which either have cut deeply into the easily eroded volcanic rocks forming precipitous canyons for tens of kilometers or, where rock types are more resistant, formed waterfalls with vertical drops as large as 250 m.

(2.2) Bedrock Geology

Physiography and drainage patterns of the Bella Coola basin are influenced by a diversity of rock types and geologic structures characteristic of the Coast Plutonic Complex and Intermontane Volcanic Belt of the western Canadian Cordillera. The geologic setting of the study area described below fits the general pattern of the east slope of the Coast Mountains (*cf.* Roddick and Hutchinson, 1972).

The geology of the Bella Coola area has been mapped in detail by Baer (1973) and more generally by Tipper (1979) (figure 2.3). The oldest rocks are middle Triassic plutons of quartz-diorite and granodiorite in the southeastern portion of the catchment and on the north flank of Bella Coola River near Stuie. Metamorphism of varying intensities has generated gneissic and schistose foliation in some of these rocks, particularly in

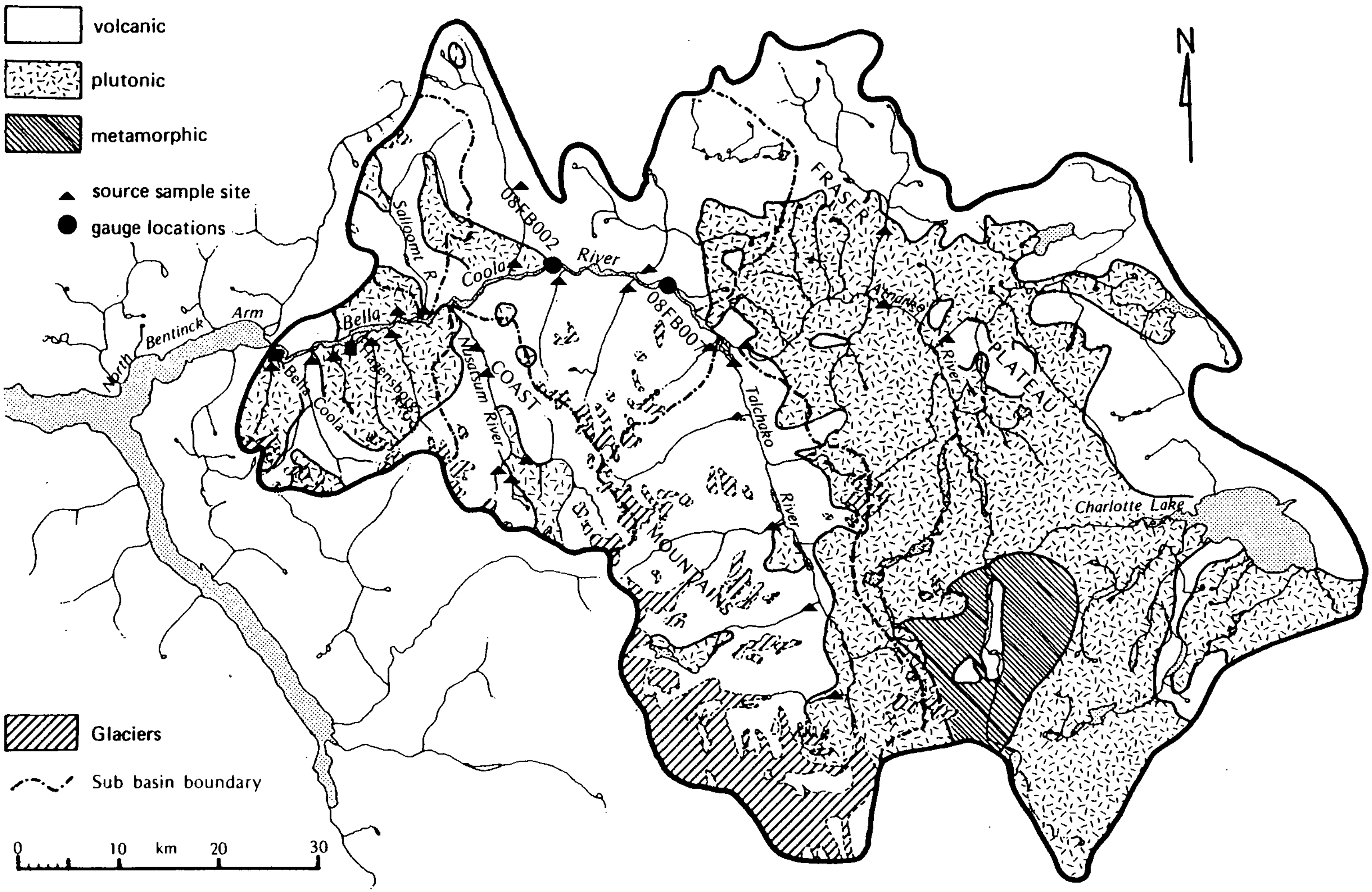


Figure 2.3 Distribution of bedrock lithologies in the Bella Coola basin (after Baer (1973) and Tipper (1979)). Triangles are sampling locations used to characterize lithological and mineralogical attributes of sediments derived from tributary basins.

the headwaters of Atnarko River. Volcanic, metavolcanic and sedimentary rocks (Triassic to lower Cretaceous) characterize the central and southern portions of the catchment comprising chlorite-rich greenstone and andesite. The volcanics exhibit widely scattered shear zones occupied by slate and argillite.

Eocene quartz-monzonites and granodiorites have intruded these older volcanics and form competent, steepened slopes where present. Joint spacing is wide and fracture density is low in these younger plutonic rocks; thus, the quantity of weathered debris is low and the texture is usually large blocks of angular debris. Miocene vesicular basalts and rhyolites are found in the extreme northeast marking the boundary between recent plateau basalts derived from a set of three shield volcanoes of the interior region to the east, and the older volcanic and plutonic rocks which dominate the west (Souther, 1986). Table 2.1 is a summary of the principal formations, rock types and constituent mineral assemblages found in the Bella Coola basin.

Topographic expression of structural features and formation boundaries is not well defined in the region. Baer (1973) attributes this to the similarity of mechanical properties (*e.g.* hardness, jointing) for most rock types. Orientation of individual ranges is not consistent throughout the area except for the regional northwest trend of the Coast Plutonic Complex. Faulting does not appear to be common although relative displacements in volcanic rocks are difficult to observe. Baer (1973) mapped two dominant fault directions, northerly and northeasterly, both of which separate the older chloritized greenstone and granodiorite from the younger andesites.

Uplift has occurred during two orogenic events: a less severe Lower Jurassic event and a stronger period of uplift in the late Tertiary. Baer

Table 2.1. Summary of formations and constituent mineral assemblages of rock types found in the Bella Coola basin after Baer (1973) and Tipper (1979).

Era	Period	Lithology	Minerals
Cenozoic	Lower/ Pliocene	rhyolitic lava	- very fine-grained mafic minerals - occasional obsidian
	Upper/ Miocene	vesicular basalt	- fine-grained mafic minerals - small amygdaloids
	Eocene	quartz monzonite	- 20-30% quartz and 35-45% pagioclase - k-spar as high as 35% - biotite, hornblende and muscovite
	Eocene/ Paleocene	granodiorite	- 25-30% quartz and 40-60% plagioclase - minor k-spar, biotite is the only mafic mineral
Mesozoic	Cretaceous/ Jurassic	andesite flows and agglomerates	- fine-grained plagioclase and minor k-spar - plagioclase phenocrysts, 1-5 mm long - minor epidote and quartz infilling
	Middle Jurassic	black slate/ argillite	- fine-grained phyllosilicates with occasional microconglomerate or grit with pebbles of quartz monzonite
	Lower Jurassic	greenstone	- subhedral to anhedral plagioclase (<0.5 mm) - fine grained matrix of chlorite, hornblende, epidote
	Middle Triassic	quartz diorite	- up to 35% anhedral quartz and small amounts of k-spar - plagioclase, biotite, hornblende, epidote and minor chlorite

(1973) and Parrish (1982) have attributed most of the present geologic structures to the latter event. Rock shoulders on the eastern flanks of the larger massifs along Talchako River have been interpreted as remnants of a Tertiary erosion surface which extends eastward as the Fraser Plateau. Uplift, estimated to be as much as 1800 m (Souther, 1977, 1986), warped the erosion surface and realigned the drainage such that the Atnarko and Talchako Rivers, which were thought to flow eastward from a divide centered along the axis of the Coast Mountains (Tipper, 1971a, 1971b), are supposed to have been captured by knick point incision of a rapidly growing, orographically-fed, Bella Coola River (Baer, 1973).

(2.3) Late Quaternary Geology

The major valley/fiord systems of British Columbia have been deepened by multiple glaciations. The severity of Wisconsinan glacial activity near the end of the Pleistocene has precluded the preservation of ice-contact materials dating from events earlier in the Quaternary. However, the entire B.C. coast probably experienced ice-advances essentially synchronous with the chronology identified for the southern coast (Armstrong, 1981).

Deteriorating climate following the Olympia non-glacial period resulted in the advance of ice (28,000 - 15,000 years B.P.; Clague, 1981). Ice flow was from major accumulation areas along the axis of the Coast Mountains. From here ice spread both east and west, initially occupying major valleys and fiords until ice thickness exceeded the local relief, resulting in flow directions less dependent on the underlying topography (Clague, 1981). Erratics found on the shoulders of extinct active shield volcanoes (Rainbow, Ilgachuz and Itcha Mountains) to the east of Bella Coola, along with ice-beveled platforms on some of the highest peaks in the Monarch Icefield, indicate that ice thicknesses exceeded 2000 m (Tipper, 1971a). Ice flow directions estimated by striae, drumlins, eskers, and fluted terrain show northeasterly flow away from Monarch Icefield towards Charlotte Lake and westerly down the Bella Coola Valley (Baer, 1973; Tipper, 1971b). During the maximum advance, about 15,000 years B.P., ice from the Coast Mountains probably extended as far west as the continental shelf (Clague, 1981) and as far east as Williams Lake, there merging with west-flowing ice from the Cariboo Mountains (Tipper, 1971a).

Evidence from the Prince Rupert-Kitimat and Bella Bella areas indicates that deglaciation along the outer coastal lowlands was rapid with glaciers retreating to their respective fiord head positions by 12,500 years B.P. (Clague, 1981; Andrews and Retherford, 1978). Although no dating control is available to the east of the divide, the greater elevation and ice thicknesses there may have resulted in a less rapid retreat. Coastal and valley bottom deposits laid down by the eastward retreating ice were related to the position of sea-level, controlled by isostatic and eustatic responses of the land and ocean to melting ice. Relative sea levels were 150 to 200 m higher until approximately 9000 years B.P. (Andrews and Retherford, 1978; Clague *et al*., 1982), which resulted in ice fronts terminating in the sea, producing conditions similar to those found today in Glacier Bay, Alaska (Powell, 1983). Glacigenic sediments deposited in proglacial fiords comprise a variety of materials indicative of these changing environmental conditions. The characteristics of the Bella Coola valley-fill have been identified using local borehole and regional data and sedimentary evidence found in outcrops (figure 2.4).

Overall the valley-fill represents facies produced by glacial, glaciomarine, marine and alluvial processes. Thick deposits of glaciomarine silty-clay were deposited during the marine inundation phase between 12,500 and 9,000 years B.P. At the same time, tributary streams were depositing

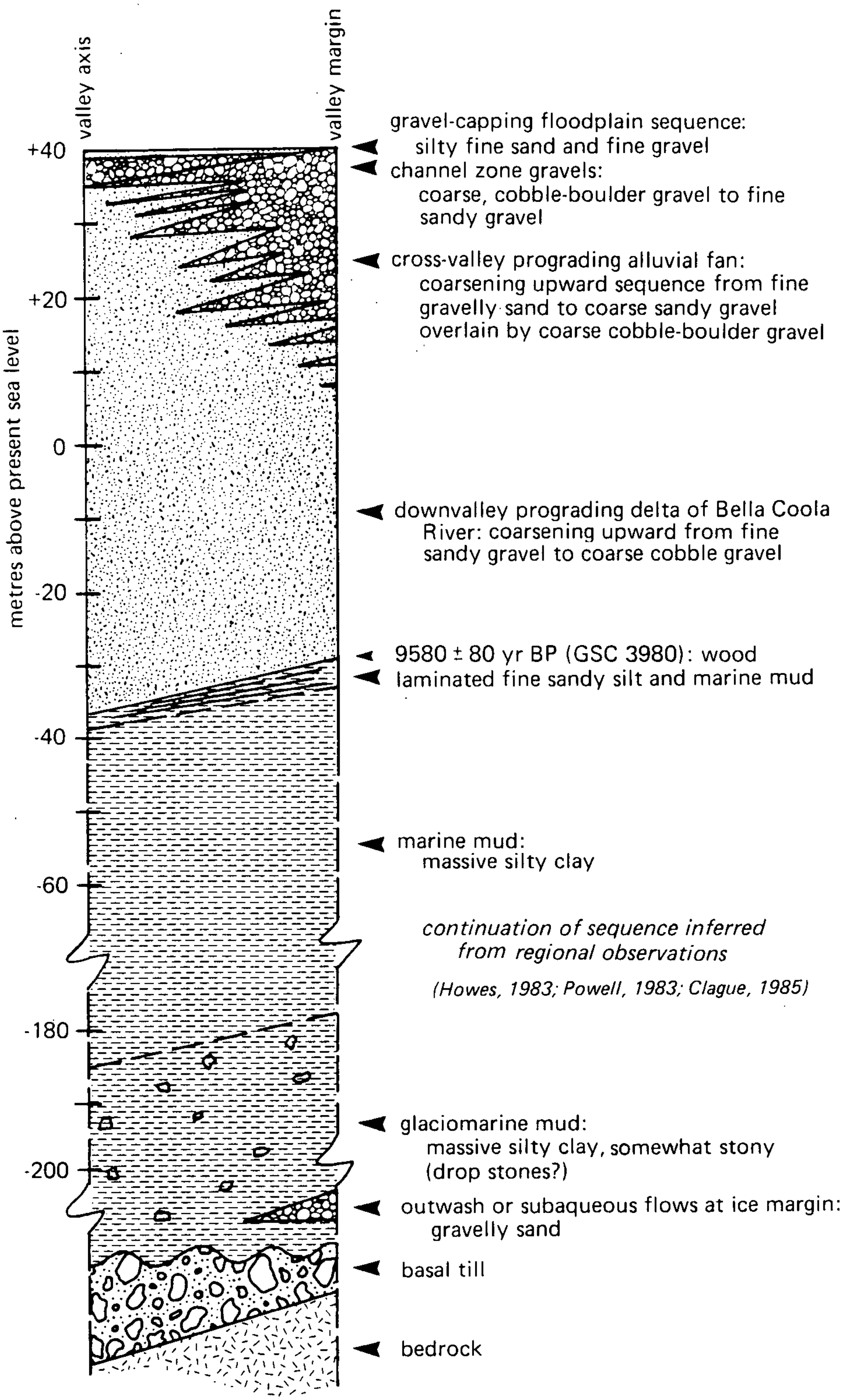


Figure 2.4 Observed and inferred valley-fill in the lower reaches of Bella Coola River near Snootli Creek.

material in deltas along the margin of Bella Coola valley. Continued ice recession and progradation of the main valley delta produced thick deposits of sand and gravel in a coarsening upwards sequence. Concomitantly, alluvial fans along the valley margin prograded across the valley depositing sandy gravels and coarser cobble material.

Till of varying thickness mantles most slopes above the valley floor. Prominent late-glacial features such as terminal or recessional moraines are rare in this catchment as in many other Coast Mountain valleys. Ryder (1981) has attributed this scarcity of deglaciation deposits to one of two possible mechanisms: (1) rapid but orderly retreat of ice fronts allowing for nothing more than brief stationary periods and minimal deposition or (2) *in situ* ablation of main valley ice which has been cut off from the headwater ice source.

A hummocky, kettled ridge extending part way across the narrow Bella Coola valley above Nusatsum River may be a subdued terminal or medial moraine that was formed during a brief pause in late glacial ice retreat. Although morphology indicates that the ridge is ice-contact in origin, exposures of stratigraphy along margins of the deposit are mostly deltaic, and dip directions of foreset beds and constituent lithologies suggest that sediment and water were derived mostly from the upper Bella Coola valley, rather than from the adjacent Nusatsum River. In any event, the bedrock-controlled restriction of the main valley and confluences of the Salloomt and Nusatsum Rivers at this position, appear to have formed a complex depositional environment characterized by glacial, marine and alluvial processes.

With the exception of deltaic deposits and a high terrace of marine deposited clayey-silts, late-glacial deposition is found mostly in the upper Bella Coola valley comprising marine deltas, valley bottom kames, till benches, lacustrine silts, eskers and some eolian transported sand deposits. However, the number and size of these deposits are limited.

Clague *et al*. (1982) and Clague (1985) demonstrate that sea levels were within 11 m of present by at least 9720 years B.P. for the Bella Bella area. A date of 9,550 ± 80 years (GSC 3964) was determined for wood sampled near the contact of subareally exposed marine clay and overlying alluvial deltaic sands near the outlet of Salloomt River (elevation is 61 m.s.a.l.). The elevation of this contact indicates that relative sea level in the Bella Coola area was still higher at that time than at present, although marine regression at the site occurred soon after. Evidence from both north and south coast regions indicates that isostatic uplift, sea level adjustments, and ice retreat were all complete by 8,000 years B.P. (Armstrong, 1981; Clague, 1985).

Glacier fluctuations during the Holocene Epoch have been documented for a number of different locations along the northeast Pacific rim from Washington to Alaska (Mathews, 1951; Miller, 1969; Easterbrook and Burke, 1972; Denton and Karlen, 1977; Mann and Ugolini, 1985; Ryder and Thomson, 1986). Early Holocene advances are recognized only in the Aleutians (Black, 1981; 1983), Copper River (Sirkin *et al*.,1971) and Lituya Bay districts (Mann and Ugolini, 1985) of Alaska, and even then appear to be asynchronous suggesting that they were localized events, generally not representative of the warmer and drier conditions which dominated the northeast Pacific between 6,000 and 10,000 years B.P. (Clague, 1981).

Miller (1969) and Ryder and Thomson (1986) document evidence for an early Neoglacial advance (Garibaldi Stade) approximately 5,000 years before present. Similar evidence has not yet been found in mountain regions to the north of the Garibaldi Park region in southwestern British Columbia.

Several investigators have indicated that an advance of mountain glaciers occurred between 1800 and 3000 years B.P., which varied in intensity, duration and timing between regions, but is generally recognized throughout the coastal region from Alaska to Oregon. Tiedemann Glacier to the south of Bella Coola basin reached its maximum Neoglacial extent during this period, unlike most other glaciers within the region (Ryder and Thomson, 1986).

In accord with chronologies developed for Dome Peak and Mt. Rainier in Washington (Crandell and Miller, 1964), Garibaldi Park (Mathews, 1951) and southeast Alaska (Mann and Ugolini, 1985), Ryder and Thomson (1986) found evidence within the Pacific Ranges for ice advances commencing around 1,000 years B.P. with maximum positions achieved by 300 to 400 years B.P. In most regions this was the most pronounced excursion of alpine glaciers throughout the entire Holocene Epoch. A regionally synchronous response of glaciers appears to have occurred during recession from the Little Ice Age maximum, beginning 100 to 350 years B.P. (Field, 1975). Evidence presented in Chapter 7 indicates that a Little Ice Age chronology, similar to that developed by Ryder *et al*. (unpublished) for regions to the south of the study area, is also applicable to the Bella Coola basin.

(2.4) Climate Setting

Bella Coola River and the rest of the British Columbia coast fall within the Pacific coast maritime climate regime, characterized by the on-shore movement of cyclones and anticyclones generated in the moist, and usually warm, North Pacific region. The intensity and frequency of cyclones vary with latitude and season, and annual climate along the coast is controlled by the seasonal pattern of fronts and air masses. Kendrew and Kerr (1955) recognized the importance of several air masses and their associated fronts. During summer a northward shift of the maritime arctic front towards southeast Alaska is accompanied by the increasing influence of mild Pacific air and the occasional incursion of maritime tropical air along the central and southern B.C. coast. Warm and dry conditions prevail, although the pattern is not invariable, particularly if high pressure ridging is poorly developed, and moisture bearing cyclones migrate southward.

The transitional seasons (spring/autumn) are marked by rapid shifts in mean frontal positions and changes in zonal circulation intensity. As the winter season progresses, convergence of major air streams along the north coast bisects the region into a northern half, influenced by maritime arctic air, and a southern half, dominated by warmer and moister maritime Pacific air (Wendland and Bryson, 1981; Yarnel, 1985). Throughout this season the continental arctic front tends to occupy a position along the axis of the Coast Mountains forming a strong ridge of high pressure extending vertically to mid-tropospheric levels and horizontally as far west as the outer coast (Hare and Hay, 1974). Hence, the eastward movement of cyclonic disturbances can be impeded. Fronts nearly always occlude and the warm sector aloft yields high precipitation along the coast. Outbreaks of arctic air are not uncommon early in the winter season (Baudat and Wright, 1969), and it is the timing and duration of these anomalously cold and dry periods which determine snowpack stability and the potential for winter flooding and spring avalanche activity.

A succession of cyclones intersects the coast during the winter season and the separation and intensity of these systems are mostly dependent on origin. Kendrew and Kerr (1955) suggest that three source areas for cyclone development are important: (1) a southern group which start as waves on the Polar front in the west Pacific, tracking northeastward towards the B.C. coast; (2) a northern group generated by the effects of Polar air undercutting Maritime Arctic air off Japan and then tracking east into the Gulf of Alaska; and (3) a transitional group forming between fronts in the western and central Pacific. Occlusion and instability through a considerable depth are common to all groups of cyclones as they approach the coast. With decreasing storm separation there is an increase in cumulative rainfall leading to a higher snowpack volume and water content and thus greater potential for high-magnitude winter and spring runoff.

Compounding the effects of synoptic-scale circulation features are local and regional topographic controls on temperature and precipitation. Variability of mean temperature, temperature extremes and total precipitation, including the proportions of snowfall and rainfall, all show marked horizontal and vertical gradients. For a series of three stations forming a west-to-east transect from the outer coast to the interior, mean annual temperatures decline from $8.1^{0}C$ to $3.9^{0}C$, and total annual precipitation decreases by an order of magnitude from 4390 mm to 410 mm (figure 2.5). This strong inland gradient from a mesothermal, perhumid climate to one of continental character occurs over a distance of less than 200 km. Superimposed on the horizontal precipitation gradient are large, orographically induced variations in precipitation with elevation. Using May 1 snowpack water equivalents, recorded 1800 m above the Bella Coola climate station, as a measure of total winter precipitation at this elevation, it appears that, on average, precipitation increases by approximately 100 mm per 100 m of elevation.

Approximately 60% of the Bella Coola basin lies in the lee of the Coast Mountains, which is reflected in the distribution and accumulation of snowfall during the winter season. Snow course data, collected for a

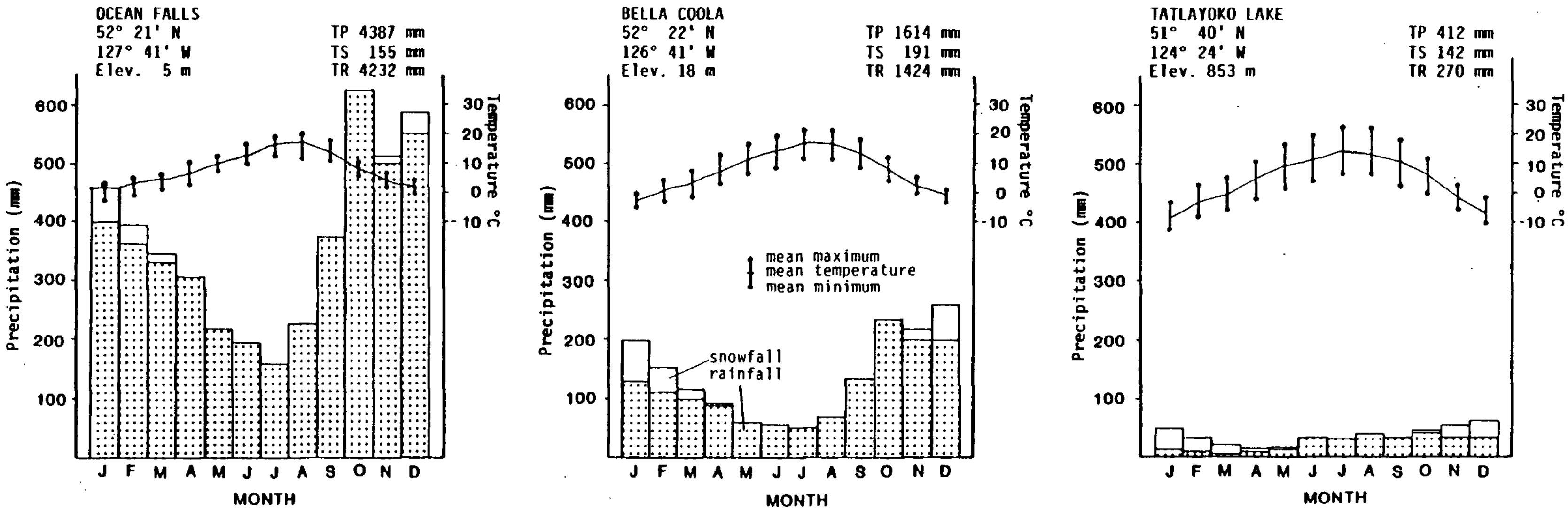


Figure 2.5 Climographs for a west to east transect through Bella Coola basin. Stations are located in figures 2.1. and 2.2. TP is total precipitation, TR is total rainfall and TS total snowfall (Source: Atmospheric Environment Service, 1981).

limited number of corresponding years from an eastern and a western site at similar elevations within the catchment show that late spring snowpack water equivalents are 8 to 10 times greater in the west (record mean of 1753 mm) than those in the east (record mean of 230 mm).

(2.5) Hydrology

Average Regime

Within the basin, nine river gauges have been operative at various times since 1930, but only five provide data for periods longer than five years after 1948. The combined data series from two stations on the main river cover the period 1948-present, while three of the tributary gauges (Atnarko, Nusatsum, Salloomt) were established during the summer of 1965 and have been in operation continuously since then. The main river gauge, relocated in 1965, is only a few kilometers below the confluence of Talchako and Atnarko Rivers, the latter of which is gauged just above the confluence, allowing for estimates of flow in the ungauged Talchako River by simple subtraction. Gauge locations and a summary of hydrometric data from each site are given in figure 2.2 and table 2.2 respectively.

Variability in basin-wide runoff is strongly affected by the distribution of snow. It also is affected by the location of short and long-term water storage sites such as lakes and glaciers. Mean annual runoff from the basin above Burnt Bridge Creek is 790 mm with a maximum of 1023 mm recorded in 1976. For partly glacier-covered tributary catchments situated in the southern and western portions of the basin (*e.g.* Nusatsum River) mean annual runoff is significantly higher, averaging over 2000 mm (table 2.2). Mean annual runoff from the larger Atnarko River draining ice-free terrain is one third of the discharge from Talchako River, which drains a 25% glacier-covered area. Inter-correlation of annual mean and

Table 2.2. Hydrometric data for stations in the Bella Coola basin. (Source: Water Survey of Canada, 1982)

W.S.C. Station name (number)	Interval of Operation (# of years)	Drainage Area (km^2)	Mean Annual Discharge ($m^3 s^{-1}$)	Mean Annual Runoff (mm)	Maximum Daily Discharge ($m^3 s^{-1}$) Spring (date)	Autumn (date)	% Difference of Autumn vs. Spring[1]
Bella Coola R. near Hagensborg (08FB002)	1947-68 (21)	4,040	119.0	930	668 (06/11/64)	963 (01/24/68)	144%
Bella Coola R. above Burnt Bridge Cr. (08FB007)	1965-present (21)	3,730	94.0	790	558 (06/11/69)	703 (01/23/68)	126%
Bella Coola R. at Bella Coola (08FB001)	1929 (1)	5,050	139.0	870	405 (07/09/29)	430 (16/10/29)	--
Bella Coola R. at Hagensborg (08FB008)	1930-32 (2)	4,800	135.0	890	561 (10/06/30)	561 (09/17/30)	--
Nusatsum R. (08FB005)	1965-present (21)	269	17.1	2,010	83 (16/05/70)	190 (09/27/73)	229%
Salloomt R. (08FB004)	1965-present (21)	161	9.3	1,820	43.3 (16/05/70)	141 (12/16/80)	326%
Atnarko R. (08FB006)	1965-present (21)	2,430	31.3	410	258 (30/05/72)	289 (01/24/68)	112%
Tastsquan Cr. (08FB003)	1946-1950 (4)	29	2.2	2,390	21.6 (06/18/50)	36.5 (24/10/47)	--
Clayton Falls Cr. (08FB009)	1980-present (5)	481	6.5	426	21.2 (05/21/82)	85.6 (08/09/82)	--
Talchako R.[2]	1965-present (21)	1,300	62.7	1,521	407 (06/11/67)	552 (01/23/68)	136%

1. Calculated as $[Q_A/Q_S]$ X 100 where Q_S is the maximum daily discharge due to spring snowmelt and Q_A is the maximum daily discharge due to an autumn rainstorm for the period of record. Not computed for stations with a limited length of record.
2. Flows estimated by subtracting Atnarko R. data from Bella Coola R. data above Burnt Bridge Creek. No major tributaries join the system between these two gauging sites. Flows may be underestimated by 2-3%.

annual maximum discharges indicates that Nusatsum/Talchako Rivers are hydrologically distinct from Salloomt/Atnarko Rivers, primarily on the basis of glacier melt contributions to runoff and higher mean elevations. However, the Atnarko River retains a somewhat singular signal due to the much lower annual precipitation and frequency of lakes, which tend to reduce peak discharges.

Several seasonal runoff regimes can be identified for Bella Coola River, which are typical of other rivers along the British Columbia coast. Beginning in late April through mid to late June, runoff due to snowmelt is somewhat variable, but steadily increasing discharges are evident. Occasionally, runoff is augmented by rainfall on the ripened snowpack. Following the snowmelt peak in early to middle June (or late May for some rivers), runoff declines until late July, when discharge again increases as glacier melt begins to contribute significant volumes of stored water (figure 2.6). Superimposed on steadily declining river discharges from late August onwards are high-magnitude synoptic runoff events derived from two sources: (1) rainfall derived floods with only minor contributions from melting snow or (2) rain-on-snow events associated with rapidly rising freezing levels following the eastward movement of autumn cyclonic disturbances. Finally, from January to April winter low flows dominate the runoff regime.

Maximum daily and instantaneous discharges almost always occur during the short-duration autumn floods (table 2.2), but the highest average runoff is during the summer snow and glacier melt period (figure 2.6). Hart (1981b) has shown that the ratio of maximum instantaneous to mean daily discharge for autumn floods in the Kitimat and Pacific Ranges varies with drainage area but is generally between 1.5 and 1.9. Maximum daily discharges from rainstorm-generated floods in the gauged catchments are on

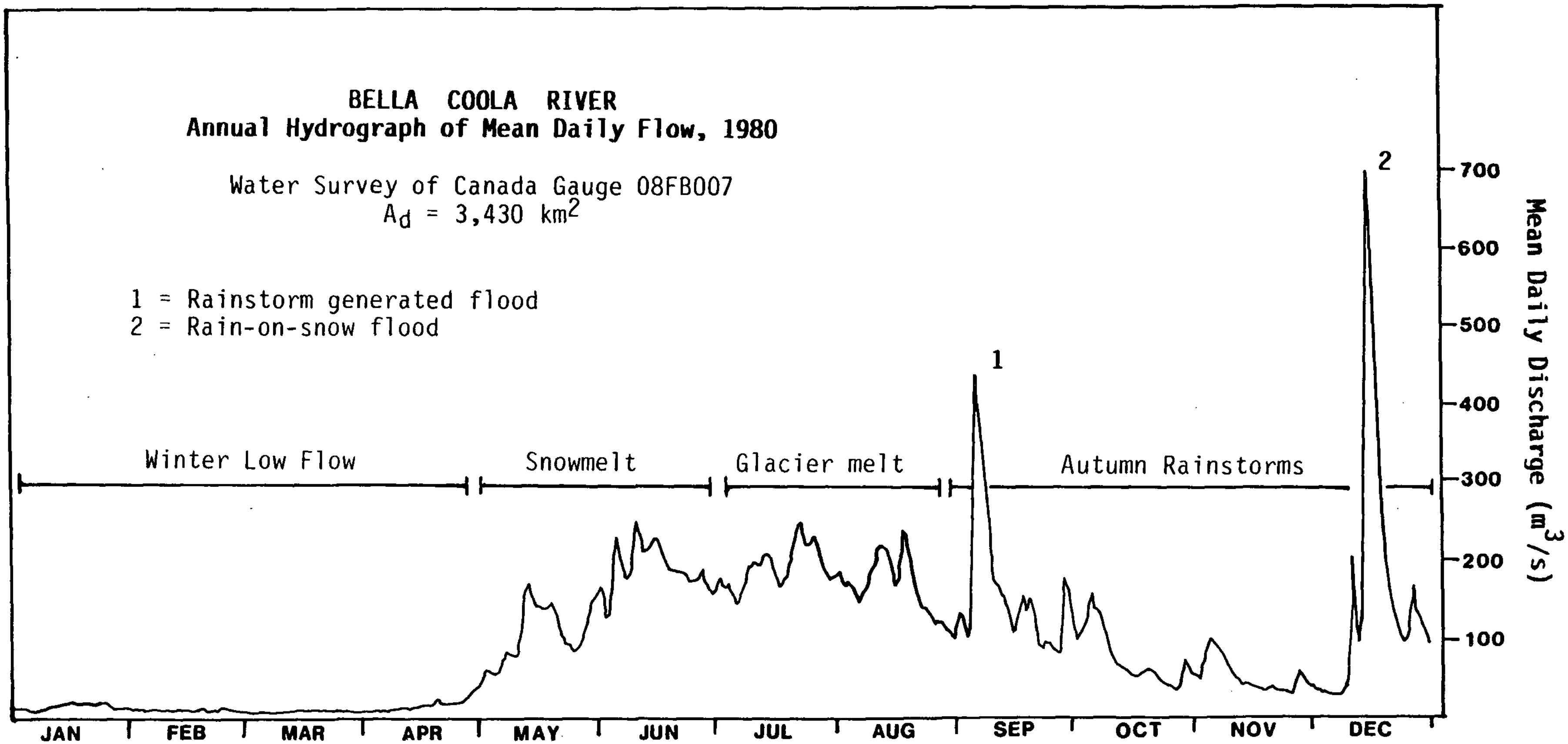


Figure 2.6 The 1980 annual hydrograph for Bella Coola River above Burnt Bridge Creek.

average 25-40% greater than spring snowmelt floods. However, as glacier cover and local relief increase and basin size decreases, this difference can be as high as 300%.

Major Floods During the 20th Century

Historical and instrument records show that there were approximately thirteen major flood events between 1896 and 1980. The pre-1948 record of less extreme flooding is probably incomplete, however; all the largest flows were documented. With the exception of a flood in the fall of 1950, all high-magnitude, autumn and spring runoff events on the Bella Coola River were associated with winter seasons (October to April) characterized by positive precipitation anomalies. Only six other seasons in the 88 year record exhibited strongly positive precipitation anomalies and no evidence of responses in the form of significant flooding. Large runoff events were not prevalent in these years because in all six cases spring and summer temperatures were well below normal promoting water storage.

The most outstanding flood events occurred in the middle 1930's and 1960's. The four floods (1934, 1936, 1965, 1968) are among the six largest measured or estimated flows on the river. As such they represent 'key events' in the long-term hydrology of the basin because in each instance, particularly the rain-on-snow events, major channel realignment or significant overbank sedimentation occurred. For this reason a brief description of the hydroclimatic circumstances associated with these two sequences is given, and the implications for effects from lower-magnitude events are considered briefly.

The flood of October 10, 1934 resulted from the occlusion of fronts associated with an intensified low pressure system (960 mb) to the west of Vancouver Island. A blocking ridge of high pressure over the western United

States led to the low stalling as it approached the coast. High-water appears to have been more localized because of significant log jams on the river. Storage of organic debris in the channel was probably significant since the last documented flood was ten years before. Flooding on November 19, 1936 was more severe than the 1934 flood for two reasons. First, a series of low pressure systems moving southward from the Gulf of Alaska, preceded the larger system yielding snowfall at middle and at lower levels in the basin. Second, a well developed warm front swept through the area raising the freezing level substantially. A melting snowpack contributed greatly to the runoff.

Synoptic conditions for the October 22, 1965 flood were similar to those in 1934: northward moving disturbances produced seven-day total precipitation of 190 mm prior to the peak discharge; one of the largest flood-related rainfall periods. Ridge development in Alberta was characteristic of the January 23, 1968 flood. Like the 1936 flood, peak discharges were much higher due to substantial snowmelt as the freezing level increased during a four day period prior to the flood. Maximum instantaneous flows from this event were probably the largest to occur this century.

With the exception of moderately high spring runoff in late May of 1948, there were no flood events of notable between 1936 and 1950. The flood of November 4, 1950 was the fourth largest estimated flow and appears to have had minimal impact in terms of channel changes. Two factors might explain this. First, both the rising and falling limb of the flood hydrograph were very steep so the peak discharge was of short duration. Second, runoff in two of the preceding three springs was significantly high (although not of major flood proportions) so that any accumulation of organic debris in the channel may have been removed. The implication is that not all large runoff events may have an impact on the river. There is some evidence for incipient instability in certain reaches of the river such that moderate floods may produce substantial localized channel changes. In most cases the instability is related to the occurrence of log jams which, when removed, modify the direction of channel flow.

(2.6) Sources and Transfer of Clastic Sediment

An important step in determining the trends of basin sediment yield and possible associations with changing hydroclimate is consideration of sediment sources and processes of sediment transfer. The purpose of this section is to document the characteristics of the sediment cascade and assess the distribution of several primary sediment sources and residence times in associated sediment sinks.

A conceptual model, shown in figure 2.7, has been adapted from those given in Simons *et al.* (1982) and Roberts and Church (1986), and is used here to illustrate the primary sources, transfer and storage of clastic sediment in the Bella Coola River basin. Three primary sediment sources can be identified: (1) *in situ* weathered bedrock; (2) glacial deposits in proglacial and ice-marginal areas of contemporary glaciers; and (3) clastic materials, including soil, which mantle hillslopes or infilled gullies, the bulk of which was deposited during the late Pleistocene and early Holocene epochs. These materials are mobilized by mass wasting (rockfall/ rockslide, avalanche, debris flow, gully wash, soil creep) and fluvial processes.

Storage of recently produced sediment in the upland or headwater areas is in the form of colluvial deposits or partly infilled proglacial lakes which were formed during ice retreat. Residence times will vary as a function of reservoir capacity, slope stability and sediment transfer rates. Fluvial erosion and mass wasting of slope deposits may introduce

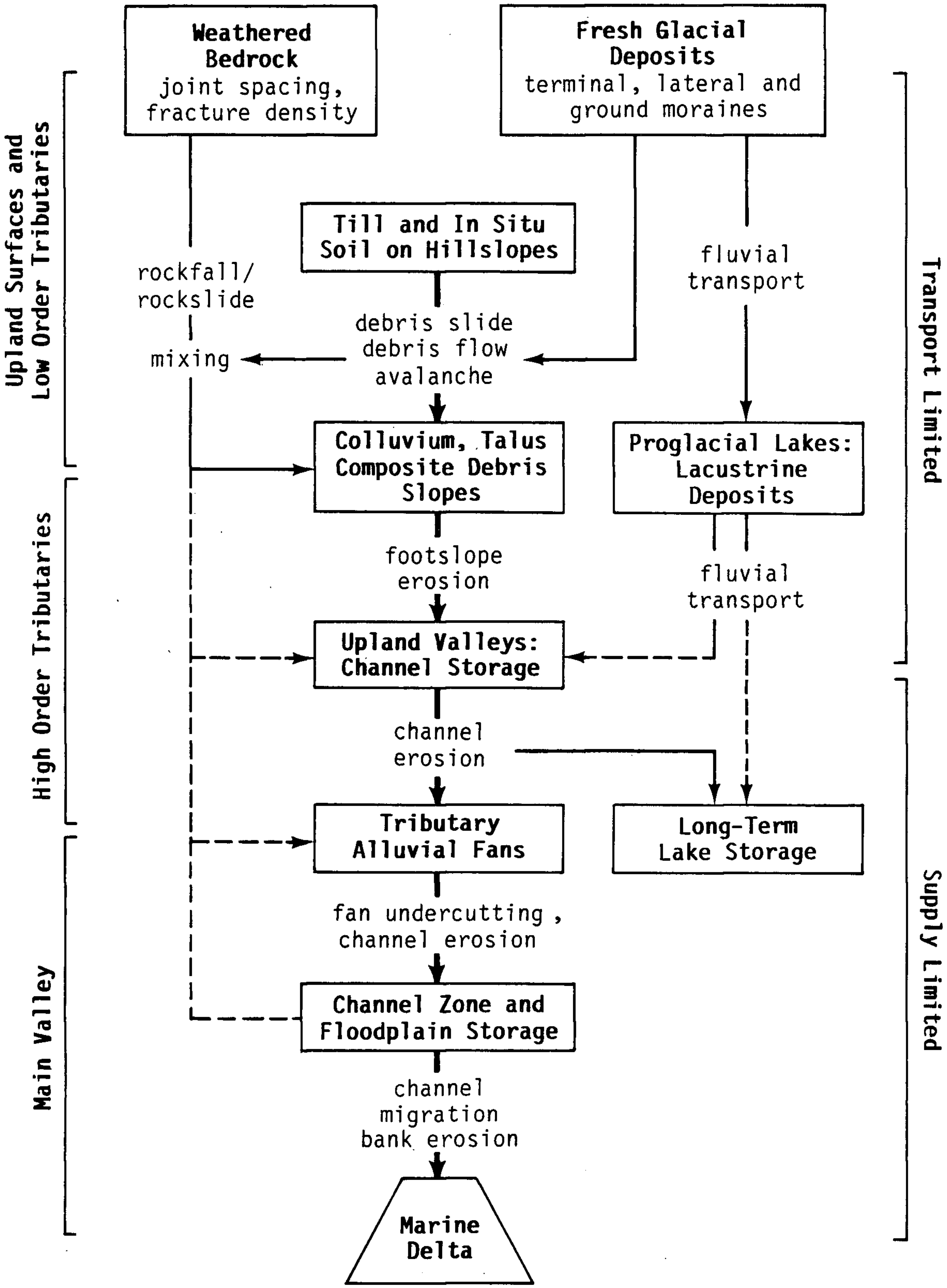


Figure 2.7 Generalized clastic sediment routing and primary transport processes in Bella Coola basin.

clastic debris directly to the channel zone of low-order tributaries. Grain sizes are usually very coarse (bouldery to blocky) and thus transport-limited conditions prevail. Fluvial transport moves some of the sediment into downstream storage sites such as tributary alluvial fans, upland valley trains, lakes along principal drainage lines or ultimately to the channel and floodplains of higher order streams. If storage capacities are high and sediment transfer rates low, a transition to supply-limited conditions is likely to occur downstream. Finally, fluvial processes along the main stem of the system will pass sediment through to the basin outlet where sediments are deposited in a growing marine delta.

Interpretation from aerial photographs offers the only practical method of evaluating clastic sediment sources and differential movement rates, even though the technique yields limited quantitative information on storage volumes and deposit stability. Two scales of mapping were undertaken to evaluate sediment sources: (1) interpretation of basin-wide features from medium scale (1:21,000 to 1:40,000) black and white air photography and (2) detailed air photograph and field mapping of three subcatchments: Salloomt River (169 km^2), Burnt Bridge Creek (215 km^2), and Nusatsum River (269 km^2), which are representative of the spatial variations in hydrologic, physiographic and geologic conditions within the basin.

Upland Sediment Sources

Seven types of sediment sources were identified: composite debris slopes, talus slopes (gravity sorted), alluvial fan or talluvial debris cones, rockfall/rockslide deposits, *in situ* mountain top detritus or felsenmeer, lateral and proglacial till deposits, and blankets or veneers of till along forested slopes.

Composite debris slopes are polygenetic landforms produced by any combination of rockfall, avalanche and debris flow (*cf.* Rapp, 1960; Chandler, 1973; Church *et al.*, 1979). Highly fractured and jointed greenstone rocks in the basin are an important source for these blocky deposits which form below open rock slopes dissected by numerous avalanche and rockfall chutes. They are also found on confined rock slopes above the flanks of active cirque glaciers, where removal of ice by melting has promoted the failure of rock surfaces. Although widely distributed and of considerable volume throughout the volcanic terrain, these slopes are not overall an important source of onward moving clastic sediment due to the coarse-grained nature of the debris and transport limited conditions which result.

Talus slopes are gravity-sorted, rockfall dominated landforms, forming discrete but small sheets or multiple coalescing cones, primarily above treeline, or below steep bedrock cliffs at lower elevations. These deposits occur frequently along the edges of former lava flows and below extension faults in the plateau areas of the northeastern catchment, but are poorly connected to the fluvial network.

High magnitude rockfalls/rockslides are defined on the basis of criteria suggested by Mudge (1965), the most important of which are the slope-base position of coarse blocky debris and a visible upslope failure surface. The largest deposits occur on terrain underlain by plutonic or metamorphic rock types. For example, of the twelve deposits identified in the Nusatsum basin, eight are derived from intrusive rocks and four from volcanic/sedimentary sources. This is significant given that the distribution of rock-type is about the reverse (figure 2.3). Most recognized deposits are at or above treeline and in some cases have impounded small lakes. Estimates of volume are difficult to make because of unknown depths, but appears to range downward from 10^6 m^3. The largest rockslide mapped occurred in approximately 1951 in the headwaters of the Atnarko River as a result of a foliation plane failure in gneissic rocks (J.J. Clague, personal communication, 1986). The volume is estimated at 10^8 m^3. The deposit resulted in minor realignment of the drainage divide between the Atnarko and North Klinaklini Rivers.

Alluvial fans and talluvial debris cones are found below low-gradient and steep basin outlets, respectively, and usually contain a perennial surface stream along one margin of the deposit. Fluvial processes, snow avalanches and rockfalls deliver sediment to the surface of these deposits. Perennial streamflow on the surface and active toe erosion by higher order streams indicate that these landforms are an important source of sediment.

Block fields are found on elevated rock platforms, ice eroded benches and cols. Sediments are derived by *in situ* mechanical and chemical weathering of well jointed bedrock, particularly along non-conformable contacts of various volcanic rock types. They are perhaps the least important sediment sources because of limited volumes and stable upland positions.

Lateral and proglacial till deposits formed by the recession of valley, hanging valley, cirque and niche glaciers are the most important sources of sediment in glacierized portions of the catchment, particularly where the deposits are on steep terrain drained by avalanche and debris torrent channels (figure 2.8a and b). Although the volume of material delivered to the stream is low in relation to the amount available for transport (*i.e.* low sediment delivery ratio), aggradation and channel braiding are obvious impacts on the fluvial network immediately downstream from these sources. On a basin-wide average, lateral deposits are of less overall importance.

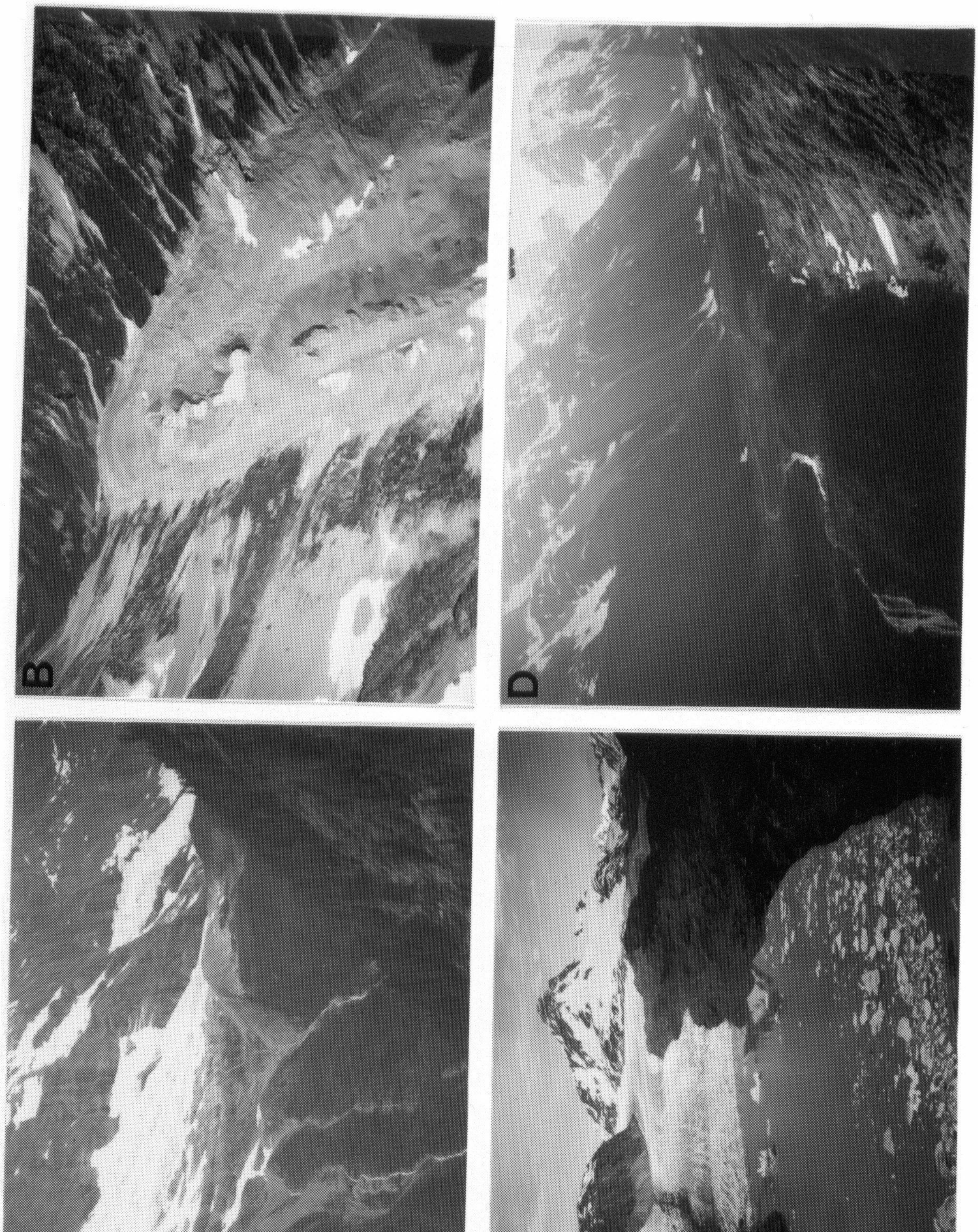
Figure 2.8 Illustrations of primary clastic sediment sources for the higher-order fluvial network in Bella Coola basin.

(A) Niche glacier on the western flank of the Nusatsum basin supplies a significant amount of sediment to channels below the glacier. The debris torrents are well connected to the high-order fluvial network.

(B) Rock glacier/ablation debris in the south fork of Gyllenspetz Creek is dissected by numerous avalanche chutes on easily eroded volcanic rocks. This site is a significant source of sediment for Talchako River.

(C) Proglacial lake of Jacobsen Glacier acts as an important storage site for ablation sediments and debris transported by ice marginal streams.

(D) Upland plain in the east Nusatsum valley is one of several storage sites for glacial derived sediments.


A
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D

The final sources of clastic materials are surface wash, soil creep and bank erosion. Roberts and Church (1986) estimate that surface wash and soil creep, on average, constitute less than 10-20% of the total sediment volume delivered to small, partly logged watersheds in the Queen Charlotte Islands. Relative delivery rates in large, unlogged tributaries, such as most of those in the Bella Coola basin, would be proportionally lower due to the stabilizing effects of the vegetation and a higher frequency of lower-gradient slope segments. Streambanks are an important source for fluvial entrainment of sediment but, estimates of contribution to sediment yield integrated over an entire watershed are difficult to make.

Four major types of sediment storage have been identified in the Bella Coola basin. First, relatively deep lakes along principal drainage lines, particularly in the Atnarko watershed, effectively trap all the clastic sediment derived from above each lake site. The largest, Charlotte Lake (65 km^2), is fed by creeks which drain an area of approximately 750 km^2, thereby eliminating any significant contribution from this area of the eastern catchment (see figure 2.2 for locations). Residence times in these sites are comparable in length to the interval of deglaciation (10^4 years). Shallow, proglacial lakes formed during recent glacial retreat (figure 2.8c) are less efficient sinks but equally important over a shorter time scale (10^2 years) (*cf.* Smith *et al.*, 1982).

Second, upland valley trains form on low-gradient, bedrock-controlled, valley flats at higher elevations in proglacial areas of the southwestern catchment, or on plateau areas of the eastern catchment. Glacial and fluvial sediments derived areas as large as 30 km^2 are stored at these sites (figure 2.8d). Trap efficiency is less than that of lakes but residence times are equally long (estimated at 10^3 - 10^4 years). Third, rockslide dams of various magnitudes also form effective barriers to the throughput of clastic materials. Their frequency of occurrence is much lower, but residence times are also long (10^3 - 10^4 years). Finally, colluvium and till, stored on many of the forested hillslopes, are probably the largest sediment sinks in the basin. Ultimate turnover rates are most likely in excess of 10^4 years.

Aerial photographs and field evidence indicate that much of the sediment in the upland tributaries is derived from well-defined, unstable slopes and bank erosion of material already stored within the valley flat. The distribution and volume of the major sediment sources, along with the locations of major storage sites, are plotted in figure 2.9. In several tributaries there are one or two primary sources, such as unstable proglacial deposits (*e.g.* Nusatsum, Nordschow, Thorsen Creeks) or composite debris slopes (*e.g.* Burnt Bridge, Nordschow, Noosgulch Creeks). The combined area of the sediment sources in figure 2.9 constitutes less than 16% of the entire watershed. These results suggest that the transfer of sediment to the Bella Coola River over a time scale of 10^2 - 10^3 years is associated with a relatively small number of upland sources and fluvial activity within several tributary channels.

(2.7) Pedologic Setting

Soil development in the Bella Coola basin reflects the strong variability in topography, frequent sediment additions from river flooding in the valley and, at higher elevations, slow decomposition of organic material. Within the confines of valley bottoms most soils vary between poorly developed orthic regosols and eutric brunisols (British Columbia Ministry of Agriculture, 1973). Thin Ah horizons (3 to 5 cm) over fine sand are common on many alluvial surfaces where, additions of mineral matter may exceed organic matter accumulation significantly. Under thick coniferous

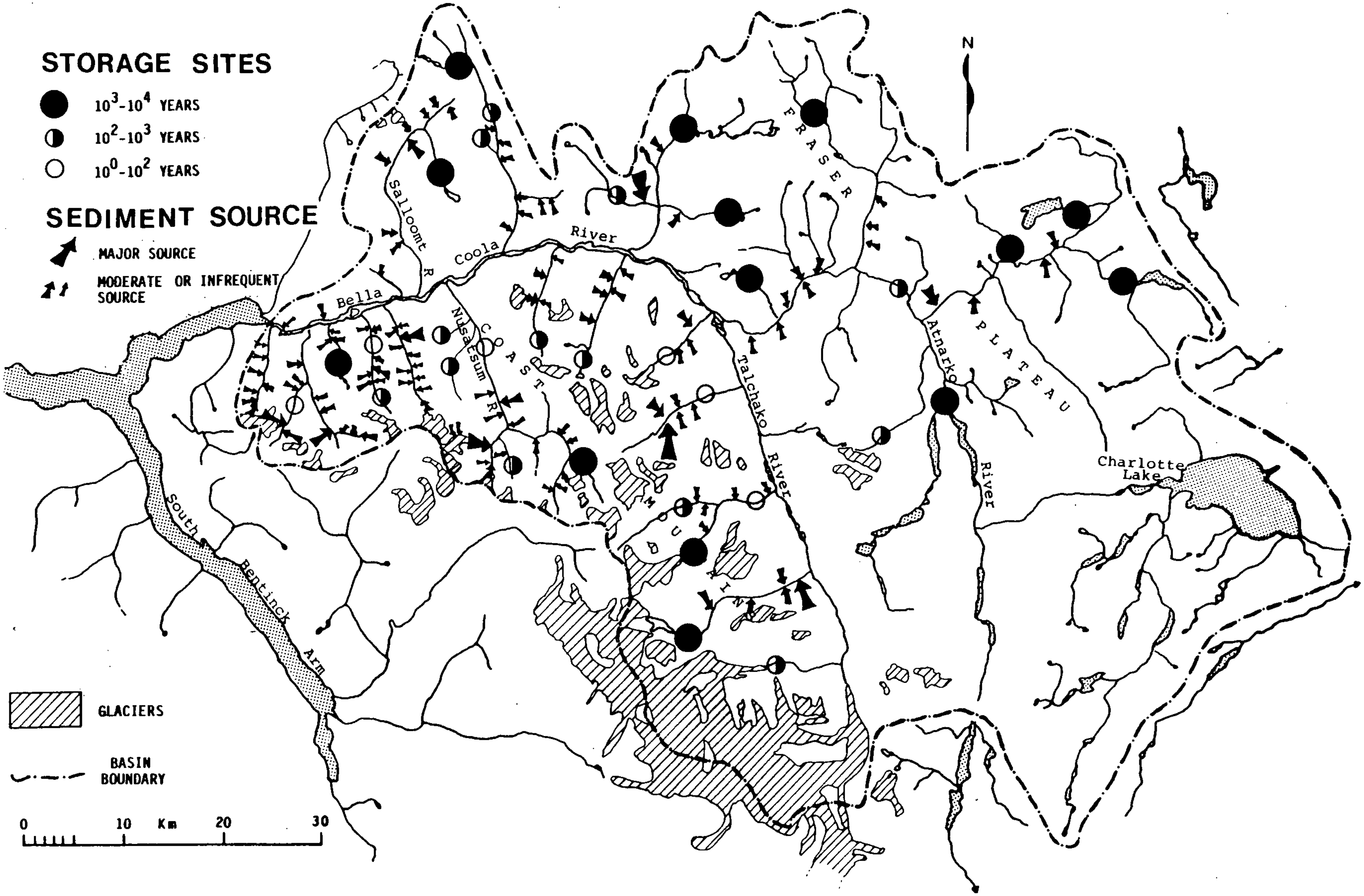


Figure 2.9 Major clastic sediment sources for the higher-order fluvial network and estimates of residence time in important storage sites. Arrows represent the location of both the sediment sources and primary routes through which sediment is transferred (see text for discussion).

forest cover, dystric brunisols may develop into orthic humo-ferric podzols if the site has remained stable for some time. Organic soils are restricted to poorly drained locations surrounded by both coniferous and deciduous vegetation. Typic mesisols and hydric fibrisols occupy many of the water-saturated alpine meadows.

(2.8) Flora

The floristic component of the landscape provides an essential key to geomorphic activity and hydrologic change within certain growth-time boundaries. The capacity of many species to withstand environmental extremes yet preserve evidence of stressful periods, provides correlations with geomorphic/hydrologic events, as well as a basis for absolute and relative dating. For this purpose some of the dominant vegetation communities are noted.

Six biogeoclimatic zones and several subzones have been identified in the Bella Coola basin, each based on the dominance or co-dominance of climax tree species (Robinson and Pojar, 1981; Leaney and Morris, 1981). Along the lower-elevation western periphery where mild, moist, maritime conditions prevail, a thick cover of western red cedar (*Thuja plicata* Donn) and western hemlock (*Tsuga heterophylla* (Raf.) Sarg.) is found. The oldest and tallest timber grows on comparatively organic-rich soils of the floodplain and steep colluvial slopes. At elevations below 500 m between Bella Coola and Firvale a mixture of Douglas fir (*Pseudotsuga menziesii* (Mirb.) Franco) and western hemlock overlie an understory of mosses and devil's club (*Oplopanax horridus*). This association grades quickly into a Douglas fir/red cedar/moss ecosystem east of Firvale.

The Bella Coola River floodplain is dominated by a forest cover indicative of early successional phases on recently flooded surfaces or areas with high water tables, and late successional species on higher and more stable surfaces. Black cottonwood (*Populus trichocarpa* Torr. & Gray), red alder (*Alnus Rubra* Bong.) and white birch (*Betula papyrifera Marsh.*) are most evident on recently colonized portions of the floodplain while Western red cedar, Sitka spruce (*Picea sitchensis* (Bong.) Carr.), Douglas fir and western hemlock dominate older stable areas.

At middle elevations throughout most of the basin (150 to 1000 m) there exists a climax ecosystem comprised of hemlock, amabilis fir (*Abies amabilis* (Dougl.) Forbes) and occasionally subalpine fir (*Abies lasiocarpa* (Hook.) Nutt.). Above 1000 m there is a continuous cover of mountain hemlock, amabilis fir and yellow cedar (*Chamaecyparis nootkatensis* (D. Don) Spach) although the latter, which occupies a mesic habitat, is present in the east only. A zone of Engelmann spruce (*Picea engelmannii* Parry) and subalpine fir occurs between the alpine tundra and mountain hemlock zone. Lodgepole pine (*Pinus contorta* Dougl.) is a co-dominant species where fires occur frequently, such as the drier Interior Plateau, prohibiting the development of a climax spruce-fir forest .

The elevation of the contemporary tree line averages 1600 to 1650 m above sea level, but fluctuates spatially with changes in aspect, slope stability and microclimate. Within the tundra are a variety of alpine meadow species such as mosses, lichen and heather.

(2.9) Settlement and Logging History

Historical observations and factors influencing settlement patterns provide a potentially valuable source of both direct and indirect evidence of environmental change. Prior to European contact in the late 18th century the Bella Coola valley was populated by Indians of the Salish nation, a splinter group from tribes of the interior and southern coast regions (Kopas, 1970). Estimates by McIlwraith (1948) and more recently by Lepofsky (1985) show that at least 2,700 people, concentrated in 27 known village sites, occupied the valley between Stuie and Bella Coola. Since fishing was the primary economic resource, village locations were mostly determined by accessibility to river habitats favorable for salmon. Flooding and channel shifting are known to have influenced settlement locations and the type of structures constructed.

First white contact was made in 1793 by a ship from the fleet of Captain Vancouver making explorations of Burke Channel and North and South Bentinck Arms. One month after the first direct contact, Alexander MacKenzie, travelling by land from Fort Chipewyan on Lake Athabasca, descended into the valley along the east flank of Burnt Bridge Creek, the first caucasian to reach the North American west coast (north of Mexico) by land. No permanent white settlements were established until 1867, four years after Lieutenant Henry S. Palmer made the first survey of the route between Bella Coola and Fort Alexandria on the Fraser River. Several of the village sites were still seasonally occupied when the largest influx of Europeans, mostly Norwegians, settled in the valley in the autumn of 1896.

The preliminary land survey by Palmer in 1863, the legal survey of 1889-1893 and geological mapping by G.M. Dawson (1878) are the first systematic records of the Bella Coola environment. Exploration and survey notes, journals, letters and diaries from these sources along with limited archaeological evidence from the pre-contact period provide some data regarding environmental conditions over the last several centuries.

Interpretation of aerial photographs demonstrates that prior to the early 1950s timber removal was restricted to small plots on the valley bottom and was mostly related to land clearance. Between 1946 and 1954 only about 3 km^2 had been commercially logged, all of it below the Nusatsum River confluence. Up until 1968 an additional 13 km^2 had been logged mostly along the steeper slopes above Bella Coola River and in the Salloomt, Nusatsum, Noosgulch and Cacoohtin tributaries. From 1968 to 1974 another 8 km^2 was logged, much of it on the floodplain above Burnt Bridge Creek and partly on valley slopes of the lower Talchako River. Declining timber prices in the last 5 years have resulted in renewed activity in the more accessible tributaries in the Bella Coola basin. Thus over 18 km^2 has been logged since 1974, much of it concentrated in the Nusatsum and Noomst Creeks and upper Bella Coola River. The total of 43 km^2 represents less than 1% of the basin area but the concentration of commercial timber removal on lower slopes and at tributary junctions in the southwestern catchment would suggest a greater potential for impact on sediment yield.

CHAPTER III
Hydrophysical Records of Environmental Change: Tests Within the Instrument Period

(3.0) Introduction

Variations in regional climate are ultimately related to fluctuations in atmospheric circulation. These can occur as persistent departures in circulation intensity, shifts in the mean position of atmospheric features, or large-scale changes in circulation pattern. When making inferences about the nature of atmospheric processes and environmental change using predominantly biogeophysical evidence, the functional relationships between various indicators of climate change and climate itself need to be examined for a period in which both the climatic perturbations and environmental responses are known. It is the purpose of this chapter to assess the strength and form of these cause and effect relationships.

(3.1) Post-1945 Synoptic Climatology of the N.E. Pacific Sector

Precipitation and temperature are commonly available measures of climate, and provide a means for assessing the effects of changing atmospheric circulation. Establishing the nature of secular variations in temperature and precipitation for a given region can be difficult for three reasons: (1) insufficient sampling density for an area which exhibits large spatial variations - this may be due to real differences in local climate, or alternatively, to apparent differences produced by various sampling devices; (2) lack of temporal homogeneity at a station due to relocation or changes in data collection procedures and equipment; and (3) inadequate methods of analyzing and presenting data. Keeping in mind these potential sources of error, data from several climate stations in and around Bella Coola were examined in order to identify the regional response.

Eight climate stations, extending as far west as the outer coastal plain and as far east as the central Chilcotin River basin, were used to compile seasonal averages of monthly mean precipitation and temperature (see figure 2.1 for station locations and table 3.1). Winter is defined as the interval October to April and summer as May to August. September was not included in either seasonal average because of the significant negative correlation between summer temperature and September temperature (*i.e.* tends to dampen the signal) and relatively low total precipitation during this month. All seasonal data series were entered into a factor analysis which yielded two distinct groups based primarily on variations in precipitation (table 3.1). They are referred to here as coastal and interior regions and are split between the lower elevation sites to the west and higher elevation stations in the lee of the Coast Mountains. Secular trends in temperatures were similar enough to allow consideration of all stations as one group, but, for consistency, temperature trends are also examined in the two regions.

To maximize the regional signal seasonal variables for a given region are standardized, averaged and then plotted as adjusted partial sums using the following scheme:

$$P_{tn},\ T_{tn} = \sum_{j=1}^{n} \left[\sum_{t=1}^{t} \left[\frac{X_{tj} - \bar{X}_j}{|\bar{X}_j|} \right] \right] / n$$

P_{tn}, T_{tn} = cumulative sum of standardized seasonal precipitation (P_{nt}) or temperature (T_{nt}) for n stations and t years
X_{tj} = observed seasonal value for year t and station j
$\bar{X}_j$ = 1951-1980 seasonal mean for station j

Adjusted partial sums have advantages over simple moving average techniques. These include giving equal weight to generally wetter and

Table 3.1: Climate Stations in and near the Bella Coola River Basin.[1]

Station	Latitude	Longitude	Elevation (m)	Years of Record	Variables[2]	number of station moves (year)	Rotated Components[3] 1	2
COASTAL								
Bella Coola	52°20'	126°38'	3	1895-Present	T,P	1 (1959)	0.712	0.431
Bella Coola B.C. Hydro	52°22'	126°49'	14	1961-Present	T,P	0	0.699	0.395
Ocean Falls	52°21'	127°41'	5	1924-Present	T,P	0	0.812	0.298
Bella Bella	52°10'	128°09'	12	1933-1977	P only	1 (1966)	0.839	0.169
INTERIOR								
Kleena Kleene	51°59'	124°56'	899	1942-1969	T,P	0	0.313	0.746
Tatlayoko Lake	51°40'	124°24'	847	1928-Present	T,P	1 (1979)	0.453	0.865
Tatla Lake	51°54'	124°36'	945	1973-Present	T,P	0	0.239	0.731
Big Creek	51°10'	128°48'	1134	1904-Present	T,P	1 (1978)	0.121	0.649
						percent total variance	54.9%	25.1%
OTHER STATIONS								
Stuie	52°20'	126°15'	140	1933-1947	T,P	0		
Anahim Lake	52°21'	125°20'	1097	1954-1967	T,P	0		
	52°28'	125°18'	1054	1975-1980	T,P	0		

1. See figure 2.1 for locations.
2. T is temperature. P is precipitation.
3. Rotated component loadings for precipitation stations. Stations are grouped according to the strength of their correlation with each factor. See text for discussion.

generally drier stations within a region as well as displaying both high frequency (year to year) and low frequency (decennial) trends common to all stations. Secular variations in seasonal precipitation and temperature between 1945 and 1984 are shown in figure 3.1. Positive slope segments indicate persistently above average conditions whereas negative slopes are indicative of below average conditions. The interval 1945 to 1984 was selected to facilitate comparisons with other hydroclimatic data which were collected for this period only. The 30 year mean (1951-1980) is considered to be a representative of the period average and is also the adopted standard by various atmospheric agencies.

Secular Trends in Seasonal Temperature and Precipitation

Winter temperatures show the greatest year-to-year variance over the common period, and long-term trends show a high spatial coherence between the coastal and interior regions (figure 3.1a). Between 1948 and 1957, consistently below normal temperatures prevailed in both regions, whereas from 1958-1965 this trend reversed itself with above normal winter temperatures dominating. Fluctuations which followed were of lower magnitude, maintaining near average conditions until 1975 when winter temperatures were once again mostly above normal. Summer temperature departures (figure 3.1c) are less pronounced and more persistent within the Interior region.

Total October to April precipitation trends after 1960 show a strong inter-regional coherence with above normal precipitation from 1961 to 1968, average conditions between 1969 and 1976, and below normal after 1977 (figure 3.1b). Prior to 1960 the trends between regions are inversely related, with the exception of a few years characterized by average precipitation everywhere. Summer precipitation (figure 3.1d) has lower

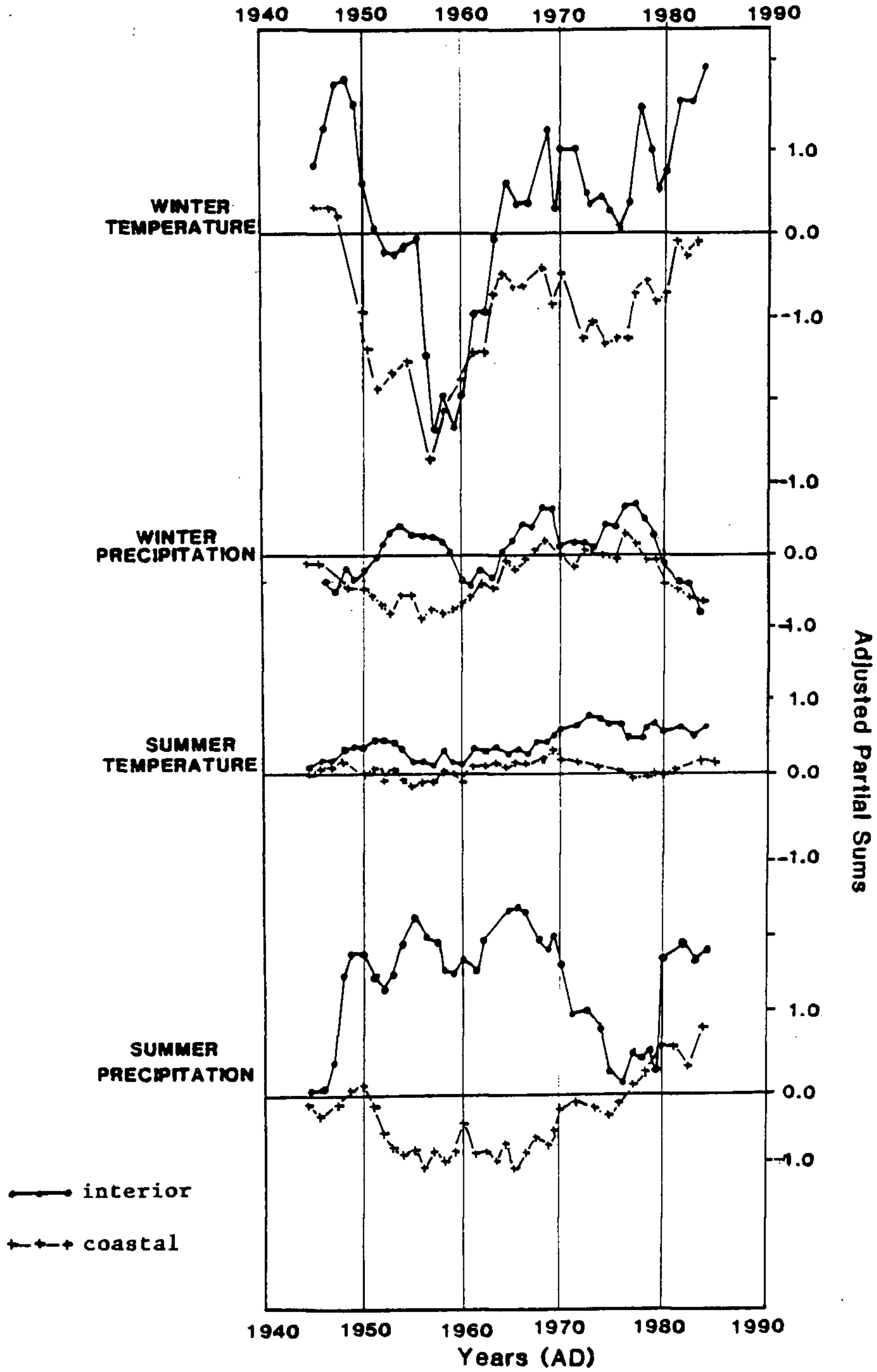


Figure 3.1 Adjusted partial sums of seasonal temperature and precipitation for coastal and interior regions of west-central British Columbia. Departures are from the 1951-1980 station normals and are summed from 1945 to 1983. Positive slope segments indicate persistent above average departures and negative slopes below average departures.

year-to-year variance and the greatest inter-regional differences, which is most likely related to differences in precipitation generating mechanisms during the summer. Convective thunderstorms occur frequently in the Interior region whereas summer rainfall along the coast is mostly derived from cyclonic disturbances. Since 1965, summer precipitation along the coast generally has been above normal and normal to below normal for the Interior. Near average conditions are evident in both regions prior to 1965 except for increases in the late 1950s.

Overall, trends in winter climate are towards well below normal coastal precipitation and regional temperatures to 1957, above normal precipitation and temperatures to 1964/65, average conditions everywhere until 1976 and then below normal precipitation and above average temperatures to 1983. Secular variations in summer climate are more variable because of the increasing importance of localized atmospheric instability.

Comparison With Other Coastal Regions of British Columbia

Temperature and precipitation trends for most of the province have been examined by Crowe (1963), Powell (1965), Thomas (1975) and Thomson *et al.* (1984). In each analysis, variably-lengthened, unweighted moving averages were used, a method which is poorly suited for detailed analysis of secular variations in climate when more precise determinations of climate shifts are required (Mitchell, 1966). More recently, investigations of winter temperature and precipitation fluctuations have been undertaken on the Queen Charlotte Islands (Karanka, 1986) and various stations along the British Columbia and United States west coast (McGuirk, 1982; Yarnel, 1985). A generalized summary of winter precipitation trends for the north, central and southern coasts is presented in figure 3.2.

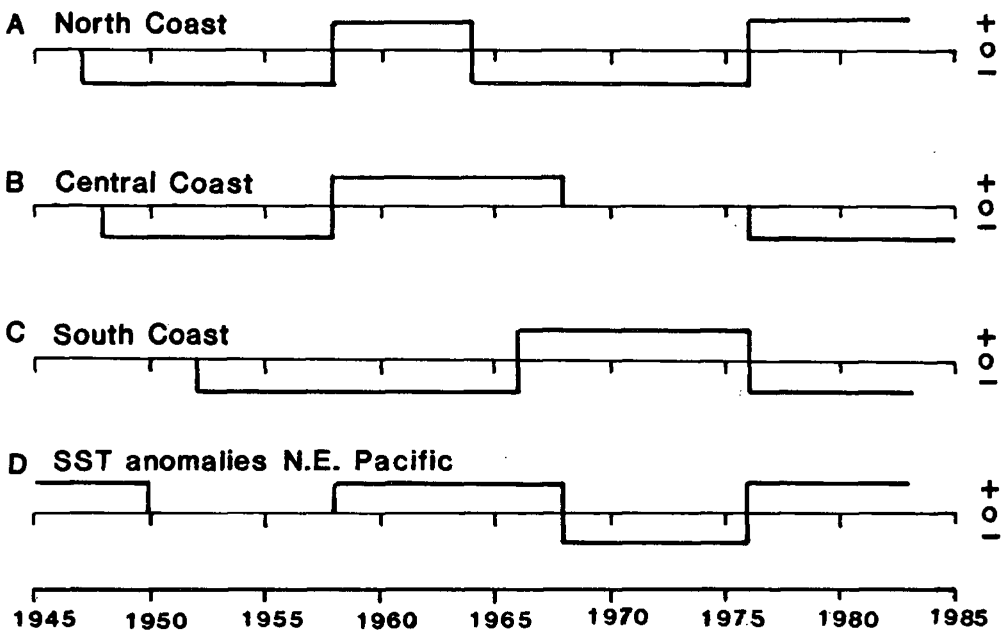


Figure 3.2 Generalized departures of winter precipitation from post-1945 normals. Bottom figure is generalized trends in sea surface temperatures. Areas above and below the zero line define periods of above and below normal precipitation, respectively. The magnitudes of departure are not indicated. A) north coast trends from Karanka (1986) (normal period 1900-1983) and Yarnel (1985) (normal period 1948-1980); B) central coast (this study); C) south coast from Powell (1965) (normal period 1900-1960) and Yarnel (1985); D) sea surface temperature anomalies for the northeast Pacific (Chelton, 1984) (normal period 1947-1984).

The most obvious feature of climate change along the northeast Pacific sector is the synchronous shift in 1976 to above average precipitation along the north coast and southeast Alaska, and below average precipitation in the central and south coast regions. As indicated in figure 3.2, these changes were associated with a shift to above normal sea surface temperatures (SST) in the northeast Pacific. Prior to 1976 precipitation trends along the central coast appear to represent a transitional phase between the almost inversely related trends of the northern and southern coasts. For example, above average precipitation along the north coast between 1958 and 1964 persisted until 1969 along the central coast and until 1976 along the south coast, suggesting a southward shift in moisture bearing disturbances through this period (*c.f.* Namias, 1983; Karanka, 1986). Between 1945 and 1957 when winter precipitation was average to below average for much of coastal British Columbia, SSTs were average to above average.

Unlike precipitation, winter temperature trends show a much greater spatial and temporal coherence similar to those in figure 3.1a. Above average (1958-1964, 1976-1983) and below average temperatures (1948-1957, 1965-1975) for all coastal areas correspond closely to positive and negative SST anomalies, respectively. Secular variations in summer temperature and precipitation have not been examined closely except for the earlier investigations of Crowe (1963) and Powell (1965). There is some indication of regionally consistent trends with above average temperatures prior to 1952 and average to below average temperatures between 1953 and 1960.

Atmospheric Circulation

Usually, no one feature of the atmospheric circulation can explain adequately variations in surface climates. Sea level pressure patterns, positioning of hemispheric semi-permanent pressure systems, intensity and direction of flow aloft, mean position of seasonal fronts, frequency of storms and orientation of storm tracks are all important components. However, there is increasing evidence of interactive linkages between atmospheric circulation and SST anomalies. It appears from figure 3.2 that much of the recent climatic variability along the B.C. coast is associated with these anomalies. Horel and Wallace (1981) demonstrated that above normal SSTs in the eastern north Pacific result in the development of a high pressure ridge over western North America and a trough in the central Pacific. Ridging along the coast is often associated with a westward shift in the Aleutian Low (Angell and Korshover, 1982).

If ridging extends out into the Pacific then there are two possible outcomes: cool/dry conditions as northwesterly airflow aloft is advected southward down the trailing limb of the ridge (figure 3.3a) or warm/dry conditions if the ridge axis is centreed over the coast (figure 3.3b) (Yarnel, 1985). Wet conditions will prevail if ridging is poorly developed, *i.e.* zonal flow occurs (figure 3.3c), or if the ridge is displaced eastward by an intensified Aleutian Low, in which case the coast is under the broad, upsweeping, southeastern quadrant of the low (figure 3.3d). Analyses of post-1945 synoptic-scale circulation features by Balling and Lawson (1982) and McGuirk (1982) indicate that flow patterns have been dominantly meridional, characterized by a low Rossby wave number leading to persistent and sometimes extreme climatic departures and abrupt shifts in the character of the climate.

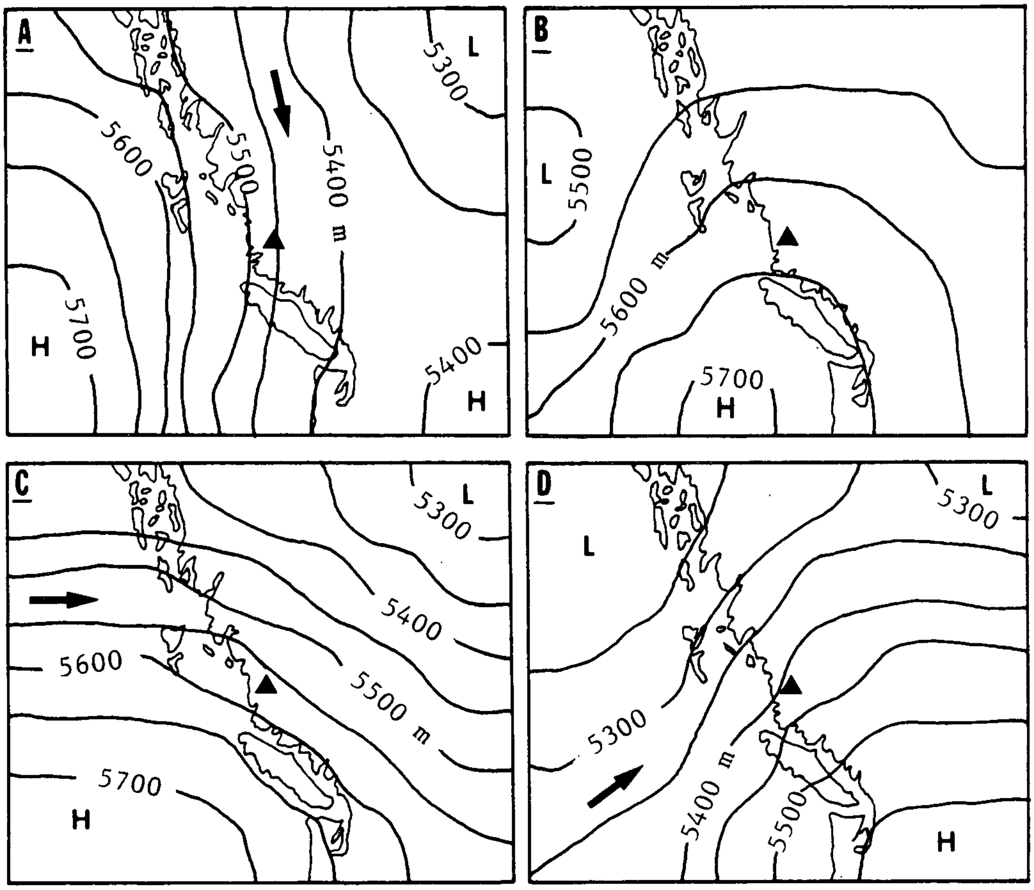


Figure 3.3 Synoptic-scale pressure patterns (500 mb) leading to distinct departures of temperature and precipitation in southwestern British Columbia (from Yarnel, 1985). (A) High pressure ridging extends beyond the coast to produce cool/dry conditions. (B) Ridging is centred over the coast resulting in warm/dry weather. (C) Cooler and wetter conditions occur when ridging is not dominant producing zones of intense 500 mb flow. (D) Ridging occurs to the east of the coast resulting in warm/wet conditions.

Based on the trends in figures 3.1 and 3.2 and the results reviewed above, positive SST anomalies between 1958 and 1968, associated with ridging over the west coast, promoted above average precipitation along the central and north coasts. Although statistical confirmations cannot be made with the available storm track data, this would correspond to a northward displacement of cyclonic trajectories. Below average precipitation after 1976 for the central and southern coasts occurred when the Aleutian Low shifted westward significantly further enhancing the northward displacement of storm tracks.

Frequency and Seasonal Persistence of Synoptic Types

Periods of persistent departures in seasonal temperature or precipitation are thought to be characterized by a set of unique synoptic conditions which produce the observed trends (*i.e.* wet, dry, cool, warm). Each season is comprised of a large number of synoptic types. However, it is likely the frequency or within-season persistence of only certain synoptic types drives the mean departure. This hypothesis is tested here using an objectively defined set of synoptic types which are known to yield certain climatic characteristics.

Synoptic circulation types for a particular season and sequence of years were identified using a catalogue of daily 500 mb and daily surface pressure patterns developed by Yarnel (1983) and Barry *et al.* (1982) for the northeastern Pacific. Both catalogues were constructed using the objective classification scheme of Kirchhofer (1973) in which estimates of daily atmospheric pressure at predefined hemispheric grid points are subjected to a sequence of selection and grouping procedures. Full details of the catalogues and grouping schemes are given in Kirchhofer (1973) and

Yarnel (1984). Only potential problems with the technique are considered here.

Because of the length of the study period (>10,000 days in both studies) a portion of the pressure/elevation data set is used to define 'key days' representative of specific synoptic groups. The major assumption here is that the smaller data set is representative of the range of synoptic conditions over the longer study period. Problems with the Kirchhofer method include: (1) selection of a cutoff threshold for grouping types - a low threshold produces a larger set of synoptic types and leaves few days unclassified, whereas a high threshold produces fewer types but leaves many unclassified days (Key and Crane, 1986); (2) grid size selection - size will influence the spatial resolution of synoptic patterns with large grids yielding poor definition of mesoscale circulation phenomena; (3) within-group variability - even small changes in the inflection point of a trough or ridge can alter the pattern of temperature or precipitation over an area with no apparent differences in the zonality of circulation; and (4) several key synoptic types may not be classified by the daily analysis during a period when circulation patterns are changing rapidly and when there is a high frequency of missing data. Although a less stringent grouping of synoptic types results from the use of 'key days' and specification of a subjectively defined grouping threshold (Key and Crane, 1986), the same methods were employed in both studies, which enhances the compatibility of interpretations derived from the catalogues.

The Yarnel (1983) catalogue, for the period 1946-1978, is based on eighteen, 500 mb pressure-elevation patterns for a 2.5° X 2.5°, 30-point grid covering the coastal regions of British Columbia and the far eastern Pacific. The Barry *et al.* (1982) catalogue is based on 31 surface pressure patterns for a 35-point, 5° X 5° diamond-shaped grid over western North

America and is for the period 1899 to 1980. Only 6 grid-points in the latter catalogue fall within the coastal B.C. and Gulf of Alaska region. The larger number of synoptic types in the surface pressure catalogue is due to the greater variability in surface pressure patterns and the larger spatial coverage of the grid. Characteristics of synoptic types in each catalogue are summarized in table 3.2.

Difference of means tests were carried out to determine if there are significant changes in the frequency of surface and 500 mb synoptic types associated with winter precipitation for: (1) periods of above average and below average precipitation indices along the central coast and (2) groups of years comprised of the most extreme departures in winter precipitation irrespective of ordering in time.[1] The results presented in table 3.3a show that the frequency of "wet" surface and 500 mb synoptic types are not significantly different between the dry period (1948/49 to 1957/58) and wet period (1959/60 to 1968/69). In contrast, when groups of years characterized by the most extreme departures are tested (table 3.3b), there is a significant difference between wet and dry intervals with an average of ten more days per winter season dominated by "wet" 500 mb types and fifteen more days per season dominated by moisture bearing, surface lows.

The same sequence of tests was conducted for cool and warm summer periods, but due to the comparatively weak circulation intensities during the summer, there is greater difficulty in identifying distinct surface pressure patterns associated with extremes in summer temperature. Thus, only the 500 mb patterns are considered (table 3.4). Similar conclusions can be made for summer temperature regimes as for winter precipitation in that there is no significant difference in the dominance of a particular

1. Extreme years were defined as those years with climatic departures in excess of one standard deviation of the mean. Sample sizes were determined by the group with the fewest years fitting this criterion.

Table 3.2. Synoptic Types from Yarnel (1983), Barry et al. (1982).

Group or Type[1] (Yarnel - 500 mb) (Barry - surface)	Synoptic Circulation	Climate Effect
Wet Y=(1,3,4,8,10) B=(3,4,7,11,12, 15,23,31)	- southwesterly cyclonic component and advection of moist air due to positive 500mb pressure anomalies over the western USA and central Canada and an intensified Aleutian Low - surface lows are mostly from the west and northwest but also migrate from the south	- above average precipitation and temperature for the entire B.C. coast (upper air types 1,4) and the central and southern coast (upper air types 3,8,10)
Dry Winter Y=(5,9,14,15,17) B=(8,14,18,26,30)	- north or northwest flow aloft and strong ridging centered on the coast - surface lows diverted into Gulf of Alaska or southern US west coast	- below average temperatures and precipitation for the entire coast (upper air types 5,9) and the southern coast (upper air types 14,15,17)
Dry Summer Y=(2,6,11,13,18) B=(5,6,9,13,21)	- strong zonal flow or ridging centered west of Vancouver Island	- dry and warm for most of the central and southern coasts (upper air types 2,6) occasional cold lows (11,13,18)
Warm Summer Y=(1,4,6,7,15,18) B=(3,4,6,11,23,25)	- ridging centered over the coast or inland with advection of warm air aloft from the southwest	- well above average temperatures for the entire coast (upper air types 1,4) and above average for the central and southern coasts (6,7,15,18)

1. Type numbers refer to the 18 objectively classified daily synoptic-scale patterns (500 mb) of Yarnel (1983) (Y=) and 31 surface pressure types of Barry et al. (1982) (B=).

Table 3.3. Winter Precipitation and Synoptic Type Frequencies.

A. Significance test between a dry period and a wet period along the central B.C. coast

	Mean winter precipitation (mean departure)[1]	Wet synoptic type frequency[2]	Wet synoptic type frequency[3]
dry: 1948/49 to 1957/58	-0.04 ± 0.03	82.2 ± 8.3	42.5 ± 10.5
wet: 1959/60 to 1968/69	0.07 ± 0.05	86.6 ± 8.8	44.4 ± 7.8
One-tailed t-test (α = 0.10, df = 18)[4]	t = 2.73 significant	t = 1.11 not significant	t = 0.46 not significant

B. Significance test between wet years and dry years

dry years: 1978,71,70, 56,49	-0.19 ± 0.08	78.8 ± 7.0	33.4 ± 7.23
wet years: 1976,74,68, 64,54	0.27 ± 0.08	89.4 ± 5.8	48.0 ± 8.9
One-tailed t-test (α = 0.10, df = 8)[4]	t = 8.48 significant	t = 3.38 significant	t = 2.80 significant

1. Calculated as the average departure of winter precipitation (Oct-Apr) for the central coast precipitation index from the 1951-1980 normal.
2. Mean number of days dominated by "wet", 500 mb, synoptic types from October to April after Yarnel (1983).
3. Mean number of days dominated by "wet" surface synoptic types from October to April from Barry et al. (1982).
4. A low significance level was selected because of the exploratory nature of the analysis.

Table 3.4. Summer Temperature and Synoptic Type Frequency.

A. Significance test between a cool period and a warm period in Bella Coola basin

	Mean Summer temperature (mean departure)	Warm Synoptic type frequency[1]
cool: 1948/49 to 1957/58	-1.7 ± 0.8	20.5 ± 7.0
warm: 1959/60 to 1971/72	1.9 ± 0.4	22.4 ± 8.0
One-tailed t-test (α = 0.10, df = 21)	t = 4.62 significant	t = 0.56 not significant

B. Significance test between cool years and warm years

cool years: 1975,57,56, 54,49	-2.3 ± 1.0	14.5 ± 6.1
warm years: 1972,70,69, 67,61	1.9 ± 0.7	26.5 ± 3.9
One-tailed t-test (α = 0.10, df = 8)	t = 5.31 significant	t = 3.00 significant

1. Warm synoptic types from Yarnel (1983).

group of synoptic types between a warm interval (1959/60 to 1971/72) and cool period (1948/49 to 1957/58) (table 3.4a), but for groups of years characterized by extreme departures a significant difference is evident (table 3.4b).

Unlike winter precipitation, extreme departures in summer temperature tend to fall within periods of the same tendency. This would suggest that an interval of persistent departures in summer temperature has a higher probability of extreme occurrences and it is these extreme, within-period, years which are characterized by significantly different frequencies of synoptic types. Conversely, anomalous winter precipitation does not appear to be restricted to these longer more persistent intervals of above or below average departures. One or two exceptional years may occur during an interval of otherwise average conditions (*e.g.* 1974, 1976).

In addition to significant changes in synoptic type frequency positive anomalies in winter precipitation can be distinguished by the within-season persistence of synoptic types. For example, during the very wet winter season of 1963/64, 36 days between mid-October and mid-November (with the exception of one day) were classified as "wet" 500 mb synoptic types - clearly the most exceptional signal in the entire 33 year record. During that same winter season and for others between 1960 and 1969 intervals of between 10 and 12 days were not uncommon whereas during wetter winters within dry periods these within-season runs were mostly less than 5 to 8 days.

The conclusion drawn here is that the synoptic signature between drier and wetter intervals is related to the within-season persistence of "wet" synoptic types whereas anomalously wet years are characterized by both a higher frequency and greater persistence of these same types. Extreme departures in precipitation do not appear to be restricted to a period of given tendency. This distinction is an important one particularly if there are well-defined response thresholds in the biogeophysical environment below which the impact of changing weather is not significant.

(3.2) Responses of Hydrologic Variables to Climate Fluctuations

Large scale motions of the atmosphere are manifested not only in regional precipitation and temperature indices, but also in hydrological phenomena such as winter snowpack, river runoff and glacier fluctuations. Linkages between these components of the terrestrial phase of the hydrological cycle and atmospheric circulation can be spatially non-homogeneous and lagged in time (Barry, 1981). This occurs because the atmosphere, a system with low heat and moisture capacities, adjusts rapidly to changing energy levels (Bradley, 1985). The high water storage capacity of the terrestrial environment (*e.g.* snowcover, glaciers, lakes, groundwater), spatial variability in actual storage and the existence of response thresholds all tend to dampen the response to climatic change. For this reason it is necessary to understand how feedback mechanisms might be important in the system. The spatial and temporal trends in water storage and runoff along the southwestern British Columbia coast are examined here.

Regional Winter Snowpack

Snowpack data have been collected for a number of sites in southern British Columbia but few have records longer than 40 years. The majority of solid precipitation falls between early November and late March. Thus April 1 snow water equivalents were compiled from a number of stations (see figure 2.1 for locations) and grouped into homogeneous regions using factor analysis (table 3.5). Two statistically distinct groups exhibiting similar long term trends but minor interannual differences were identified: a

Table 3.5. Winter Snow Courses in southwestern British Columbia.[1]

Station	Latitude	Longitude	Elevation (m)	Years of Record	Rotated Components[2] 1	2
COASTAL						
Bella Coola	52°31'	126°38'	1380	1953-1954	--	--
Newcastle Rdg.	50°24'	126°03'	1170	1961-Present	0.746	0.290
Powell (upper)	50°16'	124°18'	1040	1938-Present	0.753	0.541
Dog Mountain	49°23'	122°58'	1080	1945-Present	0.823	0.525
INTERIOR						
Precipice	52°26'	125°38'	1220	1972-1973 1970-1971	--	--
Tatlayoko Lake	51°36'	124°20'	1710	1952-Present	0.280	0.848
Tahtsa Lake	53°34'	127°38'	1300	1952-Present	0.319	0.872
Tenquille Lake	50°32'	122°56'	1680	1953-Present	0.332	0.895

1. See figure 2.1 for locations.
2. Rotated component loadings for snow course stations. Stations are grouped according to factor scores.

coastal zone to the west of the Coast Mountain crest and a zone in the lee of the Coast Mountains towards the drier interior. Secular variations in snowpack are plotted as adjusted partial sums in figure 3.4. Again, positive slope segments represent above average departures and negative slopes below average departures.

A greater interannual variability is evident for the regional snowpack index representative of the interior but both series exhibit three periods of persistent departures: normal to below normal snowpack between 1957 and 1963, well above normal snowpack between 1964 and 1976, and finally strongly negative departures after 1976. The occurrence of below normal spring snowpacks after 1976 corresponds well with below normal winter precipitation along the central and southern coast of B.C. coupled with normal to above normal winter temperatures (figure 3.2). Between the mid-1960s and 1975 normal to above normal winter precipitation along this same coastal zone is reflected as persistent positive departures of winter snowpack for both coastal and interior regions. Winter temperatures were on average below normal during this same interval. Below normal winter precipitation and above normal winter temperatures dominated the southern coast between 1957 and 1964 resulting in below normal snowpacks.

The association between regional snowpack volume and frequency of "wet" winter circulation patterns for periods and groups of years characterized by anomalous departures is tested using difference of means. Unlike the strong association between positive snowpack departures and above normal precipitation, four of the six extreme low snowpack years were not the same years with low winter precipitation. For three of these four years winter temperatures were above average which probably led to greater melting near the beginning and end of each season. Hence, average precipitation and above normal temperatures resulted in low snowpack

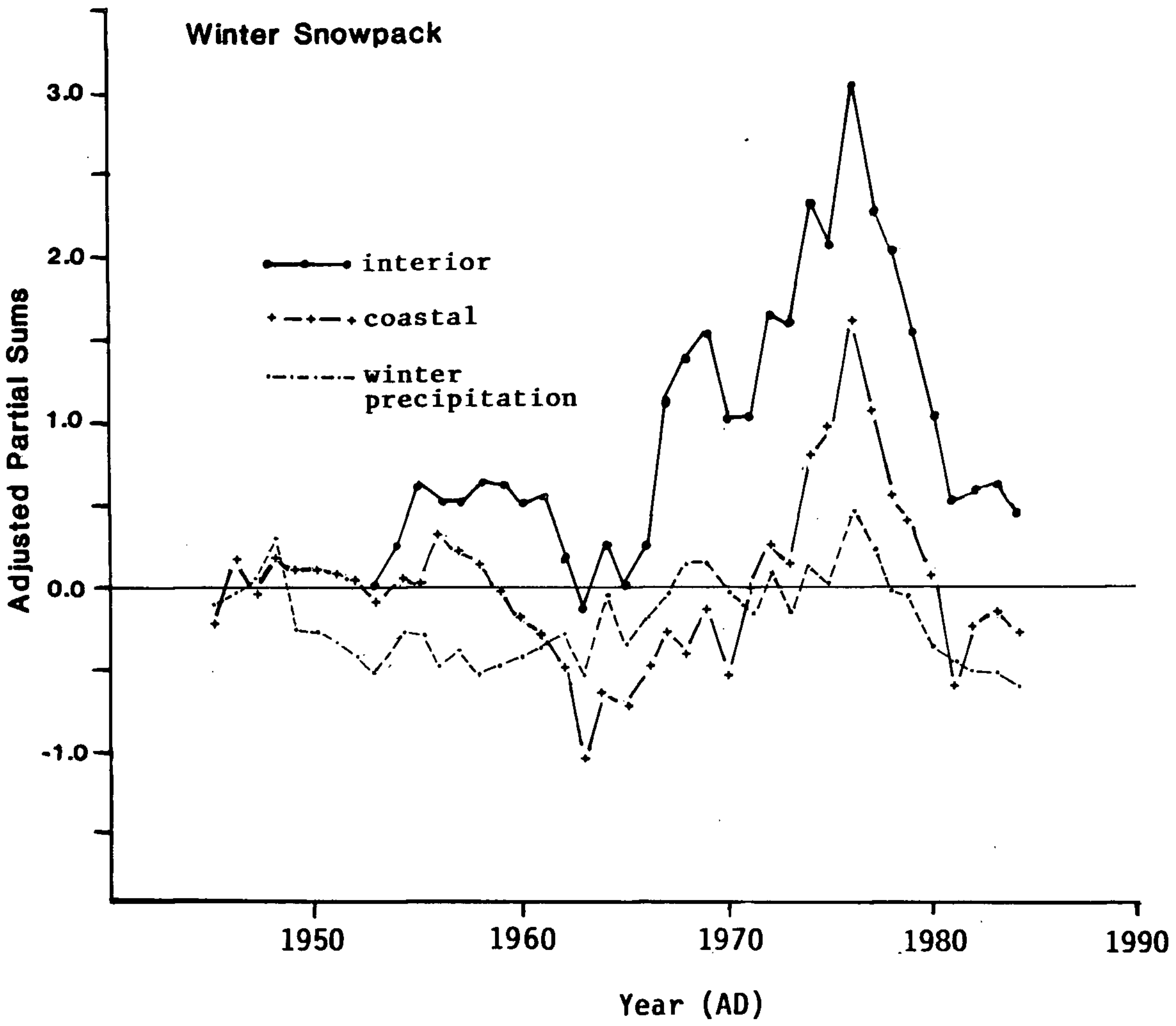


Figure 3.4 Adjusted partial sums for April 1st snowpack (water equivalent). Normal period is 1951-1980. Coastal winter precipitation taken from figure 3.1 is included for comparison. Positive slope segments represent persistent above average snowpack years whereas negative slopes represent below average snowpack.

volumes. Even with these complications the results in table 3.6 show that variations in spring snowpack are positively linked to the mean number of days dominated by "wet" synoptic types for groups of anomalous years and also for the two periods characterized by above average (1966-1976) and below average (1977-1980) departures. The latter result is partly related to the limited number of data during 1977-1980 and the above temperatures which occurred then, but does indicate that seasonally integrated hydrologic parameters such as snowpack are sensitive to changes in the frequency of specific circulation patterns not only for extreme years but also for sequences of years in which a tendency for persistent departures occur.

Spring Runoff

Spring runoff is likely to be greater following winters characterized by anomalously high precipitation and snowpack accumulation but spring and summer temperature conditions may enhance or depress actual water yield. Furthermore, spring runoff should be less spatially homogeneous than snowpack distributions because of the large spatial variability in runoff generating mechanisms and storage conditions within and between basins. However, as basin size increases, water yield becomes less responsive to isolated storms or to snowmelt periods of short duration. Thus, large scale synoptic trends may be reflected more directly in temporal runoff variability within larger basins (Bruce, 1974).

Most hydrometric records for streams draining the west slope of the southern Coast Mountains are shorter than 40 years and the number of medium sized basins (3,000 - 6,000 km^2) for which discharge data exist is small. Time series of spring snowmelt runoff for the Bella Coola, Homathko and Squamish Rivers, three moderate size catchments draining partly glacierized

Table 3.6. Winter Snowpack and Synoptic Type Frequency.

	Mean April 1st water equivalent (mean departure)[1]	Wet synoptic type frequency[2]	Wet synoptic type frequency[3]
A. Significance between a high snowpack period and a low snowpack period			
high: 1966-1976	0.69 ± 1.26		40.3 ± 8.9
low: 1977-1980	-1.01 ± 1.22	catalogues only to 1978	32.3 ± 12.1
One-tailed t-test (α = 0.10, df = 13)	t = 2.94 significant		t = 1.61 significant
B. Significance test between high snowpack years and low snowpack years			
high snowpack years: (1976,74,68,67,64,54)	1.32 ± 0.93	91.5 ± 4.13	46.5 ± 8.8
low snowpack years: (1978,77,75,70,62,60)	-1.28 ± 0.75	79.2 ± 6.3	29.8 ± 15.2
One-tailed t-test (α = 0.10, df = 10)	t = 5.22 significant	t = 4.01 significant	t = 2.32 significant

1. Calculated as the average departure for the three lee side snow courses: Tashta, Tatlayoko and Tenquille.
2. Frequencies of 500 mb patterns after Yarnel (1983).
3. Frequencies of surface pressure patterns after Barry et al. (1982).

mountains of southwestern British Columbia, provide some evidence of secular runoff trends during the post-1945 period (see figure 2.1 for locations). Results are plotted as adjusted partial sums in figure 3.5. The interval May to mid-July was defined as the spring runoff period because hydrographs from all three rivers indicate that discharge peaks related to snowmelt occurred during this time (see figure 2.6). While there are minor interannual differences between each catchment, similar long-term secular trends are evident. Fluctuating, but near average discharges occurred between 1947 and 1958 in the Bella Coola River. Following well above average spring runoff during 1958, below average discharges in Bella Coola and Squamish Rivers prevailed until 1966. Homathko River discharges were more variable through this period. After 1966, discharges for most of the coast were well above average until 1972 when this trend was sharply reversed for the remainder of the decade. Post-1980 flows are near average.

A comparison of runoff trends with snowpack water equivalents and winter precipitation (figures 3.1, 3.4 and 3.5) reveals some important differences. During the interval 1973-1976, when wet winter conditions prevailed, spring runoff in all three basins was below average. This divergence can be explained by the occurrence of well below average winter and summer temperatures during all four of these years resulting in lower runoff and increased water storage. Similarly, wet but cool conditions between 1963 and 1966 may explain the below average spring runoff, particularly in the Squamish River.

Spring runoff data from the north coast were not examined but Royer (1982) has modeled secular trends in annual freshwater discharges in southeast Alaska. The estimates are based on seasonal trends in precipitation, temperature, and water storage and are calibrated against various catchment runoff data. The annual trends appear to fit quite well

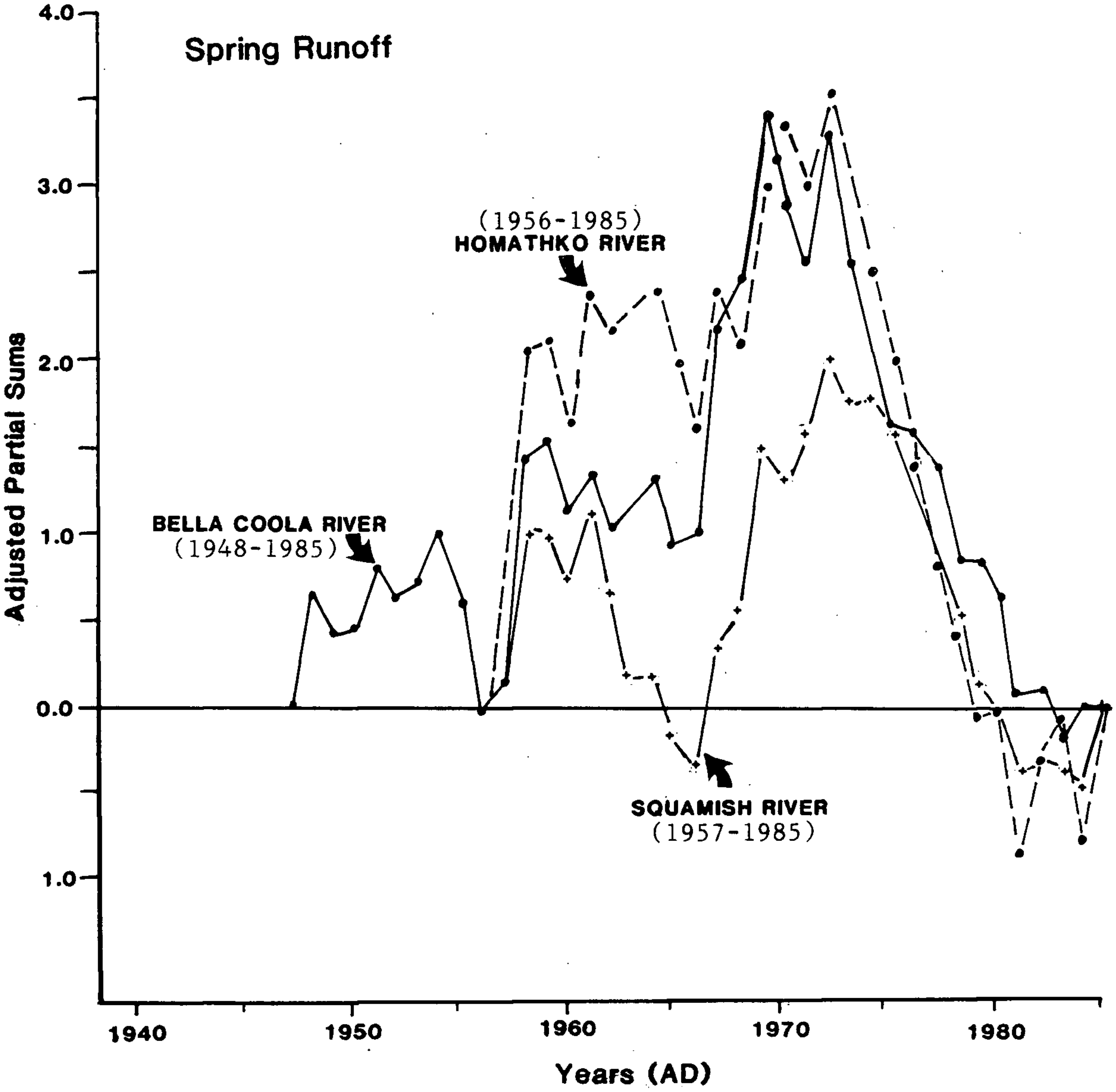


Figure 3.5 Adjusted partial sums of spring runoff (May - mid July) for three coastal rivers. Normal periods for each station correspond to the length of record. Positive slope segments represent persistent above normal runoff. Note the the regional response of below average runoff after 1972.

the variations in winter precipitation along the north coast (figure 3.2a), suggesting that precipitation is the dominant factor in determining the temporal variation of runoff. Above average runoff occurred during the interval 1958 to 1963 and after 1976, and below average runoff between 1964 and 1976. Below average runoff and winter precipitation prior to 1958 have been documented for Prince of Wales and Revillagigedo Islands in the Alaska Panhandle (Karanka, 1986) which contrast with the above average annual runoff identified by Royer (1982). These differences for the earlier period may be due to variations in the definition of the water year, the much more limited area of the latter study and the sensitivity of the runoff model used by Royer. In general, the spatial coherence of winter precipitation and spring runoff is strongly evident but spring and early summer temperatures also contribute to runoff variability.

Spring and Summer Runoff - Bella Coola River

Factors other than winter precipitation have been incorporated into a number of empirical and physically based models for estimating spring flood discharges and high summer water yields (Ostrem *et al.*,1967; Tangborn and Rasmussen, 1976; Power and Young, 1979; Collins and Young, 1981; Tangborn, 1984; Young 1985). The most important are the distribution and magnitude of temperature departures and the contributions of glacier melt to total runoff. Glacier melt generated runoff can be as high as 50% of the total water yield even when glacial cover is relatively limited (Young, 1985). Estimates of glacier melt contributions to summer discharge in the Homathko River (21% glacial covered) range as high as 21% (British Columbia Hydro, 1984).

While accumulation season precipitation and ablation season temperatures are important model components, these studies of ice-covered basins have shown the importance of more spatially variable parameters such as relative humidity, incident solar radiation and wind in estimating summer runoff from individual glaciers. Direct measurement, or even indirect estimates, of these additional parameters are generally not possible for paleohydrological modeling. Models based on primary hydrometeorological variables such as temperature and precipitation are likely to explain less of the year-to-year variance in runoff. An assessment of simple runoff models for the Bella Coola River is made here in order to document the level of explained variance using primary meteorological variables only.

Analysis of ablation season runoff in all four gauged tributaries of the Bella Coola River reveals some interesting differences. Interannual variability of early season runoff (May-mid-July) is similar throughout the catchment, presumably due to the contribution of water from a basin-wide melting snowpack. As the melt season progresses and icemelt contributions increase, two distinct runoff regimes become apparent: (1) a strongly diurnal runoff regime for the southern catchment (*e.g.* Talchako/Nusatsum Rivers) and (2) more uniform, but steadily declining flows, in the eastern and northern segments of the catchment (*e.g.* Atnarko/Salloomt Rivers). During high snowpack years, and normal to below-normal snowpack years in which spring temperatures (April - May) are depressed, high seasonal runoff is common to all tributaries. It is low snowpack years coupled with high summer temperatures which accentuate the differences in runoff regimes. Since this latter situation occurs relatively infrequently, a summer runoff index was computed for the Bella Coola River below Burnt Bridge Cr., which integrates discharges from both glacierized and unglacierized sources. Although this may reduce the climate-runoff relationship by mixing populations, the index incorporates a basin-wide response to changing hydroclimates and more properly estimates total summer discharge through the lower Bella Coola River.

Several empirically-based linear statistical models, incorporating seasonally and regionally averaged climate data, were evaluated using an interactive version of all-possible-subsets regression. Indices of seasonal and monthly temperature, precipitation and spring snowpack for both coastal and interior regions comprise a group of independent variables used to estimate spring runoff. Preliminary models were constructed using a 33 year data set (1952-1984), limited by the length of spring snowpack records. Early season (May to mid-July), late season (mid-July to August) and summer (May to August) runoff indices were considered separately. Results are presented in table 3.7. Scatter plots of bivariate relationships, probability plots of residuals and a test for serial correlation for each model indicate that assumptions of the regression technique have not been violated.

All three preliminary models relating Bella Coola River runoff to climate are significant but the strongest is summer runoff ($Ra^2 = 0.58$) in which winter precipitation along the coast, spring snowpack volumes measured in the lee of the Coast Mountains and summer temperature indices derived for the interior, enter the equation. This combination of variables suggests that summer runoff is controlled by snowpack conditions and melt energy in the headwaters of the southern and eastern catchment.

The three preliminary models were tested by using independent data from the Nusatsum and Salloomt Rivers. Runoff data from these two downstream tributaries were combined and compared with predicted runoff in Bella Coola River. The use of Nusatsum/Salloomt data was possible because of the high correlation between these data and observed discharge in the Bella Coola River ($r \geq 0.79$). Verification procedures for regression models

Table 3.7. Significant variables and regression coefficients for climate-runoff models.

Climate Indices	Runoff Period: Preliminary Models: May to mid-Jul	mid-July to Aug	Summer	Final Model: Summer
	Coefficients			
WP(I)				
WP(C)	0.62	0.49	0.56	0.68
SS(I)	0.23		0.18	
SS(C)				
AT(I)	-0.09			
AT(C)				
ST(I)	1.21	0.49	0.78	0.78
ST(C)				
Intercept	0.11	0.01	0.01	0.01
Summary Statistics:				
R^2	= 0.52	= 0.29	= 0.62	= 0.55
R_a^2	= 0.45	= 0.24	= 0.58	= 0.52
D	= 1.50	= 2.22	= 1.51	= 1.94
F	= 7.15*	= 5.8*	= 14.8*	= 17.16*
SE	= 0.45	= 0.12	= 0.09	= 0.10
Confirmatory Statistics:				
r	= 0.74	= 0.41	= 0.89	= 0.84
RE	= 0.19	= 0.05	= 0.41	= 0.36

WP = winter precipitation (Oct-Ap)
SS = April 1st snowpack (water equivalent)
AT = April temperature
ST = summer temperature (May-Aug)
(I) = Interior Region
(C) = Coastal Region

R^2 = coefficient of determination
R_a^2 = adjusted R^2
D = Durbin-Watson d (reject H_o - no significant autocorrelation)
F = F ratio (* significant; α = 0.025)
SE = standard error
RE = reduction of error statistic
r = correlation coefficient

The coefficient of determination (R^2) is adjusted (yielding Ra^2) to account for the number of predictor variables in each equation.

evaluate how well a model predicts a new value of the independent variable. Two statistics reported in table 3.7 (confirmatory statistics) are used to test the significance of the predicted runoff values: the correlation coefficient (r) and RE statistic. In all three models the correlation coefficients are significant suggesting a reasonable correspondence between actual and predicted runoff. The reduction of error statistic (RE) (Fritts, 1976) has theoretical limits of $-\infty$ and 1.0 where a positive value indicates that the predictions made by the model are better than a prediction based on the mean of the observed data. Again all three models have positive and significant RE values suggesting reasonable predictive ability of the models.

The model which incorporates spring snowpack, a variable with the lowest partial correlation coefficient, was removed from further consideration because snowpack data prior to 1952 are not available. The calibration period was extended to 1948 (earliest available runoff) and all outliers which might have had a significant influence on the regression coefficients were tested using Cook's D statistic (Draper and Smith, 1981). None were found. The final model is used to reconstruct spring runoff prior to 1948 and these results are discussed in chapter 6.

Analysis of spring runoff hydrographs shows that extreme runoff events in May or June are related to years with persistent positive departures in monthly winter precipitation and negative spring temperature anomalies. For this reason, the final model presented in table 3.7 may approximate the potential for high-magnitude spring runoff but, because there is no spring temperature index in the final model, is apt not to provide a good estimate of peak spring discharge. Hence, residual variance in the final model is highest during years characterized by monthly climatic anomalies which significantly influence runoff (e.g. depressed

April temperatures) but are not incorporated into any of the seasonal climate indices.

Autumn Floods

As noted in chapter 2 the autumn runoff period is hydrologically dissimilar to the summer discharge regime. Steadily declining freezing levels promote reduced contributions of glacier meltwater as the season progresses. Superimposed on these declining flows are short-duration, high-magnitude floods which have two distinct sources: (1) runoff due to heavy precipitation only and (2) runoff due to high precipitation combined with rapidly rising freezing levels and significant snowmelt at middle elevations. Typically, the former events occur early in the autumn (August to October) while the latter may occur as late as January following accumulation of snow throughout much of the basin.

Analysis of daily precipitation records prior to autumn flood events in the Bella Coola River suggests that for most high-magnitude flows, antecedent precipitation is important. Flood generating storms are often preceded by a sequence of one or two day rainfall events which saturate the snowpack or soil so that subsequent runoff is maximized. Figure 3.6 demonstrates a mean separation of 4 to 5 days between storms, beyond which the potential for heavy storm runoff is greatly reduced. This separation of storm peaks appears to be valid for peak 24 hour precipitation events with recurrence intervals of 4 years or less; that is, flood generating storms with one-day recurrence intervals greater than 4 years generally produce enough precipitation to yield significant runoff and are not necessarily preceded by hydrologically (climatologically) distinct events (*e.g.* autumn floods of 1950, 1973).

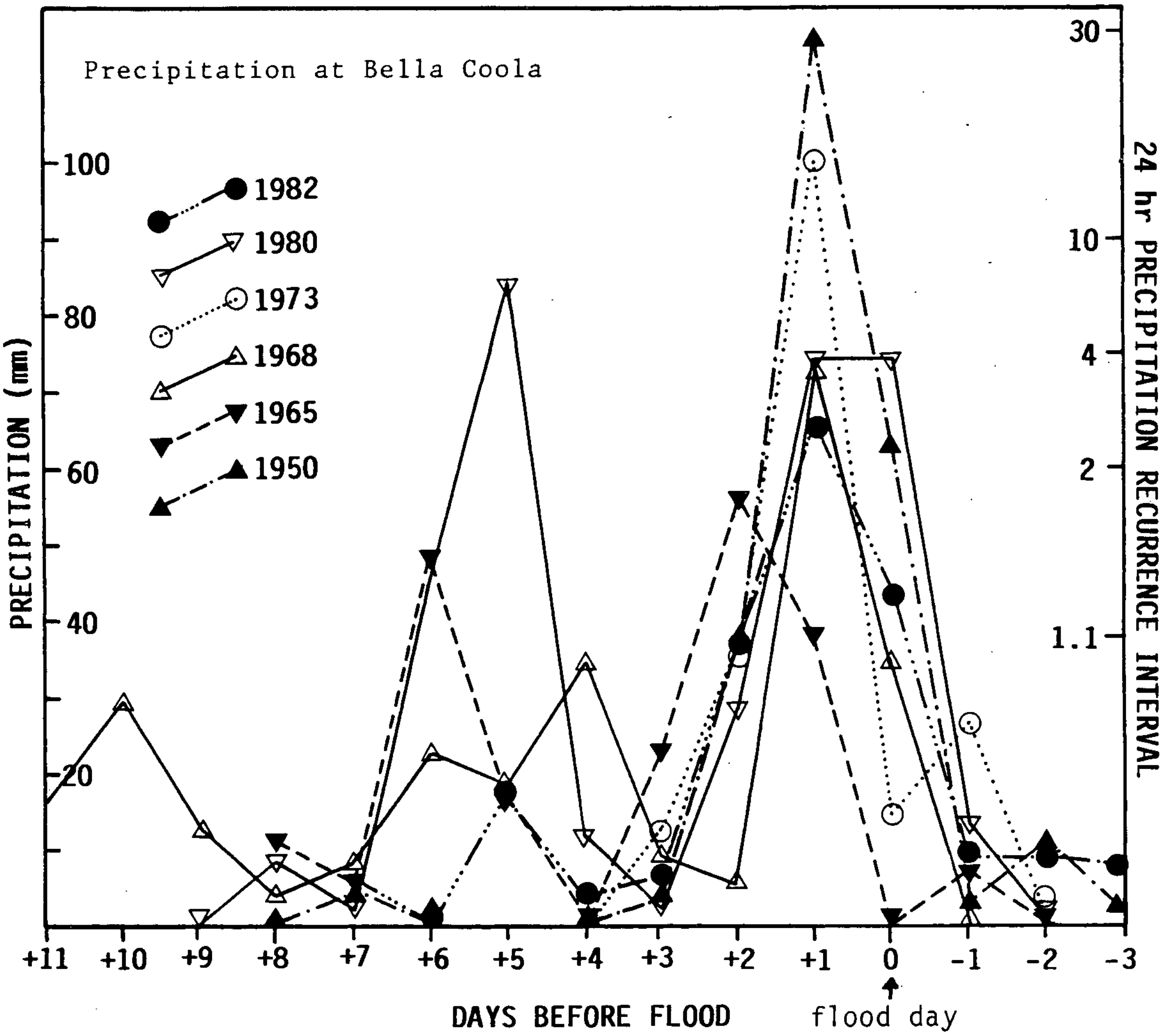


Figure 3.6 Antecedent autumn precipitation for the six largest period-of-record floods on Bella Coola River. 24 hour precipitation recurrence intervals are calculated using 80 years of record at the Bella Coola climate station. Mean separation of storm precipitation peaks is 4 to 4.5 days.

Analysis of synoptic charts for ten-day sequences prior to each major flood on the Bella Coola River suggests a number of synoptic conditions must be met in order to produce significant runoff. The most important is the rapid reduction in the velocity of eastward moving low pressure systems and in several cases a switch to a more northerly trajectory into the Gulf of Alaska. This is facilitated by high pressure ridging over the interior of the province. Warm and cold fronts become fully occluded as the centre of low remains over the coastal boundary. The warm sector aloft produces heavy rainfall and snowmelt at higher elevations. With the exception of the smaller 1982 flood, cyclogenesis is almost always centered south of the Aleutian Low between 155° and 170° W and between 45° and 55° N.

Storm precipitation plots in figure 3.6 also reveal a major difference in synoptic climatology for floods within a sequence of years of above average precipitation (*e.g.* 1965, 1968). These events are characterized by a closely spaced sequence of lows intersecting the coast under a synoptic regime which persists for at least ten days prior to the storm and in some cases longer. Floods during other intervals do not have strong antecedents and are generally singular events (*e.g.* 1950, 1973).

The spatial variability of autumn floods requires consideration if inferences regarding regional hydrology drawn from site specific evidence are to be valid. The intensity or size of cyclonic disturbances, storm track orientation and source areas for cyclogenesis may significantly affect one area without any appreciable effect in another.

Evidence presented earlier suggests different winter precipitation trends along the north and south coasts, where the central coast is a transitional zone alternately exhibiting characteristics of one region or the other. For this reason the magnitude and timing of fall flood events in the Bella Coola basin are compared with Homathko and Squamish Rivers, catchments of similar size and glacier cover in the central and southern coastal environments, respectively (see figure 2.1 for locations). Figure 3.7 is a plot of peak autumn discharges above a base discharge in each river for period-of-record data.

It is apparent from figure 3.7 that autumn flood magnitudes for a given storm vary spatially. For example, the period-of-record maximum flood in October 1957 on Homathko and Squamish Rivers was only the sixth largest on the Bella Coola River. Although magnitudes may vary, the significance of autumn storms can be seen in the coast-wide response to rainfall and rain-on-snow events. The major autumn flood of late October, 1965 on Bella Coola River corresponds with significant runoff events in the other coastal rivers, particularly Squamish River. The less precisely matched record with Homathko River may be due to the higher baseflow for that river and differences in physiographic or glaciologic effects. Runoff events in the Bella Coola basin during 1967, 1980 and 1982 were also recorded in these other rivers. This suggests that there is potential for using hydrological records of any one of these rivers as an index of major synoptic-scale storms for southwestern British Columbia. These synoptic-scale storms then become 'key' events in the longer term departure of climate along the central and southern coasts.

Another feature common to all three rivers is the increased number of floods during the period 1957-1968. This is coincident with above average October - April precipitation along the coast associated with an increased frequency of ridging along the Pacific Northwest specifically, and the northern Hemisphere in general (Treidl *et al.*, 1981; Knox and Hay, 1985). The evidence indicates that positive SST anomalies for the northeast Pacific Ocean and an intensified Aleutian Low were major factors.

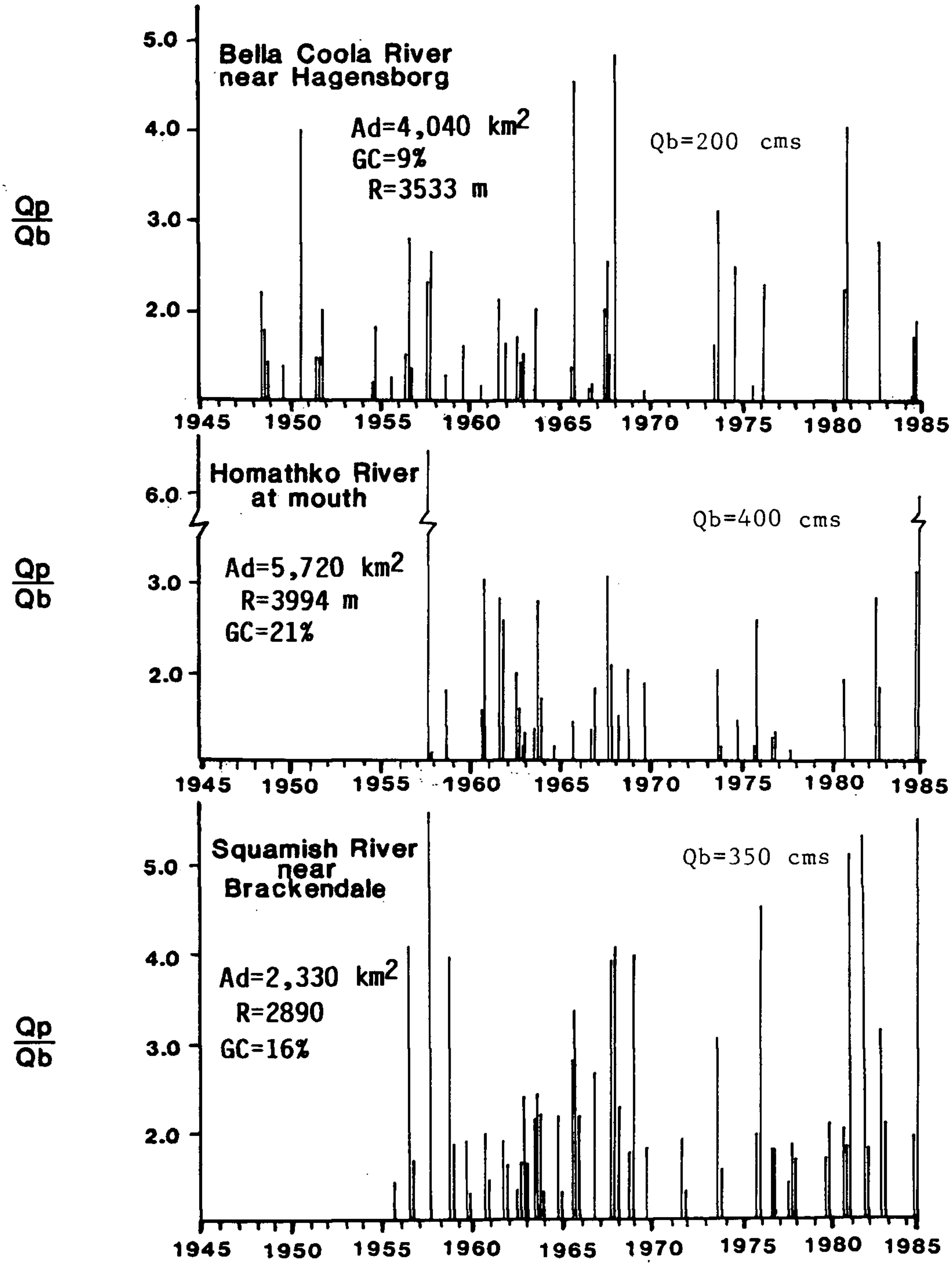


Figure 3.7 Frequency of autumn (August to January) rainstorm generated floods of the Bella Coola, Homathko and Squamish Rivers. Q_p is the maximum mean daily flow above a base discharge Q_b. Q_b was selected as the mean August-September daily discharge to emphasize flood peaks above the maximum glacier melt contribution. Statistics include drainage area (A_d), relative relief (R) and % glacial cover (%GC).

In summary, flooding on the Bella Coola River is related to several seasonally-averaged climate conditions as well as singular anomalies in the general circulation. High magnitude spring runoff is a function of accumulated snowfall throughout the winter season and spring temperatures: a short-term response to seasonally integrated climate. A higher probability of spring flooding occurs during intervals of increased winter precipitation (*e.g.* 1964-1969). Peak autumn floods, which generally produce the largest instantaneous discharges, are not restricted to these same intervals, although there is a higher frequency of flows above a given base discharge during years of above average winter precipitation. The major implication for inferences regarding seasonally-averaged hydroclimates is that certain flood-related evidence may not be exclusively associated with a particular runoff regime or interval of persistent climatic departures.

Flood Frequency

Five different runoff regimes can be identified for the Bella Coola River: spring snowmelt, glacier melt, rainstorm only, and spring and autumn rain-on-snow. Rain-on-snow during the autumn produces the highest discharges while spring snowmelt and rainstorm augmented glacier melt are probably the next most important in terms of peak discharge and sustained high runoff. In hydrological terms each could be treated as a separate population with distinct distributions and individual statistical properties (*cf.* Waylen and Woo, 1983).

In terms of impact on river morphology, channel and bank erosion, sediment transport and related alluvial processes, distinguishing between these several distributions is less important. All flows above the threshold for significant sediment motion should be considered. Therefore, the partial duration series is used here because it is based on all flows above a given discharge regardless of their ordering in time. The partial duration series of all four gauged catchments and inferred flows of the Talchako River are shown in figure 3.8.

The curves in figure 3.8 are ordered on the basis of catchment size with the exception of the Talchako River which produces higher-magnitude floods compared with the larger Atnarko River. Those basins with higher percentages of ice-cover have proportionately higher flood magnitudes with increasing recurrence intervals (*i.e.* the ratio of Q_{20}/Q_2 is larger for the Talchako, Nusatsum and Salloomt Rivers). At recurrence intervals less than 1.5 all plots exhibit similar curvature. The partial duration series for Bella Coola R. begins to depart from the extreme value series at flows near Q_3 which corresponds with the mean flood calculated using the partial duration series. Thus, the frequency of flows with moderate magnitudes on Bella Coola River (approximately 400 - 600 m^3s^{-1}) is notably higher on the partial duration series.

This distinction remains an arbitrary one if only high-magnitude floods (Q_3 or perhaps even as high as Q_{10}) produce significant changes in the alluvial system. However, if in addition to high-magnitude floods sequences of moderate floods have a noticeable impact on the alluvial system, then an accurate estimate of their frequency is important. Changes in hydroclimatology which favor a higher frequency of flood flows in both autumn and spring seasons are apt to be better represented by the partial duration sequence. Possible shifts in the long-term flood frequency curves in response to changes in the hydroclimatic regime are considered below and in chapter seven.

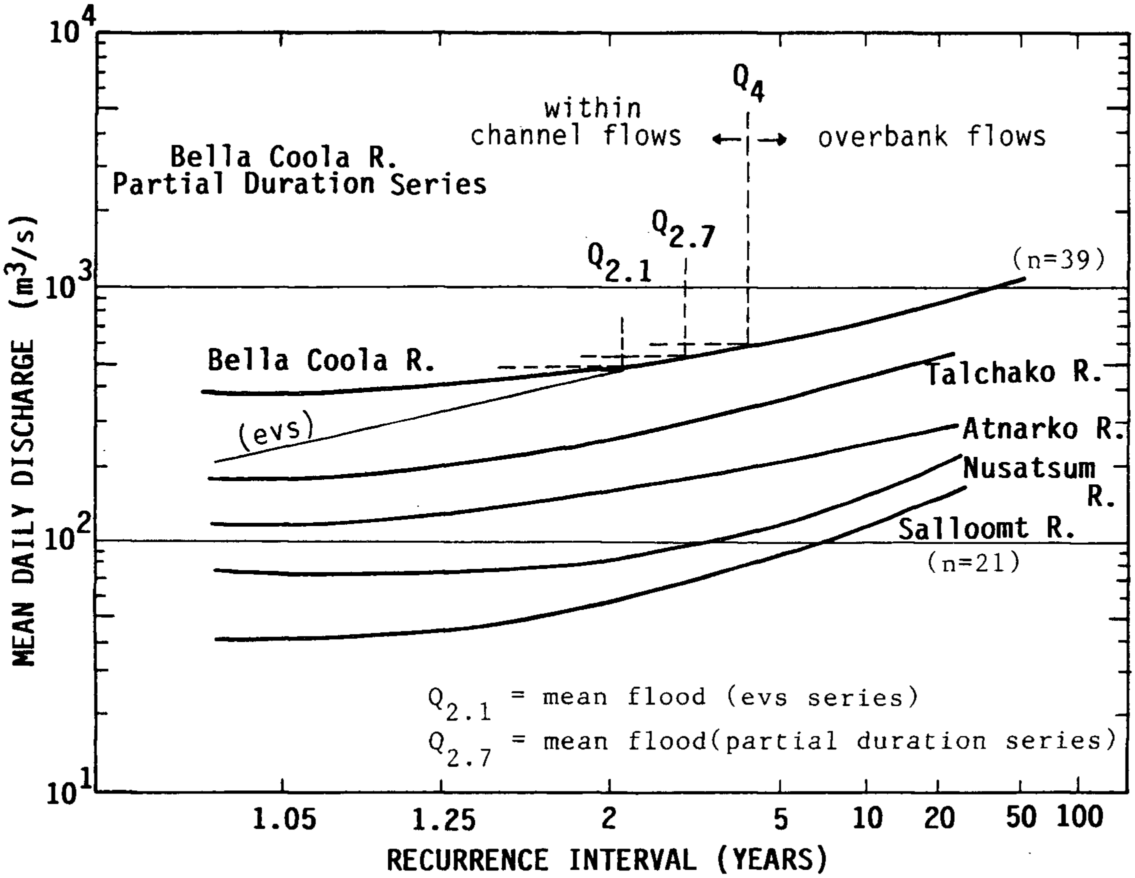


Figure 3.8 Partial duration series for runoff from all five gauges in the Bella Coola River watershed. Within-channel and overbank flows are defined for the stable reach in which the main gauge is located (Bella Coola River above Burnt Bridge Creek).

(3.3) Longer-Term, High-Resolution Climate Analysis

The frequency of both extreme and persistent departures in climate during the 20th century can be inferred from only a small number of climate stations in the study area. Although inferences of basin-wide climate will be biased towards these spatially-restricted records, comparison with other regional records and notable hemispheric changes in dominant circulation patterns provide a basis for assessing the utility of the records as sources of regional climate information. Monthly data are examined here for three reasons: 1) to augment and extend the short-term, regional records; 2) to provide a basis for comparison with biological and geophysical responses in the basin and; 3) to identify the occurrence of extreme intra- and interannual variations in the longer-term climate record.

Neilson (1986) has displayed records of precipitation and temperature as contoured, two-dimensional surfaces. This technique has been adopted here for the two longer-term stations in the vicinity of the study area. The Bella Coola climate station is assumed to be representative of environmental conditions towards the west side of the basin specifically, and coastal conditions in general. Big Creek climate data are used to characterize trends in the eastern basin or Fraser Plateau area (refer to figure 2.1 for locations and table 3.1 for factor groupings).

Contour surfaces of total monthly precipitation and monthly averages of mean daily temperature for the period 1904 to 1983 are illustrated in figures 3.9 and 3.10. Fifteen months of data are presented along the horizontal axes representing the current year and the last three months of the previous year. The most obvious features of figure 3.9 are the winter maximum in precipitation for Bella Coola and a trend towards above average precipitation in the latter half of the century (with the exception of high October to November precipitation between 1934 and 1940). There is also

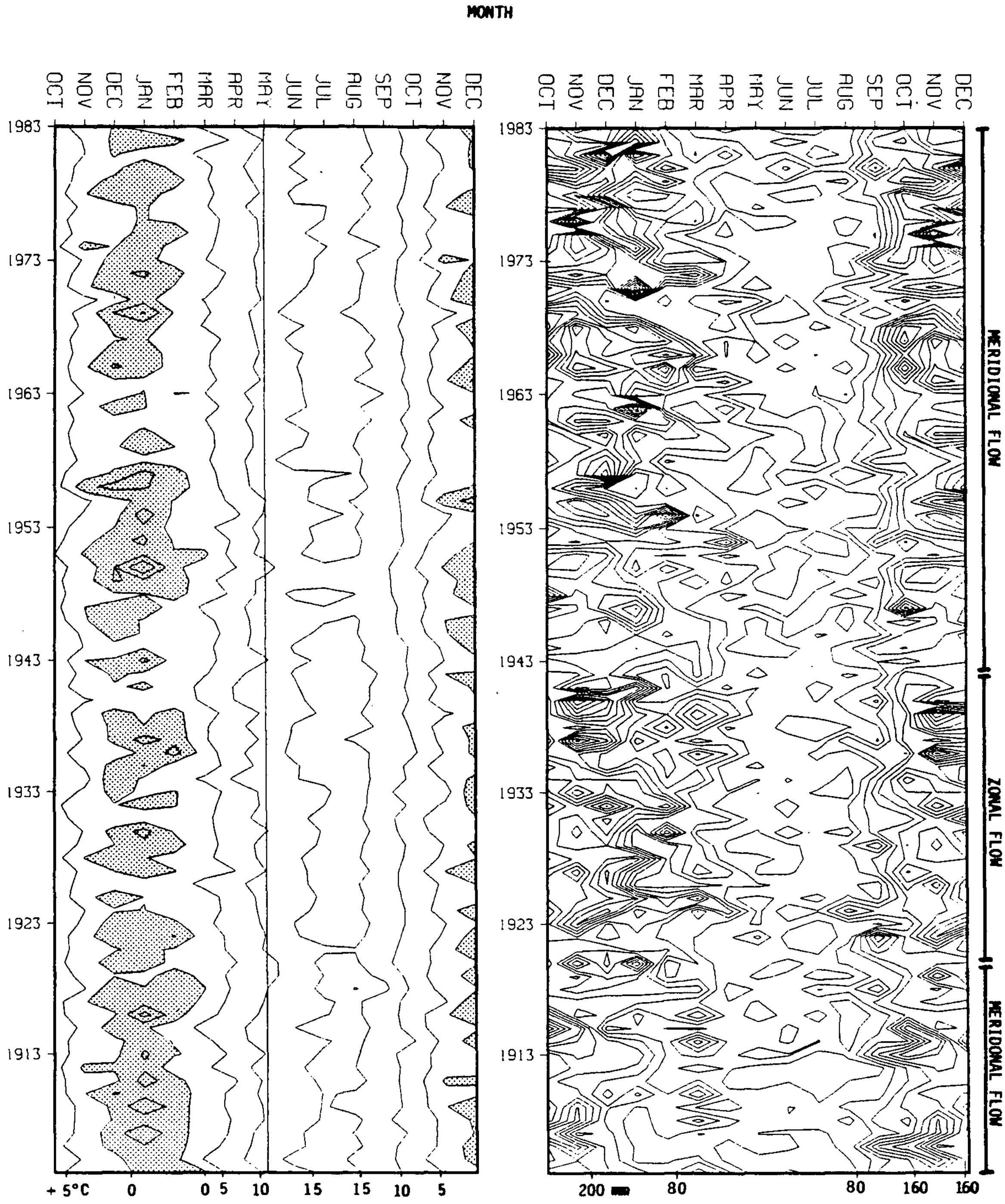


Figure 3.9 Contoured surfaces of intra- and interannual temperature and precipitation for Bella Coola climate station, 1904-1983 (see figure 2.2 for station location). October to December values on the left are from the previous year. Temperature contour interval is 5° C and precipitation contour interval is 40 mm. The shaded areas on the temperature plot represent extreme departures.

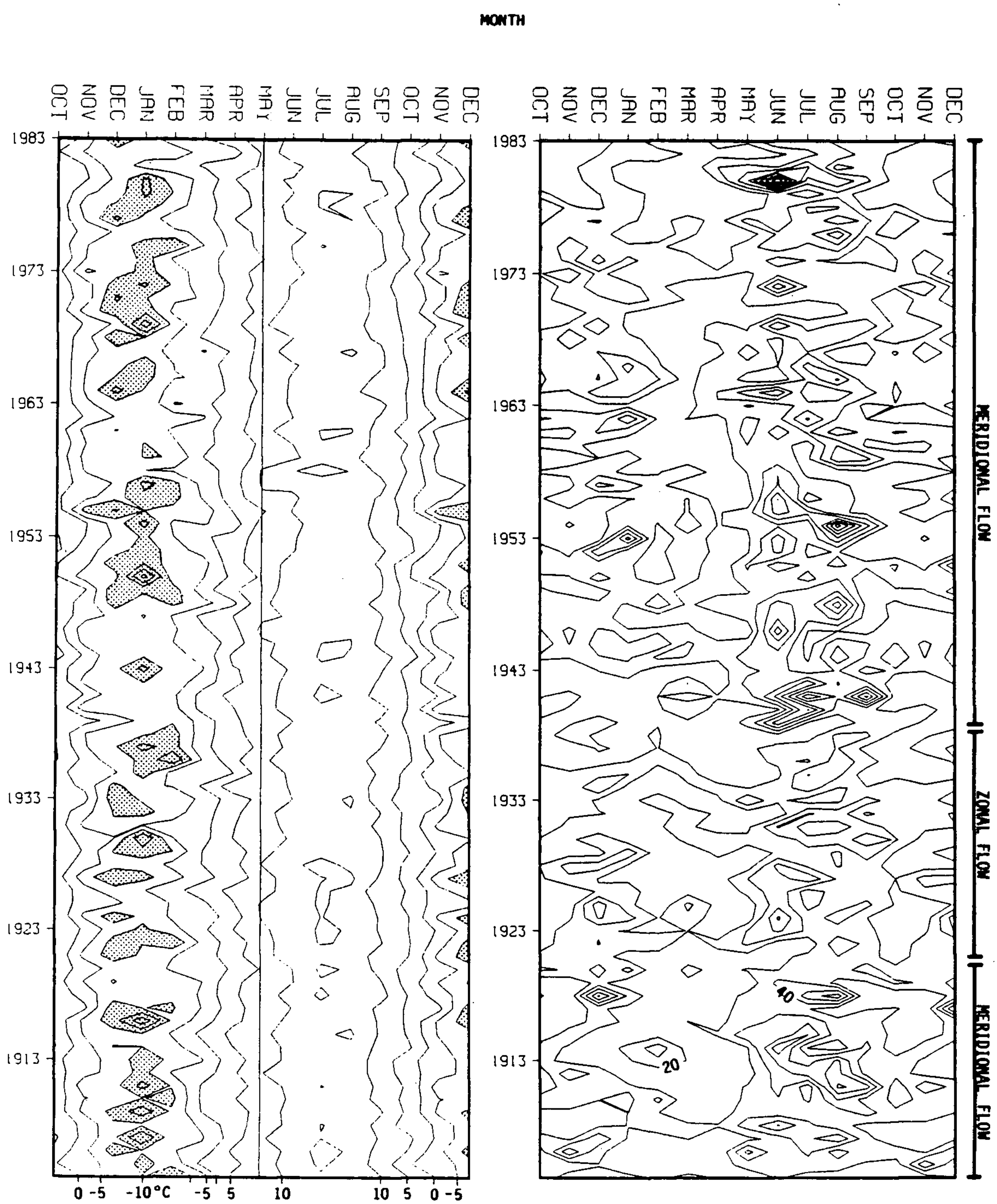


Figure 3.10 Contoured surfaces of intra- and interannual temperature and precipitation for Big Creek climate station, 1904-1983 (see figure 2.1 for station location). October to December values on the left are from the previous year. Temperature contour interval is 5° C and precipitation contour interval is 20 mm. The shaded areas on the temperature plot represent extreme departures.

evidence for higher and more persistent October to January precipitation after 1953. Summer precipitation was lowest between 1924 and approximately 1945 and during the interval 1963 to 1976.

Mean December to February temperatures were consistently below 0°C until 1920 after which gradual warming occurred until at least 1945 (note the frequency of closed contours < -5°C prior to 1920). Post-1945 winter temperatures, particularly 1948 to 1958 and 1962 to 1976, were more similar to those recorded prior to 1920. A tendency towards warmer springs in the latter half of the century is also discernible (see the slope of the 5 and 10°C contours near March and April). Cooler summers prevailed until approximately 1930 after which higher temperatures were evident with the exception of 1945-1957 and 1968-1976.

Some interesting differences and similarities are associated with climate trends at Big Creek (figure 3.10). A summer maximum in precipitation dominates there with lowest summer totals occurring between 1920 and 1940. Summer and winter precipitation is generally higher after 1940 compared to the first half of the century. Like Bella Coola, pre-1920 winter temperatures were amongst the lowest in the century, although unlike Bella Coola warming was not so evident until after 1940. The cooler winters of 1948-1955 are most similar to the pre-1920 temperature departures. Early century summer temperatures at Big Creek were comparatively warm but maximum temperatures were not recorded until after 1920, ending around 1930 with a cooling trend which lasted for the next decade. Summer temperatures after 1958 are the lowest of the century with the exceptions of 1978 and 1979.

With the exception of the interval 1930 to 1940, the pre-1950 period at Bella Coola is characterized by years with less persistent departures of intra-annual monthly precipitation when compared with the post-1950 record (figure 3.9b). This is particularly true for the winter months. At Big Creek, summer and early autumn months exhibit persistent positive departures in the pre-1920, 1940-1965 and post-1980 records. Extremes in monthly precipitation tend to cluster within these periods of positive departure but are not exclusively found there.

The contoured surfaces would suggest a non-random ordering of climate departures through the 20th century. It has been widely recognized that three distinct periods of characteristically different circulation regimes have occurred over the last 80 years related to transitions from meridional flow patterns (pre-1920 and post-1945) to zonal flow patterns (1920-1945; see table 3.8). When the energy gradient between the pole and equator is steep, a series of high amplitude waves generate stronger longitudinal contrasts in climate than would be experienced under a zonal flow regime (Makrogannis *et al.*, 1982). Persistent and often extreme departures occur when the meridional wave pattern remains entrenched. Even small longitudinal shifts in the mean position of blocking ridges and associated troughs can lead to abrupt changes in the direction of departure. Under a zonal flow regime, year-to-year variance is enhanced and the frequency of persistent departures is greatly reduced.

Blasing and Lofgren (1980) have analyzed monthly sea-level pressure data for the N.E. Pacific sector covering the period 1899 to 1971. Using the map correlation method, they were able to classify objectively dominant circulation types characteristic of each period. During winters of the first two decades of this century above normal pressures in northern Canada resulted in an anomalous flow of cold arctic air, presumably associated with a southward displacement of the arctic front, resulting in cooler and moister conditions along the outer British Columbia coast. A strong

Table 3.8. Recognized Trends in 20th Century Synoptic Conditions in the N.E. Pacific Sector.

INTERVAL	SYNOPTIC CONDITIONS	METHODOLOGY	SOURCE
1900 -1920's	- meridional flow (some zonal)	classification of N. hemisphere surface pressure maps by (1) and eigenvector analysis by (2)	(1) Dzerdzeevski (1966, 69)
1920's -1950's	- zonal flow		(2) Kalinicky (1974)
post 1950's	- meridional flow		
1900 -1920's	- decreased intensity of the January Aleutian Low	eigenvector analysis of N. hemisphere sea level pressure maps for both January and July	Kutzbach (1970)
1920's - 1950's	- intermediate conditions		
1950's -1970	- increased intensity of the January Aleutian Low and an eastward extension		
1899 - 1920's	- above normal pressure in high latitudes of the Pacific (cool)	map correlation method using sea level pressure data	Blasing and Lofgren (1980)
1920's - 1940's	- below normal pressure in N. Pacific sector (warm/dry)		
post 1940's	- more average conditions		
1893 - 1920	- cooler/wetter in U.S. West	areally averaged monthly mean temperature and precipitation for the contiguous U.S.	Diaz and Quayle (1980)
1920 - 1954	- warmer/drier		
1955 - 1977	- cooler/wetter		
1895 - 1920's	- increased frequency of cool/wet synoptic types	map correlation method of surface pressure data using Kirchhofer cutoff scores	Barry et al. (1981)
1920's - 1945	- increased frequency of warm/wet synoptic types		
post 1945	- more variable conditions but warm and wet		

1875 - 1905	- higher precipitation in the Pacific N.W.		
1905 - 1945	- reduced variability in and totals of precipitation	eigenvector analysis of precipitation records along U.S. west coast	McQuirk (1982)
1945 - 1975	- higher precipitation and greater variability		
1872-91 1902-11 1932-41 1962-71	- wet periods in northern California associated with a northward shift in the Pacific High	eigenvector analysis of precipitation records in the U.S. West	Granger (1979) Wahl and Lawson (1971)
1892-1901 1912-31 1942-61	- drier periods associated with a southward shift in the Pacific High		
early 20th C	- southward displacement of Aleutian Low and Pacific High - cooler/wetter	analysis of Pacific sector pressure data for the 20th century	Angell and Korshover (1974,82)
middle 20th C	- northward displacement of Aleutian Low and Pacific High - greater circulation intensity		
1947 - 1966	- higher 700 mb heights in Gulf of Alaska and Aleutians	analysis of 700 mb pressure data and SST anomalies	Douglas et al. (1982)
1966 - 1980	- lower 700 mb heights - increased zonal index		
1951-54 1966-71	- maximum blocking frequency in N. hemisphere	analysis of 500 mb pressure anomalies for N. hemisphere	Treidl et al. (1981) Knox and Hay (1985)
1960-65 1974-78	- minimum blocking frequency in N. hemisphere		
1948 - 1965	- slightly higher 500 mb heights for N.E. Pacific	upper air pressure data grouped into several synoptic types	Yarnel (1985)
1965 - 1978	- slightly lower 500 mb heights for N.E. Pacific		

negative pressure anomaly in the Aleutians led to an increased southwesterly flow aloft and above normal summer precipitation.

Reduced precipitation and higher temperatures after 1920 followed a northward displacement of cyclonic activity as positive pressure anomalies dominated the central Pacific, particularly during the summer. Circulation patterns between 1934 and 1937 best exemplify this period (Blasing and Lofgren, 1980: 717) especially in the sequence of spring and summer surface pressure anomalies; a feature which is most evident as warmer summer temperatures recorded at Bella Coola (figure 3.9) and significant because two of the largest autumn floods on the Bella Coola River occurred in 1934 and 1936. After 1945 a return to meridional flow, although with higher amplitudes in the Rossby wave pattern, was associated with positive pressure anomalies in the western United States (Blasing and Lofgren, 1980).

Monthly and seasonal variance of temperature and precipitation at Bella Coola and Big Creek were evaluated for these three periods (pre-1920, 1920-1945, post-1945) along with an index of year to year variability of monthly values. Although some seasonal differences between stations exist, within-period variances of summer and autumn temperatures were consistently higher (average of 40% greater) in the earlier and later periods. In addition, variance of summer and winter temperatures was lowest between 1920 and 1945. Similarly, autumn and spring precipitation exhibit highest within period variances (average of 65% greater) for the post-1945 period. The differences in between-period variances are partly a function of the frequency of extreme departures. Maximum 24-hour precipitation events of 80 mm or more at Bella Coola have a recurrence probability of less than 20% between 1920 and 1945 compared with 25-30% probability during the other two intervals.

Mean sensitivity is used here as an index of year to year fluctuations within each period.[2] A high mean sensitivity indicates the propensity for changes in the sign of departure between successive years within a given interval, and is thought to represent a lack of persistence. Zonal circulation regimes characterize these intervals. Precipitation at both stations for most months of the year between 1920 and 1945 exhibits the greatest year-to-year fluctuation (15 to 25% more compared with other intervals). Autumn and winter temperatures also exhibited higher mean sensitivities between 1920 and 1945; however, summer temperature, particularly at Big Creek, shows a higher mean sensitivity between 1945 and 1983. With the exception of this latter result, mean sensitivity values give some support to the secular variations in circulation regime shown in figures 3.9 and 3.10.

Comparisons With Other Regions

Although the number of long-term stations (> 80 years record) in coastal British Columbia is limited, there are some obvious similarities for the entire coastal region. Three-year running mean curves presented by Powell (1965) show a trend towards average and above average annual precipitation prior to 1920 for the outer coast (*e.g.* Clayoquot and Quatsino). This trend is less pronounced or reversed for the central and southern coasts suggesting that the non-synchronous response in winter precipitation for coastal regions prior to 1920 was driven by synoptic controls similar to those identified for the post-1945 period. Data

2. Mean sensitivity is calculated as follows:

$$M_{sx} = (1/n - 1) \sum_{t=1}^{n-1} \left| [2(x_t + 1 - x_t)]/(x_{t+1} + x_t) \right|$$

x_t = value of climate variable in year t; n = number of years presented by both Powell (1965) and Karanka (1986) show a higher mean sensitivity during the intervening period (1920-1940) with no obvious or persistent departures in precipitation.

Cooler than normal winter temperatures between 1904 and 1920 at both Bella Coola and Big Creek are also well documented in Karanka's Queen Charlotte Islands temperature index and the outer and inner coast climate stations analyzed by Powell. Although complete records in Bella Coola do not extend beyond 1904, evidence from other climate stations operating in the late 19th century (Queen Charlotte Islands, Clayoquot, Steveston) demonstrates that from approximately 1890 to 1904 winter temperature departures were mostly positive. The tendency for above average winter temperatures in Bella Coola between 1920 and 1945 is also evident for stations along the North Coast.

(3.4) Conclusions

Temperature and precipitation variability along the central British Columbia coast exhibit patterns which show some correspondence with noted 20th century changes in circulation regime. Climatic anomalies within these periods are short and not well coordinated between temperature and precipitation. The recent record shows that the synoptic signature for anomalously wet winters is both a higher frequency and greater seasonal persistence of specific synoptic types. Certain hydrologic variables such as snowpack and spring flooding respond to persistent climatic departures. Extreme departures in precipitation, which produce key hydrologic events, (*e.g.* autumn floods) are more randomly distributed across periods but have a higher probability of occurrence under a meridional flow regime. There is a definable regional signal of flood anomalies along the southwestern coast of British Columbia. Sedimentary deposits within drainage basins of this region are more likely to reflect these extreme departures. However, other elements of the biophysical systems are known to have memory. The following chapters will explore the actual patterns of response.

CHAPTER IV
Geomorphological Evidence of Response of Geophysical Systems To Environmental Change:Tests Within the Historical Record

(4.0) Introduction

In response to hydrological changes, certain effects are apt to occur within biological and geophysical subsystems. The purpose here is to examine several of these subsystem responses within the short term, using the period of instrumental record in the Bella Coola basin, and to test the utility of biogeophysical data for drawing inferences about local and regional hydrological changes. The responses examined include: glacier fluctuations, upland sediment transfers, and alluvial activity of the Bella Coola River.

(4.1) Glacier Fluctuations and Recent Climate Change

A valuable indicator of environmental change in mid and high latitude glacierized regions is the fluctuation in water storage reflected as changes in mass and ice-front positions of glaciers. Persistence in the rate of snow accumulation and glacier ablation over several seasons or more, can lead to substantial changes in the position of the ice-front and the deposition of ice-eroded materials which can then be used to infer the rate of ice-front movements. The purpose here is to establish the nature of the relationship between glacier fluctuations and prevailing climate so as to provide a basis for interpretation of glacier deposits indicative of former glacier configurations and, ultimately, associated hydroclimatic conditions.

Several factors complicate the relationship between glacier fluctuation and hydroclimate. One group of factors is related to annual changes in the water (snow/ice) mass balance of a particular glacier system. A second group is associated with glacier responses, such as fluctuations in movement integrated over the whole ice mass or indexed by changes in glacier snout position. In the first group, climatic factors such as winter precipitation and summer temperature generally account for much of the variance in the net ice-mass balance. However, they remain imperfect indices because of the importance of other parameters such as the redistribution of snow by wind and avalanching, ice-calving, and sampling and measurement errors in determining actual mass-balance for a given season (Fogarasi and Mokievsky-Zubok, 1978; Mayo and Trabant, 1984). Therefore, it is unlikely that single or even several point estimates of temperature and precipitation can fully account for fluctuations in net mass balance from year to year.

The second group of factors is related to changes in ice-front position. Several studies have demonstrated that snout retreat may not be synchronous with overall losses in total glacier volume and that glacier sensitivity to changing climate is related to glacier size and physiographic boundary conditions (Ahlmann, 1953; Hoinkes, 1968; Smith and Budd, 1981). The actual rate of movement will depend on site-specific factors such as slope, degree of confinement, aspect, total ice thickness and sub-glacial drainage features.

Several empirical and physically based models have been proposed to explain the variation in annual ice mass balance for temperate valley and cirque glaciers. Kuhn (1981) has developed a physical model which relates changes in mass to perturbations in parameters such as free atmospheric temperature near the glacier, total annual accumulation (from all sources) and the radiation balance. Alternatively, Tangborn (1980) developed an empirical model which requires estimates of temperature and precipitation measured at nearby stations to estimate fluctuations in the annual mass balance. Although derived from different assumptions, the two climate-based models of glacier fluctuation appear to yield similar results regarding changes in temperature and precipitation corresponding to specific fluctuations in mass balance.

In both studies cited above net mass balance appears to be most sensitive to changes in winter precipitation. This is in contrast with the simulation model results of Smith and Budd (1981) who found temperature to be the more important variable. This difference is probably related to climate variations at both the global scale (latitude/longitude and maritime or continental influences) and local site scale (windward/leeward effects). Hence, the relative significance of precipitation and temperature as indices of glacier fluctuations may not be consistent between regions.

Mass Balance Fluctuations of Glaciers in Southwestern B.C.

Several empirical models are examined in order to assess the relationship between climate fluctuations (temperature and precipitation) and glacier response in southwestern British Columbia. Glacier mass balance and equilibrium line altitude (ELA) data are available for three glaciers in the Squamish/Lillooet watersheds beginning in 1965, and shorter series commencing in 1976 for three glaciers in the headwaters of Bridge River. All these sites are located to the south of the Bella Coola River basin (see figure 2.1).

Monthly and seasonally averaged climate data (temperature and precipitation) beginning in 1966 were transformed into two sets of regional climate indices representing the wetter south coast (Alta Lake, Garibaldi, Whistler roundhouse stations) and the drier interior (Tatlayoko Lake, Tatla Lake, Kleena Kleene and Big Creek). The latter set of indices is that used to represent the central coast climates in section 3.1 and was selected because of the proximity of these stations to the Bridge River glaciers. Although there are other stations closer to the Bridge River basin, these were selected because of the higher spatial and temporal coherence of temperature measurements amongst interior climate stations and the greater regional significance of any derived relationships.

Independent variables consisted of winter precipitation and summer temperature for each of two regions (coastal and interior) and winter mass balance determinations for each glacier. Unlike the Tangborn model, temperature range (an indirect measure of cloud cover and radiative fluxes) was not incorporated here since this variable proved to be of low significance in his model and because such a parameter is not easily constructed using indirect or proxy data. Table 4.1 is a summary of multivariate exploratory models for each glacier.

In all three cases the strongest response model (model [1]) incorporates winter mass balance and summer temperature, accounting for 73 to 85% of the explained variance in annual net mass balance. When winter mass balance is excluded (model [2]) the explained variance remains significant for Sentinel and Place Glaciers but declines to between 52 and 59%. A lower degrees of freedom for Bridge Glacier yields a non-significant model when tested under more restrictive confidence levels ($\alpha = 0.025$). In each model [2] case, winter precipitation contributes more to the explained variance than summer temperature, especially for Sentinel and Place Glaciers which are located in a wetter, more maritime, region to the southwest.

The region from which climate indices are derived appears to be important. For example, coastal precipitation is the more significant variable in explaining annual mass balance changes on Sentinel Glacier whereas interior precipitation is of greater significance for the other two

Table 4.1 Empirical response models for glacier net mass balance and climate

Glacier	Model(1)	Winter Mass Balance	Winter Precipitation(2)	Summer Temperature(2)	d.f.	Ra^2	F
		Independent Variable (contribution to explained variance)					
Sentinel	[1]	76%		24% (i)	18	73%	25.7
	[2]		65% (c)	35% (i)	18	59%	11.4
Place	[1]	88%		12% (i)	18	73%	21.0
	[2]		79% (i)	21% (i)	18	56%	10.4
Bridge	[1]	81%		19% (i)	7	85%	15.9
	[2]		54% (i)	46% (i)	7	52%	4.8*

1. Model [1] includes all significant variables; model [2] excludes winter net mass balance from the list of independent variables
2. Indices generated using coastal stations are marked (c) and those generated from interior stations are marked (i).
3. * not significant at $\alpha = 0.025$

sites. This was not unexpected since Sentinel Glacier is situated in a belt of heavy precipitation centered on Mount Garibaldi. Temperature indices generated from interior stations are more significant for all three glaciers than are coastal indices. This may be a result of the lower overall variance in summer temperatures for the coast due to persistent cloud cover.

While only a limited number of independent variables was considered here, and the degrees of freedom are generally low, a significant portion of net annual mass balance in coastal glaciers can be explained using readily derived temperature and precipitation indices. The three glaciers examined represent a range in characteristics such as size, slope, aspect, accumulation area morphology and local hydroclimatology and it is likely that these factors contribute to the residual variance. Yet, regionally derived climate indices form a set of significant independent variables, albeit of varying importance between sites.

Primary controls of net annual balance are winter snowpack measured on the glacier (regional snow course data were not strongly significant variables although they were tested) and summer temperature. Substituting regionally estimated winter precipitation, which will usually be available, reduces the strength of the relationship but the model remains significant. Data plotted in figures 3.1 through 3.5 indicate that the interval 1966 to 1984 was characterized by some of the greatest departures in temperature, precipitation, winter snowpack and runoff. Although this extreme variability extends the calibration range of the Sentinel and Place Glacier models, a limitation of linear models is that they represent mean responses only.

The level of explained variance in these glacier-climate regression models is as high as that found in the spring runoff models. The 40 to 50%

unexplained variance using model [2] is related to at least three factors: the relative importance of temperature and precipitation at each site in determining net mass balance, the significance of extreme departures in controlling variables, and the representativeness of the regional climate indices.

Mass Balance and Climate

Since temperature and precipitation are only index climatological parameters, the linkages between glacier mass balance and synoptic scale circulation patterns were examined. Yarnel (1982) found that a significant proportion of mass balance variability for two Cordilleran glaciers during the interval 1966 to 1974 could be explained by the frequency of certain synoptic regimes. The frequency of synoptic types which produce above or below average winter precipitation and summer temperature for coastal British Columbia is tested here against groups of individual years in which winter and summer mass balances were above or below average. Because of the limited length of the data series and high interannual variance, periods of persistent positive or negative mass balance departures could not be defined with confidence. Results are presented in tables 4.2 and 4.3.

There are, on average, 11 more winter days dominated by "wet" 500 mb synoptic types and 17 more winter days characterized by moisture producing surface pressure types during strongly positive mass balance years compared with low mass balance years. In both cases these differences are significant. There is no significant difference in the frequency of "warm/dry" synoptic circulation types between years of high and low summer mass balance. This latter result is related to the difficulties in objectively identifying circulation regimes which might produce climate departures (*e.g.* warm or dry) during the summer season when the overall

Table 4.2 Winter Mass Balance on B.C. South Coastal Glaciers and Synoptic Type Frequency

Significance test between high winter mass balance years and low winter mass balance years

Mass balance years consistent over all glaciers	Mean winter mass balance (mean departure)[1]	Wet synoptic type frequency[2]	Wet synoptic type frequency[3]
high winter mass balance :(1976, 74, 1971,68,67)	1.48 ± 0.79	90 ± 4.1	46.6 ± 7.4
low winter mass balance :(1978, 77, 1975,70,72)	-0.74 ± 0.58	79 ± 5.4	30.4 ± 6.1
One-tailed t-test (α = 0.10, df = 8)	t = 4.42 significant	t = 3.63 significant	t = 3.78 significant

1. Calculated as the mean departure in winter mass balance for two (1966-1975) and five (1976-1984) coastal glaciers.
2. 500 mb synoptic types as defined by Yarnel (1983).
3. Surface pressure types as defined by Barry et al. (1982).

Table 4.3 Summer Mass Balance on B.C. Coastal glaciers and Synoptic Type Frequency.

Significance test between high summer mass balance years and low summer mass balance years

	Mean summer mass balance (mean departure)	500 mb Dry/warm synoptic type frequency
high summer mass balance years: (1977,71,70,69,67)	0.18 ± 0.17	23.6 ± 6.5
low summer mass balance years: (1978,76,75,73,72)	-0.18 ± 0.07	20.0 ± 5.2
One-tailed t-test (α = 0.10, df = 8)	t = 4.36 significant	t = 0.97 not significant

circulation intensity is low and day to day circulation variability is reduced. It may also reflect the lower significance of summer temperatures in the mass balance models due to the dominance of maritime conditions in the lee of the Coast Mountains.

In general, high mass balance years are characterized by an increased frequency of circulations with strong cyclonic curvature favoring the accumulation of snow along the south coast. Summer ablation does not appear to be simply related to the frequency of 500 mb synoptic types which are thought to favor warm/dry conditions. Hence, glaciers form on the coast only where winter precipitation is so high as to dominate the mass balance. Inland, where summer temperatures may dominate, fluctuations in mass balance are less predictable.

Glacier Snout Fluctuations

An alternative method for assessing the response characteristics of glaciers to climate change is to examine the record of fluctuations in the mean frontal, or in some cases lateral, position of several glaciers representative of a particular area. Paterson (1981) has outlined many of the key variables controlling glacier snout changes which ultimately respond to perturbations in mass balance. The most important geophysical factors are glacier size, bed slope, ice velocity, ice thickness and the geometry of accumulation and ablation areas. Smith and Budd (1981) have attempted to model the response of both small and large valley glaciers (3 to 30 km^2) to assumed functional changes in accumulation and ablation. This study, and others like it, have shown that glacier response in the form of length changes to some anomalous period of changing temperature or precipitation, can lag by as much as 150 years depending on the magnitude of the perturbation and glacier geometry.

Aerial photographs offer the only practical means for estimating the temporal variability in glacier snout changes over large remote areas. Photographs taken in 1947, 1954, 1964 and 1978/79 are available for the Bella Coola basin in addition to field-based measurements made between 1983 and 1985. While the coverage is not annual, the data do allow for at least a decennial resolution of changes. From an inventory of ice lowering and frontal retreat of 50 glaciers, six were selected as representative of the size and geometry of glaciers present in the basin. The selection was made on the basis of measurements available from those glaciers which were investigated in the field. Relative base line distances were established using prominent terrain features near each ice front and were measured using stereopairs correctly scaled and oriented on an Wild A6 stereo plotter. For simple and very regular ice fronts, recessional distances are the average of several measured line segments between two ice front positions. Where the ice margin geometry is more complex, surface area changes were computed and divided by the length of the ice margin to yield a mean rate of advance or retreat. The data were then standardized to reflect the magnitude and direction of ice front movements for each glacier. Figure 4.1 is a summary of ice front fluctuations for six Bella Coola glaciers during each period of measurement.

The major trend in the post-1945 period for all six glaciers is the switch from regionally persistent recession prior to 1964, followed by either a reduction in the rate of retreat or actual advances between 1964 and 1979. Examination of the 1964 photography did not show evidence for ice advance at that time (*e.g.* sediment free lobes with steep, crevassed fronts) suggesting that advances between 1964 and 1979 commenced sometime after 1964. For the larger valley glaciers such as the Fyles, Jacobsen and Talchako (the latter two of which are not shown in figure 4.1) recession

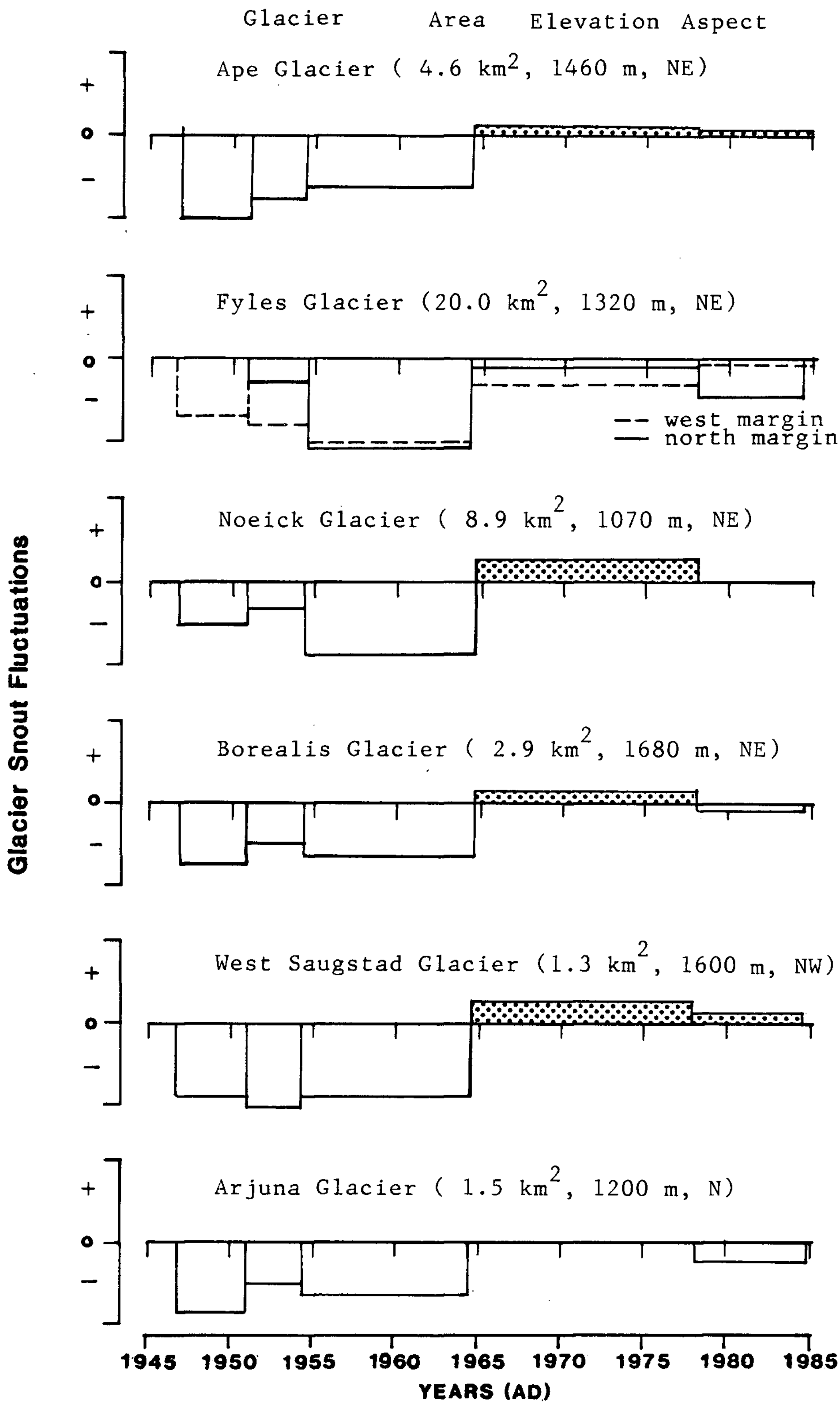


Figure 4.1 Relative rates of advance (+) and recession (-) of selected glaciers in the Bella Coola basin between 1945 and 1984.

was continuous but slower after 1964. Between 1954 and 1964 recession rates were average to above average for many glaciers, but for the two larger glaciers actual rates of retreat were increasing up to at least the early 1960s. Some of the smaller glaciers show declining recession rates throughout the 1947 to 1964 period. Between 1979 and 1983-85 the rates of advance declined or recession once again prevailed (figure 4.2).

Although boundaries separating the periods of different frontal movement rates are arbitrarily fixed by the dates of air photography, some general similarities between ice front fluctuations and prevailing climate are evident. It appears that for glaciers in this study area larger than approximately 10 km^2, continuous recession has occurred throughout the post-1945 period. This is a continuation of 20th century recession from Little Ice Age maxima documented for many parts of the northern hemisphere (Grove, 1979; Porter, 1981). The decline in recessional rates for these large ice masses and actual advances of some glaciers smaller than approximately 10 km^2 between 1964 and 1979 appear to be associated with the persistent above average winter precipitation along the coast, commencing in 1956 and lasting until at least 1968 (followed by several individual higher snowfall years, *e.g.* 1974, 1976). Winter precipitation was 7 to 10% above the post-1945 average during this interval. Increased recession rates or declines in the rate of advance between 1979 and 1983-85 are most likely related to the decreased winter precipitation and increased summer temperatures after 1976 for both the central coastal and central interior regions (see figure 3.1).

The above evidence indicates that there is an element of sensitivity in both large and small glaciers related to the magnitude of climate changes in the post-1945 period. An interval of between 5 and 10 years is an estimate of glacier snout response time to departures in regional

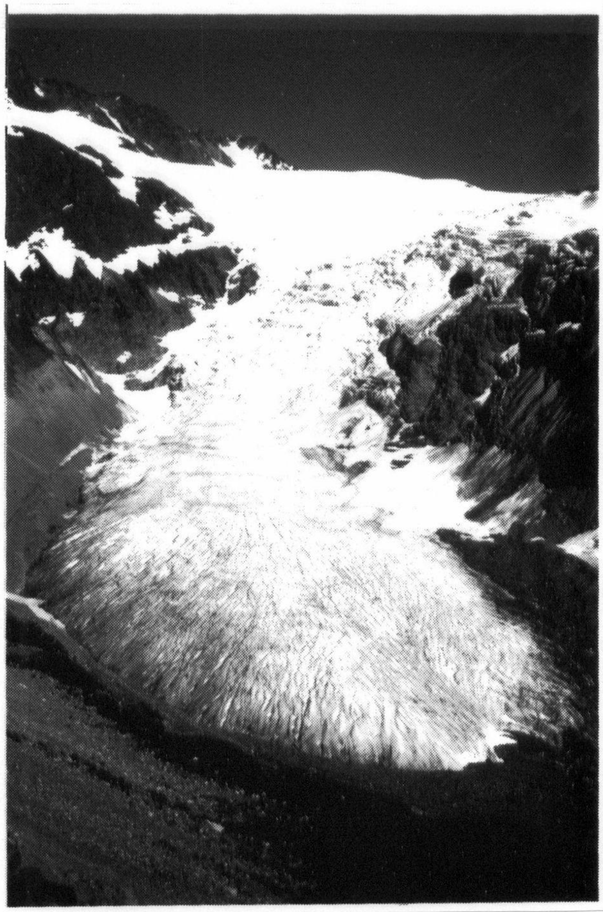


Figure 4.2 West Saugstad Glacier in late July of 1985. Note the lack of ablation debris on the ice front and the comparatively steep surface along the margin of the snout indicative of forward ice-movement.

precipitation and temperature which persist for at least 8 to 12 years. It is difficult to evaluate what magnitude of climate departure is necessary for a given interval to ensure a response in the glacial system but the evidence here seems to indicate not more than an average of 7 to 10% in winter precipitation for a decade. Interpretations of temperature and precipitation records prior to 1945 (see chapter 3) do not yield any evidence which would support the contention that the post-1964 ice front advances are due to climatic perturbations prior to 1945. Thus, the magnitude and persistence of climatic departures in the post-1945 period have resulted in significant responses in glacier systems of the Bella Coola basin and past climate changes of similar magnitude and duration would have affected rates of ice movement.

(4.2) Recent Fluctuations in Upland Sediment Transfers

Several methods are available to assess the magnitude and frequency of clastic sediment transfer from upland sources to the higher order fluvial network. The most reliable is repeated surveys of aggradation or degradation in natural hillslope gullies or tributary alluvial channels (Dietrich *et al.*, 1982) or in artificially constructed sediment traps or debris basins (Leopold and Emmett, 1976; Gardner, 1979). Unfortunately, several years or decades of monitoring may be required to produce meaningful results because of the lower sensitivity of hillslope elements to climatic perturbations (Brunsden and Thornes, 1979). Alternatively, measurements from air photographs taken at several different times can yield data on the frequency of mass wasting and, when the airphotos are of a large enough scale and some ground based measurements are available, reliable estimates of magnitude also can be made. Finally, indirect evidence in the form of damage to vegetation or detailed historical records may be useful.

Two methods were adopted to ascertain if there was any significant response of clastic sediment yields to changes in recent climate and runoff. These include (1) airphoto measurements of changes in channel width on tributary alluvial fans and (2) damage or burial of vegetation on these tributary fans which also provide some chronological data. The morphological analysis here is restricted to the post-1945 interval because aerial photographs were not available until 1946.

Aerial photographs taken in 1946, 1948, 1954, 1961, 1974 and 1982 were used to evaluate the temporal variability in channel widths on major tributary alluvial fans. These dates represent the end or beginning of shifts in hydrological conditions in the Bella Coola area (see chapter 3). Active channel areas on alluvial fans were computed for eight tributaries of the Bella Coola River. These include Thorsen, Snootli, Nooklikonnik, Noosgulch, Cacoohtin and Burnt Bridge Creeks and Salloomt and Nusatsum Rivers. Thorsen, Nooklikonnik and Salloomt basins are characterized by an absence of long-term storage sites. Most of the headwater reaches of Burnt Bridge Creek and the east fork of the Nusatsum River are long-term storage areas, whereas the remaining tributaries have isolated short-term storage sites (see figure 2.8). Other tributary alluvial fans, particularly those in headwater sediment-source areas (*e.g.* Talchako River) were either too confined or of limited longitudinal extent to reveal consistent trends.

Each stereo model was reduced to a common scale using a zoom-transfer stereoscope. Although not corrected for tilt or swing, base lengths for each model were calibrated against at least two distances measured in the field which were identifiable on each set of air photographs. Linear measurements are accurate to within ± 5 m. Channel areas were computed for the region which extends from the apex to the toe of the fan, excluding stable (vegetated) mid-channel islands. These areas are divided by channel length along the thalweg to yield an estimate of mean channel width. Figure 4.3a is a plot of standardized secular changes in mean channel widths between 1948 and 1982 for eight tributary fans.

Increases in channel width occurred during the interval 1948 to 1954 and 1961 to 1974 while decreases are noted during 1954 to 1961 and 1974 to 1982. Major exceptions to this overall trend are Snootli Creek which had continuously decreasing widths after 1954 and Salloomt River where widths continued to increase after 1974. Increasing widths are associated with bank erosion due to increased peak runoff and persistence of flooding as well as channel braiding which is related to increased sediment accumulations within the channel zone. Decreasing widths (channel recovery) occur as a single thread channel becomes dominant and channel margins are stabilized by vegetation.

Channel widening on most tributary fans between 1948 and 1954 appears to be related to the November 1950 flood, therefore, channel recovery following the flood probably began prior to 1954. In four of the eight sites channel widths in 1961 were less than those in 1948 suggesting a continuation of a longer term trend towards channel stability. The response of these largely unconfined alluvial channels to high magnitude autumn floods in 1965, 1968 and 1973 and spring floods in 1964 and 1969 is evident as a trend towards increasing widths until 1974. Reduced or less variable widths occurred in most channels by 1982. Part of the recovery during this latter interval is related to training works (dykes) along the middle and lower reaches of four of the tributaries.

A similar trend can be identified using damage to vegetation along channel margins as an indirect indication of flooding and sediment

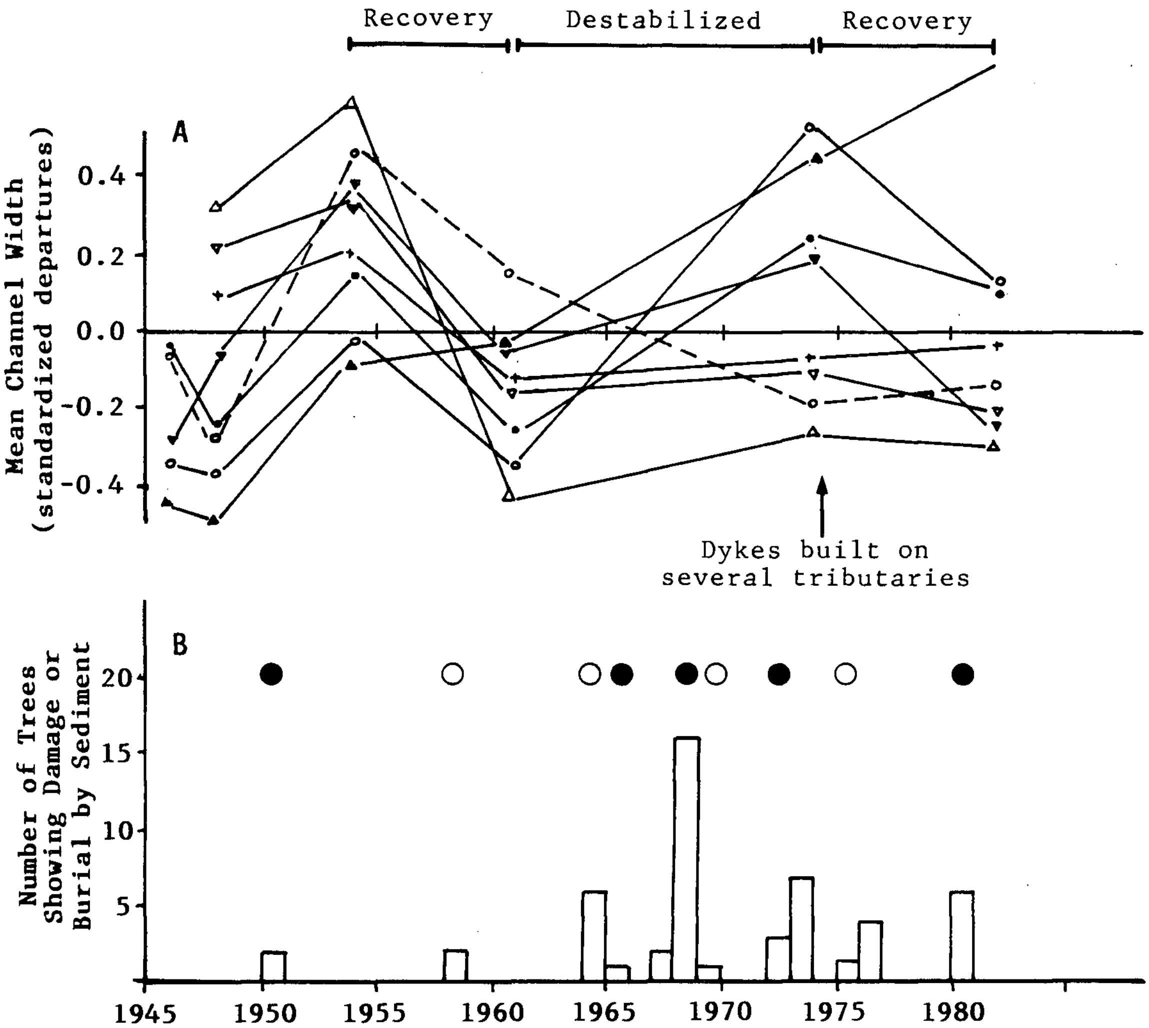


Figure 4.3 Indirect evidence for secular changes in sediment delivery to outlet tributary channels of Bella Coola River. A) Mean width of alluvial channels on several tributary fans. B) Number of damaged or buried trees sampled along the the margins of the same tributary channels shown in A. Dates of spring (○) and autumn (●) floods and major channel training works are also plotted. Damage from the 1968 flood was the most widespread.

transport (Yanosky, 1982,1983). Approximately 200 trees including alder, cottonwood, spruce and fir were examined for the presence of reaction wood indicative of wood stress due to bending or scarring. A sample of 55 trees from all eight sites showing significant damage is plotted in figure 4.3b. The histogram suggests a higher frequency of damage and burial after 1964. There is some bias in the data due to preservation problems of trees prior to the high magnitude floods of 1965 and 1968 but most of the sampled surfaces were occupied by mature vegetation older than 40 years and generally at elevations low enough to have been inundated by most post-1945 peak flows.

In addition to channel morphology and tributary fans, the frequency and magnitude of new landslides throughout the period were estimated from photographs for the lower catchment areas of all eight tributaries. While there was a notable increase in the number of small slides (less than 300 m^3) in hillslope materials between 1961 and 1982, most seem to have occurred in recently logged areas below access roads, particularly in Noosgulch and Cacoohtin Creeks. These small slides may have contributed locally to sediment transfers onto the outlet fans but are not considered as major sediment sources.

Changes in headwater sediment sources were difficult to evaluate because basin-wide airphoto coverage was not available for each interval and access is restricted to only a few sites. A qualitative assessment of the degree of channel braiding downstream from glacial sources (*e.g.* west fork of Nusatsum River, lower Talchako River) and below major composite debris slopes indicates less temporal variability in channel conditions because in many instances transport limited conditions prevailed. Additional contributions of sediment from persistent flooding would be less noticeable because of the already unstable (braided, shallow, wide) nature of these reaches.

The results presented here provide some insight into the connection between runoff variability and sediment yield from major tributaries of the Bella Coola River over the last 40 years. Evidence presented in section 2.6 demonstrates that a significant amount of material is stored in the floodplain and channel zones of major tributaries. Therefore, clastic sediment yield over the short-term is linked to available sediment in higher-order channel segments of these tributaries and influenced less by the frequency and distribution of headwater sediment sources. Aerial photographs support the contention that increased runoff and frequency of flooding resulted in the movement of this sediment and destabilization of alluvial fan channels between the early 1960s and mid-1970s. Periods of channel recovery preceded and followed this interval.

The aerial photographs are temporally discrete samples and therefore these data do not clearly distinguish between the proportional impact of single flood events versus a period of increased flood frequency. It is likely that the threshold for channel recovery (*cf*. Wolman and Gerson, 1978) is more closely tied to the highest magnitude events which introduce substantial volumes of clastic material. In addition, damage and burial of vegetation is the consequence of discrete events only. However, the distribution identified here supports the observed changes in flood frequency over the post-1945 interval. Sampling was restricted to a small number of surfaces along channel margins some of which may have been inundated preferentially (*i.e.* well-defined overflow channels), therefore the sample and interpretations remain imperfect. However, the methodologies appear to be valid for making inferences regarding long-term flow variability in tributary basins.

(4.3) Sediment Sources and Processes of Floodplain Development

Inferences about environmental change from geomorphic evidence, such as channel pattern, hydraulic geometry and flood deposits, require an understanding of the processes involved in the transportation and deposition of alluvial material and primary sources of sediment. The purpose of this section is to document the processes of floodplain development and characterize the alluvial sediment sources for Bella Coola River.

Processes of Floodplain Development

The Bella Coola River exhibits an irregularly sinuous channel, sometimes split about channel islands and in some places braided: seasonal or perennial 'side' channels - subordinate anabranches of the river - are common. This river type has been referred to as a "wandering" gravel-bed system (Neill, 1973; Church, 1983).

The alluvial sediments consist of 0 to 3 meters of sandy channel-fill or overbank silty sands deposited above several meters of channel cobble-gravel. Similar sequences have been described for sinuous gravel channels by Bluck (1971, 1976), Jackson (1978) and Forbes (1983). Lateral accretion is the dominant mode of deposition. This river type is widespread in stable mountain valley and foreland settings and is typical of several rivers draining the west slope of the southern Coast Mountains of British Columbia. Church (1983) has considered recent river changes in some detail.

Details of sediments comprising the alluvial portion of the Bella Coola River have been described elsewhere (Desloges and Church, 1987) so only the most salient points are given here. Storage of sediment in the contemporary channel zone varies along the river. In several reaches, large volumes of resident gravel influence channel pattern producing multiple channels and associated gravel bar development. Intervening reaches have greater lateral stability and occur where the valley width is restricted by bedrock or tributary alluvial fans. Hence, sediment accumulation and resulting facies associations within the active channel zone can be divided into those within unstable and those within stable reaches.

Medial bars and islands dominate the channel within unstable reaches. Medial bar complexes are composed of several mid-channel bars forming clusters upstream from confluences with major tributary channels (figures 4.4a and b). Islands typically are wooded and gravel is exposed only on the upstream side of the island. Channel shifting occurs by avulsion. The islands represent old floodplain surfaces which have been isolated by avulsion. Approximately 40 percent (22 km) of Bella Coola River is dominated by channel macroforms typical of laterally unstable channels.

Bank-attached lateral bars or point-bars dominate the more sinuous stable reaches of the river (figures 4.4c and d). A low-flow back channel or chute, active only at high flow, is almost always present containing a cobble-gravel bed (probably crudely stratified) and structureless sand levees on both flanks. These subordinate channels are important sites for the preservation of sediments indicative of varying flow conditions through time.

Abandonment and infilling of back channels lead to floodplain development. The overall importance of the sandy facies in terms of contemporary floodplain sediment volumes can be determined by assuming that the depth of scour (thalweg depth) in the active channel represents the bottom of the gravel basement or 'bar platform'. This assumption can be made because of the apparent vertical stability of the river (Church, 1983). Thicknesses of the sandy floodplain deposits range from a few

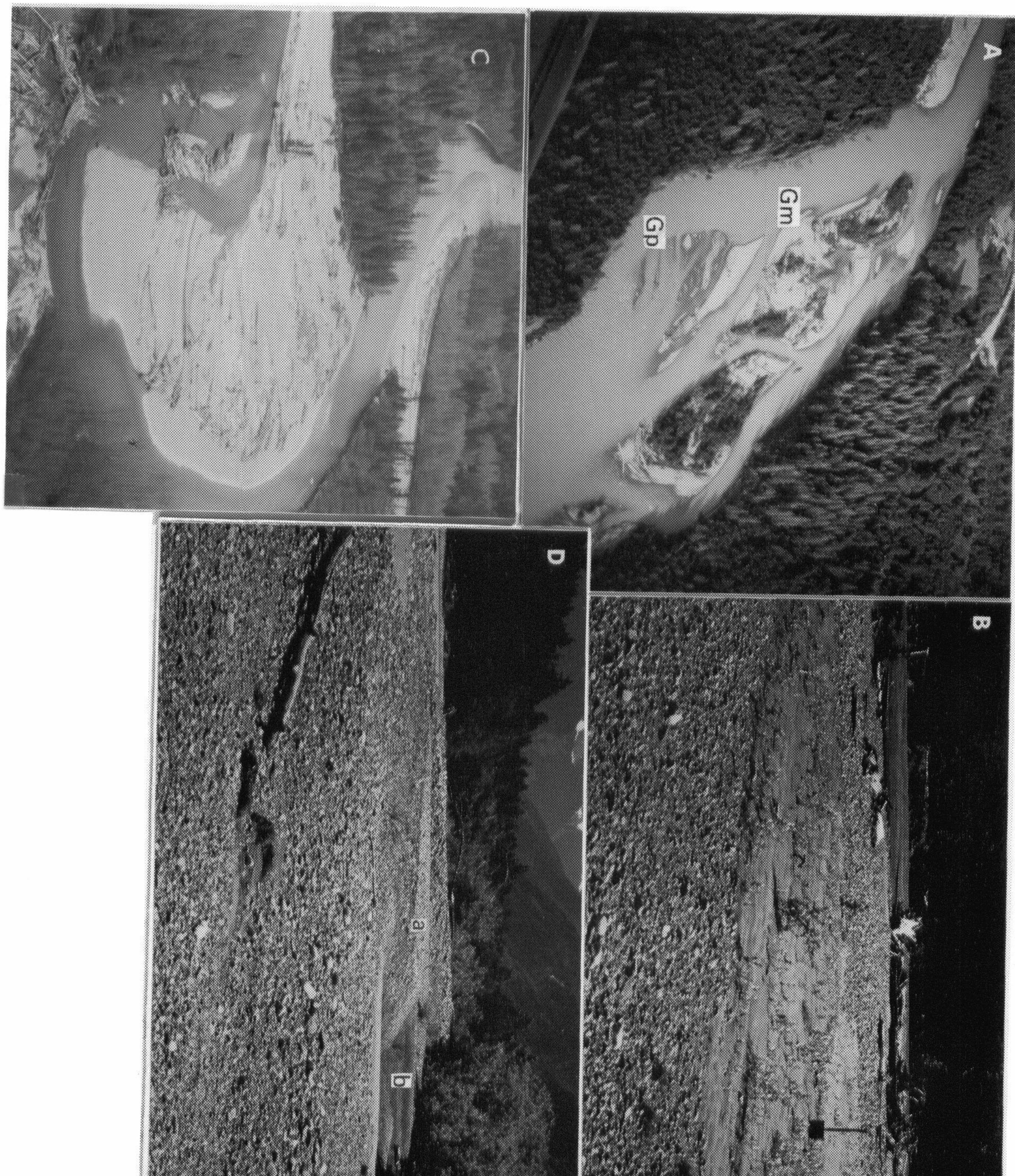
Figure 4.4 Morphology and channel zone sediments of Bella Coola River.

(A) An unstable reach of the river with medial bar deposits. Massive gravel (Gm) or planar stratified sediments (Gp) dominate the channel zone gravels.

(B) Bar topography in unstable reaches is subdued and finer grained sediment wedges cover the gravel surface. Note the two year old growth of alder in finer sediments.

(C) A sinuous reach of the river and accumulation of organic debris on a point bar surface.

(D) Point bar deposits some times exhibit gravel deltas (a) which migrate into finer back channel deposits (b).


A
Gm
Gp
B
C
D
a
b

centimeters to over 3 m of coarse gravelly sand to fine sandy silt and represent 31 to 36% of the total 'active' sediment pile (see table 4.4). Nowhere does clay constitute more than 15% of any sample: most frequently it is much less.

The sedimentary sequences in the alluvium reflect the integrated processes of floodplain development over time. In unstable reaches of the river, rapid channel avulsions, the occurrence of multiple channels with varying capacities for sediment and water discharge, and more variable flow conditions result in a diversified set of facies units within the channel and on the the vegetated floodplain surface. With increasing stability, the spatial variation in flow conditions declines and the probability of transitions between facies units becomes lower than in less stable reaches of the river. Progressive lateral accretion and the vertical stability of the river appear to favor the simpler facies sequence, although infilled back channels may still exhibit the wider variety of sedimentary structures. The greater incidence of overbank flooding in stable reaches produces vertical accretion of the floodplain here and development of the finer-grained facies (Desloges and Church, 1987). However, lateral accretion deposits dominate the overall sequence in just the manner described by Wolman and Leopold (1957).

Channel and point bar development is accompanied by the cross-valley migration of the channel, resulting in the formation of a bar platform and, in addition to the main channel, several active slough channels or a single, poorly connected back channel.

Sediment volumes (see table 4.4) indicate that lateral accretion of gravels accounts for approximately 65% of sedimentation during floodplain construction. As the vegetation develops, sands are trapped on bar tops and overbank, so the floodplain rises by vertical accretion to its final level.

Table 4.4. Volume estimates of alluvial-fill for the Bella Coola River.

Reach Type[1]	Valley Length (km)	Floodplain Area (km^2)	Overbank (sand)(2) Volumes X $10^7 m^3$	Basement Gravel(3) Volumes X $10^7 m^3$	Proportion of fill as overbank sands
Unstable	22.1 (39%)	32.6 (42%)	4.2 (60%)	7.6 (59%)	36%
Transitional	23.1 (40%)	30.6 (39%)	2.4 (34%)	4.3 (33%)	36%
Stable	12.2 (21%)	14.4 (19%)	0.4 (6%)	0.9 (8%)	31%
Total	57.4 (100%)	77.6 (100%)	7.0 (100%)	12.8 (100%)	35%

1. Reach type classification is based on the frequency of channels and characteristics of floodplain and channel sediments. Unstable reaches are high-gradient sediment storage zones with numerous sloughs, back channels and medial bars. Stable reaches are characterized by more sinuous single-thread channels. Transitional reaches exhibit a combination of morphological features.
2. Volume of fine-grained (sand and silt) vertical and lateral accretion deposits.
3. Volume of coarse-grained (gravels and cobbles) lateral accretion sediments above maximum thalweg depth.

Sands are very mobile in this high-energy, cobble-gravel river, moving predominantly in suspension. Once re-entrained, the sand may be flushed from the system or redeposited a substantial distance downstream. Much of the sand probably moves through the system in only one or two stages.

The actual modes of channel migration in unstable reaches are progressive erosion on the cutbank opposite a growing channel bar, and rapid channel avulsion, or reduction of flow in the channel around one side of a medial bar, leaving behind a series of active, "subordinate" sloughs. Sloughs along the inner margins of point bars may remain active, sediment transporting channels for periods in excess of 50 to 80 years following subordination. Complete infilling is likely to occur only if the slough entrance becomes blocked by large organic debris. Sedimentation would be episodic and indicative of flood events only.

In sinuous reaches of lower gradient, sloughs become poorly connected at most flow stages and a back channel develops with deposition of fine sands, silts and organic material. Infilling may never be complete as some of the older sloughs and back channels are preferentially occupied during high magnitude flows. As a result, channel avulsions frequently lead to the occupation of older channels rather than the development of new ones. This leads to a reduction in the potential number of sites with sequences representative of long-term sedimentation. Although avulsion produces abrupt realignment of the channel, progressive lateral accretion and erosion may be substantially more significant in the history of floodplain development and sediment exchange.

Alluvial Sediment Sources

The principal sources for alluvial sediments deposited within the Bella Coola valley are as follows:

(a) headwater tributaries introducing sediment to the active channel;

(b) downstream tributaries entering along the river reach contributing sediment directly to the floodplain surface or to the active channel;

(c) bank erosion and re-deposition of previously deposited alluvial material; and

(d) fluvial erosion of non-alluvial deposits resident within the zone of channel activity.

Sources (a) and (b) are directly connected with the uplands. Sources (c) and (d) represent remobilization of sediment previously introduced into the valley. Since less than 13% of the channel of Bella Coola River is in contact with non-alluvial deposits, the majority of which are either well-vegetated, stable colluvial slopes, bedrock or deposits of glaciomarine silt and clay, they are not important sources for the sand and gravel that dominate the alluvial valley-fill.

Klages and Hsieh (1975), Mazzullo (1986) and Statteger *et al.* (1987) have demonstrated that the mineralogy of alluvial and littoral sand deposits may reflect provenance provided that source areas yield sediments dominated by distinctive lithologies, characteristic proportions of certain minerals, or contain unique minerals not found in other source areas; and that these properties are not appreciably attenuated by mixing during transport to a particular depositional site.

To determine which source areas are most important for delivery of material to the Bella Coola River, the composition of sediments from several major tributaries was compared with channel zone and floodplain sediments sampled within the Bella Coola valley. Samples extracted from the active channel zone and floodplain were classified on the basis of discriminant functions derived for samples taken at the outlet of major tributaries. The sampling design and data analysis are discussed in Appendix B and the results are plotted in figure 4.5.

These results indicate that fine sediment (sands) along the active channel and in the floodplain derives mainly from the glacierized volcanic terrain of upper Talchako River. Downstream, there is in the channel a detectable admixture of sands from lateral tributary basins with plutonic sources. However, sands dominated by lateral sources are restricted to the vicinity of the tributary junction. Mixed sediments originate from tributaries which produce material that truly reflects a mixed bedrock source (*e.g.* Nusatsum River) and from mixing of plutonic and volcanic sediments on the floodplain.

Larger clasts appear to reflect more uniformly the overall distribution of source terrain in the basin. About this several things need be said: certain plutonic lithologies - though by no means all of those in the basin - resist attrition and may become over represented in lag deposits; plutonic clasts may be derived disproportionately from eroded Quaternary deposits; finally, size-related effects may influence the results (size was not controlled in the surface clast sample).

Sediments are delivered from volcanic headwater sources to the downstream channel and floodplain during basin-wide high flows, and during summer glacial runoff. Sediments originating from plutonic terrain are delivered to the margins of the floodplain during basin-wide floods, and also during storms in which the contribution of runoff from the lower basin is much greater than from the headwater areas. This is not an unusual occurrence because of the eastward decline of freezing levels during late autumn storms, and because of the marked gradient in precipitation across the basin. This pattern of sediment delivery influences channel morphology and the character of alluvial deposits of Bella Coola River.

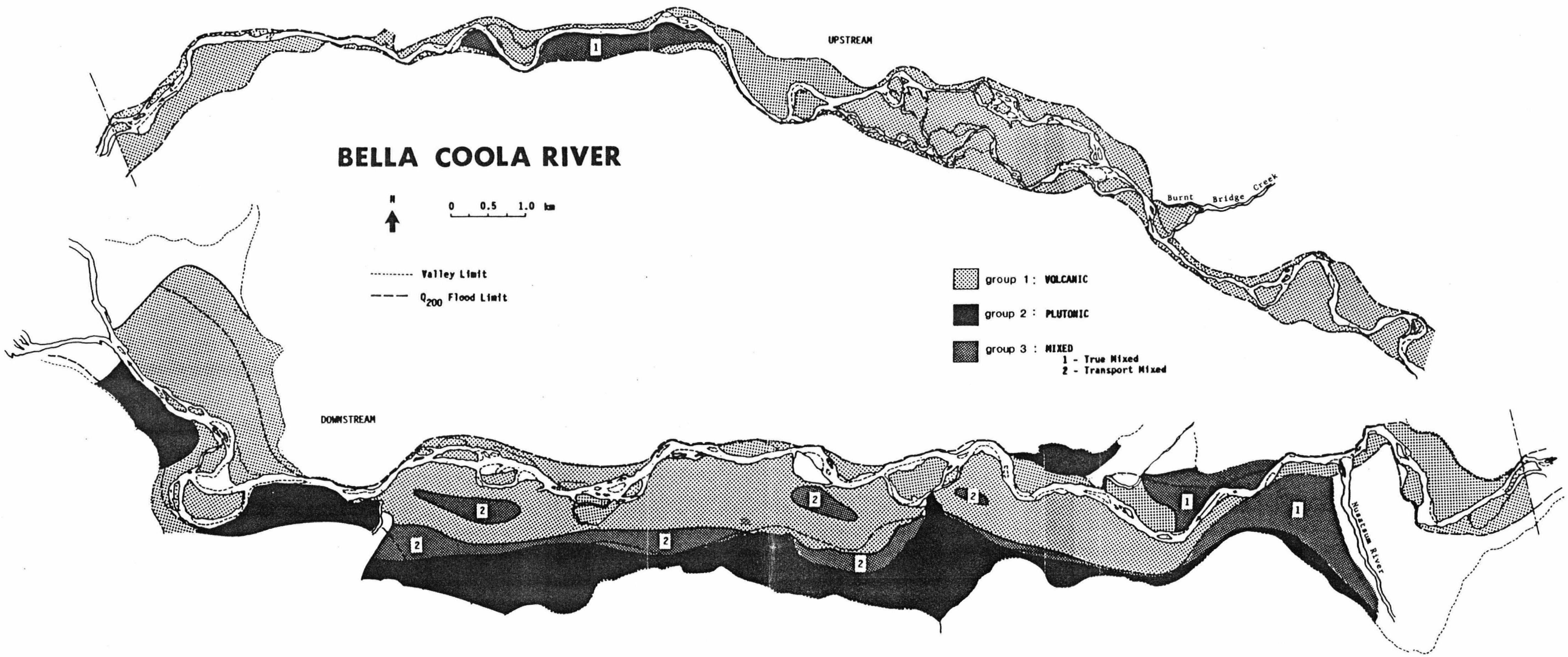


Figure 4.5 Alluvial sediment sources for the Bella Coola River floodplain. Grouping corresponds to proportions of minerals in the sediments which reflect bedrock source conditions. Mixed sediments can be from truly mixed sources (1) (i.e. basins underlain by volcanic and plutonic sources) or may represent mixing once introduced into the higher order alluvial system (2). See the text and Appendix B for discussion.

(4.4) Equilibrium Conditions and Inferences from Channel Morphology

Morphological adjustments of a stream channel depend on the magnitude and timing of water and sediment discharge to the river and the relaxation time of the system (Mackin, 1948; Brunsden and Thornes, 1979). Alluvial channel morphology therefore can be used to infer fluctuations in the discharge of sediment and water provided that the equilibrium status of the river can be evaluated. Howard (1982) has defined equilibrium as a

> "single-valued, temporally invariant functional relationship between the values of an output variable and the values of the input variable(s) in a geomorphic system."

Disequilibrium occurs when the output variable no longer resembles that predicted by the functional relationship previously observed. If the interval of time required to achieve a new equilibrium is longer than the frequency of significant perturbations then a complex sequence of morphological adjustment may occur and the functional relationship is no longer easily definable. Such a system is intransitive and the possibility of continuous disequilibrium makes inferences about the independent variables from morphological evidence unreliable.

However, as Schumm and Lichty (1963) point out, equilibrium is scale dependent. Hickin (1983) and Howard (1982) give several examples of this scale dependency but three divisions seem appropriate:

1) for very short time scales (10^{-2} to 10^{0} years) and individual reaches or cross-sections of the river, channel gradient becomes a fixed value and sediment discharge is a dependent variable;

2) over longer time scales (10^{0} to 10^{2} years) and several reaches of the river, slope adjustments become dependent on the sediment and discharge regime; and

3) for geologic time scales (10^3 to 10^5 years) and entire fluvial networks, both slope and sediment discharge become dependent on various geological processes such as isostatic or tectonic adjustments.

The first and third divisions are mostly for engineering and geological time scales, respectively, while the second is best suited for the geomorphic (and engineering) time scales of interest here.

The generalized functional relation between variables which influence river morphology have been identified by Lane (1955) and re-interpreted by Schumm (1977) as:

$$Q_s \propto b, \lambda, s/d, P \qquad \text{and} \qquad Q_w \propto \lambda, b, d/s$$

Channel width (b) and meander wavelength (λ) are directly proportional to discharges of water (Q_w) and sediment (Q_s). Channel gradient (s) is inversely related to water discharge and directly proportional to sediment discharge. The opposite association is true for channel depth (d). Sinuosity (P) decreases as sediment discharge increases.

Several variables have been selected to evaluate the equilibrium status of Bella Coola River and the impact of changing water and sediment discharge. These include indicators of vertical and lateral channel stability, channel width, width/depth ratio, meander wavelength, sinuosity and braid index.

Vertical and Lateral Stability of Bella Coola River

River bed and water surface profiles for a moderate flow in the Bella Coola River between the Water Survey of Canada gauge at kilometer 56 and river mouth are plotted in figure 4.6. Below the Nusatsum River confluence a concave-up profile dominates with average bed-gradients of 0.0038 above Nooklikonnik Creek, declining to an average of 0.0018 near the mouth of the river. Locally steep gradients can be found in proximity to several tributary confluences yielding a stepped profile. From Nusatsum River up to Noosgulch Creek gradients are comparatively steep (0.0032) and then become much lower until the Burnt Bridge Creek confluence (0.0016). The overall upstream profile is convex-up.

Changing bed-gradients in the vicinity of tributary alluvial fans suggests that constriction of the river by these features is an important factor in determining the river profile. Sediments accumulate in storage or "sedimentation zones" (*cf.* Church, 1983) above the alluvial fan constrictions and gradients adjust to accommodate the increased volume and calibre of sediment stored there (see Appendix B for evidence of grain-size variations). This promotes increased lateral instability in many of these reaches.

Vertical stability of the river has been inferred for the interval 1948-1983 using hydrometric surveys from the two upstream gauging sites. Vertical changes in bed-elevation at each site have not exceeded 1 m between any particular flood of magnitude Q_5 or larger, and absolute elevation changes over the entire interval are less than 0.5 m. In other reaches, vegetation and stable, cobble-paved sediments also indicate long-term vertical stability. Thus, the overall vertical stability and lack of evidence for significant aggradation or degradation suggest an equilibrium between water and sediment supply on the river scale for at least the last 35 years and perhaps much longer. However, localized adjustments to increased sediment storage are made by lateral channel shifting and an increase in the number of subordinate anabranches.

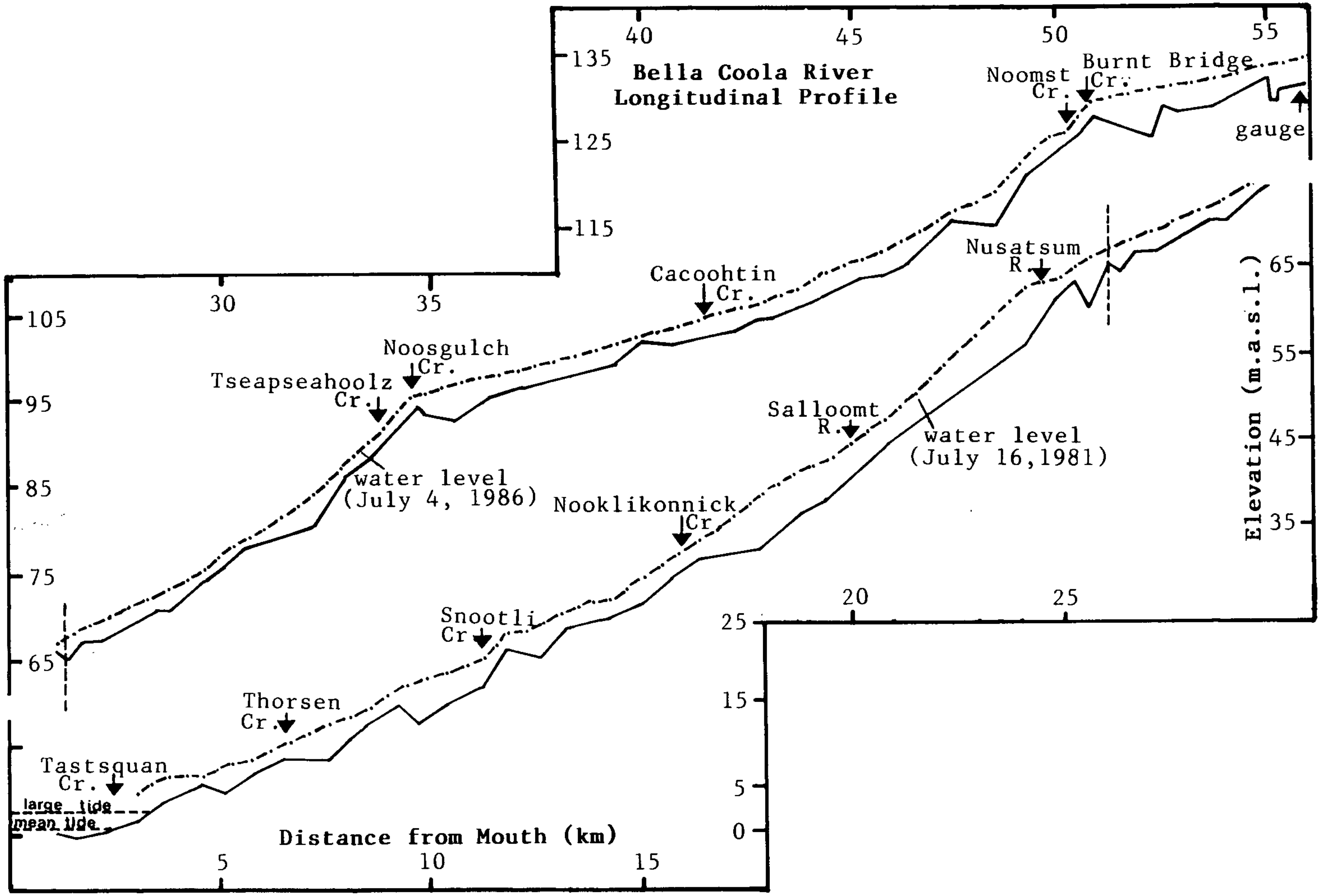


Figure 4.6 Water surface and bed longitudinal profiles of Bella Coola River between Water Survey of Canada gauge above Burnt Bridge Creek and river mouth. Arrows indicate the position of tributary creeks. Water profile survey for the lower river was done on July 31, 1981 and for the upper river on July 4, 1986 Source: Water Management Branch, Ministry of the Environment, British Columbia (unpublished data).

Analysis of maps, aerial photographs and field measurements indicates that lateral channel instability coincides with zones of increased sedimentation. Unstable reaches consist of several channels separated by mid-channel bars and the surrounding floodplain is dissected by back-channels and sloughs. Stable reaches exhibit a single, well-defined channel course.

Channel morphology data were compiled for a number of reaches and grouped into laterally stable or unstable categories. Group assignments were made on the basis of observed channel pattern, floodplain sediment facies and vegetation characteristics. Thalweg slope, river gradient, channel width, width/depth ratio and hydraulic radius were measured at a number of sections in each reach and then averaged across each group. Results presented in table 4.5 demonstrate that the cross-sectional data are significantly different between stable and unstable reaches, which would confirm that the processes of channel development (*e.g.* bank erosion, bar development, channel migration) vary significantly in different reaches of the river.

Lateral channel stability for several reaches was evaluated by considering the correlation of reach-averaged slope and channel forming discharges. The results of this analysis (not shown here) and the above discussion revealed several things: 1) that the river must be viewed as a sequence of hydraulically distinct reaches which have very different morphological and sedimentological characteristics; 2) no one reach is likely to be indicative of river or basin scale fluctuations in sediment or water discharge; and 3) significant changes in runoff and sediment flux are likely to influence first those reaches near the threshold of change (i.e. transition from stable to unstable pattern) and leave others less affected.

Table 4.5. Morphological data for selected reaches of Bella Coola River

Reach[1]	N	Width (m)	Width/Depth Ratio	Hydraulic Radius(m)	Thalweg Gradient	Valley Gradient
Stable	7	104 ± 13	41 ± 14	2.12 ± 0.39	0.0019 ± 0.0008	0.0026 ± 0.0010
Unstable	6	171 ± 39	109 ± 59	1.38 ± 0.41	0.0033 ± 0.0010	0.0034 ± 0.0012
one-tailed t-test (d.f. = 11, α = 0.10)		t = 4.60 significant	t = 3.17 significant	t = 3.89 significant	t = 3.22 significant	t =1.31 not significant

1. Reaches vary in length and were sampled between the Antnarko/Talchako confluence and river mouth

Inferences from Channel Morphology

The above results suggest that variations in morphological features of a river are potentially useful indicators of changing sediment and water discharge. Width, slope and channel pattern are functionally related to the discharge regime and adjustments therein may demonstrate the magnitude of change. Three indices were selected to assess the degree of association between secular changes in Bella Coola River morphology and changes in runoff and/or sediment supply: regime equations applied to width fluctuations in stable/unstable reaches, channel sinuosity and channel braid index. Sinuosity is the ratio of channel length to valley length while braid index is the ratio of total channel length, including all back channels and sloughs, to main stem length. The propensity for a river to occupy single, well-defined meandering channels when runoff magnitude, runoff variability and sediment discharge are reduced is reflected in increasing sinuosities and decreasing braid index. Secular variations in sinuosity and braid index, measured from several sets of air photographs, are plotted in figure 4.7a.

Sinuosities are highly variable (range 1.05 to 1.27) for the lower Bella Coola River. No strong trend is evident but there is a tendency for decreasing channel curvature in both upstream and downstream river segments over the last 10 to 15 years in contrast to steady or slightly increasing sinuosities prior to 1968. The braid index is similar for both segments showing decreases up until 1968 followed by increases to 1982, particularly for the upstream reach. Although long-term trends are evident, significant channel changes occurred following major floods in 1950, 1968, 1973 and 1980 suggesting the importance of flood frequency and magnitude in determining channel pattern.

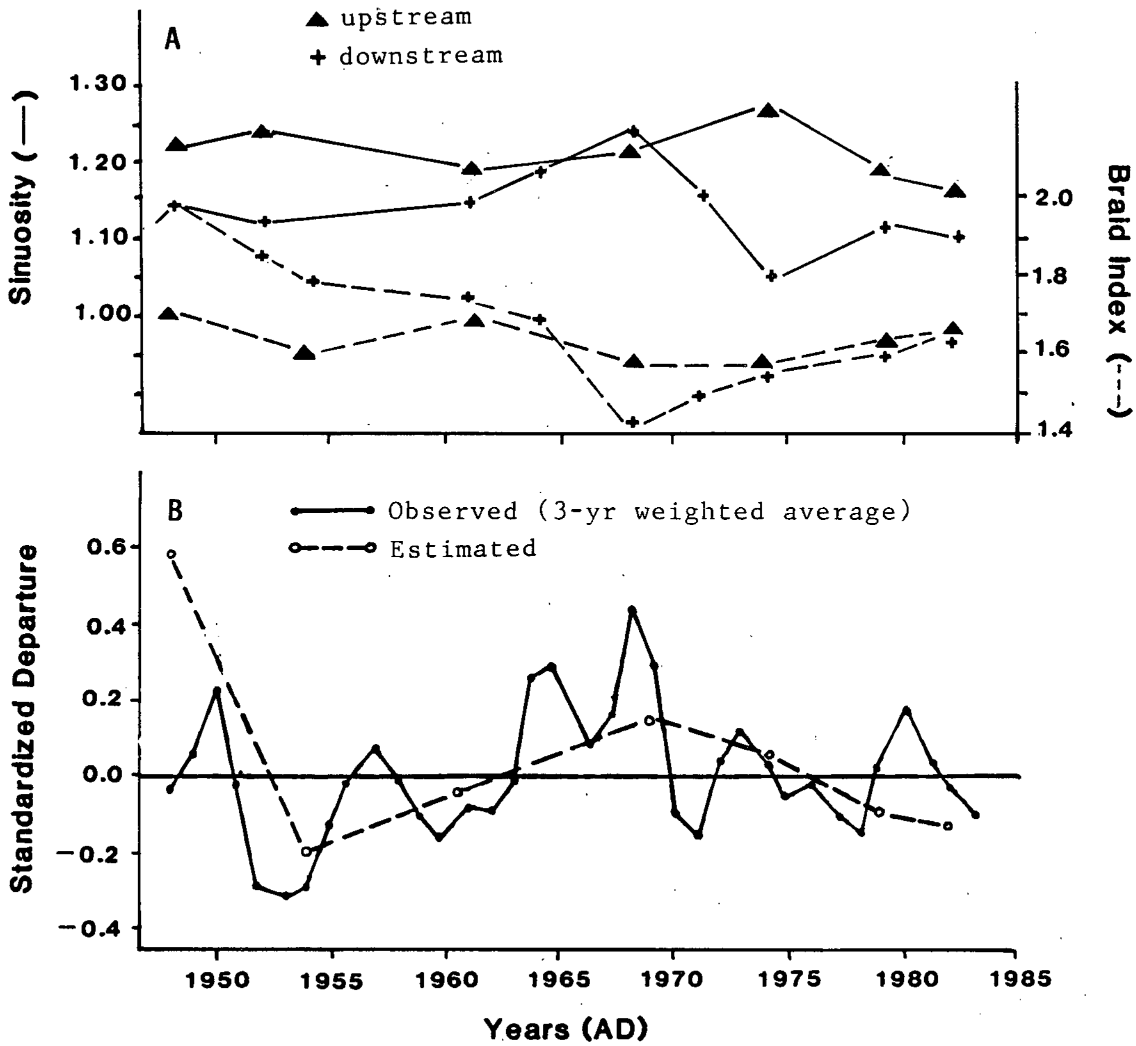


Figure 4.7 A) Secular changes in sinuosity and braid index for Bella Coola River above and below the Nusatsum River confluence. Decline in the braid index (dashed line) suggests increasing lateral stability. Sinuosity (solid line) shows less overall trend. B) Actual and estimated secular variations in annual maximum discharge at the Bella Coola gauge above Burnt Bridge Creek. Solid line is a three year moving average and dashed line is estimated flow using regime equations (Bray, 1982).

Channel modifications within individual reaches can be important; for example a loop extension cutoff in the autumn flood of 1973 accounts for much of the decline in sinuosity of the lower Bella Coola after 1973. However, it is usually adjustments within several reaches that control the variance. River training works installed since 1968, mostly in the form of dykes, have probably influenced the less dramatic changes than might have been expected following the 1973 and 1980 floods. However, their limited distribution does not appear to have affected the overall trend of increasing channel stability to 1968, followed by reduced lateral stability after several floods in the late 1960s and early 1970s as sediment from upstream and tributary fans was deposited in the main stem of the river.

Measurements of channel width in the three stable/unstable reaches were taken from several sets of aerial photographs available for selected years between 1948 and 1982. The highest probability of adjustments to changing discharge is in these threshold reaches. Bray (1982) has developed a set of regime equations from a sample of a Alberta gravel-bed river which relate hydraulic features such as bankfull width to some dominant discharge (in this case Q_2). In order to test the strength of inferences drawn from changing channel widths, annual peak, mean-daily discharges measured at the upstream gauge were compared with estimated departures in discharge using the Bray regime equations. Since the original equations were computed with water discharge (Q_w) as the independent variable, the relations were reformulated with width (w) as an independent variable, an exercise which more appropriately estimates the error associated with the dependent variable (*cf.* Williams, 1984). The resulting equation takes the form:

$$\log Q_w = -1.51 + 1.797\log w \qquad r^2 = 0.97$$

$$\text{s.e.} = 0.19 \text{ log-log}$$

The results plotted in figure 4.7b indicate some similarity between measured (3 year weighted mean departures) and estimated flows using channel widths in these reaches. Widths appear to have been increasing during the interval of higher-magnitude flows between 1964 and 1976. Decreasing maximum mean-daily discharges (with the exception of 1980) have resulted in decreased channel widths since the early or middle 1970s as vegetation begins to colonize lateral bar surfaces. It should be noted that the relative proportion of width changes due to flood events and the less significant inter-event bank erosion cannot be distinguished with these data. However, air photographs taken only four months after a late January flood in 1968 indicate that flood-related channel changes are the most significant.

The legal survey of 1889-93 fixed channel boundaries fairly precisely although features within the channel were estimated only. Estimates of channel width in the vicinity of the gauges taken from 1:31,680 scale maps are between 20 and 30% greater than recent measurements. The early 20th century trend then, is towards declining channel widths and discharge.

(4.5) Recent Floodplain Sedimentation

The distribution and composition of sedimentary facies in the Bella Coola River floodplain indicate that rates of sedimentation during and between flood events are highly variable (see Desloges and Church, 1987). This variance is related to the frequency of channels in a particular reach of the river and how well connected a depositional site is to primary sources of alluvial sediment. Inferences regarding temporal variability in rates of sedimentation from alluvial sedimentary sequences require that these conditions be known with some confidence.

Rarely is it possible to calibrate sedimentation processes against recent alluvial activity because of low contemporary rates of sedimentation and the high variability in boundary conditions. Stratigraphical interpretations from a number of sites usually provide the only means of assessing temporal variability (*cf.* Brakenridge, 1985). Common difficulties in achieving accurate inferences about the changing alluvial environment have been enumerated by Butzer (1980) who demonstrates that chronological control and between-site correlation are most problematic.

An attempt was made to determine alluvial sedimentation rates in the Bella Coola River in order to assess the linkage between recent runoff variability and sedimentation. Sediments in several active or recently active sloughs, back channels and ponds were sampled for this purpose. From an initial set of 16 sampling sites only six yielded sediments with sufficient textural, organic and structural variability to warrant consideration. Pollen, organic matter content and textural characteristics were used as primary indicators of sedimentation variability. Sampling locations, field methods and laboratory techniques are given in Appendix C. Interpretations from one particularly interesting site are given here along with a summary of results from all sites investigated.

Four cores were recovered from standing water located in a former back channel area of the Bella Coola River approximately 500 m downstream from the contemporary stream gauging site above Burnt Bridge Creek (figure 4.8). Air photographs taken in 1946 indicate that the back channel was well connected to the Bella Coola River prior to this time. Over the last several decades the backwater has become an isolated series of ponds in which water and sediment are derived from two sources. During non-flood conditions water enters via a stream situated in a back eddy area on the concave bend of the meander loop. Much of the channel is presently

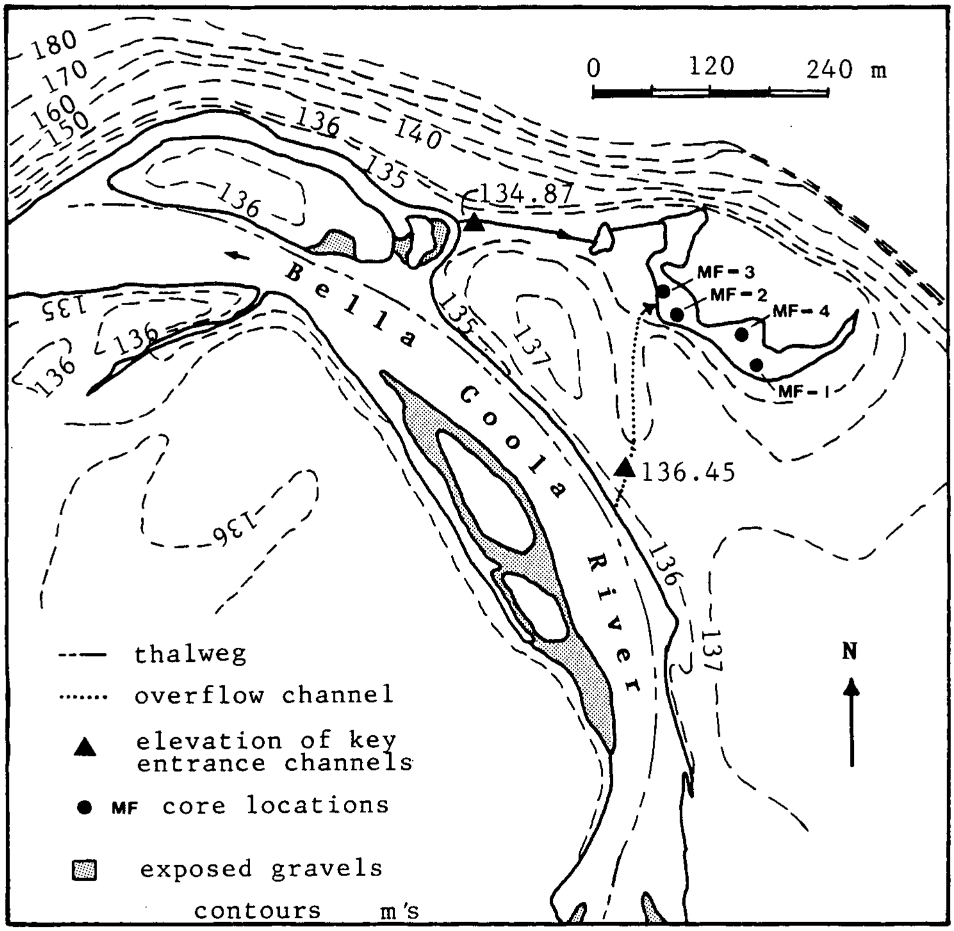


Figure 4.8 Topographic setting for the McCall Flats backwater ponds just below the Water Survey of Canada Gauge near Burnt Bridge Creek. Note the location of the overflow depression between the river and ponds. Vegetation and the legal survey of 1889-93 both indicate that the general curvature of the river in this reach has been stable for at least the last 125 years. Backwater areas and the size of the concave bench opposite the point bar have changed significantly.

overgrown with aquatic plants and low-flow sediment delivery is minimal (suspended sediment concentrations in the ponds were as low as 0.2 mg/l). During flood, water and sediment are delivered through the same stream entrance as well as an overflow channel which becomes active only during high water stages when sediments are deposited directly into the ponds (figure 4.8). Concentrations of suspended particulate matter sediment measured during a moderate flood in the fall of 1984 were as high as 5 mg/l, although the overflow channel was not active.

Core stratigraphy shown in figure 4.9a indicates the presence of several coarse layers (50 to 75% sand) which apparently relate to flood events over the last several decades. To test this hypothesis fine-grained layers in core MF-2 were sampled and analyzed for pollen type and concentration (see Appendix C for analytical procedures). Airphotos indicated that much of the surrounding forest vegetation (*Pseudotsuga menziesii*, *Thuja plicata*) was removed by commercial logging operations during the summers of 1969 to 1970 leading to the potential for establishing a chronostratigraphic marker using pollen. Pollen percentages, sediment texture, organic matter content and stratigraphy in core MF-2 are shown in figure 4.9b.

Three distinct zones can be identified on the basis of changes in the percentage of pollen (4.9b). In zone I sediments, Lodgepole pine and the firs are dominant accompanied by low percentages of alder and grasses. With the exception of Douglas fir these species were not well represented in the surrounding vegetation community prior to logging suggesting that the pollen was derived from riverine sources. Lodgepole pine is a co-dominant species in middle elevation areas of the eastern catchment and its pollen is known for long distance airborne transport (Mannion, 1980) but based on analyses of pollen types and concentrations in Bella Coola River water

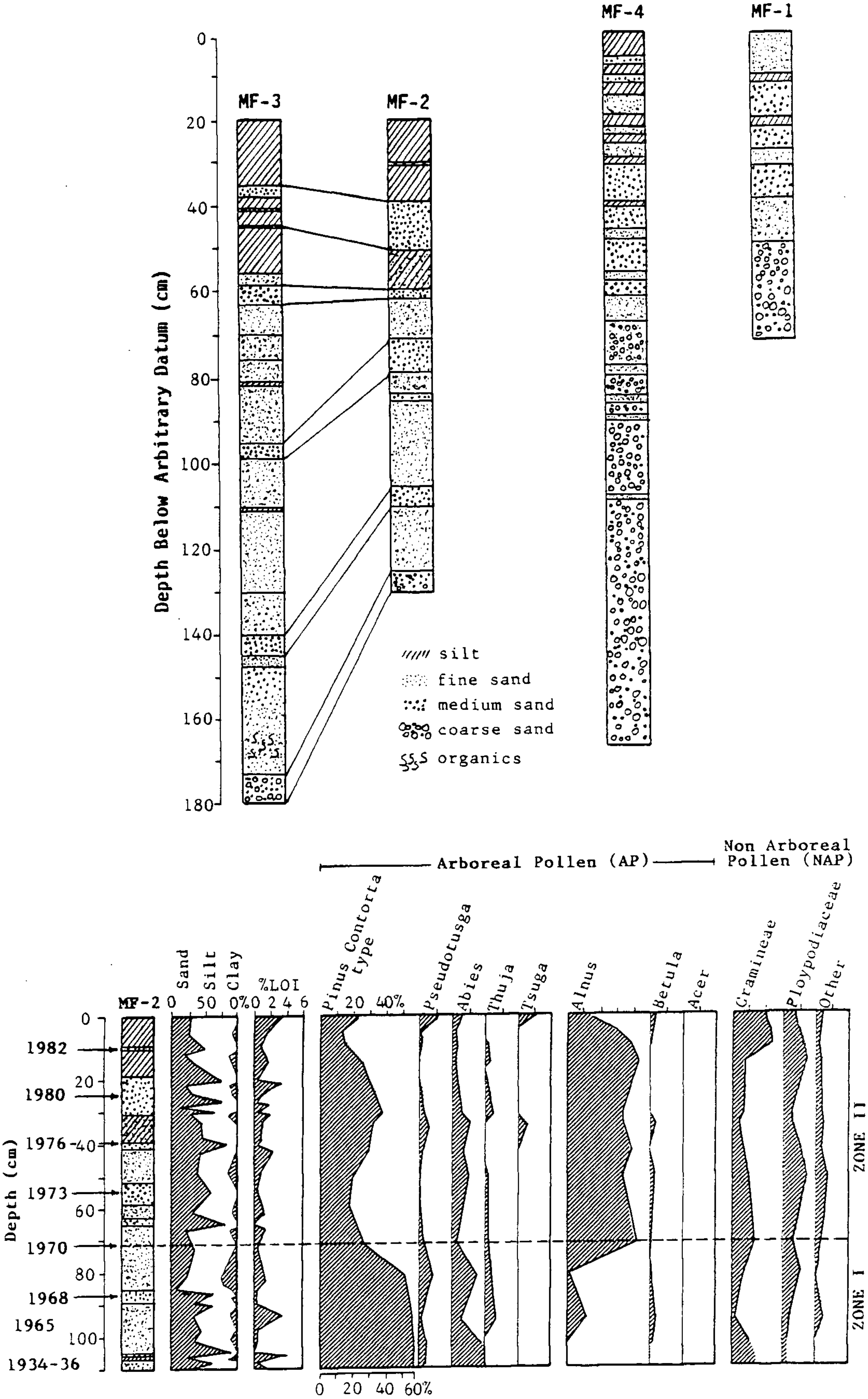


Figure 4.9 (A) Stratigraphy of McCall Flats cores. See figure 4.8 for sampling locations. (B) Pollen concentrations and inferred chronology for flood deposits in core MF-2. See text for discussion.

(normal summer discharge) and pond water (during the October 1984 high water event) significant waterborne contributions are also likely (see Appendix C).

Above 70 cm in zone II, a very sharp transition to alder-dominated sediments is evident along with relative decreases in Douglas fir and cedar in the arboreal component. These quite dramatic shifts are related to the influx of alder following logging in 1969 and 1970. Alder pollen is less prone to long distance transport thus the percentages shown in figure 4.10b probably represent the importance of alder in the local vegetation community. Alder percentages remain very high until the 15 cm level above which an increase in some of the conifers and grasses occurs (zone III). Absolute concentrations of pollen are highest in zone II which further reflects the importance of alder, a species which is a prolific producer of pollen grains.

Based on water level data and cross-sectional morphology of the Bella Coola River adjacent to the overflow channel, flows in excess of Q_5 (500 m^3s^{-1}) would have sufficient depth to occupy the overflow channel. Although the configuration and height of channel divide will have been modified after each flood, the age and structure of vegetation adjacent to the overflow channel suggest that the modifications have not been significant. The airphoto evidence indicates that the back water area was not decoupled from primary river flows until sometime before 1946, so it is assumed that the basal gravels represent deposition from the 1934/36 sequence of floods. Between these basal gravels and the stratigraphic marker, which is assumed to represent 1970, two coarse "flood" units can be identified (figure 4.9a and b). The lowest unit may represent the 1950 or 1965 flood, although the latter is more likely since all aerial photographs taken prior to 1961 show that the ponds were very shallow and occupied by gravels in some places.

The second coarse unit is assigned to the 1968 flood, the largest measured flow over the last 40 years. Since the change from pine to alder occurs at the top of the fine-grained material which caps this flood unit, the whole sequence from 65-90 cm is probably related to the 1968 event (basal flood sands and post-flood siltation). Above the chronostratigraphic marker three prominent units can be identified and correspond with known high flows in the autumns of 1973 and 1980 and the spring of 1976.

A fourth thinner unit, composed of finer material than the other three, might be associated with high water in 1982 but it is more likely material deposited during the waning stages of the 1980 flood. The flood units assigned in core MF-2 correlate with layers in core MF-3 but are more difficult to cross-correlate with sediments taken in an adjacent pond which is not directly connected with the overflow channel (figure 4.9a).

Evidence from the lower Bella Coola River is not as conclusive as the interpretations made from this site but two locations in particular (East Hagensborg Slough and Snootli Slough; see Appendix C) give clear evidence of above average sedimentation in the 1960s and early 1970s followed by lower depositional rates. Dykes constructed after the 1965/68 series of floods in parts of the lower Bella Coola River has reduced the sedimentation effects of floods during the 1970s and in 1980 hence, most of the infilling sediment is related to flooding during the 1960s (e.g. Rambeau Slough). Sediment accumulations in the Mclellan Rd. slough exceeds 120 cm and is probably related to a combination of lateral and longitudinal infilling. At least the top 30 cm of sediment (25%) is the result of post 1965 flooding. A similar trend is inferred for the Walker Island slough (see Appendix C).

The evidence presented here suggests that, while most backwater or sloughs selected for this investigation probably became effective sediment traps sometime after the two high magnitude floods in 1934 and 1936, significant accumulations of sediment did not occur until 1968 (average between 1968 and 1976 is 6.25 cm/yr). Sedimentation is highly episodic and mostly related to flood events greater than Q_5.

(4.6) Conclusions

Glaciological evidence provides a low-resolution record of climatic variability over the last several decades. The turning of glaciers in the Bella Coola basin was a response to persistently positive precipitation anomalies in the middle 1960s and early 1970s. Glacier size, basin aspect and other morphological parameters complicate actual responses and are independent of hydrological variations. Individual years characterized by extreme departures in glacier mass balance correspond to significant changes in the seasonal frequency of "wet" synoptic regimes.

Variability of channel morphology on alluvial fans reveals a secular pattern similar to that for the basin glaciers. Above average precipitation and increased runoff between 1964 and 1974 lead to increasing channel width and promoted channel instability. However, closer examination shows that individual flood events are responsible for most of the channel changes. Floodplain sedimentation is clearly episodic and related to extreme events only. The 1957-1976 period was clearly exceptional in terms of sedimentation in back channels and sloughs.

CHAPTER V
Sedimentological Evidence of Environmental Change: Tests Using Lake Sedimentation Rates

(5.0) Introduction

Long-term lake sedimentation rates can provide an indirect measure of environmental change. While the functional relationship between source area sediment yield and rates of deposition in lake settings may be complex, there is an increasing understanding of linkages and controlling variables in both glacio-lacustrine (Church and Gilbert, 1975) and non-glacial lake systems (Oldfield, 1981; O'Sullivan, 1983). The presence of proglacial, ice-marginal and ice-dammed lakes in the Bella Coola basin provided the opportunity to extend the record of recent environmental change.

There are three important categories of variables which control the erosion, transportation and deposition of material in a glacio-lacustrine system: erodibility of terrestrial materials (*i.e.* sediment sources), hydroglaciological factors and limnological factors. Sediment flux rates depend on the character of sediment sources and the timing, frequency and magnitude of sediment transport processes (Karlen, 1981; Grunell and Clarke, 1987). The interannual and seasonal variability of temperature and precipitation in mid-to-high latitude alpine regions provide for discontinuous fluxes of water and sediment of varying magnitudes from the surrounding terrain. Ablation season temperatures, winter and summer precipitation, spring snowpack volumes and near surface winds are important hydroclimatic controls of the rate of sediment yield from the glaciated terrain to the lake (Granar, 1956; Rainwater and Guy, 1961; Shaw *et al.*, 1982; Simola and Tolonen, 1981; Ringberg, 1984; Leonard, 1985).

Certain morphological, chemical, biological and dynamic components of the lake will control the distribution and deposition of material. Flat, extensive lake bottoms, free from internal disturbances such as slumps or other sub-aqueous slope failures, are most likely to preserve annual layers or varved sediments (Hakanson, 1983). Thermal stratification, wind-generated lake currents and seasonal biological activity interact to provide a range of depositional conditions each yielding different structural characteristics (Sturm, 1979; O´Sullivan, 1983). Finally, disturbance by physical or biological processes and post-depositional diagenesis of both clastic and organic materials can alter pre-existing structures. Despite all these physical and chemical interactions annually laminated lake sediments indicative of prevailing hydrological and glaciological conditions are found frequently in glacio-lacustrine environments and provide geochronological evidence for environmental change on time scales ranging from days to millennia (Renberg, 1982; Morner, 1984).

Reconnaissance of several lakes in the Bella Coola basin revealed the potential for the formation and preservation of annually-laminated sediments. Due to limitations of time and accessibility only Ape Lake, an ice-dammed, proglacial lake in the southern catchment, was selected for examination. A complete summary of field and laboratory techniques is given in Appendix E. The following is a discussion of the analytical results of this investigation.

(5.1) Patterns of Sedimentation in Ape Lake

Physical Setting

Ape Lake, situated at an elevation of 1395 m.a.s.l., is located in the southwestern portion of the catchment straddling the divide between the Noeick and Talchako-Bella Coola watersheds (figure 5.1). The lake occupies an area of 2.47 km^2 and has a contributing watershed of approximately 38 km^2 of which 20 km^2 (53%) is covered by valley glaciers and upland ice.

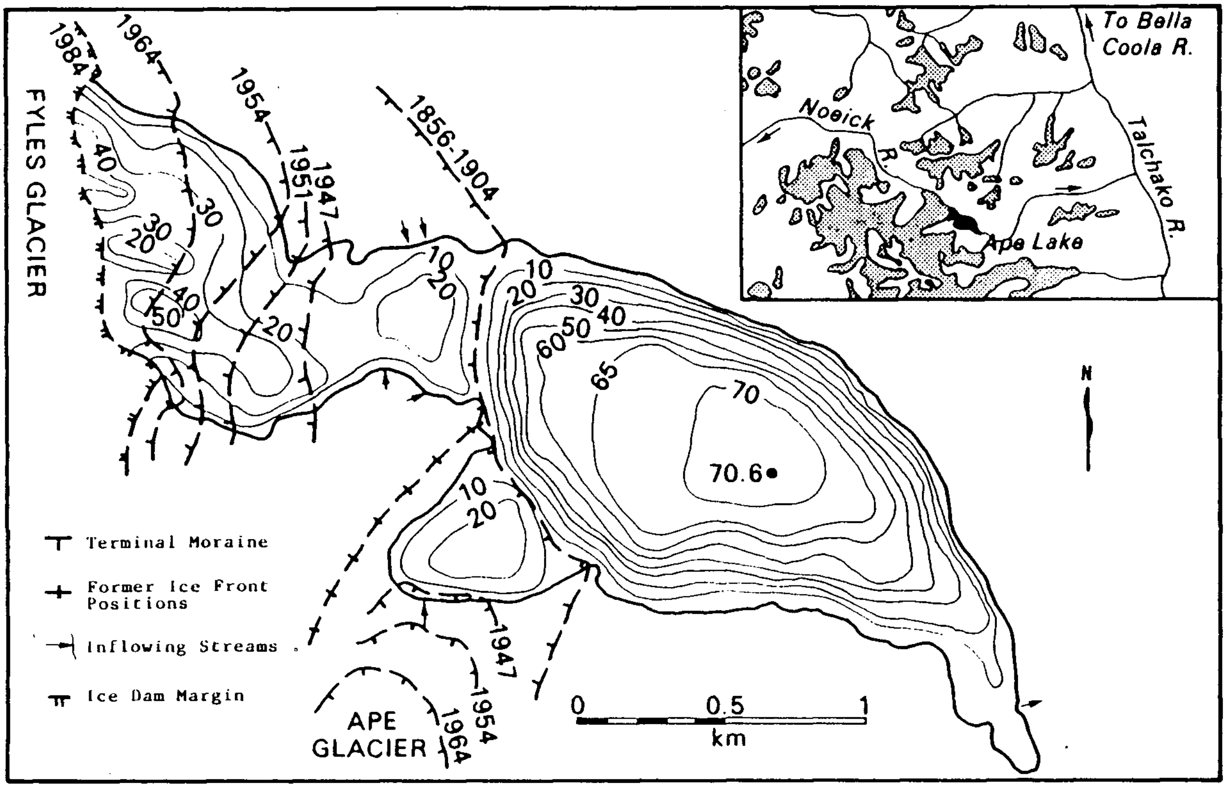


Figure 5.1 Location and bathymetry of Ape Lake. Fyles Glacier forms the ice dam to the west. Dates for glacier margins are taken from air photographs which were available beginning in 1947. Terminal moraine dates are derived from tree-ring evidence (see chapter 7). Bathymetry data were collected during the summer of 1984 with a 50 kHz echo sounder.

Fyles Glacier, a northeast flowing valley glacier which originates in the basin circumscribed by West Jacobsen Mtn., Mongol Mtn. and Mt. Fyles, forms an ice dam along the western margin of Ape Lake. Water levels have been controlled by an outlet cut into bedrock at the east end of the lake where waters discharge into Ape Creek, a tributary of Talchako River. Failure of the Fyles Glacier ice dam in October 1984 released approximately one half of the lake volume into the Noeick watershed (Jones *et al*., 1985) allowing for examination of bottom sediments in limited portions of the basin.

Ape Lake can be divided into four sub-basins (figure 5.1, table 5.1). Maximum water depths of 70 m occur in the larger, bowl-shaped, east basin which is separated from the rest of the lake by a sub-aqueous sill forming a continuous arch across the mid-basin (figure 5.2a and b). The small embayment at the foot of Ape Glacier (Ape Glacier Bay) has maximum depths of just under 25 m. A shallow depression ($\cong$ 23 m) in the middle of the lake is bounded to the east by the sub-aqueous sill and to the west by lake delta sediments which originate from tributaries on both the north and south lake margins. Lake floor gradients then increase towards the Fyles Glacier ice dam where depths at the ice-front vary between 16 and 54 m.

Ape Lake is monomictic to polymictic with weak thermal stratification developing during the summer. Although no measurements were made, overturning of the water column in the east basin is not likely to occur because the vertical thermal gradient is steep and the epilimnion is comparatively shallow (Gilbert and Desloges, 1987). Surface sediments in the lake are dominantly allochthonous clastic material with organic contents, estimated by weight loss on ignition, generally below 1 percent. Concentrations of suspended particulate matter in lake waters vary between 10 and 30 mg/l with higher concentrations found closer to the Fyles Glacier ice dam and at depth. These concentrations are half to one sixth those

Table 5.1. Ape Lake Morphology.

Sub-basin[1]	Area km^2	Max Depth m	Volume[2] $\times 10^6 m^3$ full	drained
West				
E Shoal	0.17	ca. 30	2.077	0.033
W Basin	0.61	57.6	15.752	0.372
Subtotal	0.78	---	17.829	0.405
East	1.50	70.6	64.947	38.768
Ape Glacier Bay	0.19	ca. 30	2.808	0.632
Totals	2.47	---	85.584	39.805
Percent of Totals	100	---	100	46.5
Surface Area km^2			2.47	1.27

1. For locations of sub-basins see figure 5.1.
2. Volumes determined using bathymetric data collected August, 1984. Drained volume is amount released during the jokulhlaups of October 20, 1984, and August 2, 1986.

Figure 5.2 A) Ape Lake basin two days after the October 20, 1984 outflow. Emergent climax moraines of Ape Glacier (left) and Fyles Glacier (center) divide the half-emptied lake into four sub-basins. Water levels in the larger east basin (foreground) have been lowered by 20 m. Photo courtesy R. Lenci. B) A breached climax end moraine (foreground) separates the east basin (-20 m from full lake level) from the west basin. Sand and gravel deltas are prograding into a shallow mid-basin pond (-22m) immediately behind the terminal moraine. Standing water adjacent to Fyles Glacier is at -38 m.

measured in inflowing waters and are low in comparison to those measured in other glacial lakes (*cf*. Campbell, 1973; Gustavson, 1975; Pickrill and Irwin, 1983). The majority of the water and sediment are derived from four ice-marginal and proglacial streams in the west basin (figure 5.3). All four of these streams enter behind the mid-basin sill.

Sediment Facies

Four lacustrine facies can be recognized based on surface (Ekman) sediment samples taken from all sub-basins of Ape Lake. The distribution of these facies is shown in figure 5.3 and a complete description is given in Gilbert and Desloges (1987). Facies I is light, medium-to-fine sand regularly interlayered with darker silt and is found in proximity to Ape Glacier (Ape Glacier Bay) and Fyles Glacier ice dam. Regular bedding in the sediments and the occurrence of isothermal water temperatures in this area of the lake suggest that deposition of the finer material is from suspension in a well mixed water column. Continuous, within-layer laminae of fine sand indicate the possible influence of underflow currents for sediment dispersal. Sediments for facies I are primarily derived from ice marginal and proglacial streams which enter directly into proximal sub-basins where sediments are effectively trapped behind the mid-basin sill.

Within the shallow depression of the middle lake coarse sands and fine gravels of facies II accumulate. Sand and fine gravel show poorly defined layers and occur at the front of deltas that have been building out onto this portion of the lake for at least the last 50 years (figure 5.2b). Facies III deposits are associated with sedimentation processes in the deeper and larger eastern basin (figure 5.3). Silty-clay laminae, separated from overlying silty-fine sand layers by a well-defined contact, alternate regularly with silty layers forming parallel sediment couplets which have a

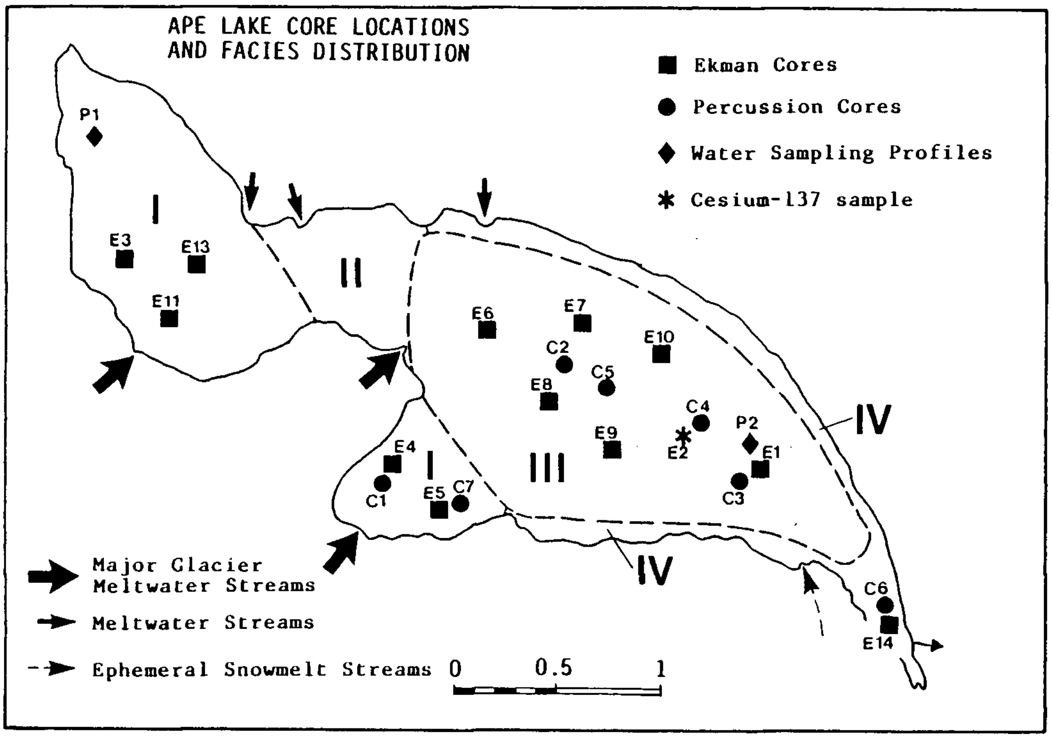


Figure 5.3 Major water and sediment sources, facies distribution and sampling locations for ekman and percussion cores in Ape Lake. The four lacustrine facies are described in the text. Large arrows represent runoff and suspended sediment from predominantly glacial sources. Small arrows represent contributions from mostly nonglacial sources. Dashes lines are ephemeral stream sources.

true varve appearance (figure 5.4). Multiple sub-layers or micro-laminations within the lighter, coarse fraction are not common but several coarser laminae (coarse silt - fine sand) can be found towards the base of some of the silty layers.

Smith (1978) and Leonard (1981) found structures similar to facies III in Hector Lake, Alberta, and suggested that for annual rhythmites these coarse basal layers would represent an influx of sediment during spring snowmelt and summer icemelt peaks respectively. Like facies I and II no current-related structures are found. With the exception of one very small delta being formed along the north shore, direct sediment sources for the east basin consist only of small ephemeral streams. Since most underflow sediments are trapped behind the sills, material is delivered primarily in homopycnal flows. Correlation of laminae amongst all 7 east-basin Ekman cores is obvious even across distances of up to 1 km (figure 5.4).

In the most distal portions of the lake and on benches which flank the east basin where water depths are considerably shallower (5-10 m), more massive clay-rich facies IV sediments are deposited (figure 5.3). Core 84E14 shows 2-3 cm of massive to weakly layered clay overlying thin, regularly layered silty-clay indicative of reduced sedimentation rates. Deposition is primarily from suspension.

Recent and Long-Term Sedimentation Rates

Estimated sedimentation rates, using measurements from all Ekman cores, vary according to the dominance of specific sediment sources. In the west basin and Ape Glacier Bay, both of which are behind the mid-lake sill and are fed by major meltwater streams, sedimentation rates average 2.56 ± 0.65 mm per couplet (n = 4 cores). In contrast, the average rate of deposition in the east basin is lower (1.81 ± 0.45 mm per couplet; n = 7

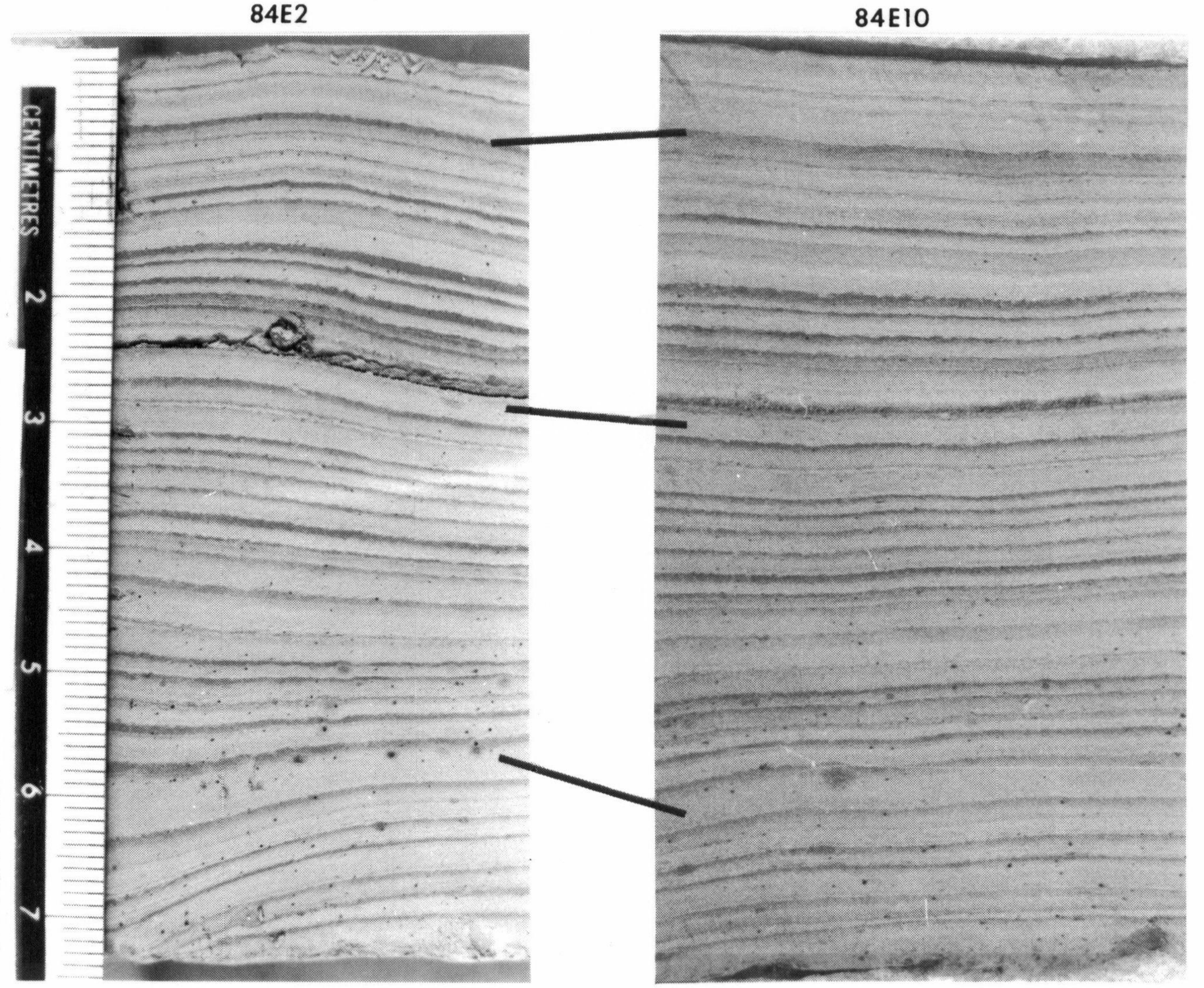


Figure 5.4 Bottom sediments of the east Ape Lake basin. Sampling locations are shown in figure 5.3. Distance between the more distal sample (84E2) and proximal sample (84E10) is approximately 225 m. Correlation between laminae is indicated by the lines.

cores) due to the larger lake bottom area and more distal location from primary sediment sources. A plot of laminae thickness with depth in figure 5.5 indicates that there are no strong spatial trends in sedimentation rate within the eastern basin itself suggesting sediments are derived from suspended load.

Seven percussion cores were extracted in order to assess longer term sedimentary processes and rates of sedimentation (see figure 5.3 for sampling locations). Core lengths vary between 30.5 cm and 73 cm. Due to some distortion (wavy, sub-parallel laminae) interpretations are based on the comparatively undisturbed segments from the four deep water cores of the east basin. Layering in the upper 15-20 cm of all four cores is similar to that in the Ekman samples, showing light-coloured, clayey-silt material alternating regularly with darker silt-clay laminae which are bounded on the top by sharp horizontal contacts. Below 20 cm the light-coloured laminae become coarser (sandy-silt) and thicker, and several contain microlaminae of medium-to-fine sand. Structures and grain sizes are similar to recently deposited sediments in the west basin. Comparatively thick sedimentary layers also appear at several levels in the lower sedimentary sequence (figure 5.6). It is possible that part, or all, of these massive layers formed by the redistribution of sediments from sub-aqueous slope failures, although they lack strong grading which is typical of turbidity currents (Harrison, 1975; Gilbert and Shaw, 1981).

Rates of sedimentation in the eastern basin are substantially higher in the lower core below 20 cm (*cf.* figure 5.6). An average rate of deposition is 4.91 ± 1.07 mm per couplet compared with a rate of 1.81 ± 0.45 mm for surface sediments. This higher rate is comparable with contemporary sedimentation rates in front of the Fyles Glacier ice dam which average over 3.00 mm per couplet.

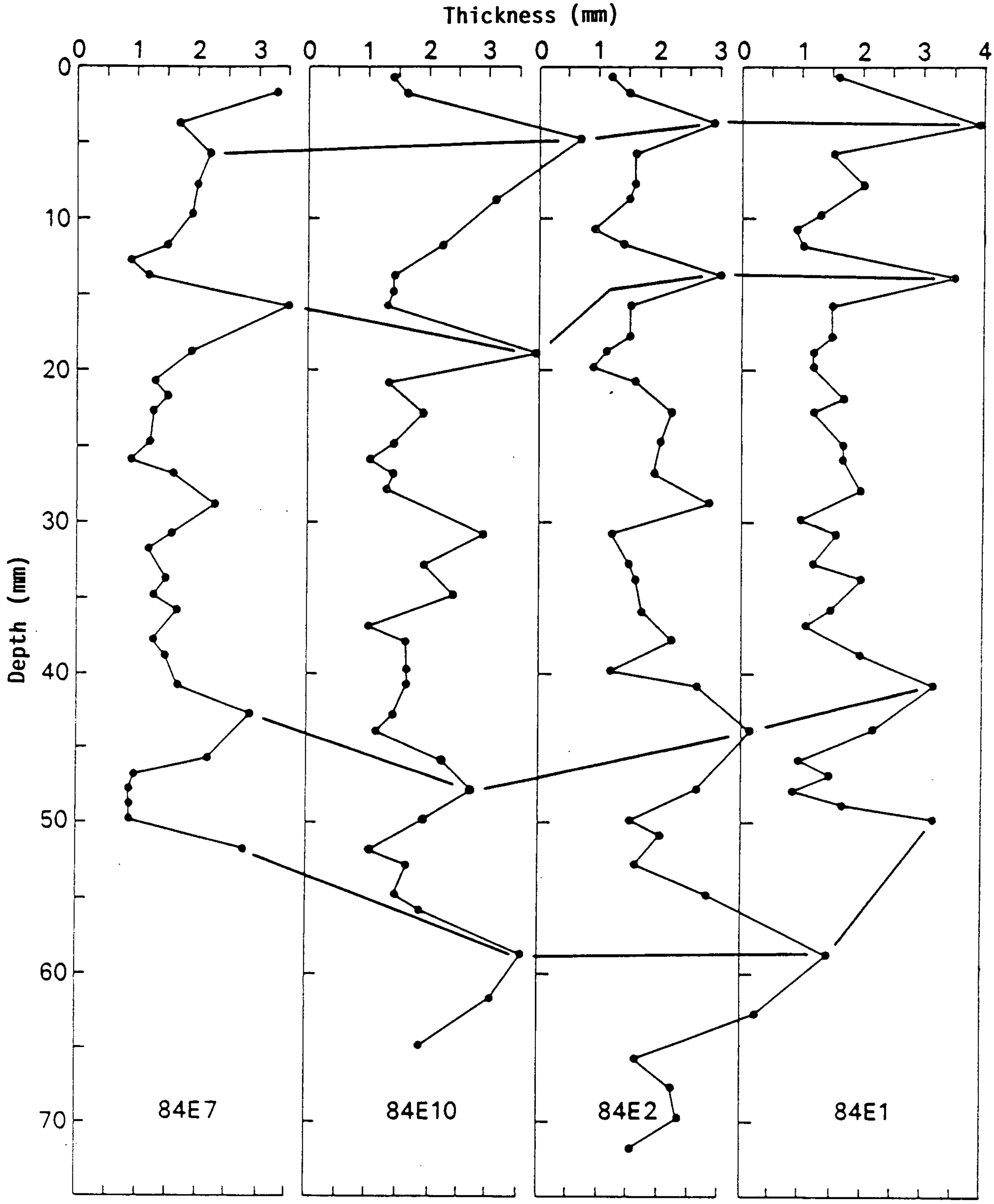


Figure 5.5 Couplet thickness versus depth in selected Ekman cores from eastern Ape Lake. Correlated laminae are connected by lines. Cores are plotted with increasing distance from water and sediment sources (see figure 5.3).

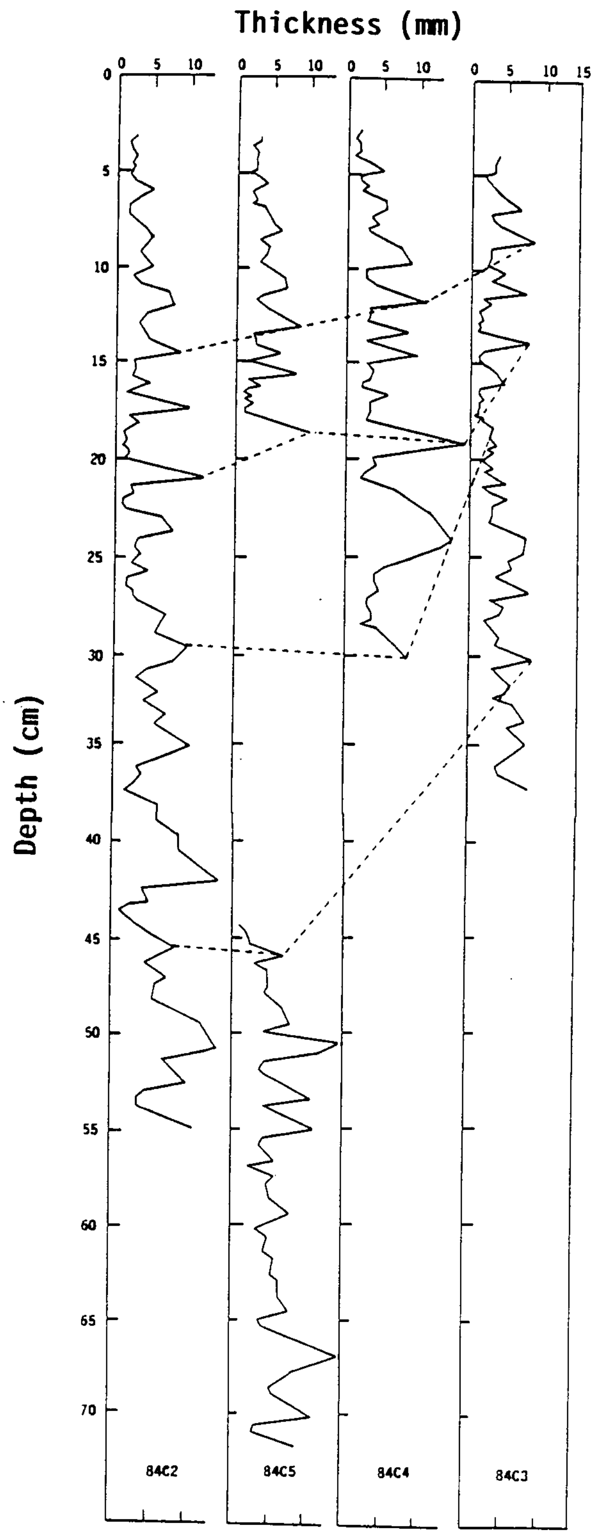


Figure 5.6 Couplet thickness versus depth for percussion cores extracted from eastern Ape Lake. Correlated laminae are connected by dashed lines. Cores are plotted with increasing distance from water and sediment sources (see figure 5.3).

Grain size variation in core 84C2 was determined using 57 samples taken as representative of grain sizes observed under macroscopic and microscopic examination of the core (Gilbert and Desloges, 1987). In the upper 20 cm there are only moderate differences in grain size of the lighter "summer" layers (63% silt/37% clay; silt-clay boundary at 4 m) and darker "winter" layers (46% silt/54% clay). Both indicate a low-energy depositional environment. Significant differences in grain size between light layers of the upper and lower core are evident. Below 20 cm the lighter layers are coarser (fine sand) and are characterized by silt/clay ratios that are four times those of the upper core. These significant textural variations suggest different sources for sediment and processes of deposition between the more recent and older sediments.

Thicker, more massive appearing layers in the upper core contain higher percentages of silt when compared with other light laminae from the same level. This coarser texture is similar to that found in lighter layers of the lower core and indicates that more energetic processes were involved in the transport of this sediment. Internal grading was not visible in thin sections but the anomalous thickness of these layers may be related to different processes than those of the surrounding laminae.

On the basis of changes in grain size and structure three types of depositional conditions might prevail throughout the year if these rhythmites are annual. During winter, following lake freeze-over, fine silt and clay are deposited from suspension to form the dark "winter" layers. Commencing with spring snowmelt runoff, coarser sediments are delivered to the lake and, later in the summer, icemelt derived materials contribute to total accumulation in the "summer" layer. Finally an influx of sediment from the surrounding basin and redistribution of sediment on submerged slopes may contribute to the sediment pile following a high magnitude runoff event (*e.g.* autumn or early winter storm).

Down core changes in grain size, sedimentary structures and rates of deposition indicate a change in the processes of sediment delivery during the interval represented by cores 84C2 to 84C5. During the Neoglacial maximum Fyles and Ape glaciers coalesced and terminated in what is now mid-Ape Lake forming a terminal moraine (figure 5.1). At the time of maximum ice advance Ape Lake would have consisted of the eastern basin with water and sediment derived directly from ice marginal and subglacial sources of the two glaciers. It is probable that facies I sediments found today in Ape Glacier Bay and in the west basin can serve as analogs for past conditions in the eastern basin, a hypothesis which is supported by similarities in grain size and structures found in the older east-basin sediments.

Each of these streams would carry coarse sediments into the lake. The occurrence of multiple sandy layers within the "summer" unit, higher sedimentation rates for both "summer" and "winter" layers and a higher variance in laminae thickness all suggest that sediment delivery was dominated by hyperpycnal flows (underflow currents) during the interval of increased ice extent. A plot of mean grain size versus sorting in figure 5.7 demonstrates the distinctiveness of these sediments.

Following retreat of the glacier ice fronts, ice marginal streams would have entered at positions behind the terminal moraine. Initially, for the larger Fyles Glacier, it is likely that the glacier remained grounded in the shallow mid-lake basin and that streams along the ice margin still entered directly into the east basin. After substantial vertical thinning and lateral retreat a greater proportion of sediments transported by meltwater streams and associated underflow currents would have become trapped behind the terminal moraine. Mixing interflow and overflow currents

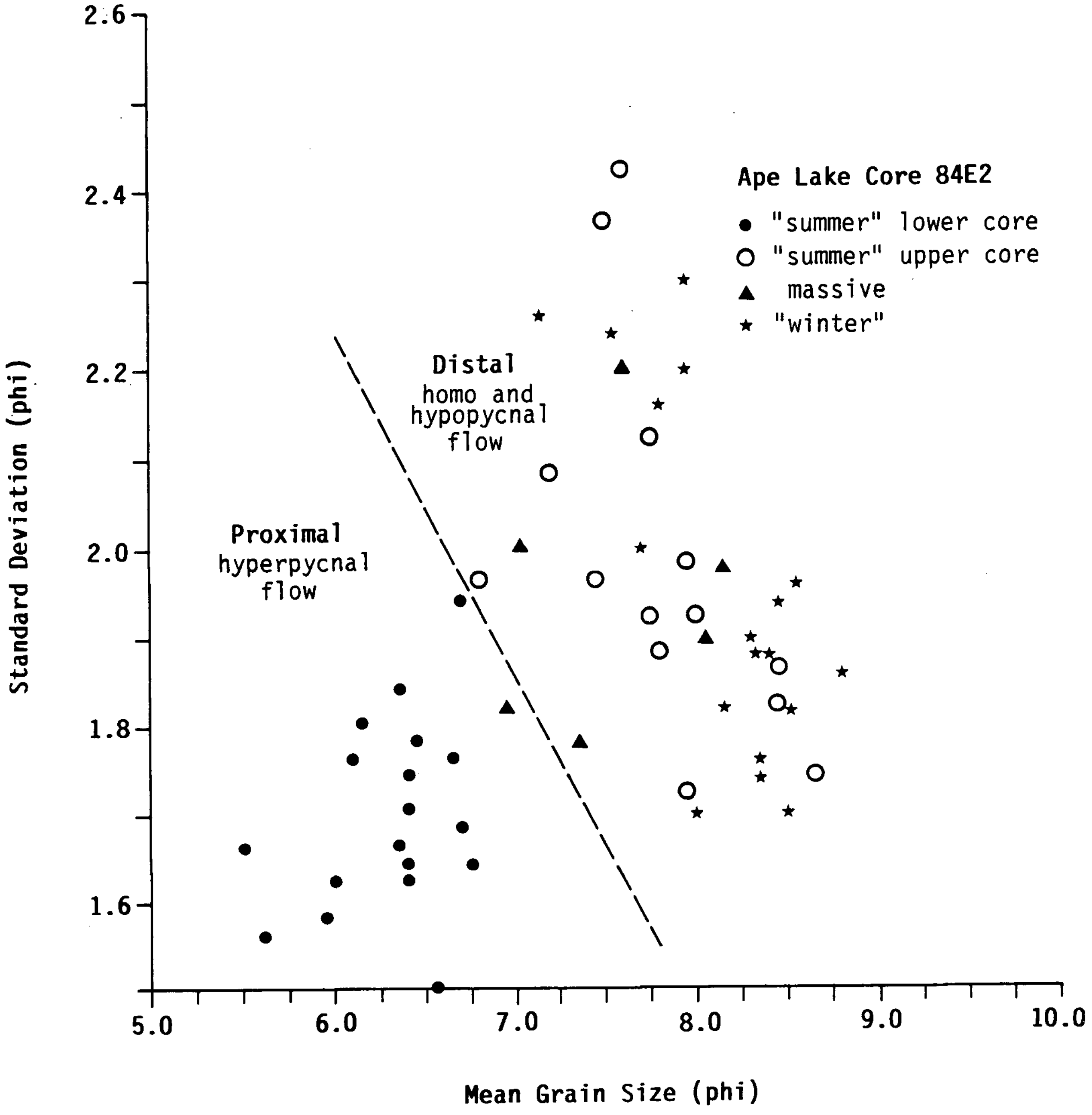


Figure 5.7 Standard deviation versus mean grain size for samples taken from core 84C2. Coarser "summer" sediments (lower core) were deposited in proximity to Fyles/Ape ice fronts when these two glaciers drained directly into the east basin. Hyperpycnal flows (underflow) dominated. Finer "summer" sediments (upper core) are deposited after retreat of the two glaciers via homo- or hypopycnal flows.

(homopycnal and hypopycnal flows) would deliver sediment to the now distal eastern basin where deposition from suspension would occur (figure 5.7).

The implications for interpretations of hydrologic/climatic controls of sedimentation rates are two-fold. First, sediments delivered by meltwater streams directly are controlled partly by those factors governing icemelt rates and subglacial drainage, partly by sediment exchange and loss in the proglacial area and partly by the frequency of subaqueous slope failures at the lake margin where streams enter the basin. Single high-discharge events can be important and therefore the sediments may not represent the overall seasonal trend in runoff and sediment delivery. Secondly, for sediments transported in homopycnal flows the rates of sedimentation may be directly linked to the degree of lake water mixing at the point of stream entry and, because sedimentation rates are lower, more easily influenced by contributions from other non-glacial sources such as turbidity currents generated by isolated slope-failures. Thus careful consideration must be given to the possible origin of individual laminae and associated hydrologic inferences.

Inferences regarding changing hydrologic conditions from changes in sedimentation rates can be made only if absolute dates can be associated with each lamina. If it can be shown that the sediment couplets referred to here are of an annual nature then the record of sedimentation rates can provide a potentially useful means for reconstructing the longer term hydroclimatology of the Bella Coola basin.

(5.2) Periodicity of Sedimentation in Ape Lake

Extremely regular layering and textural contrasts of Ape Lake sediment couplets in both east and west basins provided the first indication that these might be annual. To test this hypothesis ^{137}Cs

was selected to verify the annual periodicity of these sediments. Cesium is particularly attractive because of its reasonably short half-life (30.5 years), precisely known fallout history and global distribution (McHendry *et al.*, 1973). Cesium has a particularly high affinity for the fine-grained, organic-rich components of micaceous sediments; therefore, textures of silt and clay, such as those found in lake sediments, are most likely to mirror the ^{137}Cs atmospheric flux rate (Francis and Brinkley, 1976).

Several factors may influence the actual relationship between atmospheric fallout and observed ^{137}Cs concentrations in lake sediments. Fallout at any given latitude is influenced by the frequency and amount of rainfall (Davis, 1963), therefore, a strong gradient in isotope concentration might be expected across the Coast Mountains. Apart from the flux rate to the lake surface itself, ^{137}Cs may be temporarily stored in soils and vegetation in the surrounding catchment (McHendry *et al.*, 1973). Although data are limited, Wise (1980) indicates a lag time of 2 to 4 years between peak fallout concentrations and the transport of cesium-enriched sediments to the lake bottom. Actual lag times would reflect the frequency and efficiency of storage sites in the routing system. Leonard (1981) suggests that there may be a lag of several years or more if cesium is absorbed onto the finer (<2 cm), unflocculated sediment, thereby remaining in suspension and possibly being transported out of the system. In either case it is probable that for a partly ice and snow covered catchment, such as that which surrounds Ape Lake, a lag of several years can be expected. Other factors controlling the distribution of ^{137}Cs are related to physical, biological and chemical characteristics of the lake (see Appendix E for the implications of these factors).

Despite these problems cesium has been used successfully to establish the periodicity of sediments in Pitt Lake, B.C. (Ashley and Moritz, 1979), Bow Lake, Alberta (Leonard, 1981) and Hector Lake, Alberta (Leonard, 1986). Samples collected from Ape Lake were submitted for cesium determination to establish the periodicity of sedimentation and to assist in assigning calendar dates to rhythmites.

Results

Sampling was restricted to one core only in the east basin (84E2) (figure 5.3). The coarse-grained nature of sediments in the west basin makes cesium dating difficult and thus reliable calendar dates cannot be assigned. Although it would have been advantageous to combine laminae from a number of cores in the east basin, initial inspection of the samples indicated potential problems with intra-annual layers and cross-correlation of units. The relatively thin laminae of core 84E2 precluded sampling on a couplet-by-couplet basis, hence, three or four couplets were included in each sample to ensure sufficient sample sizes even though a lower temporal resolution in ^{137}Cs activity would result. The sediment was trimmed to exclude contaminated material along the edge of the core and samples were then shaved using a sharp thin knife. Boundaries separating each sample were sharp and usually free of distortion so that mixing of sediment between samples was unlikely. A summary of sample depths, sample sizes and measurement procedures is included in Appendix E.

Concentrations of ^{137}Cs in core 84E2 are plotted and compared with the atmospheric flux recorded at Lake Michigan in figure 5.8. The smoothed profile recorded in core 84E2 is similar to the atmospheric fallout rate suggesting that laminae in 84E2 are annual in nature and therefore are true varves. Very low level concentrations of ^{137}Cs (0.11 pCi/g) are detected in

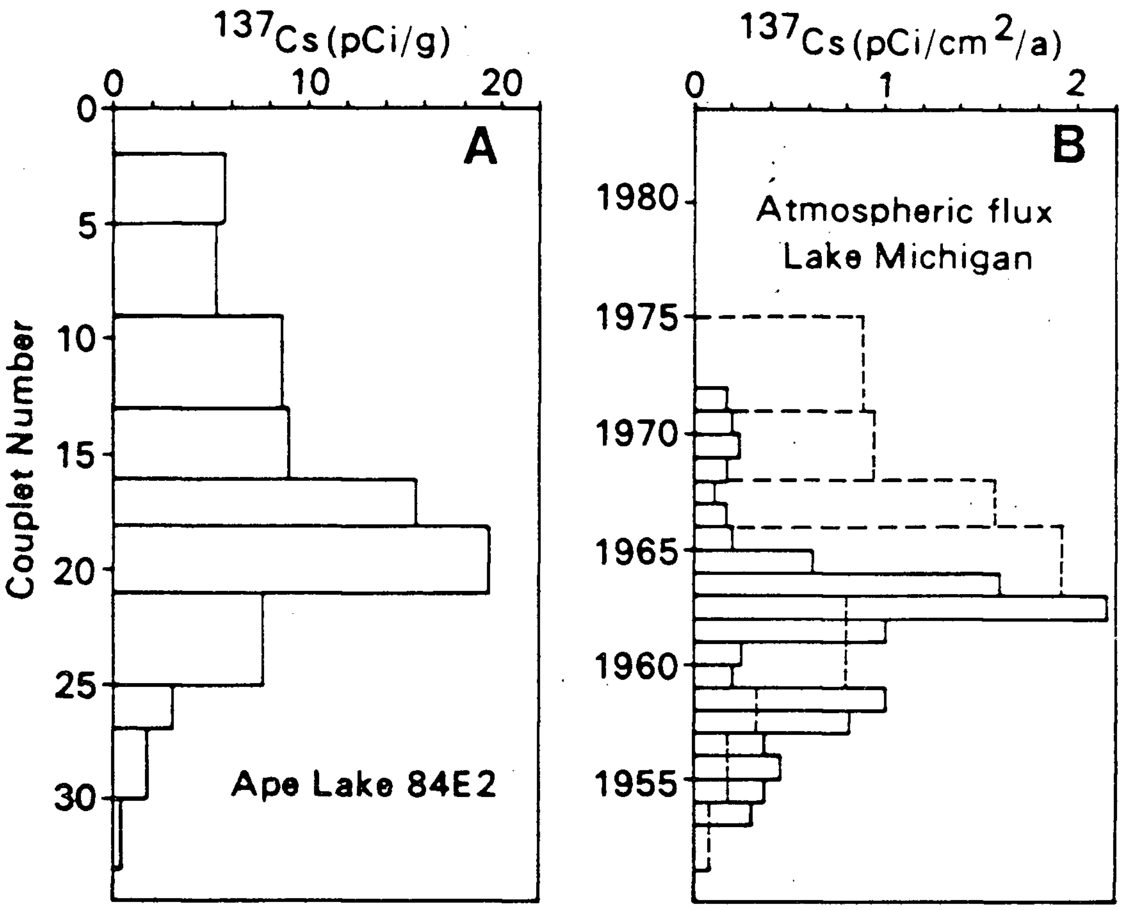


Figure 5.8 A) ^{137}Cs profile in core 84E2 from eastern Ape Lake (see figure 5.3 for sampling location). Concentrations were determined on a 2-4 couplet basis. B) Estimated annual atmospheric flux of cesium-137 in eastern North America (solid line is from Robbins and Eddington, 1975). Dashed line is superimposed profile from A.

couplets 31-33 indicating that this is the lower boundary of the cesium profile. Peak concentrations are found in couplets 19-21 and then decline rapidly higher in the profile but do not return to the low levels measured at the bottom of the core (couplets 1-2 were not measured for ^{137}Cs activity). The peak values found in couplets 19-21 suggest that the high atmospheric fallout rates recorded in 1963 are incorporated in this level of the core. Under the assumption that each couplet represents a single year of deposition and that the surface layer recorded in core 84E2 represents sedimentation in 1984 then the sample containing couplet number 22 should correspond with the 1963 peak. Since this is not the case a lag of at least one year and as many as four years may be present which is a reasonable estimate given the evidence presented earlier.

High values of ^{137}Cs activity shown in figure 5.8 several years after the sharp decline in atmospheric fallout suggest one of two things. First, storage of cesium in the surrounding watershed is high and continuing contributions of ^{137}Cs from snow, ice and sediments to the lacustrine environment are significant. Evidence from both Ashley and Moritz (1979) and Leonard (1986) suggest that this is possible but lake bottom sediments from both these studies show a more rapid decline in cesium activity following the reduction in atmospheric fallout. The lag and smoothing of the profile is more pronounced in glaciated terrain which would suggest significant storage in snow and ice. A second possibility is that recycling of ^{137}Cs is occurring at the lake bottom, thus maintaining these elevated values (*cf.* Albert *et al.*, 1979).

Cesium isotope activity in Ape Lake sediments, then, provides good evidence for the annual periodicity of the rhythmite couplets. Though absolute dates based on ^{137}Cs activity cannot be assigned, the results do indicate that counting of well-defined couplets is a reasonable means of establishing a history of sedimentation. The surface layer of most Ekman cores sampled in August 1984 consisted of thin light-coloured silt with no evidence of overlying clay deposits suggesting that this layer is comprised of sediments derived from spring snowmelt and summer ice melt during 1984. Hence, for the varve chronologies discussed in the next section it is assumed that surface layers, found in all cores were deposited during the summer of 1984 and that couplet number 22 was deposited in 1963.

(5.3) Chronology and Controls of Recent Sedimentation Rates

Measurements from Ekman cores taken in Ape Lake were used to construct a time series of recent (post-1948) sedimentation rates. To ensure that the chronology reflected lake-wide changes and to smooth anomalously high or low differences in sedimentation rates several cores were combined to produce a composite series. Samples from the east basin only were selected because of the morphological similarity of the lake bottom from which the cores were extracted and because sediments in this part of the lake are derived from sources which are representative of changes integrated over the entire catchment. Additional factors include high frequency of samples and the potentially longer record preserved in the east half of the lake.

Typically, Ekman cores covered a period of 30 years occasionally extending beyond 35 years (1948) and all but one (84E1) contained sediments deposited during the summer of 1984. Problems associated with the presence of possible intra-annual summer layers in three Ekman cores reduced the number of samples used in the composite chronology to four, the same number of samples available for construction of the longer chronology.

Recent Sedimentation Chronology

Total varve thicknesses in the four Ekman samples (84E1, E2, E7, E10) were standardized in order to reduce first order autocorrelation and give equal weight to each core. Each varve series was standardized by dividing the actual varve thickness by the sequence mean which yields a mean of 1.0 for the transformed data set below which sedimentation rates are considered low and above which sedimentation rates are high. Evidence presented earlier showed that declining rates of sedimentation occurred after the mid-1940s as the shallow middle-basin formed and then again after the early 1960s as the deeper west basin became fully developed. Thus, two sequence means were used for generating indices of secular variations in sedimentation rates, one for the earlier period (1948-1960) and one for the latter period (1961-1984). All four standardized series were then combined as an unweighted average to form the composite chronology. Estimated lake-wide sedimentation prior to 1953 is based on fewer than four samples. The composite sedimentation chronology is plotted in figure 5.9.

Sedimentation rates show some variance about the record means with well above average values in five individual years (1952, 1953, 1959, 1976, 1982) and below average sedimentation in the middle 50s, early 60s and 70s. For comparison, post-1948 generalized departures in winter precipitation for interior stations are also plotted in figure 5.9. While there is some agreement, particularly in the late 1960s and early 1950s when both precipitation and sedimentation rates were above average, the two series are not closely matched. This suggests other mechanisms may be important determinants of sediment delivery to the lake.

It was thought that the influence of early autumn storms might be reflected in the sedimentary sequence. Significant rainfall and snowmelt runoff from both glaciated and non-glaciated terrain is apt to occur when

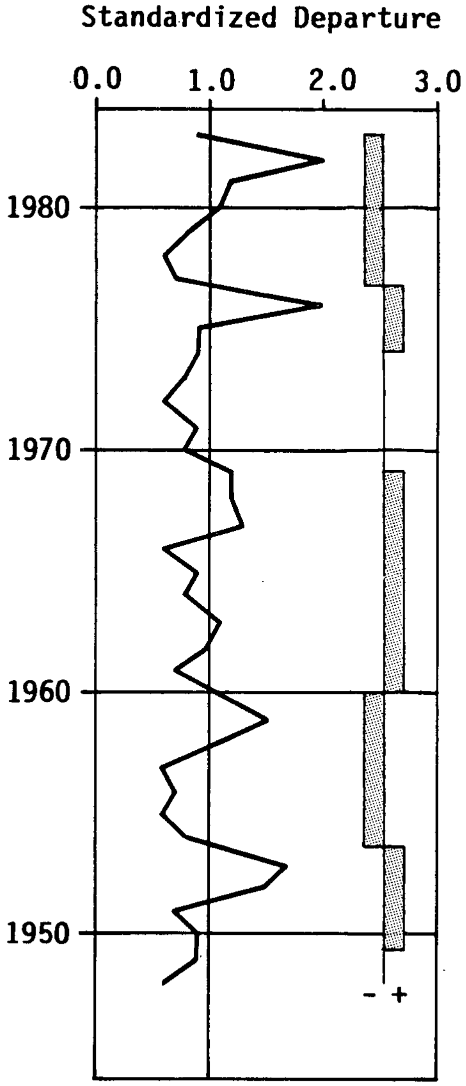


Figure 5.9 Composite sedimentation chronology for cores 84E1, 84E2, 84E7 and 84E10. Original curves shown in figure 5.5 were standardized by dividing observed sedimentation rates by the respective period mean. Two period means were used to reflect known fluctuations in the position of Fyles Glacier: 1948-1960 and 1961-1984. The standardized curves were then combined as an unweighted average. For comparison generalized departures of winter precipitation for interior stations are shown on the right.

the freezing level rises during autumn storms. Heavy rainfalls observed in late August of 1984 produced visible evidence of underflow currents and sediment redistribution in proximity to proglacial and ice-marginal stream outlets. However, careful inspection of prepared thin sections for east-basin varves associated with years in which several high magnitude autumn storms occurred revealed no systematic departures in sedimentation rates.

Controls of Glaciolacustrine Sedimentation Rates in Ape Lake

Inferences regarding changing environmental conditions from lake sedimentation rates require that the linkages between sediment yield and controlling glacio-hydroclimatological variables be known. Standardized values of varve thickness for the post-1948 series (figure 5.9) are compared with annually and seasonally measured hydroclimatic variables using correlation analysis. A set of 13 independent variables includes: winter precipitation (Oct.-Apr.) for coastal and interior regions; mean annual temperature; summer (May-Sept.) and late season (Aug.-Sept.) temperature; and runoff in the Talchako and Bella Coola Rivers (total summer, early summer (May-June) and late summer (Jul.-Aug.)). Results are presented in table 5.2.

In general, most comparisons using the entire 32 year data set yield low and non-significant correlation coefficients with the exception of winter snowpack measured in the interior region ($r = +0.39$), late summer season temperatures ($r = +0.44$) and runoff in the Talchako River ($r = +0.45$). Scattergrams of varve thickness with precipitation, snowpack and runoff indices revealed a substantial amount of "noise" during years characterized by low sedimentation rates. This effect was much less pronounced in comparisons with temperature indices. Examination of multiple regression models (sedimentation rate as a function of hydroclimatic

Table 5.2. Pearson's r for Varve Thickness and Hydroclimatic Variables

	Winter Precip. (i)	Winter Precip. (c)	Snowpack (i)	Snowpack (c)	Temperature Annual	Temperature Summer	Temperature Late Summer
[1] (df = 32)	0.03	0.10	0.39*	0.21	0.15	-0.12	0.44*

	Runoff Summer	Runoff Early	Runoff Late
[1] Bella Coola R. (df = 32)	0.31	0.21	0.30
[2] Talchako R. (df = 19)	0.39*	0.29	0.45*

* significant at $\alpha = 0.10$.

(i) refers to index derived for interior stations and (c) for coastal stations

[1] Entire post-1945 data set

[2] Post-1965 data set

variability) indicated that confounding effects between temperature and precipitation exist. For example, removing years with anomalously low summer temperatures improves the relation with snowpack ($r = +0.51$) and with summer runoff in the Talchako River ($r = +0.55$).

Perkins and Sims (1983) and Leonard (1985) have found a positive and significant association between annual or summer temperature and sedimentation in proglacial lakes. Correlations were much higher than those for Ape lake and confounding effects were not reported. With increasing ablation season temperatures a greater proportion of ice melt and discharge into the lake is thought to lead to increasing concentrations of suspended particulate matter. A similar conclusion cannot be made from the results shown here. Low and even negative correlations between annual or summer temperature and sedimentation rates suggest that seasonally averaged temperatures from nearby stations are poorly related to conditions found at Ape Lake. However, during the late summer season (August-September), when the maximum snow-free glacier surface is exposed and glacier melt is at a peak, temperature and sedimentation exhibit a strongly positive relationship.

April 1st snowpack and varve thickness also show a strong positive association particularly during years of average or above average late season temperatures. Increased snow cover (and temperatures) would lead to a higher discharge volume to the lake from the surrounding watershed and potentially higher concentrations of sediment in the lake if this inflow carries a significant suspended sediment load. Perkins and Sims (1983) found a strong <u>negative</u> correlation between snowpack and varve thickness in a southern Alaska lake and suggested that reduced glacier melt during years of heavy snowfall (increased snow cover and albedo of the glacier surface) produces lower sediment yield and thus reduced rates of deposition.

The opposite relationship found for recent Ape Lake sediments may relate to sources and transfer of sediment. A substantial amount of fine grained material (silt, fine-sand) can be found in the proglacial areas of Fyles and Ape Glaciers. During years of high snowpack, runoff from these areas may contribute more sediment than during low snowpack years. This hypothesis, however, would dictate that these areas be at least co-dominant sources of the suspended sediment found in Ape Lake. Inspection of non-glacial streams in July of 1985 and 1986 indicates that sediment concentration in these inflowing streams is small in relation to those draining Fyles and Ape Glaciers directly.

Although correlations with regional runoff indices are generally low, a higher and statistically significant association occurs in years when temperature and snowpack conditions favour heavy runoff. This may suggest that certain minimum concentrations of suspended sediment are introduced to the lake each year but vary significantly only during moderate and high runoff years. Gilbert (1975) found that sedimentation in Lillooet Lake, British Columbia was significantly related to discharge during higher discharge years only, further supporting the contention that snowmelt generated runoff may be an important mechanism in delivery of sediment to the east basin. The higher correlation with Talchako River flows would support this conclusion.

Recent depositional processes then, are related to several morphological and hydroclimatic factors which promote increased sediment concentration in Ape Lake. The potential for these proxy data to reveal a record of long-term hydrologic change is limited by the fact that no simple hydroclimate-sedimentation model can be specified. These complications are associated with several conditions: the strong maritime influence on runoff processes in coastal British Columbia; changing basin morphology, sediment sources and glacier configurations over the last century; and the processes of sediment dispersal in the lake. Although a quantitative model cannot be explicitly specified for the period of detailed instrumental data, above average sedimentation in Ape Lake is associated with years characterized by above average snowpack, late summer season temperatures and regional runoff. The long-term sedimentation record was investigated in an attempt to verify hypotheses regarding controls of glacio-lacustrine deposition and to possibly make qualitative inferences about earlier hydrological environments.

(5.4) Inferences from Long-term Variations in Glacio-Lacustrine Deposition

The four deep-water, east-basin percussion cores (84C2, C3, C4, C5) were used to construct a standardized long-term chronology of sedimentation rates in Ape Lake prior to 1948. Due to increasing sedimentation rates lower in the sequence, the much longer percussion cores spanned a proportionally shorter interval when compared with the Ekman cores, the longest being 143 years (84C5). Thus, a chronology extending back to 1836 A.D. was available. Distorted sections of 84C3 and 84C5 were excluded so the number of cores used in the composite chronology varies between two and three in the period prior to 1904 (see figure 5.9). Prior to 1865 only one core (84C5) was used; therefore, there is a greater potential for bias in the earliest part of the record. Each sequence was standardized by dividing actual sedimentation rates by the 1843-1940 mean. The mean for this interval was selected because prior to 1940 sediment delivery was directly into the east basin. The unweighted series were then combined to produce an average chronology (figure 5.10a). To enhance the lower frequency variation and to facilitate interpretations of longer-term trends the composite

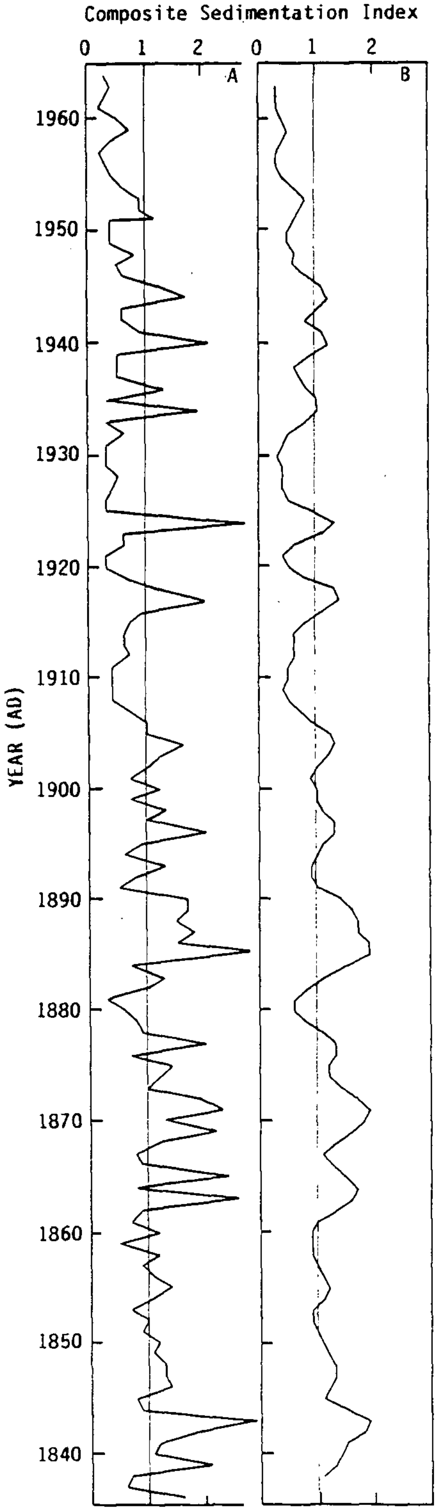


Figure 5.10 Composite sedimentation chronology for cores 84C2, 84C3, 84C4 and 84C5. Original curves shown in figure 5.7 were standardized by dividing observed sedimentation rates by the 1836-1940 period mean. The standardized curves were then combined as an unweighted average. Unsmoothed data in A and a 5-year weighted moving average in B.

series was filtered with a five year weighted moving average (figure 5.10b).

The long-term trend shows declining sedimentation rates from the early 19th century through the middle of the 20th century. The decreased rate of deposition after 1940 is related to positional changes of Fyles Glacier rather than actual sediment yield but is included in figure 5.10 for comparison. Below average rates of sedimentation are evident after 1905 until 1940 and local minima occur between 1873 and 1884 and between 1845 and 1862. Comparatively high annual depositional rates are found in 1843, 1863, 1865, 1869, 1885, 1917, 1924, 1934 and 1940. The intervals 1836-1844, 1862-1885 and 1917-1945 are characterized by relatively high year to year variance in depositional rates. Mean sensitivity is approximately 20-40% greater than in other intervals.

Key Events

Examination of sedimentary structure and varve thickness in the record prior to 1940 indicates a more complex sequence of intra-annual sedimentation than the more recent sequence. Micro-laminations of silt and clay alternate with fine sand in several layers suggesting a sequence of discrete events (*e.g.* turbidity currents) might be responsible for the transfer of some sediment within the eastern basin. It was thought that sediment yield would reflect variation in discharge of meltwater streams as they entered directly into the east basin in addition to high-magnitude runoff events, particularly autumn storms which produce the greatest discharges in the regional drainage network.

To facilitate comparison an annually calculated index of cumulative autumn discharge (summation of discharges above 250 m^3s^{-1} in the Bella Coola river between late August and January) is plotted against positive sedimentation anomalies for Ape Lake in figure 5.11. The less detailed record of flows prior to 1940 reflects the limited data available especially regarding flood events of low to moderate magnitude. The records do not match particularly well after 1940 when much of the sediment derived from Fyles Glacier was trapped behind the terminal moraine. A comparatively good association exists in the early 20th century record, particularly in the years 1896, 1917, 1924, 1934, 1936. Major floods were recorded in the Bella Coola River during each of these years. The correspondence indicates the significance of autumn rainstorm-generated runoff in the earlier sedimentary record.

The timing of autumn rainstorms also appears to be an important factor governing rates of sedimentation. Lacustrine deposition associated with flooding on November 19, 1936, the largest estimated flow in the Bella Coola River of this century, is lower when compared with other sedimentary units all of which correspond to runoff events earlier in the season (August to mid-October). It is probable that once lake freeze-over has commenced (November-December) rainstorm runoff has less effect on contributions of sediment to the lake. A greater proportion of precipitation which falls as snow at higher elevations later in the season will also result in lower runoff contributions to the lake.

In addition to positive sedimentation anomalies, the frequency of microlaminations measured within each varve is plotted in figure 5.11. Each sub-unit may represent sedimentation associated with material derived from spring runoff, summer glacier melt or autumn rainstorms. The number of layers within any "summer" unit varies from zero (homogeneous deposit) to four. As a result of major meltwater streams entering the lake behind the mid-basin sill, the post-1945 record contains the fewest intra-varve laminations. To confirm the temporal variability in frequencies of

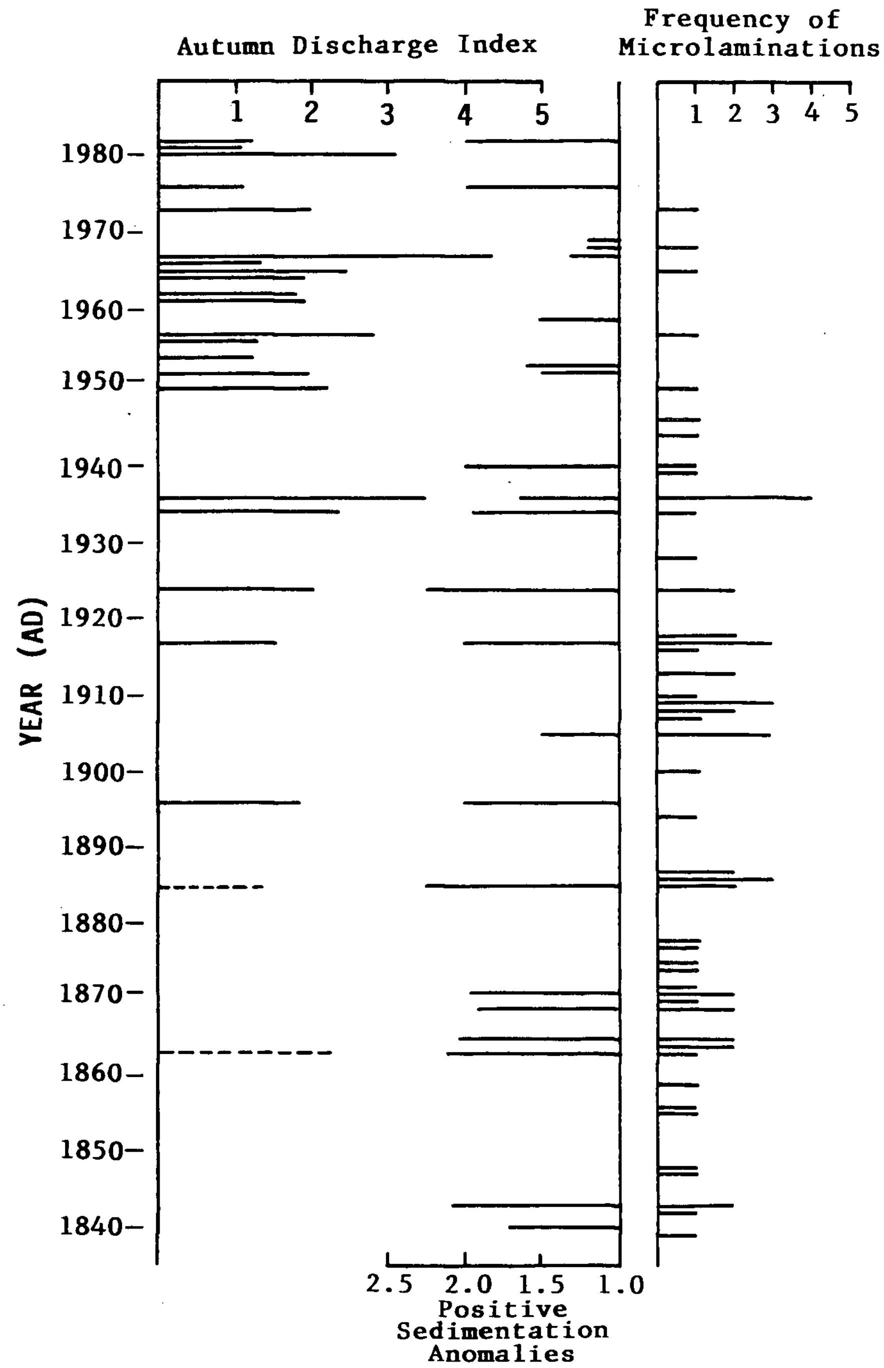


Figure 5.11 Sedimentation anomalies and frequency of microlaminations in sediments of eastern Ape Lake. For comparison an autumn discharge index between the years 1896-1984 for the Bella Coola River is included. The discharge index prior to 1948 is based on historical observations and field evidence only. Note the correspondence of flooding and sedimentation anomalies.

microlaminations a one-tailed runs test was applied under the null hypothesis that there is a random distribution of these frequencies. The null hypothesis was rejected at the 90 percent confidence level ($Z = 1.68$) demonstrating that the higher frequency of intra-annual layering noted between 1863 and 1878 and between 1905 and 1918 is significant in the long-term record. Based on the higher frequency of intra-annual sedimentation and higher rates of sedimentation during the former interval it can be inferred that some combination of above average spring runoff, longer summers and greater numbers of autumn rainstorms occurred then.

The inferred early 20th century departures can be tested qualitatively by examining climate trends at Bella Coola and Big Creek. It was noted that between 1904 (limit of early 20th century records) and approximately 1920 a persistent meridional flow regime dominated the general circulation. Although summer temperatures at Bella Coola were notably below average during this time (figure 3.9), summer temperatures at Big Creek and regional winter precipitation were above average (especially 1906 to 1912 and 1918 to 1920). This is coincident with a higher frequency of intra-varve laminations.

Hydroclimatic Inferences

Further quantitative tests of the relation between hydroclimatic factors and sedimentation rates prior to 1940 (1904-1940) were considered using coastal and interior climate indices. Summer and winter precipitation, summer temperature (May-Aug.), early season temperature (May-Jun.), mid season temperature (Jul.-Aug.) and late season temperature (Aug.-Sept.) were standardized for this interval and compared with sedimentation rates in the east basin. Because of the unknown contribution of isolated autumn storms those varves which are suspected of containing sediment derived from these events (1905, 1909, 1917, 1924, 1934, 1936, 1940) and which contribute a disproportionate amount to the overall variance in the series, were removed.

Much like the recent sedimentation sequence, precipitation and summer temperature indices contribute little to the explained variance of pre-1940 sedimentation rates. Again, late season (Aug.-Sept.) temperature seems to be the most important and significant variable ($r = +0.51$) for the filtered series (which represents a substantial improvement over the unfiltered series where $r = +0.36$). It is probable that during the maximum advance of Fyles and Ape Glaciers a small or insignificant proglacial area would have been exposed, thus snowmelt-generated runoff would not have been an important mechanism for sediment delivery.

The above results suggest that controls of recent (post-1945) sedimentation in ice-dammed Ape Lake vary from those found in proglacial lakes fed by one or two major meltwater streams. A combination of stream flow from the surrounding ice-free basin and glacier melt appear to be important elements in the redistribution and delivery of sediment in the basin. The strength of the relations does not permit precise long-term reconstructions to be made but qualitative inferences regarding environmental change may be possible.

(5.5) Summary and Conclusions

Several factors appear to be important in determining the processes of sediment distribution and deposition in Ape Lake. A fairly abrupt shift in the character of deposits from microlaminated and thick summer varves containing layers of fine sand to thinner, uniform, silty summer varves is associated with changes from an underflow/interflow depositional environment to one dominated by overflow and fallout of suspended material.

In the latter case, snowpack volumes and late season ablation temperature appear to control depositional rates. This association can be only partly verified in the pre-1940 record where late summer temperature and the frequency of autumn rainstorms influence rates of deposition. This shift in depositional regimes coincided with the opening of the west basin sometime between 1940 and 1945.

Inferences regarding changes in environmental conditions over the longer-term using rates of sediment deposition as proxy evidence clearly depend upon a stable model(s) relating the two. No satisfactory quantitative model could be specified. Grain size, structure and depositional rates in the east basin appear to be consistent within the pre-1940 period suggesting that summer temperatures and the frequency of high-magnitude rainfall events are important explanatory variables for this interval. It is probable that above average sedimentation rates in 1840, 1843, 1863/64/65, 1871/72, 1885, 1887 and 1896 are associated with unusually warm summers and/or major flood-producing storms. Distinguishing between the two mechanisms remains problematic.

CHAPTER VI
Biological Evidence of Environmental Change: Tests Using Tree-Ring Evidence

(6.0) Introduction

Landforms and sediments commonly provide only limited resolution of hydroclimatic variability. Variations in tree-growth are known to be associated with climatic factors integrated over the course of a season or throughout a particular year. In order to utilize this form of proxy data it is necessary to consider response characteristics and potential sources of error involved in drawing inferences about long-term changes. The purpose here is to identify these responses, assess the utility of transfer functions developed using tree-ring data and compare inferred hydrologic changes with observed fluctuations over the 20th century.

(6.1) Tree-Growth and Climate

Annual wood increments consist of low-density, large-celled earlywood (spring) and high-density, small-celled latewood (late summer or early fall). Each year's increment is controlled by several factors, the most important of which are climate and site characteristics. The major regulators are temperature and precipitation because they control the water balance in and around the tree and regulate the metabolic growth rate (Kramer and Kozlowski, 1979). Other less direct factors include competition between species and between units of the same species, slope stresses, physical damage and depletion of soil nutrients. Growth rates from year to year will respond to all these factors but are generally dominated by the most limiting (Meyer *et al.*, 1973).

Moisture or heat stress during a significant portion of the growing season will result in reduced photosynthesis and a lower production of biomass. Actual thresholds for stress are partly species dependent and partly influenced by site conditions such as local drainage, soil depth and sources for nutrients. Between some lower and upper thresholds, each characterized by an integrated set of climatic, physiological and pedological conditions, growth rates are optimal from year to year and will be limited only if some other factor is in short supply. This situation generally produces little growth variation through time and thus a strong climatic signal is not present.

In order to isolate the effects of climate it is necessary to select sites which are host to trees that are near or at their limit of tolerance for moisture or energy. It is not uncommon for a climate factor to be limiting for only part of the growing season. For example, above average temperatures in the early spring may initiate bud opening and produce copious amounts of earlywood. However, if conditions deteriorate as the growing season progresses below average biomass production may occur and the resulting wood increment may not reflect the extremes in growing conditions. Because it is the available energy that determines the timing of bud opening, temperature is initially a more important factor (Kramer and Kozlowski, 1979). Later in the season, when the demand for water and nutrient intake is at a maximum, the availability of water is more important.

Long-term variation in tree growth is a function of: (1) a quasi-deterministic component related to the size of the tree and dynamics of the forest stand and soil environment and (2) a stochastic component related to the persistence of climatic effects in one year on the physiological status of the tree in ensuing years (Graumlich, 1985). Hence, the way in which a tree responds to climate in any give year may depend on growing conditions in previous years. The actual memory of the system may be several years but the effects may damp out quickly if above and below average growing conditions occur in close succession. Therefore, tree-ring growth indices tend to exhibit an autocorrelative structure of at least a one year lag.

Tree growth response may be compared directly with temperature and precipitation or indirectly via other hydrological parameters which are known to be sensitive to climatic change. Stockton and Fritts (1973) have shown that tree-growth may be influenced by groundwater conditions, in particular the level of the groundwater table and degree of soil saturation. Other attempts have been made at reconstructing river runoff variations using tree-rings (Stockton, 1976; Holmes *et al*., 1979; Campbell, 1982; Jones *et al*., 1984). The comparisons remain approximate since runoff and tree growth are both controlled by climate variations rather than there being a direct functional association between them.

Conifers of the Pacific Northwest take advantage of the warm, moist climate by continued, although reduced, photosynthesis and nutrient uptake throughout much of the autumn and early winter seasons. Increased moisture stress and greater continentality of temperatures towards the interior can lead to a differential seasonal response in growth characteristics for the same species (Waring and Franklin, 1979).

Tree-growth responses to climatic change have been investigated for a number of sites and species in northwestern North America. Keen (1937), Schulman (1956) and Stokes *et al*. (1973) performed much of the earlier work on drought sensitive Douglas firs in the drier regions of eastern Oregon, Washington and south-central British Columbia. More recently Fritts and Lough (1985) and Graumlich and Brubaker (1986) considered long-term temperature variations of the Pacific Northwest based on tree-growth variations for species at or near tree line. In addition, Graumlich (1985) investigated the response of drought sensitive species in the Pacific Northwest. While long-term temperature and precipitation reconstructions have been undertaken for areas further north in the Yukon (Jacoby and Cook, 1981) and further east in the southern Canadian Rockies (Josza and Oguss, 1985), little information is available for coastal and western Fraser Plateau areas of central British Columbia.

Methods

Selection of suitable sampling sites is the most important step in obtaining tree-ring data which are likely to contain a climate signal of regional significance. Graybill (1982), Graumlich (1985) and others have demonstrated that the width of a tree-ring (R) at any time t will depend upon: a biological growth trend (G) as a function of increasing age, soil dynamics and a short-term physiological memory; a macroclimatic signal (M) common to all trees at a site which can be decomposed into several frequencies; disturbance factors (D) such as fire, insect damage or slope stress which may be unique to one tree or common to all trees at a site; and some random growth signal (E) unique to each specimen; *i.e.*,

$$R_t = G_t + M_t + D_t + E_t$$

Although the breakdown of the model is not entirely realistic since some factors may become confounded at certain frequencies (*e.g.* soil nutrient status and macroclimate) it does provide a framework for decomposition of frequencies within the ring-width time series.

In order to maximize the climatic signal in this study a series of cores were extracted from several trees at a number of sites which showed no adverse reaction to the disturbance factors cited above. Usually two increment cores were extracted from each tree at breast height perpendicular to slope orientation; a sampling scheme which minimizes the possibility of sampling growth trends related to slope stresses.

Sites (trees) characterized by over-steepened slopes, poor drainage, frequent fire damage, extensive scarring, intense competition and over-mature (decaying) forest cover were avoided. Entire tree disks were taken at some sites in order to assess more completely the within-tree and between-tree variance of growth patterns. Details of sampling and acquisition of ring-width data are given in Appendix D.

A total of eight sites were selected for sampling, seven of these within the basin and one towards the drier interior of the province (see figures 6.1 and 2.1). Four of the eight sites were at or near tree line within forest communities of subalpine fir/lodgepole pine/western hemlock but chronologies were developed only for subalpine fir (*Abies lasiocarpa*). Tree line sites are thought to be primarily temperature stressed. The four other sites were in mature stands of predominantly Douglas fir (*Pseudotsuga menziesii*) located in the drier valley bottom of the eastern Bella Coola basin (three sites) and near the confluence of the Chilko and Chilcotin Rivers on the Fraser Plateau (one site). In addition to the four Douglas fir chronologies produced here, tree-ring data for several sites centered around Williams Lake (Schulman, 1956) and elsewhere in the middle Fraser River (Stokes *et al*., 1973) were also considered. Douglas fir trees within these regions are thought to be moisture stressed. A summary of site names, sampling frequency, temporal coverage of each chronology and signal-to-noise ratios is given in table 6.1. Full data series and summary statistics are in Appendix D.

A site averaged chronology was produced by first removing the deterministic portion of the long-period variance using negative exponential or low order polynomial curves fit to each series. Selection of

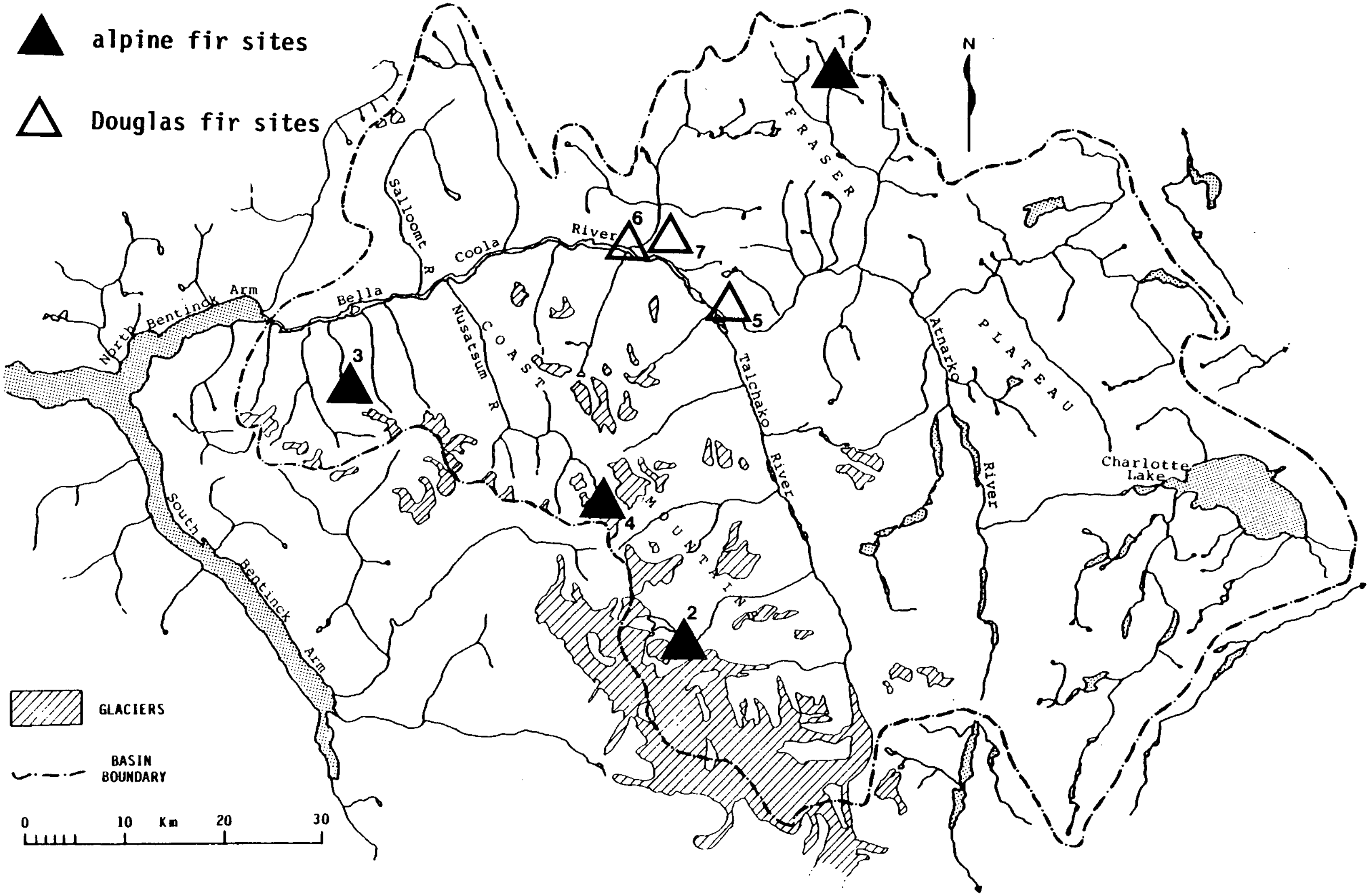


Figure 6.1 Locations of tree sampling sites in Bella Coola basin. Sites outside the basin are shown in figure 2.1. Douglas fir sites are at valley bottom locations while subalpine firs sites are near treeline.

Table 6.1. Summary of Tree Ring-Width Chronologies Used in this Study.

A. This Study

Alpine Fir	# of trees sampled	# of cores sampled	# of trees in chronology	# of cores in chronology	Period A.D.	Signal-to-noise ratio[1]	% autocorr. reduction[2]
Rainbow Range	45	83	9	15	1725-1984	4.38	3%
Ape Lake	34	61	11	20	1740-1984	5.03	6% *
Noohalk Lakes	21	34	12	18	1871-1983	3.25	2%
East Nusatsum	16	21	8	10	1711-1983	2.98	5% *
Douglas Fir							
Stuie (Tweedsmuir Lodge)	25	46	15	18	1579-1984	2.54	2%
Burnt Bridge Terrace	8	16	8	12	1704-1983	3.18	3%
Burnt Bridge Slope	6	12	6	15	1705-1983	1.67	1%
Bull Canyon	15	30	12	16	1650-1983	6.21	4%

B. Other Studies

Williams Lake[3]	--	--	--	44	1420-1944	5.50	2%
Pavilion[4]	--	--	--	20	1480-1965	6.31	3%
Kamloops[4]	--	--	--	20	1420-1965	4.25	4%

1. Calculated as the ratio between the best fit polynomial and high frequency year-to-year variance.
2. % reduction in autocorrelation using autoregressive-integrated-moving-average (ARIMA) model of Cook and Holmes (1985). The * indicates significant improvement from the original series.
3. From Schulman (1956).
4. From Stokes et al. (1973).

an appropriate growth detrending function was facilitated by plotting each series and identifying common patterns of low frequency variance. This procedure maximizes the trends shared at a particular site and reduces first-order autocorrelation. Standardized tree-ring indices were then generated for each sample by dividing observed growth by expected growth (detrending function) and then combining all index series as an unweighted average to form the master site chronology.

In an attempt to assess and remove the stochastic portion of the long-period variance in each chronology, autoregressive modeling procedures developed by Cook and Holmes (1985) were applied. The autoregressive model is calibrated using characteristics of the chronology during the period of observed climate data (1930-1983) and is then applied to the whole series assuming that the form of the persistence during the calibration interval has not changed in the past. When the modeling resulted in significant improvement (*i.e.* reduction in variance due to autocorrelation) the new chronology was used (see table 6.1). Full details of the model are given in Cook and Holmes (1985). In effect all raw data series used in the following analyses have been pre-whitened (*cf.* Graumlich, 1985).

(6.2) Response Characteristics

Although it is assumed initially that moisture and temperature stresses are the most important factors influencing yearly growth, other hydrological and micro-site factors may be significant. Graumlich and Brubaker (1986) have shown that sites near tree line may respond not only to fluctuations in available energy but also to the accumulated depth and duration of snowpack during the early part of the growing season. With increasing snowdepths below average soil temperatures are maintained for a greater proportion of the growing season resulting in stomata closure and decreased carbon assimilation (Tranquillini 1979; Tesky *et al*., 1984).

Several seasonally and annually averaged climatic and hydrologic indices given in Chapter 3 were compared with the response signal contained in the chronologies listed in table 6.1. The strength of correlation between ring-width and individual climate variables was evaluated using partial correlation coefficients, a procedure which holds constant the covariance of other variables and more accurately portrays singular response characteristics. Partial correlation coefficients for winter (Oct.-Apr.), summer (May- Aug.) and annual temperature and precipitation indices (interior stations) are plotted in figure 6.2. In addition, spring snowpack (interior stations) and Bella Coola River summer runoff indices are also correlated with each chronology (figure 6.2d).

The general pattern for Douglas fir trees is positive correlations between ring-width and precipitation (especially winter precipitation) and negative correlations with temperature. The association would suggest that years, particularly winter seasons, characterized by above average precipitation lead to greater moisture availability during the growing season and thus increased biomass production. Above average winter and summer temperatures increase potential evapotranspiration and reduce available moisture leading to negative correlations between growth and summer temperature. Chronologies developed for valley bottom locations in the eastern basin (Stuie, Burnt Bridge) are somewhat anomalous in that correlations with winter and annual temperatures are positive. The more transitional climatic conditions of these sites (warmer winters and higher annual precipitation which falls partly as rain during the winter months) might lead to positive growth responses during warm winter years. Unlike sites further to the east, Douglas firs in the valley-bottom locations

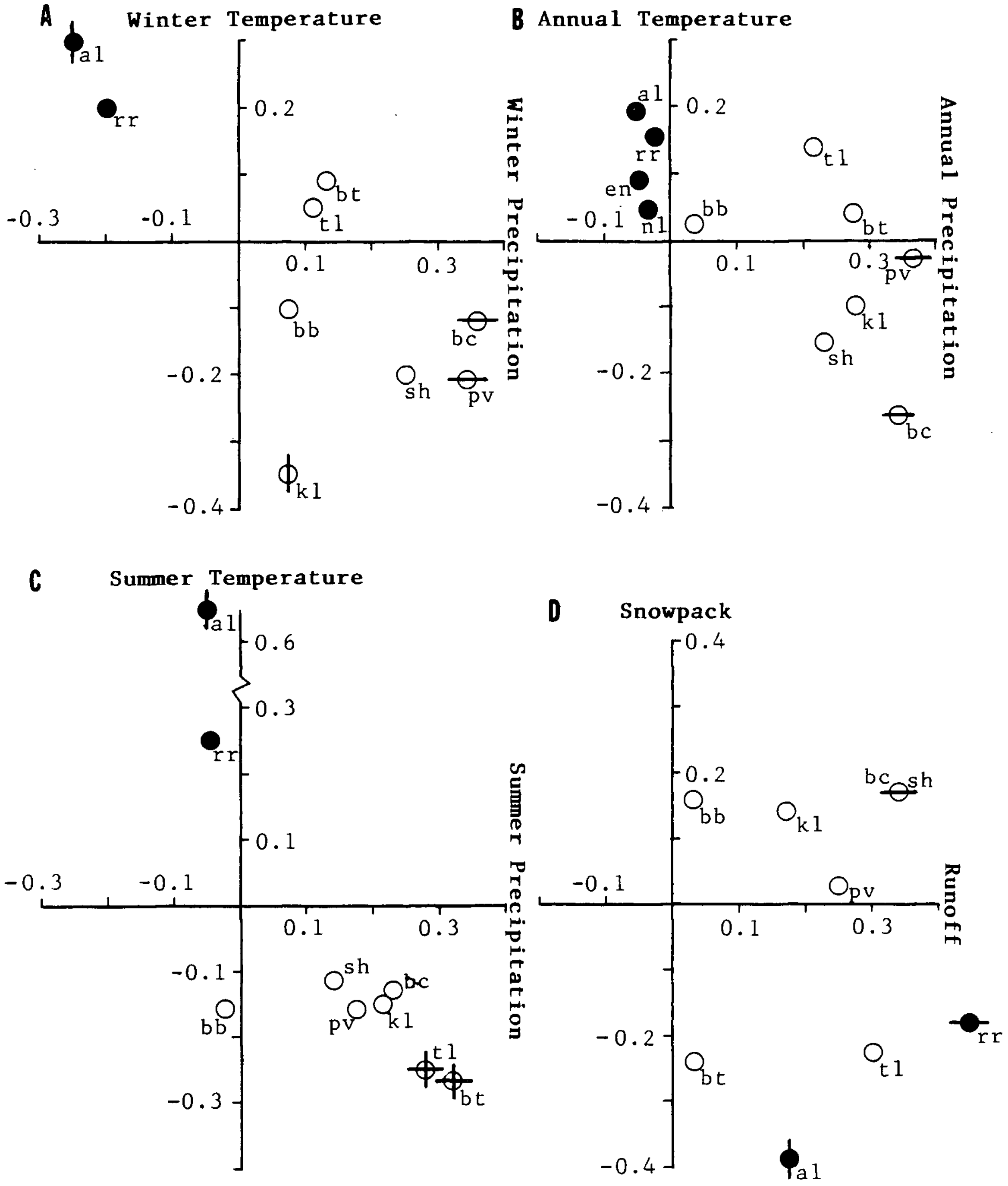


Figure 6.2 Partial correlation coefficients between ring-width for several master tree-ring chronologies and selected hydroclimatic variables. Tree line sites (●) include: Rainbow Range (rr), Ape Lake (al), East Nusatsum (en) and Noohalk Lakes (nl). Interior sites (○) are Pavilion (pv), Kamloops (kl), Williams Lake (sh), Bull Canyon (bc), Stuie-Tweedsmuir Lodge (tl), Burnt Bridge Terrace (bt) and Burnt Bridge Slope (bb). Partial correlation coefficients significant at α = 0.10 are shown by vertical and horizontal lines.

appear to be more responsive to summer precipitation conditions (figure 6.2c).

Subalpine fir generally show an insignificant response to precipitation variability and winter or annual temperatures. However, summer temperature is an important factor positively related to growth particularly for the Ape Lake site (figure 6.2c). Summer runoff indices are positively correlated with all chronologies. There are several possible causes for this association but it is most likely a combination of above average summer temperatures and above average water storage in the soil. Spring snowpack shows a negative correlation with growth in subalpine fir, which is particularly significant at the Ape Lake site. High snowpack years may retard growth at some sites involving mechanisms much like those proposed by Graumlich and Brubaker (1986). Most of the Douglas fir chronologies show a positive relationship with snowpack which is probably also related to a high soil moisture status.

Although partial correlations are generally low the results in figure 6.2 demonstrate that Douglas fir trees, particularly towards the drier interior, are most directly related to annual or winter precipitation variability. Growth at tree line sites, especially farther to the east, is closely associated with summer temperature and winter snowpack conditions.

(6.3) Growth Variations and Climate

Interior Douglas Firs

Three of the seven Douglas fir chronologies in table 6.1 exhibited similar low and high-frequency departures in growth (Bull Canyon, Pavilion, Williams Lake), hence all three were combined as an unweighted average. This produces a stronger regional signal and smooths out between-site anomalies. Since the chronologies were constructed at different times the period 1966-1983 is based on only one chronology, 1945-1965 on two and the pre-1945 interval on all three. Growth variations of Douglas fir between 1900 and 1983 are plotted in figure 6.3a.

Above average growth occurs between 1940 and 1970 whereas negative departures dominate the interval prior to 1940 and after 1970. Shorter periods of near average growth are found between 1900 and 1915 and between 1934 and 1940. Notably persistent trends in growth occur in the post-1960 interval and between 1915 and 1934. The highest year to year variance appears to cluster around the middle of the century commencing in the early 1930s and ending in the late 1950s. Also plotted in figure 6.3a are generalized departures in annual precipitation at Big Creek. These patterns correspond fairly well with growth variations in the averaged Douglas Fir chronology suggesting that available moisture is strongly associated with biomass production in this species. Although there is a strong association between generalized precipitation trends and tree growth in figure 6.3a there are notable extreme departures. These anomalies are important to consider for two reasons: (1) they represent extreme growth variations about a long term mean which may be due to a combination of environmental conditions not indexed by one or two climate variables and (2) these extremes can significantly influence linear statistical models.

To assess the temporal variability of these extremes the high frequency component of the series in figure 6.3a was extracted by first smoothing with a 13-year weighted filter. The observed growth was then divided by the smoothing function for each year to produce the high frequency series plotted in figure 6.3b. The upper and lower threshold lines shown in this figure are set at ±1 standard deviation from the mean. The frequency of crossings above the upper threshold and below the lower threshold were tabulated as runs and tested under the null hypothesis that

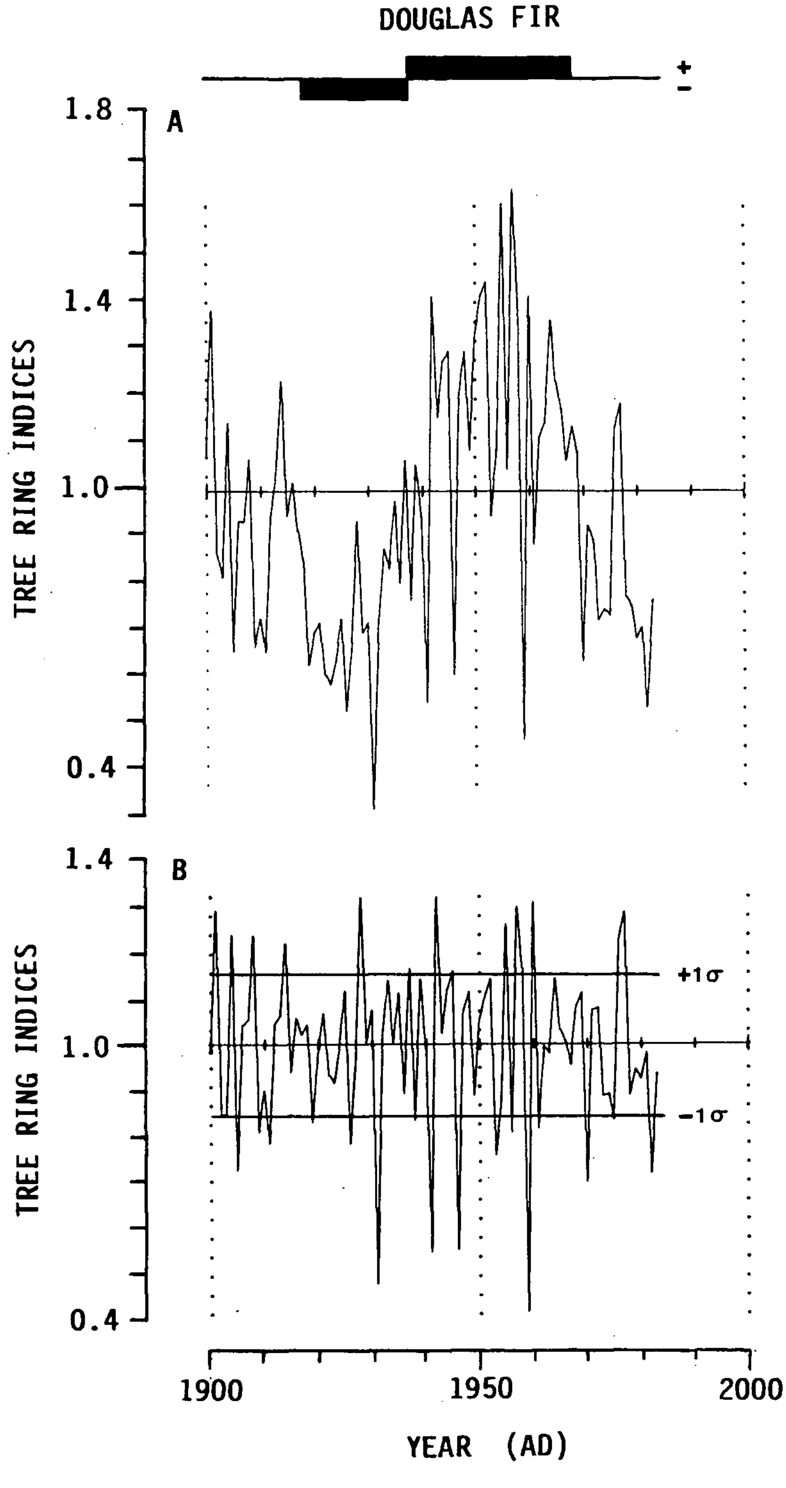


Figure 6.3 Regional Douglas fir chronology derived using three site chronologies. A) Combined low and high frequency tree-growth signals compared with generalized departures in precipitation at Big Creek (above A). B) High frequency tree-growth signal from A. Threshold lines are plus and minus one standard deviation from the mean.

the extremes are randomly distributed. Application of a one-tailed runs test resulted in rejection of the null hypothesis ($z = -2.62, \alpha = 0.10$). Thus, the clustering of extreme positive and/or negative departures during the intervals 1900-1910, 1936-1945 and 1952-1961 is statistically significant.

The higher mean sensitivity of growth during these three periods implies a higher year-to-year variance in factors controlling growth. The response relationships in figure 6.2 suggest winter or annual precipitation as a dominant control of growth at interior sites. Analyses presented in chapter 3 indicated that mean sensitivites in precipitation for most months of the year were highest between 1920 and 1945. A similar correspondence does not exist for the earlier period (1900-1910) but the period 1952 to 1961 also shows a high between year variability in precipitation departures. The implied relationship then is towards a higher mean sensitivity in Douglas fir growth during periods of less persistent departures in precipitation, a characteristic of zonal circulation regimes.

Subalpine Fir

Three of the four subalpine fir chronologies in table 6.1 exhibited similar long-term and high frequency departures as well as higher signal to noise ratios (Rainbow Range, Ape Lake, Noohalk Lakes). In order to maximize the regional signal of growth variations at tree-line the three chronologies were summed and then averaged. A plot of the averaged chronologies for a common interval 1900 to 1983 is given in figure 6.4a.

The most visible trends are average to below average growth after 1950 and above average growth prior to 1950, particularly 1900-1920. The most persistent negative departures occur in the 1970s whereas strongly positive departures occur between 1916 and 1920. The interval 1925 to 1945

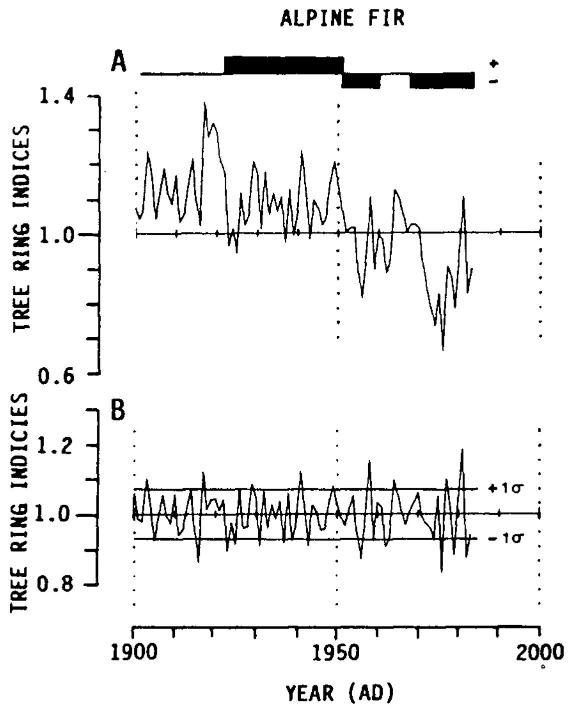


Figure 6.4 Regional subalpine fir chronology derived using three site chronologies. A) Combined low and high frequency tree-growth signals compared with generalized departures in temperature at interior stations (above A). B) High frequency tree-growth signal from A. Threshold lines are plus and minus one standard deviation from the mean.

is characterized by a high year to year variance in growth. The observed variations in growth show some correspondence with the generalized 20th century trends in summer temperature for Interior stations (figure 3.1c). The only notable divergence is above average growth during the first two decades when summer temperatures at Big Creek were mostly average.

As with the Douglas fir chronologies the temporal variability in extreme departures was evaluated (figure 6.4b). Although not as strongly clustered there is a statistically significant grouping of positive and negative departures during the intervals 1925 to 1945, 1955 to 1964 and 1976 to 1983 ($z = -1.79, \alpha = 0.10$). The first two periods overlap with those identified for the Douglas fir chronologies although there is not an exact correspondence.

This similarity would suggest a common factor controlling extreme growth responses. Examination of temperature, precipitation and snowpack records for several of the individual extreme years indicates that snowpack conditions may be important. For example, well above average winter precipitation (and snowpack) in 1957 and 1976 corresponds with vigorous growth in <u>all</u> Douglas firs sampled for this project while at the same time well below average growth was noted for subalpine fir. Summer temperature during these same years was low resulting in well above average water availability for Douglas firs (vigorous growth) and poor growth for subalpine fir due to the low soil and air temperatures at tree line sites.

These qualitative comparisons indicate that the long-period variance of growth variations in Douglas and subalpine fir trees are associated with departures in annual precipitation and summer temperature, respectively. High-frequency departures in growth have a tendency to cluster within the same intervals for both species and are related to periods of high year-to-year variance in climatic departures. Seasonal temperature variations complicate the responses of Douglas fir and spring snowpack conditions influence the early season growth in subalpine firs.

(6.4) Development of Transfer Functions

The similarity of generalized growth departures for both species of trees and significant partial correlation with certain climate variables (figure 6.2) suggest that there may be some basis for developing transfer functions using tree-ring data. Transfer functions are constructed for the period of observed climate in order to incorporate the mutual variance between the dependent (temperature or precipitation) and independent (ring-width) variables. Linear statistical models have been used for this purpose and are applied here to assess the quality of inferences drawn from this form of modeling technique and proxy data. The implication of the preceding analysis is that these linear models will reveal only the averaged trend in climatic departures.

Each chronology in table 6.1 represents site-averaged tree growth. While visual inspection of each series demonstrates a reasonably high spatial and temporal coherence amongst species, precise agreement between all chronologies during all intervals is not present nor is it reasonable to expect such agreement. In order to extract the regional signal common to all series objectively, principal component analysis was undertaken. The principal components (PCs) represent weighted mean departures which emphasize the extreme years and thus should ′track′ climate more precisely (Peters *et al.*, 1981; Josza and Oguss, 1985). Each PC represents a different weighted combination of the input series and they are formulated so that they are mutually independent, or orthogonal.

PCs were calculated separately for each species using the six chronologies considered above. PC scores are based on the entire 334 year tree-ring data set for Douglas fir and 245 year data set for subalpine fir rather than using just the interval for which climate data were available (1930-1983). This procedure incorporates the long-period variance associated with tree growth and allows for estimates of climate variations prior to the instrumental record. A separate analysis revealed that correlations between climate data and PC scores calculated for the shorter interval were only marginally stronger.

Correlations between PC scores and climate data (table 6.2) indicate the following: (1) the first PC for Douglas fir is most highly correlated with winter precipitation (October-April) whereas the second PC is associated with winter temperature and the third with summer precipitation; (2) the first PC for subalpine fir is significantly correlated with summer temperature and the second PC with spring snowpack; and (3) in both cases the first PC accounts for 65% or more of the overall variance in the original data.

Since the evidence points to the significance of annual precipitation in controlling Douglas fir growth the three PC scores for this species were used as independent variables in a stepwise multiple regression analysis against annual (Sept.-Aug.) precipitation. Data for interior stations between 1930 and 1983 were used in the calibration. Selection of the most significant principal components was made using a low inclusion level ($F = 2.0$). Addition of all three components yields a normal multiple-regression function based on the original independent variable set. Preliminary models were constructed using one half of the 1930-1983 data set (randomly selected points) and each model was tested for predictive ability (table 6.3). The entire data set was then used to develop a full model in which significant outliers were tested for. The final model in table 6.3 is the

Table 6.2. Correlation Matrix Between PC Scores and Hydroclimatic Variables

	Precipitation[1] Winter	Precipitation[1] Summer	Temperature[1] Winter	Temperature[1] Summer	Summer[2] Runoff	Spring[1] Snowpack
Douglas fir						
%VR[3]						
PC 1 (68%)	0.38*	0.19	0.21	-0.09	0.18	0.07
PC 2 (21%)	-0.09	0.04	-0.25	0.02	0.15	0.15
PC 3 (11%)	0.10	0.34*	0.14	-0.13	0.06	-0.08
Subalpine fir						
%VR[3]						
PC 1 (65%)	0.19	-0.15	0.09	0.45*	0.18	0.09
PC 2 (26%)	-0.04	0.11	-0.16	-0.21	-0.11	0.39*
PC 3 (9%)	-0.10	0.12	0.03	0.06	0.10	0.10

1. For interior stations.
2. Bella Coola River above Burnt Bridge Creek.
3. - %VR is percent variance reduction
* - indicates significant at the 0.90 confidence level

Table 6.3. Coefficients of Transfer Functions for Annual (Sept.-Aug.) Precipitation Indices.

Principal Component	Preliminary Models 1 (n=25)	2 (n=23)	3 (n=21)	Full Model (n=51)	Final Model (n=48)
1st	0.43	0.38	0.17	0.31	0.30
2nd	-0.16				
3rd		0.34	0.36	0.37	0.38
Intercepts	0.19	-0.21	0.22	0.06	0.04
Summary Statistics:					
R^2	= 0.25	= 0.37	= 0.19	= 0.22	= 0.29
R_a^2	= 0.21	= 0.33	= 0.16	= 0.18	= 0.25
D	= 1.81	= 2.10	= 1.64	= 1.78	= 1.94
F	= 3.93	= 6.20*	= 2.29	= 7.17*	= 9.48*
SE	= 0.75	= 0.67	= 0.72	= 0.71	= 0.62
Confirmatory Statistics:					
RE	= 0.35	= 0.43	= 0.10		
r	= 0.52	= 0.69	= 0.41		

R^2 = coefficient of determination
R_a^2 = adjusted R^2
D = Durbin-Watson d (reject H_o - no significant autocorrelation)
F = F ratio (* significant; α = 0.025)
SE = standard error

RE = reduction of error statistic
r = correlation coefficient

full model minus significant outliers and represents the best calibrated transfer function using the available data.

Explained variance of the dependent variable (annual precipitation) using restricted portions of the data set (preliminary models) varies between 16 and 33 percent. With the exception of the weakest model [3], reduction of error statistics (RE) confirm the reasonable predictive ability of each when tested against the withheld portion of the data set. The first and third principal components are the most significant variables in all but the first model. Using the full data set only 21% of the variance is explained but 3 outliers significantly influence the regression equation. When removed, the proportion of the explained variance increases to 25% (final model). The three estimates which yield the highest explained variance share regression coefficients which do not differ significantly and exhibit the lowest standard errors.

The final model is used to estimate annual precipitation for the interior between 1900 and 1983. Plots of the reconstructed values and filtered series are given in figure 6.5. The trend retains much of the variance in the original growth series. Below average precipitation is inferred between 1900 and 1945, fluctuating but average to above average conditions to 1970 and, with the exception of 1976 and 1977, below average precipitation to 1983. A distinct change to higher year-to-year variance occurs after approximately 1918, a tendency which ends in the mid-1960s.

With only 25% of the overall variance in the dependent variable explained, the final model in table 6.3 and reconstructed September to August precipitation series in figure 6.5 should be considered approximates at best. Extreme departures in growth do not always correspond with extreme climate which produces a high standard error. Analysis of residual variance demonstrates that the poorest fit corresponds with periods of high year-to-

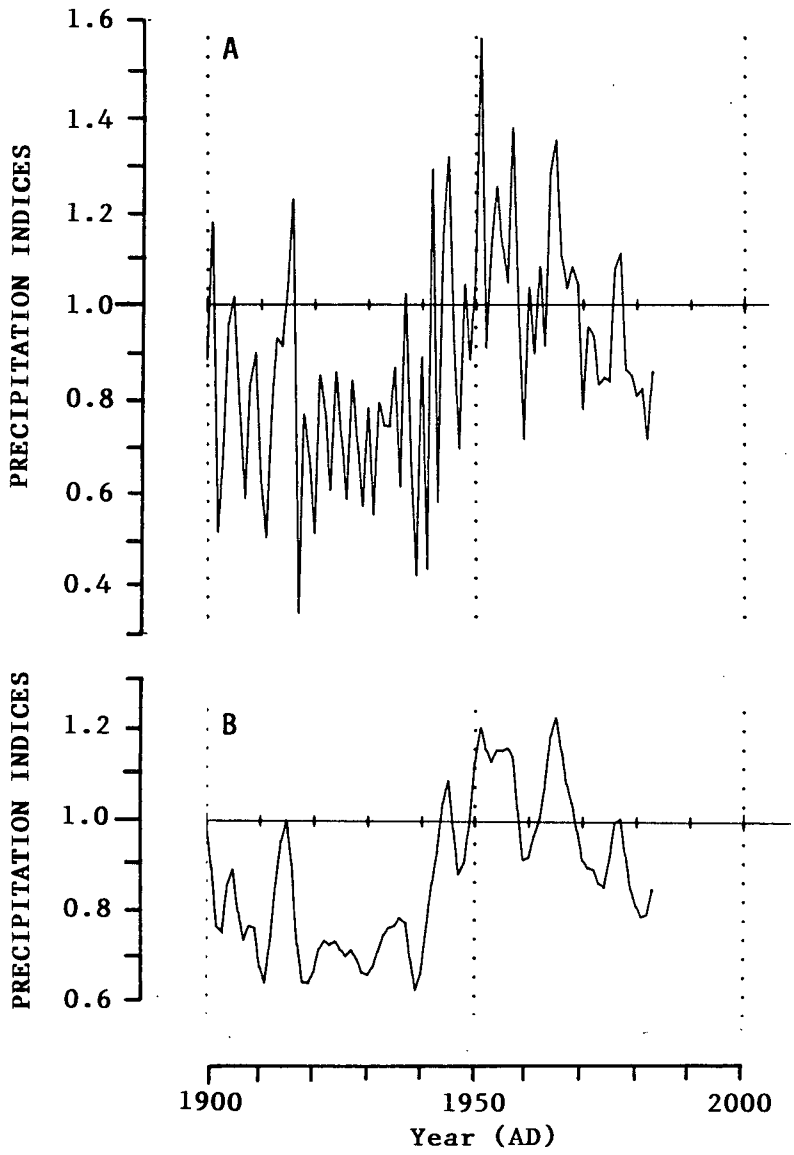


Figure 6.5 Tree-ring derived estimates of annual precipitation using Douglas fir chronologies. Low and high frequency components are shown in A. Low frequency or response of the mean in B is generated by a five-year weighted moving average of the series in A.

year variance in growth (1920-1945); a result which further supports the fact that the mean and not extreme departures are being reconstructed. Interpretations of the sequence in figure 6.5 should be made in recognition of this limitation.

A stepwise regression analysis was also undertaken using the three PCs generated from the subalpine fir chronologies. With the same inclusion level (F = 2.0) only the first principal component entered the equation explaining 45% of the dependent variable (summer temperature for interior stations). Since only one PC was important the trace of this component adequately portrays tree-ring responses and therefore is used directly as an index of temperature variability between 1900 and 1983 (figure 6.6).

The much higher level of explained variance leads to greater confidence about inferences drawn from the series in figure 6.6. The first principal component of growth variations for subalpine fir trees in the Bella Coola basin matches quite well with the overall growth patterns of mountain hemlock in the Cascade Mountains 500 km to the south (*c.f.* Graumlich and Brubaker, 1986). In addition, there is a close correspondence between the PC time series and estimates of July heating degree days reconstructed for Alaska and northwestern Canada (Jacoby *et al.*, 1985). Although there are common secular trends Graumlich and Brubaker (1986) have placed different environmental interpretations on the growth variability using information from other species. While estimates of summer temperature variations using other species were not made in this study it is recognized that the effect on growth patterns of interaction between snowfall accumulation and summer temperature is complex and therefore the interpretations given here remain approximate until more data from a variety of species are acquired.

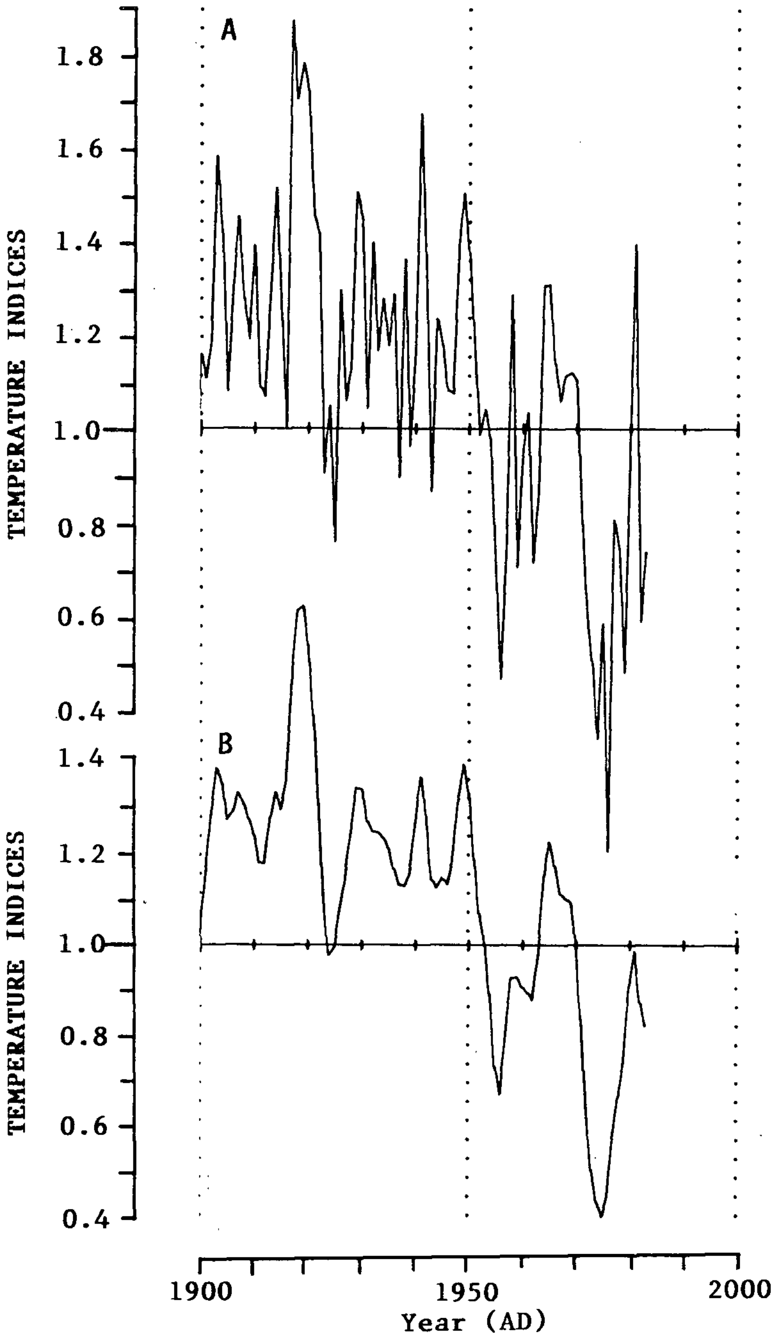


Figure 6.6 Tree-ring derived estimates of summer temperature using subalpine fir. Low and high frequency components are shown in A. Low frequency or response of the mean in B is generated by a five-year weighted moving average of the series in A.

(6.5) Inferences About 20th Century Hydrology from Tree-Rings

The temperature and precipitation trends inferred from tree-rings in the previous section are tested further using 20th century variations in observed and modeled runoff of the Bella Coola River. Two separate comparisons are made with the tree-ring data: summer runoff and autumn storm-generated runoff.

Summer Runoff

Summer runoff models for the Bella Coola River were evaluated in chapter 3 and are used here to estimate variations in summer runoff for the period 1904 to 1983 using measured temperature and precipitation data as input parameters. The final model presented in table 3.5, which includes winter precipitation and summer temperature as independent variables, accounts for 55% of explained variance in summer (May - August) runoff. This model was shown to predict reasonably well independent runoff data from Salloomt and Nusatsum Rivers.

Predictions from empirically-based models are assuredly valid only if the new input variables fall within the range of the original data set used for model calibration. Extrapolation occurs when the independent variables lie beyond this range. A prediction may be classified as an extrapolation even if each input variable is within the calibration range but they have not been observed in the combination in which they are entered into the reconstruction. These extrapolated values can be isolated by constructing the minimum covering ellipsoid (MCE) which circumscribes the observed values (Cook and Weisberg, 1978). The MCE was derived from a program given in Graumlich (1985) and applied using temperature and precipitation from the coastal and interior regions. Since indices based on multiple station data could be calculated only for the post 1930 interval (limited by the length of the shortest record) an MCE was constructed for the 1904-1930 interval separately using the more limited data base. Actual and estimated summer runoff for the Bella Coola River is given in figure 6.7a.

Prior to 1918 estimated summer runoff was well below average although three of the seasonal estimates from this period are extrapolated values. Following 1918 average to above average summer runoff characterized by higher year to year variance dominates until the mid-1930s after which there was more average summer runoff, with the exception of 1939 and 1940. Extrapolations are also noted for the years 1922 and 1934. As with most empirical regression techniques the extreme runoff years are poorly modeled (*e.g.* 1958, 1973). Thus, extremes in summer runoff are apt not to be well represented in the reconstructed record. Differences between actual and estimated runoff for extreme runoff years during the calibration interval indicate predictions of runoff vary between -15 and +30%. However, the series does indicate periods in which the potential for high magnitude runoff is highest and appears to reflect the various within-period variances of the independent variables.

For comparison, a runoff index was constructed using the tree-ring derived estimates of precipitation and temperature. Since observed precipitation accounts for 75% of the *explained* variance in summer runoff (figure 6.7a), a weighting ratio of 3:1 was adopted for averaging the precipitation series derived from Douglas fir responses (figure 6.5a) with the temperature series inferred from subalpine fir (figure 6.6a). The runoff index is plotted in figure 6.7b.

Although some similarities between the two series in figure 6.7 exist, the general correspondence is poor. Above average runoff in the late 1950s and 1960s can be inferred from the tree-ring reconstructions which is also documented in the observed runoff series. However, below average

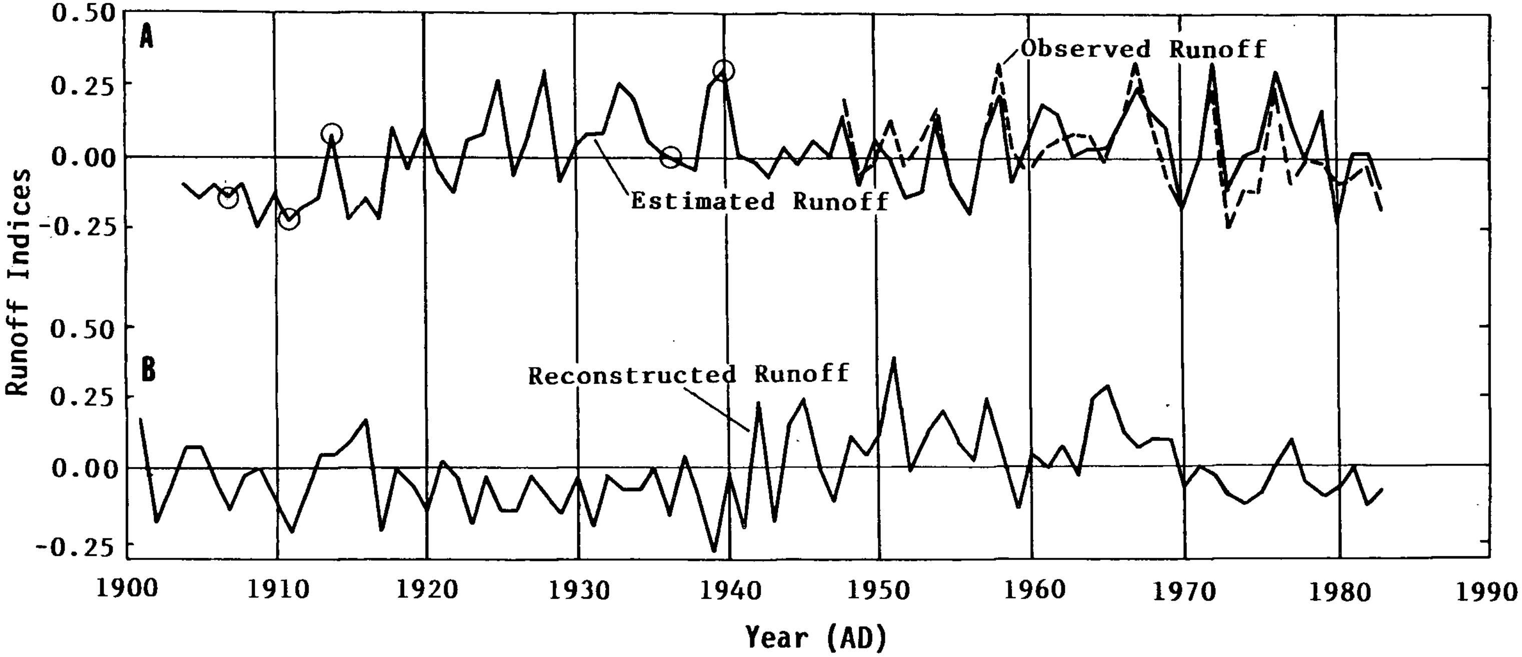


Figure 6.7 Observed, estimated and reconstructed summer runoff indices for Bella Coola River above Burnt Bridge Creek. A) Observed runoff (1948-1983 - solid line) is used to construct the equation for generating the estimated discharge (dashed line). Circles show extrapolated values. (B) Reconstructed summer runoff using precipitation and temperature transfer functions derived from tree-rings.

inferred runoff during the 1970s and between 1920 and 1940 does not appear in the observed record or in the indirectly estimated series of figure 6.7a. In addition, average runoff conditions would be assigned to the first two decades of this century based on the tree-ring derived index when it appears summer runoff was below average.

Part of this discrepancy is based in methodology and part is due to response characteristics of the trees. Modeling of the precipitation index using tree-data results in a reduction of the importance of extreme departures in the original record and hence, average departures only are documented in that series. This bias extends back to the original detrending functions used to standardize each tree specimen. Detrending and chronology development are based on the entire ring series and both are oriented towards major long-term departures. Since runoff is related to spatial as well as temporal variations in snowcover and runoff-generating mechanisms during each year the regionally-averaged climate variables remain imperfect indices and the tree-ring estimates even more so.

In addition to methodological difficulties, the trees themselves appear to be associated with strongly persistent trends (*cf.* figure 6.3) which can only partly be explained by climate fluctuations. When both species show strong positive departures in growth (*e.g.* 1960s) the system also appears to respond in a predictable manner (*e.g.* high summer runoff) but mixed responses, such as average or opposite growth trends, do not always relate to specific system effects of a predictable nature. Additional variance related to site factors or changes in temperature and precipitation over the long-term probably influence the growth patterns substantially. The modeling techniques adopted here, which are also customarily used be others, are only basic steps towards understanding a more complex interaction between tree growth and climate.

Rainstorm and Snowmelt Flooding

Analysis of climatic departures in the post-1945 period (chapter 3.0) indicated that the interval of above average winter precipitation and higher frequency of autumn and spring floods (1958-1976) was also associated with summers characterized by positive temperature anomalies. Synoptic circulation regimes which yield increased winter season precipitation and thus above average spring snowpacks are also associated with a greater potential for high-magnitude autumn storms. Above-average summer temperatures can have two effects: (1) positive, early-season temperatures (May-June) control snowmelt rates and the potential for spring flooding and (2) late-season temperatures (July-August) control ice melt rates which can contribute significantly to rainfall storm-runoff in late August and early September. Therefore, the intraannual persistence of above average winter precipitation and summer temperatures appears to be associated with an increased potential for more frequent and higher magnitude flooding.

An assessment of the 20th century variability of these intraannual trends was made by combining summer temperature and winter precipitation data from the Bella Coola climate station and contouring positive climatic departures. The contoured surface in figure 6.8 demonstrates that the most persistent winter/summer departures occurred after 1958. This same intervals is characterized by much higher frequencies of flood runoff during both the autumn and spring (recall that pre-1948 flows were not measured systematically so the threshold for considering an event as a flood is rather high). The high summer temperatures and wet winters are particularly evident between 1964 and 1968, an interval with four of the largest spring and fall flood events. Prior to 1933 the intra- and

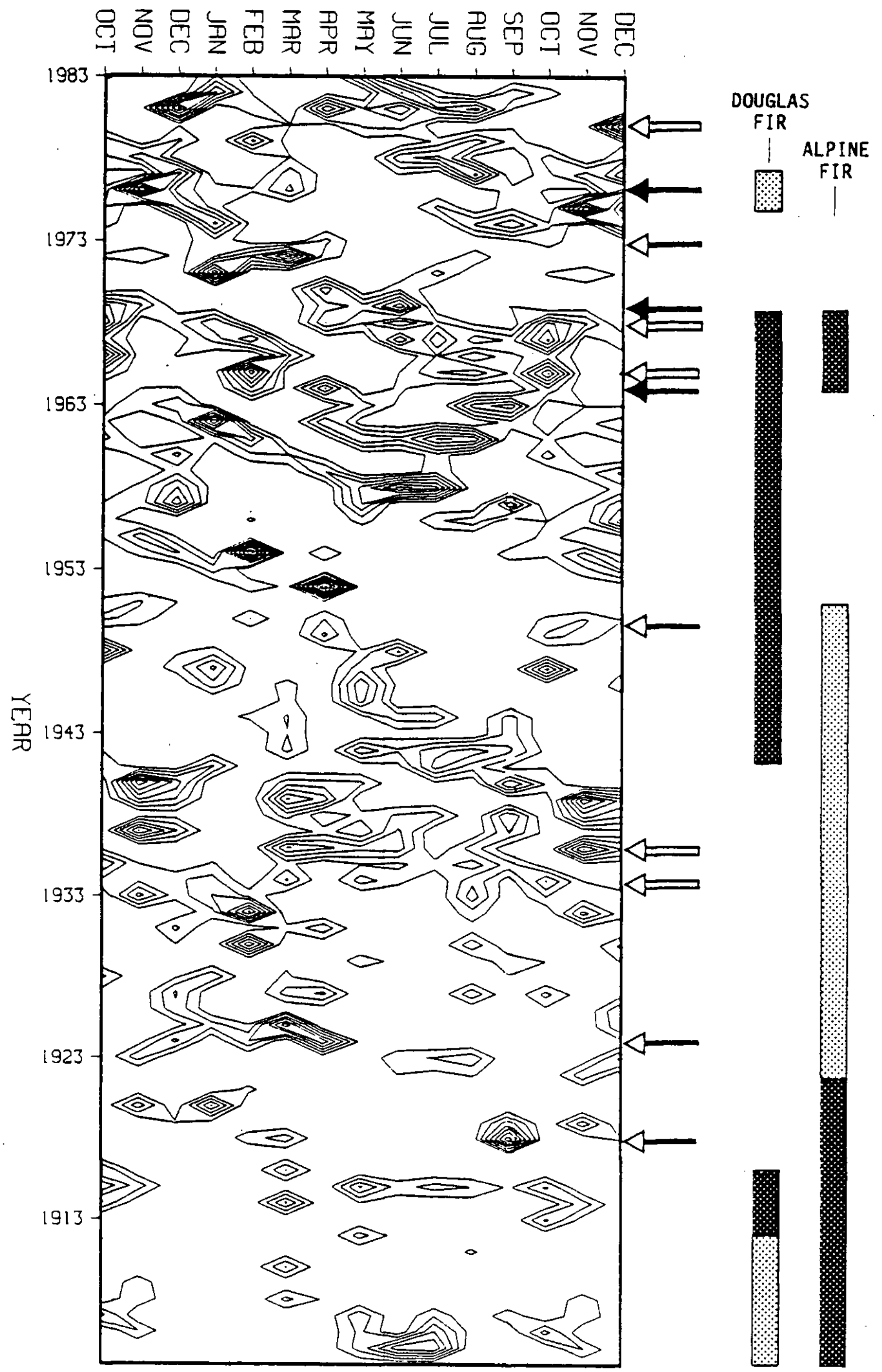


Figure 6.8 Inter- and intraannual persistence of wet winters and warm summers at Bella Coola. Winter months (September to April) are contoured positive departures in monthly precipitation (>0.50 σ). Summer months (May to August) are positive departures in monthly temperature (>0.50 σ). Standardized departures were calculated using the 1904-1983 mean for each month. Contouring interval is 0.50 σ. Horizontal lines are associated with measured and reported spring snowmelt (←), autumn rainfall (◁—) and autumn rain-on-snow floods (⇐). Vertical boxes on the right indicate intervals of persistent positive departures (dark shading is strongly postive and weak shading is marginally positive) in alpine fir and Douglas fir tree-growth.

interannual persistence of positive departures is lower and is reflected in the reduced frequency of reported (pre-instrumental) flooding along the river (figure 6.8).

It appears that the intraannual persistence of climatic anomalies which maximize water yield from the basin are closely related to the potential for flooding. Well above average summer temperatures or winter precipitation alone do not necessarily lead to significant water yield or a high potential for flooding. Prior to the 1930s persistent climatic anomalies were rare, flood occurrences low and thus the potential for important geomorphological changes reduced. Although the pre-1958 period is characterized by less persistent climatic trends the years between 1933 and 1940 exhibit greater continuity in climate departures and more significant occurrences of flood events. This corresponds with a peak in global warming (McGuirk, 1982). In terms of 20th century climate the interval between 1958 and present appears to be the most significant in terms of interannual variability and intraannual persistence of flood generating conditions.

The tree-ring record again provides some corroborating evidence for these trends (figure 6.8). The comparison is not exact because only a single climate station is shown (Bella Coola) but the interval of greatest flood frequency corresponds with vigorous growth in both Douglas fir and subalpine fir. There is also a significant correspondence between late winter and spring flood years and extreme growth responses in the two species.

The patterns revealed in figures 6.7a and 6.8 suggest that the flood frequency curves presented in chapter 3.0 would probably be shifted downwards if the same family of curves could be constructed for the interval 1904-1948. This is due to the lower frequency of reported and inferred floods of moderate and high-magnitude for the interval prior to 1948.

(6.6) Summary and Conclusions

The annual record of growth in Douglas fir and subalpine fir reveal the following. Douglas fir growth shows a strong regional signal related to moisture availability and can be used as an index of low frequency variance in annual precipitation representative of the interior or eastern portion of the Bella Coola basin. Transfer functions cannot be constructed with strong statistical confidence because of the high standard error involved in calibration. Subalpine fir growth is related more strongly to summer temperature variations but is also influenced by the degree of snowcover. A better statistical association exists between subalpine fir and temperature than between Douglas fir and precipitation. There are positive responses in growth of both species during years characterized by strong intraannual persistence of warm/wet conditions. Extreme growth responses tend to cluster and are shown here to be associated with the intraannual persistence of extreme climate. The combined records of inferred temperature and precipitation do not provide a good index of runoff variations over the 20th century.

The techniques employed here are in wide use today and the evidence indicates that both extreme responses and a longer term persistence are characteristic of these species. As a result, working backward form the record to the hydrometeorologcial antecedents becomes difficult. Inferences about climate near the Bella Coola basin using this form of proxy data can provide qualitative estimates of low-frequency variations in temperature and precipitation. Evidence from other studies indicates that more than one species needs to be considered if more precise interpretations of temperature variability are to be made.

CHAPTER VII
Inferences About Late Neoglacial Environments

(7.0) Introduction

Several of the potentially useful methods for assessing environmental change are extended here for the purpose of identifying Late Neoglacial environments of the Bella Coola basin. Specifically, biogeophysical data are considered as major indirect data sources. These include paleo-glacial evidence, tree-ring data, lake sedimentation rates and floodplain development. In order to utilize these data it is necessary to consider further the response characteristics of each attribute and potential errors involved with the application of quantitative or semi-quantitative statistical models. These sources for error are apt to have an important impact on the quality of inferences made. Hence, the chapter will demonstrate the resolution of reconstruction methods on the relatively short time scale of only a few centuries.

(7.1) Inferences of Environmental Change from Former Glacier Extent

Glacier mass balance and recent glacier front fluctuations indicate that the extent of glaciers represent prevailing temperature and precipitation regimes and, therefore, are potentially useful indicators of environmental change during the late Neoglacial period. Since direct measurement of glacier mass balance in the past is not possible, several indirect measures are evaluated here using glaciological and geomorphological evidence preserved in the Bella Coola River basin. All of these methods are associated with estimating the altitude of former and present climatic snowlines, the difference of which is generally considered to be representative of climatic change. Difficulties that arise concern errors of estimating elevations of contemporary and former snowlines and in determining the relevant climatic controls.

Glacial equilibrium line altitude (ELA), or the mean elevation of the zone which separates the accumulation and ablation areas on a glacier, represents the local climatic snowline and can be estimated easily for both contemporary and former glacier positions (Bradley, 1975). Persistent trends in temperature and precipitation are known to affect the mean position of the equilibrium line (Dorrer and Wendler, 1976; Porter, 1977; Whillans, 1981) and the spatial distribution of former ELAs can be used to infer the magnitude and sources of precipitation (Miller *et al.* 1975; Leonard, 1984).

In addition to ELA estimates, other methods for determining the position of the climatic snowline during glacial maxima can be tested. These include: (1) the altitude of cirque floors, (2) the glaciation threshold and (3) the altitude of lateral moraine deposits. Prior to a discussion of these measures and sources of error the climatic implications of a changing snowline are considered.

Snowline Fluctuations and Climate Change

ELA data from Sentinel, Place and Bridge Glaciers were used to estimate annual changes in the climatic snowline for a transect across the Coast Mountains of southwestern British Columbia. Summer temperature and winter precipitation were noted in chapter 4 as important factors influencing the glacier net mass balance. ELA data were used as independent variables against winter precipitation and summer temperature indices in order to construct a set of linear models relating climate and snowline fluctuations. Although the true functional form of these relationships has been reversed (*i.e.* ELA is actually dependent on climatic variations) the resulting transfer functions provide a means for assessing the magnitude of departure in temperature and precipitation. Linear statistical models appear to perform as well as certain physically based models (*cf.* Tangborn, 1980; Kuhn, 1981). Plots of relative ELAs versus temperature and precipitation indices for the three glaciers are given in figure 7.1.

In all three cases a significant relationship exists between ELA and winter precipitation while only at Bridge Glacier does summer temperature show a significant association with equilibrium line altitude. Hence, during low snowpack years with above average summer temperature ELAs are significantly higher. Sentinel Glacier is located in a zone of extremely high annual snowfall and is the only glacier to exhibit a persistent positive mass balance for most of the years in which measurements were made (1965-present). Summer temperature indices seem to explain only a minor amount of the variance related to fluctuations of the ELA. Towards the drier eastern side of the Coast Mountains where winter precipitation is 20-80% lower and summer cloud cover is less persistent, summer temperatures appear to have a greater influence. For a 100 m change in ELA a 20% change in winter precipitation occurs at Sentinel Glacier while only 12 and 11% differences in winter precipitation are required at Bridge and Place Glaciers respectively. The apparent higher sensitivity at the latter two sites may be related to two factors: (1) significance of extreme departures in summer temperature and (2) glacier hypsometry.

When ELA data from summers characterized by extreme positive and negative departures in temperature are removed, the sensitivity of the ELA to changes in precipitation declines to 22% for Bridge and 13% for Place Glacier. Unlike Sentinel Glacier, which exhibits a fairly uniform hypsometry through the zone in which the equilibrium line fluctuates (1600-2080 m), Bridge and Place Glacier ELA zones are characterized by a fairly

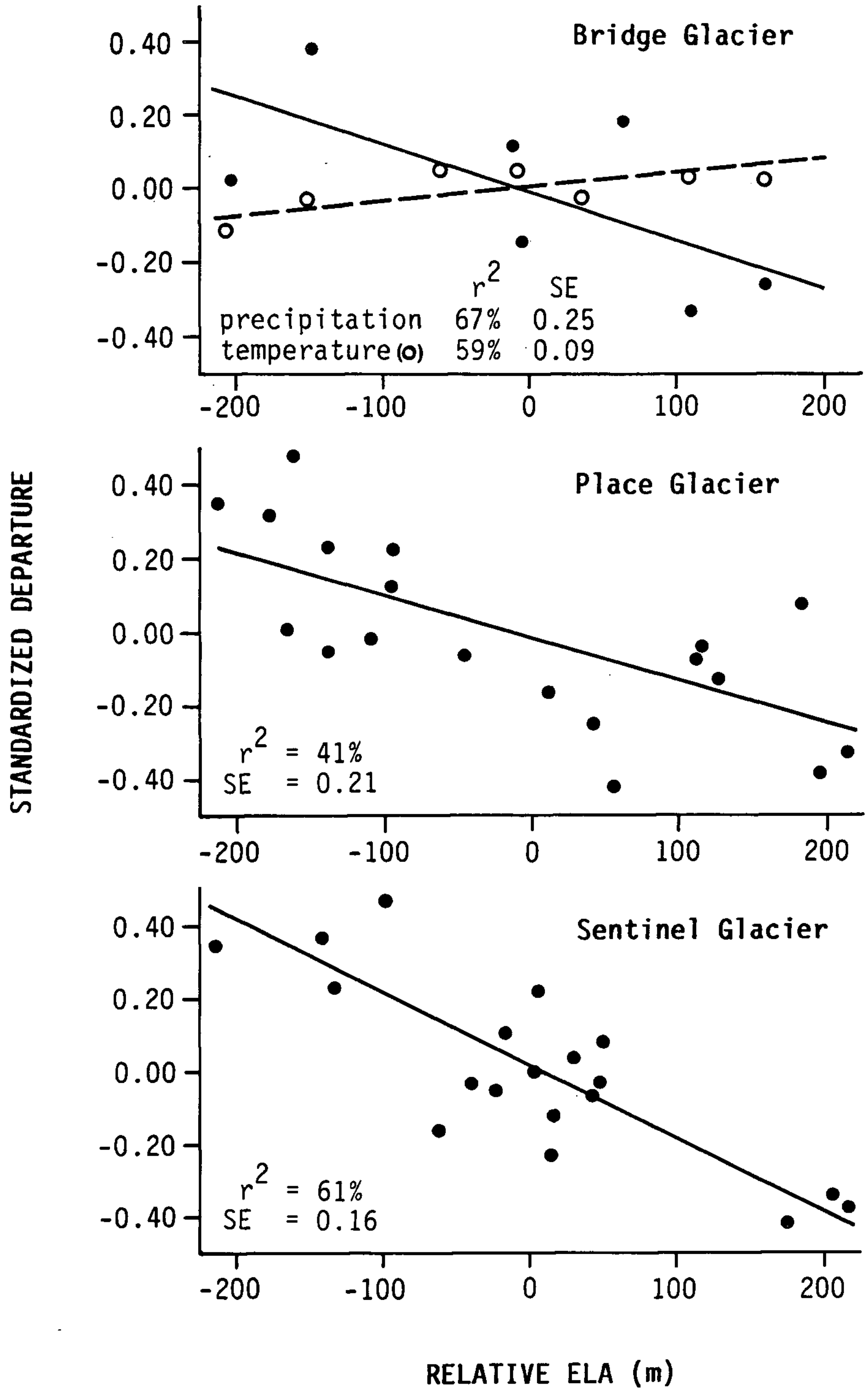


Figure 7.1 Relation between relative ELA on three glaciers in southwestern British Columbia and winter precipitation indices (● and solid line) / summer temperature indices (○ and dashed line). Significant relationships only are shown. Climate indices are standardized departures from grouped station means (chapter 3).

low gradient icefield and a steep icefall, respectively. Hence, the position of the ELA (calculated by identifying the area of accumulation as a proportion of the entire glacier area) is more sensitive to a given change in precipitation on Place Glacier (ice-fall) than Bridge Glacier (ice-field). This would suggest that the Bridge Glacier relationship is "under-sensitive" and the Place Glacier function is "over-sensitive".

Given the stability of the relationship at Sentinel Glacier, and the fact that the slope of the linear function falls in between the two estimates derived from Place and Bridge, a value of 20% is adopted as an estimate of the change in winter precipitation required for a 100 m shift in the ELA. During summers characterized by extreme temperature departures and for glaciers in which the accumulation/ablation zone boundary occurs in a steep ice-fall region this value is likely to be larger - *i.e.* more than a 20% change in precipitation is required. For an average winter precipitation of 1200-1500 mm (*eg.* Garibaldi) a 20% change equates to a precipitation increase or decrease of between 240 and 300 mm.

The difficulties encountered in this analysis suggest that glaciers with odd hypsometries (*cf.* Bridge and Place Glaciers), particularly in the zone within which the ELA fluctuates, should not be used for reconstructions. Glacier aspect is also likely to be an important factor in determining site-specific fluctuations of the ELA. For a consistent basinwide or regional indication, presumably only one aspect should be used.

Estimates of the Contemporary Climatic Snowline

Since mass balance measurements have not been made in the Bella Coola basin, several indirect lines of evidence must be used to derive estimates of changes in the climatic snowline. Three methods are used here to determine the contemporary snowline: (1) mapping of the late ablation season snowline from vertical and oblique air photographs, (2) using the height of the mean annual $0^{o}C$ isotherm and (3) identifying the ELA on several contemporary glaciers using an inferred value of the accumulation area ratio (AAR). Airphoto coverage of the Bella Coola basin for the late ablation season (Sept.-Oct.) is limited, but government photography taken in September 1944 and 1961, August 1979, and oblique photos taken in September and October 1984 provide some indication. Although there is substantial spatial variability, east and north facing slopes of moderate steepness yield a mean estimate of 1870 ± 75 m. The mean value declines to less than 1800 m on the windward side of the Coast Mountains.

The contemporary climatic snowline can also be estimated by the height of the $0^{o}C$ isotherm averaged throughout the accumulation and ablation seasons. The nearest station with upper atmospheric measurements is Port Hardy on the north tip of Vancouver Island, 210 km to the southwest (figure 2.1). The 30 year annual mean (1951-1980) height of the $0^{o}C$ isotherm is 1812 m. It was anticipated that the annual $0^{o}C$ isotherm would be biased downwards if winter temperatures were substantially below zero. However, mean daily surface temperatures during the winter months at most central coast stations are above zero. The snowline estimate from Port Hardy is a considerably higher than the value of 1550 m obtained using the moist adiabatic lapse rate of $0.6^{o}C/100$ m and station elevations from the western Chilcotin. The uncertainty of the annual average lapse rate suggests that the former estimate is more representative.

The ELA of contemporary glaciers can be estimated by identifying the elevation of a line which places a given proportion of the glacier in the accumulation area. An accumulation area ratio (AAR) of 0.65 has been used as an initial estimate (Meir and Post, 1962; Miller *et al.*, 1975;

Meierding, 1982; Hawkins, 1985), but as Leonard (1984) indicates the hypsometry of the glacier surface can have an important influence on ELA location. AARs on icefield-outlet type glaciers tend to be higher because of the increased proportional area at higher elevations while piedmont glaciers have a greater proportion of ice in the ablation area. The hypsometry of Bridge, Place and Sentinel glaciers yield period-of-record average AARs of 0.50, 0.29 and 0.72, respectively. In recognition of these limitations two groups of glaciers were selected for measurement. The analysis presented in Appendix F.1 yields a mean ELA estimate of 1836 ± 206 m for glaciers of regular hypsometry in the eastern catchment.

In summary the three methods for estimating the modern climatic snowline for the eastern portion of the catchment yield relatively similar results (1870 ± 75 m, 1812 ± 185 m and 1836 ± 206 m), although considerable spatial and temporal variance is involved depending on the technique and sampling strategy.

Estimates of Former Snowlines and Inferences of Hydrologic Change

Several methods were used to determine the elevation of the climatic snowline in the Bella Coola basin during the late Neoglacial maximum. These include the glaciation limit, cirque floor altitudes, the elevation of the highest point on lateral moraine deposits, and reconstruction of paleo-ELAs during former glacier advances (*cf*. Meierding, 1982; Leonard, 1984; Sutherland, 1984). A description of the sampling design and errors associated with each method is given in Appendix F.2 and the results from the analysis are illustrated in figure 7.2.

Ostrem (1966) proposed that the average elevation between the summit of the lowest mountain with a glacier and the highest peak capable of snow (ice) accumulation without a glacier in the region (*i.e.* glaciation limit)

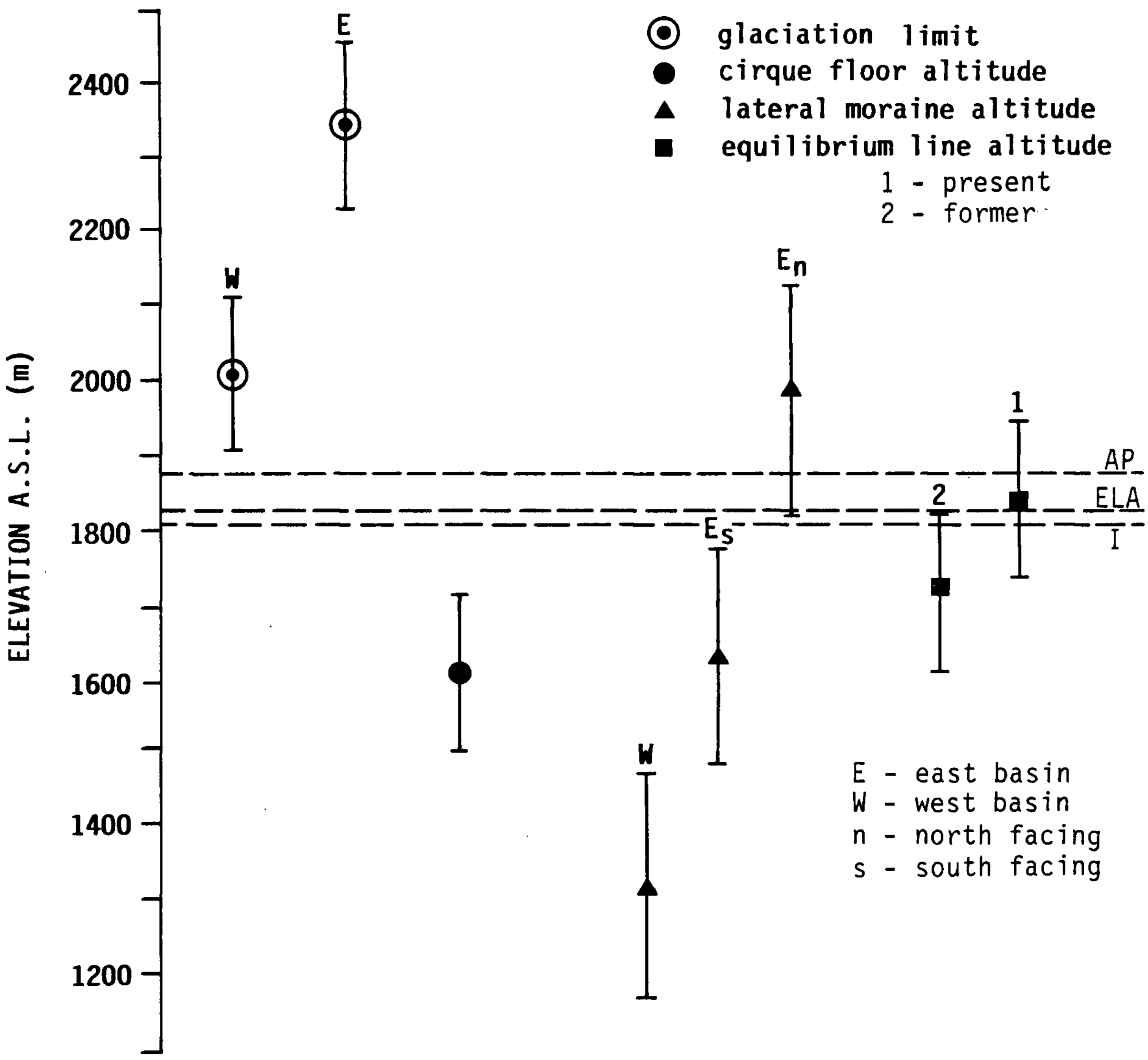


Figure 7.2 Comparison of estimates for contemporary snowline (horizontal dashed lines) and paleoclimatic snowline in the Bella Coola basin. Modern snowline estimates are from air photographs (AP), equilibrium line altitudes (ELA) and isotherm elevation (I). Points are mean estimates of former snowline made from morphological evidence (see text for discussion). Vertical lines are one standard error about the mean. Measurements were made in the east basin (E), west basin (W) and in north (n) and south (s) facing catchments.

was climatically controlled. Glaciation limit values were derived for the Bella Coola area and compared with the small scale regional map for southwestern British Columbia (Ostrem, 1966). Values plotted in figure 7.2 range from 2000 m in the western most basin to 2340 m in the lee of the Coast Mountains, somewhat higher than those suggested by Ostrem. It has been documented that the glaciation limit is typically 200 to 300 m higher than the inferred climatic snowline (Miller *et al.*, 1975; Porter, 1977). Hence, the strong east-west gradient revealed here and the subjectivity involved in computing the snowline index result in a large margin of error.

Cirque floor altitudes are thought to represent the lowest level of glacial erosion during maximum ice advance. All cirques with characteristics indicative of recent ice-occupation and subsequent withdrawal (*e.g.* fresh trim line, little or no soil development, lichen free debris) in the eastern Bella Coola basin were sampled (n = 25). Aspects between 270° and 90° only were considered. An estimate of the mean cirque floor elevation is 1610 ± 110 m (figure 7.2). Since it is not possible to ensure that cirques formed during the late Pleistocene are excluded from the sample, the mean cirque floor altitude may be biased towards a value lower than the actual Little Ice Age climatic snowline.

The maximum altitude of lateral moraine deposits is known to coincide with the boundary between accumulation and ablation zones of the glacier system and is thought to be an approximation of the climatic snowline (Meierding, 1982). From a total sample size of 36 sites, the maximum altitude of lateral moraine deposits varies from 1320 ± 150 m in the west to 1980 ± 100 m in south-facing basins of the eastern catchment (figure 7.2). North facing catchments in the eastern basin yield the most reasonable estimate of 1620 ± 110 m.

Estimates of former ELAs were made for the same set of glaciers used to estimate contemporary equilibrium line heights. The same steady-state AAR values were also used for each glacier. Moraines and erosion trim lines mark former glacier boundaries and were used for contouring paleo-glacier surfaces. From these analyses a mean ELA of 1733 ± 75 m was determined (figure 7.2). This is 103 m lower than the average contemporary ELA - a significant difference when tested at a significance level of 0.10 ($t = 2.16$).

The estimates of former ELA indicate a high spatial variability in the climatic snowline. The geographically and physically consistent method used for assessing ELA depression during the Little Ice Age suggests a decline of approximately 100 m. However, even in an attempt to control for the strong east-west climate gradient and the influence of basin aspect, the change may have been as much as 250 m (figure 7.2). Results given in Kuhn (1981) and Tangborn (1980) would indicate that the former value is the more realistic estimate of possible snowline depression between the contemporary and Little Ice Age environments in the Bella Coola basin.

Using the relationships developed for glaciers in southwestern British Columbia the climatic implication is an increase in winter precipitation during the Little Ice Age of around 20% given that precipitation is the only major factor controlling snowline variations. Concurrent decreases in summer temperature would reduce these estimates but the evidence indicates fluctuations in summer temperature, at least for the moister southwest region, are not as important. A 0.6 C° change in mean summer temperature only (1965-1984 average) would have a comparable effect on snowline elevation for sites in the drier interior. However, temperatures appear to be less important in the fluctuations of mass balance and glacier snout changes at most sites in southwestern British Columbia (see figure 7.1).

Standard errors in the climate-snowline relationships and in the determination of former snowlines are sufficiently high to make these conclusions tentative at best. In addition, there is some risk in assuming that the short-term climate-ELA relationship is applicable to changes in the mean or steady state characteristics of the system. However, the methodologies do provide some indication of the direction and magnitude of departure and are in rough agreement with those given in Smith and Budd (1981) and Tangborn (1980).

Little Ice Age Glacial Chronology

Glacier fluctuations in the basin were examined in order to document responses of glaciers to changing hydrological conditions over the last several centuries and to provide additional evidence for the timing of these changes. In total 17 glaciers were investigated ranging in size from small cirque glaciers (< 2.0 km^2) to major outlet valley glaciers (> 35km^2). Dating methods were primarily based on dendrochronology with lichenometry and soil development used as additional indicators to verify tree-ring derived ages. Details of site selection, sampling locations and methods are given in Appendix F.3

A major limitation to accurate tree-ring derived age estimates of moraines is associated with the colonization period, or ecesis interval. This is the elapsed time between stabilization of the moraine surface and establishment of tree seedlings. To document the probable colonization period, aerial photographs taken at different times between 1948 and 1982 were used to delimit recently exposed ice-marginal areas at several sites. A ground survey for the frequency and size of tree-seedlings in these areas was then conducted to assess the lag time. Except where the material comprising moraines was unusually coarse, the ecesis interval at these sites averaged between 15 and 20 years (Appendix F.3). Hence, for most sites an ecesis value of 20 years was used. Where moraine materials were unusually coarse, a value of 30 years was assumed. Additional sources of error are related to sampling methods which are discussed in Appendix F.3. An average estimate of error associated with the dendro-derived ages is ± 10 years but it may be larger at higher elevation sites, where very blocky debris is characteristic of the moraine and where valley winds are persistently strong.

In most instances, a single terminal or lateral moraine was the dominant depositional feature. Recessional moraines are generally not very well developed or are entirely absent, indicative of rapid and uniform retreat from Little Ice Age maxima. The frequency versus age distribution of terminal and recessional moraines from all 17 sites investigated hence is illustrated in figure 7.3a. The majority of terminal moraines were colonized between 1860 and 1895 while most recessional moraines formed after 1885. The dating methods adopted here indicate that no moraines formed prior to the first half of the 19th century. The comparatively limited lichen growth on the older surfaces and the absence of soil development on all but the lowest elevation sites, would support this observation.

Two sets of recessional moraines appear to have formed during the overall retreat of moderate and large sized glaciers: one set in the first decade of the 20th century, and a second set between 1930 and 1940. It is most likely that these moraines are indicative of brief pauses in ice retreat caused by a return to wetter and/or cooler climates. Analyses by McGuirk (1982) and Karanka (1986) indicate that cooler and wetter

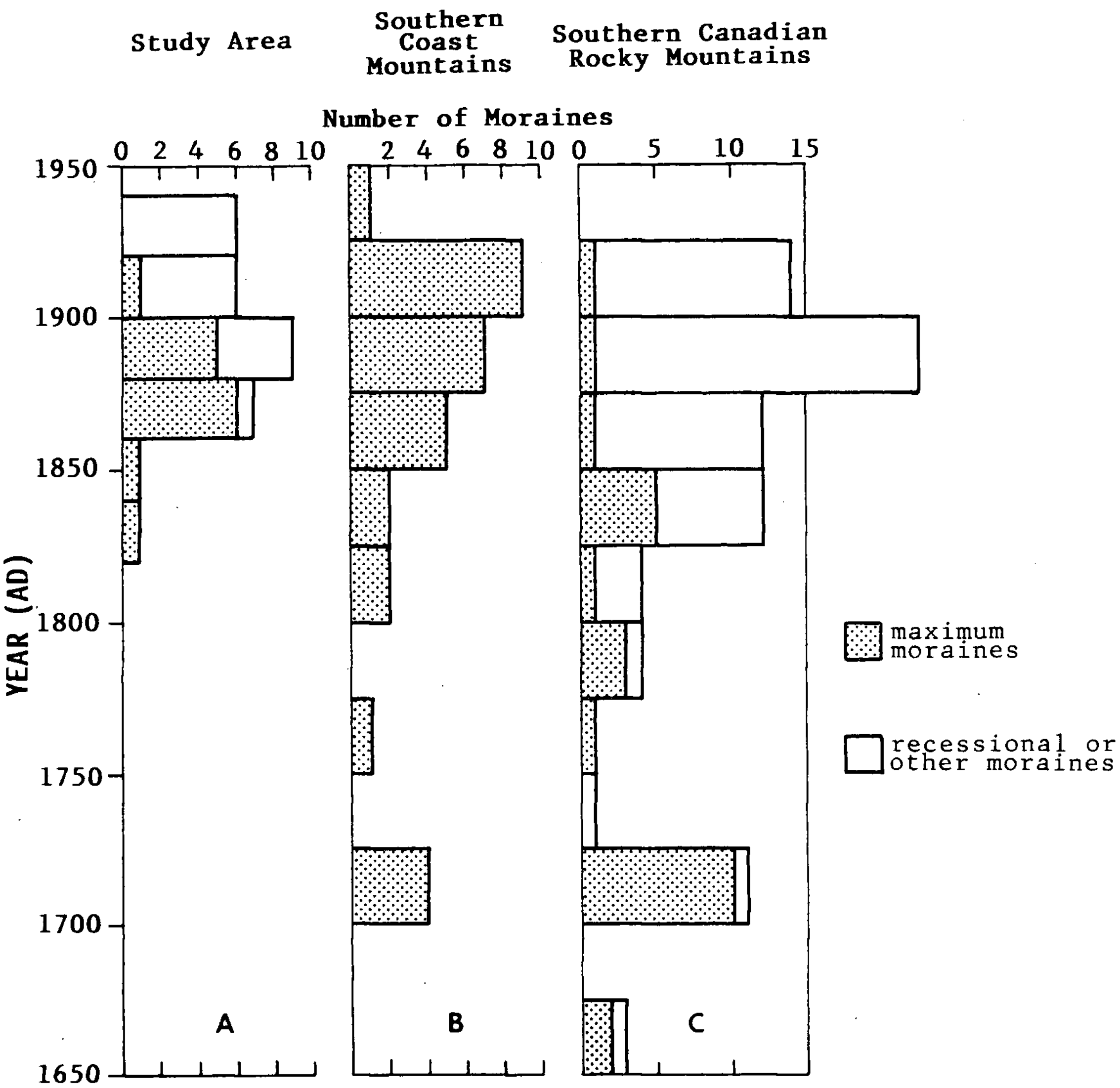


Figure 7.3 Dendro-dated, late Neoglacial terminal and recessional moraines in selected regions of the Canadian Cordillera. A) this study (see Appendix F.2 for locations). B) data given in Ryder *et al.* (unpublished). C) data from Luckman (1986) for Main Ranges of the Southern Canadian Rocky Mountains.

conditions prevailed from 1890 to 1905 for most of northern Pacific coast, which may have slowed retreat from mid and late 19th century maxima. Similarly, evidence presented in chapter 3 demonstrates a return to slightly cooler temperatures after 1930 in the interior region and wetter conditions in the coastal region.

The chronology in figure 7.3a agrees reasonably well with the Little Ice Age chronology derived for several sites in the Coast Mountains to the south of the Bella Coola basin (figure 7.4b) (Ryder *et al.*, unpublished; Mathews, 1951). Ryder *et al.* found a small number of 18th century moraines at Tiedemann and Franklin Glaciers, while Mathews (1951) found evidence for older moraines at Helm, Lava, Sphinx and Warren Glaciers in Garibaldi Park. These few sites are the only ones which indicate retreat from Little Ice Age maxima prior to the 19th century. By comparison, there is a higher frequency of moraines in the Canadian Rocky Mountains which were stabilized in the 17th and 18th centuries (figure 7.4c) (Luckman, 1986).

In part, the distribution of moraines shown in figure 7.3a will reflect the timing of linear changes in glacier margins as a function of glacier size and valley position rather than a synchronous response to a common change in climate. The absence of moraines older than the early 19th century demonstrate that the Little Ice Age advance was the most severe excursion of glaciers during the whole of the Holocene Epoch in the basin. Attendant changes in other biogeophysical attributes would also be expected to have responded significantly to these anomalous perturbations.

(7.2) Long-Term Low and High Resolution Biogeophysical Evidence

Long-term records of growth variations in Douglas and subalpine fir were compiled using the tree-ring chronologies given in Appendix E. These records are compared here with sedimentation rates in Ape Lake and glacial moraine chronologies discussed previously. The shortest site-chronology for each of the tree species limits the interval of consideration which in the case of Douglas fir is post-1650 and for subalpine fir post-1740. Neither of these records, nor that of lake sedimentation rates, is very long but they appear to overlap with the interval of substantial glacial changes in the Bella Coola watershed.

Fluctuations in Average Temperature and Precipitation

The final tree-ring climate model presented in table 6.3 (see chapter 6) for Douglas fir is used to estimate long-term variability in annual precipitation for the central interior. Since the linear transfer function for Douglas fir does not incorporate extreme climatic departures and all modeling attempts yielded relatively high standard errors, inferences regarding changes in annual precipitation are necessarily restricted to identifying period-average trends about a long-term mean. A plot of inferred trends, smoothed using a five-year weighted moving average, is presented in figure 7.4a.

With the exception of the interval 1850-1875 it appears that most of the 19th century was characterized by above average precipitation particularly in the first and last decades. The record prior to 1800 indicates less persistent trend, fluctuating above and below the mean although the variance during most of the 18th century was lower than that inferred for the last half of the 17th century and first half of the 19th century. Variance in tree-growth is generally higher than that observed during the calibration period (post-1930), except between 1700 and 1800 AD, leading to extrapolations of estimated precipitation. While there is less statistical confidence in the extrapolated values we may suppose that they are representative of generalized trends in annual precipitation.

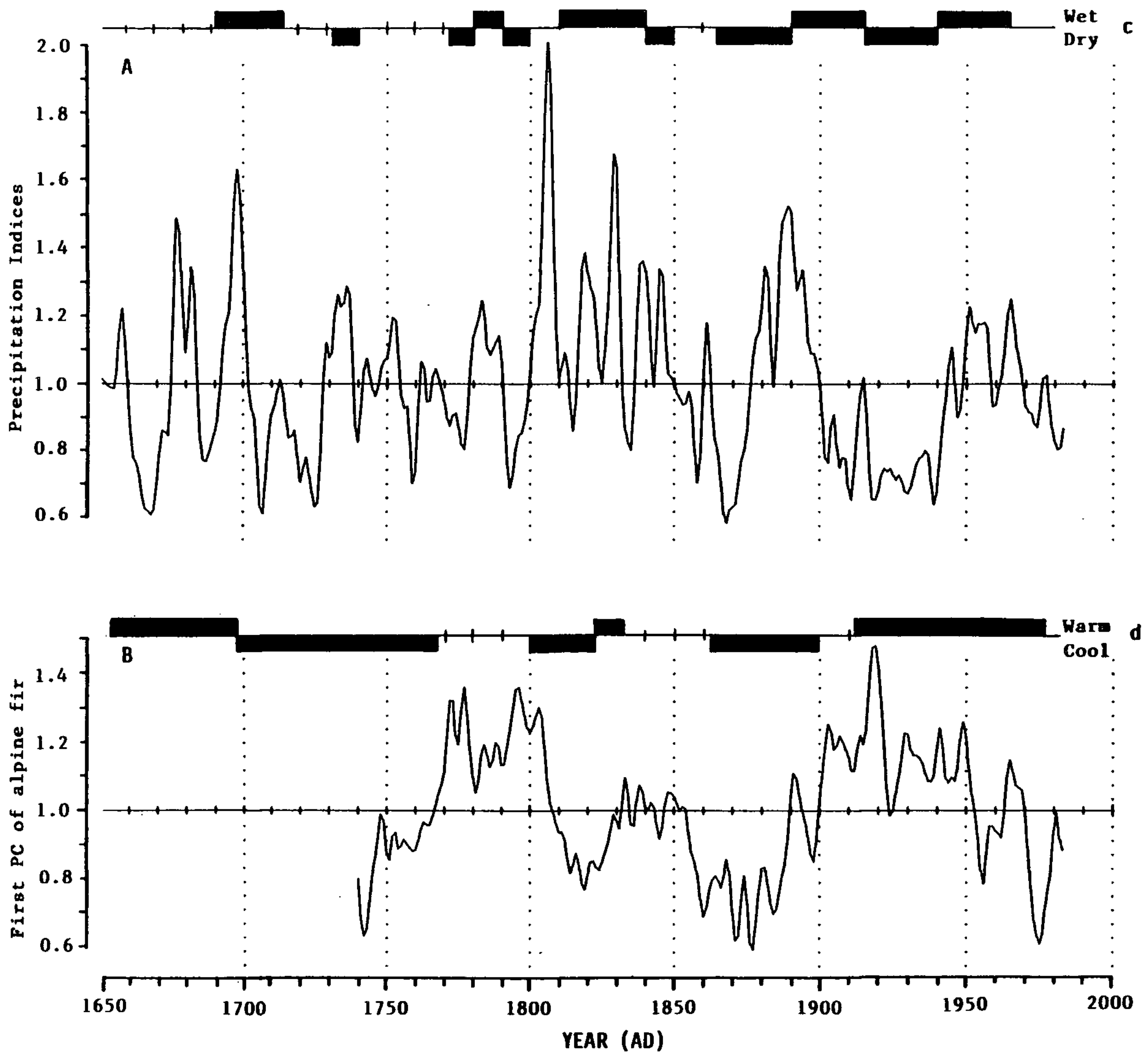


Figure 7.4 Long-term departures in reconstructed winter precipitation (A) and the first PC score of subalpine fir (B). Original data are smoothed with a five-year weighted mean. Generalized departures in precipitation for the lower Columbia basin in Washington State (c) are from Graumlich (1987) and for temperature departures inferred from various high elevation sites in the Cascade Mountains of Washington State (d) are from Graumlich and Brubaker (1986).

Using growth variations from four drought sensitive species Graumlich (1987) identified several wet and dry periods for the Columbia River basin east of the Cascade divide in Washington State. Prominent drought periods were from 1870 to 1890, the 1840s, 1770 to 1780 and around 1680. Three dominant wet periods were inferred from 1810 to 1845, 1780 to 1790 and 1690 to 1715. Comparison of this record with the reconstruction given in figure 7.4a reveals a lack of correspondence between the two records. This is consistent with the observations made in chapter 3 on spatial variations of precipitation for the north and south coastal areas of British Columbia and with the conclusions given by Namias (1978) and Yarnel (1985) in their analysis of recent precipitation variability. Over the long-term then, precipitation in the Bella Coola basin may be more closely linked with trends along the north coast.

Fluctuations in summer temperature inferred from growth variations in subalpine fir show more persistent departures than precipitation over the last 240 years (figure 7.4b). Below average temperatures dominated the last half of the 19th century preceded by average temperatures from 1830 to 1850. Above average temperatures characterized the interval 1760 to 1800 while average to below average summer temperatures were inferred between 1730 and 1750. Again, variance in growth over the calibration period (post-1930) does not cover the entire range of variance throughout the whole record. In particular, positive departures in the late 18th and early 20th centuries were higher.

Comparison between long-term temperature reconstructions for Washington State (Graumlich and Brubaker, 1986) and the record given here reveals many similarities (see figure 7.4d). Most evident is the below average summer energy conditions for much of the 19th century followed by substantial warming which began around the the turn of the century. Unlike the Washington State record, pronounced cooling is inferred for the the Bella Coola region after 1950. The longer tree-ring record from further south shows well below average temperatures were likely between 1695 and 1765: a trend which is evident in the early record from Bella Coola.

From the combined record of inferred precipitation and temperature departures several, intervals of persistently warm/dry and cool/wet conditions can be identified. These two hydroclimatic settings lead to glacier recession or expansion respectively, and were shown to be coincident with phases of decreased or increased flood activity in southwestern British Columbia. The inferred (and observed) warmer and drier climate between 1900 and 1950 is the most prevalent departure in the last 330 years. In contrast cooler and wetter conditions dominated much of the 19th century. Near average to below average precipitation was associated with above average temperatures between 1765 and 1810. There are cooler and drier periods, separated by warmer and near average precipitation prior to 1760.

Recession of some coastal glaciers from Little Ice Age maxima at the beginning of the 19th century may have been a response to the inferred warming which commenced in 1765. However, increased precipitation and cooling after 1810 produced maximum stillstands among most glaciers in the Bella Coola watershed. Below average precipitation between 1850 and 1875 signaled the end of the Little Ice Age and induced the first real trend towards glacier retreat. The response was slowed by continued cool temperatures and a return to above average precipitation until the turn of the century. Brief pauses in the retreat of ice occurred in the early 20th century in response to small increases in precipitation.

Inferences From Lake Sedimentation

Although the record of variation in sediment yield to Ape Lake is comparatively short, there is strong evidence for declining sedimentation rates over the last 150 years (see figure 5.11). Comparison between sedimentation rates for the period of direct and indirect drainage to the east basin of Ape Lake can be misleading because of changes in process (hyperpycnal versus homopycnal flow) and changes in the storage capacity of newly uncovered lake area in front of the calving Fyles Glacier. One possibility is to assume that the most recent sedimentation rates in Ape Glacier Bay and the west basin are reasonable analogs for the character of sedimentation in the east basin prior to 1940 - an assumption supported by similarities in texture between the two sets of samples.

Comparisons are made here by estimating total annual sediment volumes delivered by meltwater streams using observed mean accumulation rates and associated lake areas. For a combined total area of 0.81 km^2 and a mean deposition rate of 2.55 mm/yr (1952-1984), a total of approximately 2000 m^3 of sediment is deposited in Ape Glacier Bay and the west basin each year. In comparison, average depositional rates of 3.86 mm/yr can be established during the period 1836-1940 when Fyles and Ape Glaciers were discharging directly into the east basin. For an area of 1.3 km^2 this corresponds with a total load of 5020 m^3/yr during this earlier period. Since the two glaciers are the primary source of sediment and runoff for both periods, this two and a half times difference is large enough to suggest that declining rates of sedimentation in the east-basin are related not only to changes in intermediate storage capacity and depositional processes, but also to real declines in total sediment yield from a set of retreating glaciers.

Leonard (1981, 1986) found evidence for above average lake sedimentation rates during phases of increased ice extent and initial ice retreat, although the association was not always spatially or temporally consistent. The interval covered by the Ape Lake sedimentation chronology is too short to represent the entire Little Ice Age glacial cycle (approximately 1500-1900 A.D.), but the evidence does tend to support the findings of higher sediment yields during maximum ice extent and initial retreat, followed by a substantial decline.

Average and below average rates of sedimentation are superimposed on this long-term trend of declining rates. Above average sedimentation during 1885 to 1890, 1863 to 1877 and 1836 to 1846 shows some correspondence to inferred above average or increasing summer temperatures and inferred winter precipitation. A similar sequence of above average sedimentation rates in the late 19th century was noted by Leonard (1985) for Hector Lake, Alberta, while the data given in Perkins and Sims (1983) for Skilak Lake, Alaska, also show local increases during these periods.

One other piece of evidence sheds some light on changes over the last several centuries. Core 84C6 from the shallow lake bottom at the extreme east end of Ape Lake is capped by 6 cm of massive, fine silt and clay over approximately 170, regularly laminated, silt and clay couplets punctuated with small dropstones. The couplets have the characteristic appearance of true varves and although there is distortion in the core they can be cross-dated, in isolated sections, with the varve chronology from the deeper basin. Based on this correlation, the massive material represents deposition following retreat of Fyles Glacier in 1940-45 to a position behind the terminal moraine. A higher than normal sedimentation rate and frequency of dropstones within the sediment occurred during an estimated 25 year period at the end of the 18th century. This suggests extensive calving of Fyles Glacier and increased suspended sediment loads to the lake and is coincident with a short period of increased sedimentation in both Hector Lake and Skilak Lake.

High-Frequency Variations in Tree Growth, Lake Sedimentation Rates and Related Hydroclimatic Inferences

In addition to the long-period trends shown in figure 7.4, the high frequency or interannual variability in the first principal components for Douglas and subalpine firs was also examined. Notable year-to-year variance in Douglas fir growth occurs during the intervals 1950-1960, 1925-1945, 1825-1870, 1790-1800 and 1710-1745, (figure 7.5a). Although not as frequent, similar clusterings of extreme growth departures for subalpine fir occur during 1925-1940, 1860-1875 and 1740-1755 (figure 7.5b). Intervals of reduced annual variability in growth in both species occurred during 1960-1983, 1870-1920, 1800-1820, and 1750-1780. Schulman (1956) and Stokes *et al*. (1973) have also shown these more persistent but lower variance periods in Douglas fir growth to have occurred between 1660 and 1690 AD and between 1470 and 1490 AD .

The apparent clustering of extreme positive and negative departures in tree-growth was tested under the null hypothesis that the extremes are randomly distributed. A one-tailed runs test was conducted for both series plotted in figure 7.5. The z-scores for Douglas and subalpine fir were 3.64 and 2.10, respectively. With $\alpha = 0.10$ the null hypothesis was rejected in both tests indicating a non-random ordering of extreme growth response.

Based on the relationships established in chapters 3 and 6 it is most probable that intervals characterized by persistent above or below average departures in growth are dominated by an atmospheric circulation regime with a strong meridional component. High year-to-year variance occurs during periods of pronounced zonal flow when the mean position of the jet

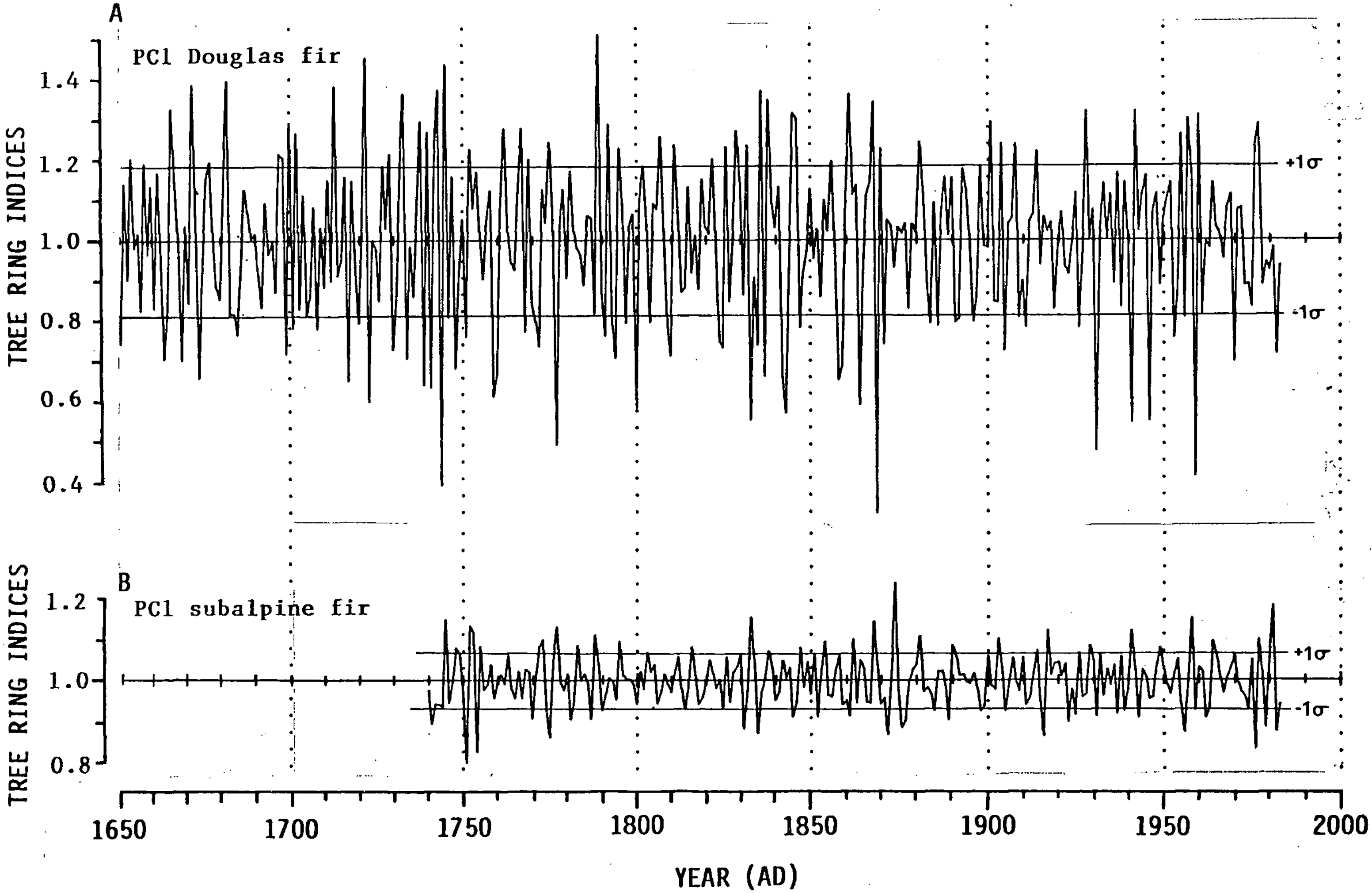


Figure 7.5 Long-term high frequency signals in the first PC score of Douglas fir (A) and subalpine fir (B). Departures are from a weighted 13-year filtered series. Horizontal line are plus and minus one standard deviation from the mean.

stream fluctuates north and south. This would yield more variable precipitation and temperature trends along the central coast of British Columbia. Lamb (1979) has demonstrated that there was minimal strength in zonal circulation between 1790 and 1829 during January for the North Atlantic area which appears to correspond with enhanced winter precipitation recorded in the Interior of British Columbia and declining summer temperatures around the study site.

Blasing and Fritts (1976) have inferred the variability of several winter pressure types for the period commencing in 1700 using some of the same tree-ring data analyzed here. They demonstrate that persistent departures of pressure leading to increased precipitation and below average temperatures for the Pacific Northwest occurred between 1745 and 1770 and the first and last decades of the 19th century. This further supports the contention that persistent departures in growth are coincident with anomalously high pressure over Alaska and northwest Canada which promotes strong ridging to the west and movement of cyclonic disturbances either into or away from the study basin. High year-to-year variance in growth is demonstrative of zonal circulation.

Comparison of sedimentation rates with growth anomalies indicates that extreme positive departures in sedimentation are coincident with extreme negative departures in Douglas fir growth. Low winter precipitation and high summer temperatures appear to be the common factors in these years, particularly in the pre-1885 record. Although some similarities are evident (see figure 7.5), the geophysical and biological indices retain certain unique characteristics demonstrative of an aggregate climate effect - over individual seasons for sedimentation rates and over one or more growing years for tree-rings.

(7.3) Effects on the Bella Coola River

Fluctuations in temperature and precipitation as well as the inferred frequency of extreme events are apt to have had an effect on the Bella Coola River. The purpose here is to assess the possibility of extending the long-term record using evidence preserved in the Bella Coola valley. Responses of floodplain vegetation and changes in sedimentation patterns over the last several centuries are used to make these inferences.

Floodplain Age

The distribution of floodplain surfaces of different ages can provide some indication of the frequency with which floodplain sediments are being recycled and the temporal variability of river channel changes. A combination of data sources was used to derive estimates of floodplain age between the mouth of the Bella Coola River and the Atnarko/Talchako confluence. In order of decreasing importance these include: 1) air photographs and early survey maps, 2) age of resident vegetation, 3) personal communication with long-term residents, 4) historical records, 5) pollen and radiocarbon dating of floodplain sediments and 6) archaeological artifacts.

Details of the available maps were augmented and extended by sampling vegetation on several disparate surfaces. Two major difficulties for inferring age using vegetation arise in ensuring that the resident forest cover is first generation growth and in establishing the ecesis interval or colonization period. Some surfaces had been cleared by early settlers (or commercially logged) and then abandoned. Many areas of the floodplain have been logged but most occurred after 1945, and therefore careful cross-checking with air photographs assisted in identifying disturbed sites. To facilitate dating, each areal vegetation unit sampled in the field was surveyed for signs of old growth (stumps, deadfall, rootwad hollows, etc.) and then verified, where possible, by checking with long-term residents (or former owners) of the property. Increment cores, tree-disks and ring counts on stumps provided the major sources of information.

Alders colonize recently flooded channel zone and floodplain areas, often in the first growing season following the flood. Seedling germination requires only a thin layer of silty-sand (see figure 4.4b). Alders and black cottonwood dominate older and less frequently inundated surfaces in which the ecesis interval for forest communities of this composition is apt to be close to zero. On the oldest surfaces, decadent black cottonwood (100+ yrs), western red cedar, Douglas fir and Sitka spruce prevail.

Determining the age of cottonwood trees was most problematic. Ring boundaries are diffuse and porous, hence difficult to distinguish and the rings which can be identified are not necessarily annual in nature (Everitt, 1968). For these reasons alder and coniferous vegetation were mostly sampled. Where cottonwood was the dominant species, wood stains were applied to the cores to enhance the ring structure. Other species were dated to verify inferred ages.

Surveys of alder/cottonwood forests at several sites showed that conifers, of similar age to the surrounding deciduous vegetation, were present, albeit in much lower numbers (average of one tree per 1000 m^2) and much smaller sizes. The implication is that the oldest members of a predominantly coniferous forest cover may be original colonizers, thus ecesis intervals may be short. Since the probability of sampling the oldest member decreases with the size of the area to be sampled, the ecesis interval is probably significantly different than zero. The sample surveys made here suggest a maximum colonization period of 25-30 years. However, complete forest succession from deciduous to coniferous may take as long as

60-80 years. Figure 7.6 illustrates the percentages of floodplain surfaces of different ages for the Bella Coola River.

Four age categories are used in figure 7.6: areas older than 1700 A.D., those areas between 1700-1900 A.D., 1901-1938 and 1938-1982. Approximately 27% of the total floodplain area (excluding the area of the active channel zone) is older than 1700 A.D. and almost 80% of this occurs below the Nusatsum confluence (marked as downstream on figure 7.6).[1] Approximately half of the floodplain area is estimated to be in the range 1700 to 1900. The data indicate that the younger areas in this category (1800 to 1900) are more frequent in the downstream reaches while the older areas are more frequent upstream. Almost 24% of the floodplain has developed since the beginning of the 20th century.

This distribution, in conjunction with other evidence reveals several things. First, inferences about rates of floodplain development over the last several centuries are possible because of the long-term vertical stability of the river. The absence of significant aggradation or degradation makes the available floodplain area approximately constant. Secondly, areas older than 1700 A.D. are in some cases significantly older. Age estimates of buried organics within some shallow overbank deposits are in excess of 1700 years B.P. (see Appendix A; samples S-2698 and S-2730).

Over the last 280 years, a proportionally larger area of the floodplain has been constructed during the post-1900 interval. This is manifested in two trends; a decline in the number of anabranches or sloughs occupied by the river and a decrease in the width of the dominant active

1 This estimate is thought to be reasonable because of offsetting errors. In the first instance it may be an underestimate because of the problem of determining if the growth is first generation. Secondly, it may be an overestimate because of difficulties in delimiting the floodplain boundaries. A fairly generous boundary was used where the flood limit was not know precisely.

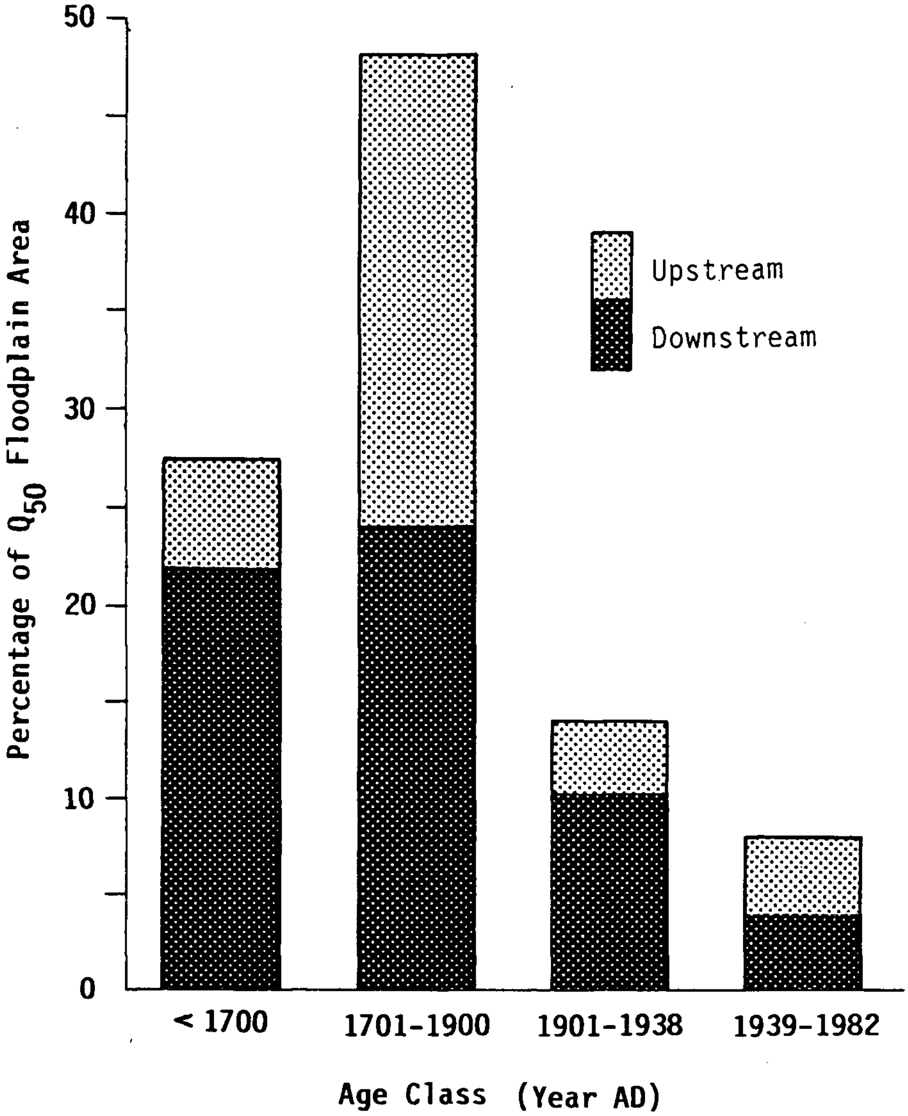


Figure 7.6 Frequency distribution of floodplain age based primarily on estimates from surface vegetation below the Water Survey of Canada gauge above Burnt Bridge Creek. Note that age class sizes are not equal. Upstream and downstream segments are separated by the Nusatsum River confluence. Total Q_{200} floodplain area is approximately 30 km^2 of which 38% is upstream of the Nusatsum River and 62% below. Area estimates are subject to greater error for the two older age classes but errors are probably no larger than ± 5%.

channel in at least the stable reaches. Approximately 14% of the floodplain was constructed between 1900 to 1938 and 10% between 1939 and 1979. Differences between the two periods cannot be considered significant but during the latter interval channel avulsion has been less frequent resulting in continued stabilization of early 20th century surfaces and the development of no new channels.

Once established, vegetation type and age provide a chronology for channel migration. The succession of vegetation on point bar segments in laterally stable reaches of the river indicates the mean rate of channel migration to be in the range of 3.0 to 3.5 m/yr (Desloges and Church, 1987). Channel migration rates in sinuous reaches upstream of the Nusatsum confluence indicate that the replacement or turnover rate for sediment along a unit length of the 0.5 km wide floodplain would be between 145 and 165 years. Once deposited there, floodplain sediments remain undisturbed for periods of the order one to a few centuries.

Downstream of the Nusatsum confluence the turnover rate is longer because of the increased valley width but migration rates in stable reaches are comparable. While regular cross-valley migration of the channel is important, the presence of floodplain remnants which date to the previous millenia indicate that over the long-term regular lateral progression does not reach all areas of the floodplain.

From the generalized interpretations given here it appears that a significant amount of sediment turnover and floodplain development which began around the turn of the century was commensurate with predominant changes in prevailing hydrology and related sediment transfers. Although a decrease in precipitation and an increase in summer temperature, which became fully apparent by 1920, promoted increased channel stability, the river and associated sediment transfers still respond in a significant way to the higher-magnitude flood events. These events appear not to be exclusively linked to long-term persistent climatic trends. The apparent higher frequency of surfaces which date to the 19th and 20th centuries suggests that the most significant changes to the alluvial environment occurred during and after the peak of the inferred Little Ice Age interval.

Evidence for a Long-Term Flood History

Sampling of the vegetation in several areas of the floodplain provided the opportunity to assess the effects of flooding on the structure and growth of woody plants. This exercise was undertaken because damage and burial of vegetation along the margins of contemporary alluvial fans provided good evidence of recent (instrumental period) flood events (see chapter 4). There are several effects flooding might have on woody vegetation which survives inundation. A juvenile may be bent and partially buried yielding differential stress on the trunk thereby producing cell damage or reaction wood (Yanosky, 1983). On older individuals damage to the upstream side of the trunk may produce scarring and localized distortion to the ring-pattern. Finally, burial of trunks or branches may produce new stems which can then be used to date the event. Damage patterns then will depend on the age and species of tree and the position of the site with respect to flooding.

In addition to the cores and stumps used to date the floodplain surface, tree disks and 2 or 3 increment cores/tree (radii) were sampled from other vegetation which was thought to have been affected by flooding. Vegetated channel banks in stable reaches of the river are the preferred sites because the discharge-stage relationship would likely have been constant during the recent past. The sampling design and locations produced mixed results in which only a very small number of consistent dates could be assigned.

During the 20th century the effects of the 1936 flood were the most widespread and dramatic. Vegetation sampled on alluvial fans, several areas of the floodplain and even on several hillslopes contained some evidence of scarring or reaction wood from this autumn flood event. Less frequent, but equally obvious at certain sites, was damage during 1924. The only other major event recorded at several sites was a flood either in the autumn of 1805 or the spring of 1806. Less significant damage occurred in 1896, 1886, 1826 and 1785. Evidence from earlier in the 18th century is more scarce because of the smaller number of older trees and the frequency of heartrot in the older individuals. Coniferous trees, in particular spruce and cedar, were the most valuable data sources.

There was some indication of severe flooding at some sites, particularly from the 1980 and 1968 floods, but earlier dates could not be confirmed because either very few individuals responded or adjacent lower-elevation sites, which should have been more susceptible to damage, showed no evidence of flood damage. The somewhat singular response is not so surprising given that flooding can occur locally as a result of log jams and channel avulsion. Some damage was also the product of disturbance from adjacent individuals (falling trees). The attempt to reconstruct a long-term flood chronology using woody vegetation cannot be considered successful but there is an indication that a more rigorous sampling design in several stable reaches of the river might yield more consistent results.

Long-Term Floodplain Sedimentation

Examination of recent floodplain sediments in back channels and sloughs was undertaken to assist in interpretations of older floodplain sequences. Results presented earlier demonstrate that floodplain sedimentation is episodic, dependent on channel configuration near the site and associated with a variety of depositional processes. Pollen and spores are incorporated in a variety of ways and represent airborne, riverborne and local sources. For these reasons only a selected number of sites, in which vertical accretion was thought to be the dominant mode of deposition, were examined for variations in stratigraphy, mineralogy, grain size variations and pollen concentrations. Site selection was based on the distribution of inferred floodplain age (see figure 7.6) and inspection of sediments from boreholes in each of the cross-valley transects. Seven sites were selected but after examination of grain size and pollen data four were thought to have formed by lateral accretion of the river channel and thus were less likely to be of any significance. All three remaining sites were in stable reaches of the river (locations are plotted in Appendix figure C.1). Only one of the remaining three sites is discussed here but reference is made to two others.

The Brekke site is on the upstream side of the Salloomt River fan in a confined portion of the Bella Coola valley (figure C.1). Due to river training works on the south side of the valley, Bella Coola River is currently undercutting a portion of the floodplain which has been stable for at least the last 70 years. Bankfull width of the river here (256 m) accounts for approximately 37% of the maximum valley width and, given that the height of the bank above mean summer flow is only 1.1 m, it is probable that overbank flows are frequent. A monolith was extracted from just behind the currently eroded bank to determine the physical and organic properties of the sediment. Figure 7.7 is a summary of these characteristics.

Grain sizes are mostly fine sand and silt with a minor amount of clay indicative of a low energy depositional environment. Charcoal is found in

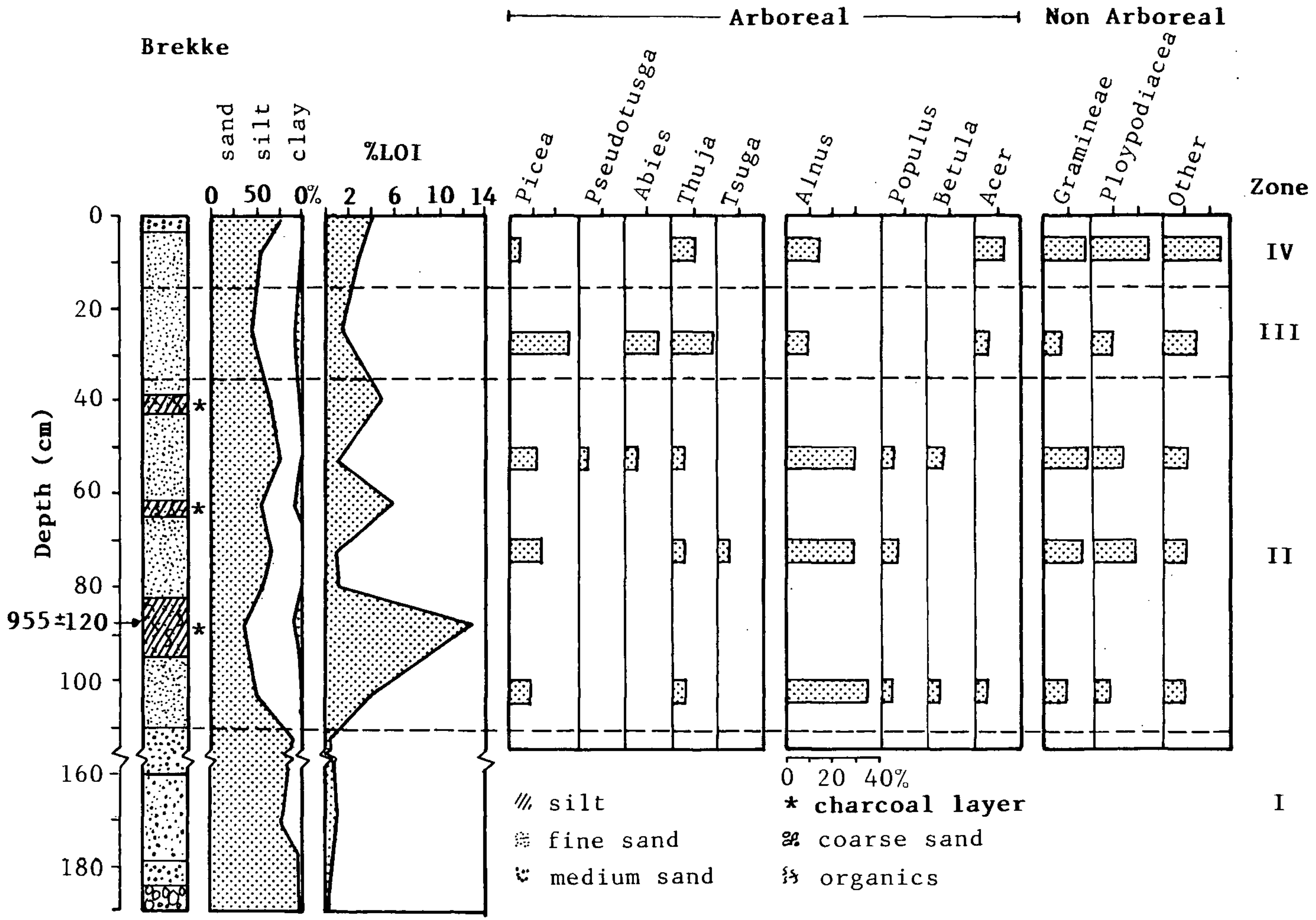


Figure 7.7 Stratigraphy, grain size variations, organic matter content and pollen concentrations in vertical accretion sediments of the Brekke site on the Bella Coola floodplain.

three distinct horizons and the lowest is the thickest containing whole log fragments. The ring structure is that of cedar. Below 110 cm there is medium sand with faint bedding structures which in turn overlies gravel at 190 cm. The summer water table is coincident with the transition between fine and medium sand.

The original surface vegetation of mature cedar and spruce was logged in the 1920s. From ring counts on standing stumps an estimate of the age of the surface is 220 ± 20 yrs B.P. and a radiocarbon date on the lowest charcoal layer in the sequence was 955 ± 120 radiocarbon yrs B.P. Based on pollen concentrations four zones can be assigned (figure 7.7). Zone I is lateral accretion sediment and contains no pollen. From the micro topography of the site it is possible that this was a former slough or back channel and the basal sands represent the bar top facies. Zone II is layered silts and fine sands with three distinct charcoal horizons. The lowest horizon can be traced laterally for at least 15 m in all directions indicating that it was an *in situ* forest burn. Fragments of charcoal constitute the other two horizons but they are also found at consistent depths in the surrounding terrain. Organic matter content between the organic layers is low.

Pollen in zone II is dominantly alder, grasses and ferns with most of the floodplain conifers represented as well. The dominance of alder pollen in the lower 10 cm of this zone suggests that the site was a recently active slough channel or else very near to frequently flooded surfaces of the main channel. Above this zone the three charcoal horizons and abundance of successional forest pollen types (spruce/cedar) indicate that overbank flooding (vertical accretion) occurred episodically.

Although textures do not change, zone III is marked by a substantial increase in spruce and other conifer pollen and a decrease in the nonarboreals. This shift may represent accretion of the floodplain to its final level. The increased absolute and relative importance of the grasses, ferns and several other species not found in the lower horizons is demonstrative of recent disturbance in zone IV. Human disturbances to this site in the form of logging and cultivation began as early as 1920.

Evidence from this site suggests that vertical accretion of the floodplain has proceeded slowly over the last millenium. Even though a persistent channel course along the south side of the valley would reduce the rate of sedimentation, the river is sufficiently wide in relation to the valley width that it is unlikely that the larger flows did not go overbank at this site. There are no structural or mineralogical indications that the site has undergone continual scour and fill. A recent accumulation of 3.5 cm of medium sands can be attributed to post-1960 overbank flows.

Inferences about long-term rates of sedimentation can be made given the following assumptions: 1) the top 3.5 cm of deposition is related to floods between 1965 and 1980; 2) significant changes in the pollen assemblage above 15 cm is related to forest clearance around 1920 AD: 3) sediments deposited above the upper most charcoal layer (i.e. top 40 cm of sediment) represent flood events which post-date the establishment of the mature, 220 year old, cedar/spruce forest (rooting depths of the trees generally correspond with this horizon which further supports the claim); 4) the principle channel of Bella Coola was not persistently entrenched on the south side of the valley and 5) no erosional hiatuses exist between dated horizons. Under these assumptions the rate of vertical accretion averages 18 mm/yr over the last 20 years, 19 mm/yr over the last 80 years and around 18 mm/yr over the last 220.

The radiocarbon dated horizon of 955 ± 120 years (figure 7.7) corresponds to a wider range in calendar dates due to increased fluctuations in atmospheric carbon content around this period (*cf.* Stuiver, 1982). A minimum estimate for the age of the lower horizon is 1210 AD or approximately 550 years older than the upper most organic layer. For a total accumulation of 43 cm between charcoal horizons this relates to a maximum estimate of the mean depositional rate of 8 mm/yr or half the rate of the recent sedimentation. Using mean and maximum age estimates for the lower horizon of 1030 AD and 910 AD the depositional rate is reduced to 6 and 5 mm/yr, respectively.

Dating of older organic horizons at two other sites (Canoe Crossing and Firvale) could not be made with any confidence but again it appears that recent depositional rates, based on pollen changes following land clearance and recent flood sediments, are significantly higher in relation to the total amount of accumulated sediment. Within the limitations of the techniques adopted here a tentative inference would be that floodplain development over the last few centuries, or Little Ice Age interval, has proceeded more rapidly than over the last millenium. This conclusion is supported by inferences made from the distribution of dated floodplain surfaces.

Clearly then, the resolution of long-term changes in alluvial sedimentation in the Bella Coola valley is low and limited by the availability of suitable sites. Substantial reworking of the floodplain over the last three centuries has removed much of the evidence and difficulties in establishing long-term dating control are pervasive. Charcoal is an abundant constituent of vertical and lateral accretion sediments but the available fragments have a high potential of being reworked and thus may yield anomalous estimates of age. An even higher frequency of sampling than that adopted here may be required to verify and extend further these limited conclusions.

Chapter VIII
SYNTHESIS, CONCLUSIONS AND FUTURE DIRECTIONS

The major objectives of this study were: 1) to characterize the response of selected biogeophysical attributes to recent climatic change; 2) to determine the resolution and length of paleoenvironmental records in the study area and; 3) to ascertain the regional significance of inferred pre-instrument environments. The overriding goal is to establish the basis for inferences about past environments. Paleohydrological studies of individual drainage basins in coastal British Columbia have not been made, yet there is increasing evidence to suggest that such studies provide a better understanding of integrated physical processes (*cf*. Helley and LaMarche (1973) and Coghlan (1984) for California studies). This study was an attempt to determine the feasibility of systematically documenting recent environmental change on a spatial scale commensurate with factors that are apt to control regional climate and hydrology. It was recognized that the nature of results from the first two objectives would influence what could be said about long-term environmental history.

The following discussion is divided into three sections: a summary with conclusions about the methodological implications of the study and characteristics of the response records; a discussion of inferred Little Ice Age environments of the study area; and directions for future research.

(8.1) Methodological Implications and Biogeophysical Response Records

Paleohydrological models are based commonly on interpretation of cause and effect relationships in an environmental system containing both hydroclimatic and biogeophysical components, such as that defined in figure 1.1. The attempt here was to consider how functional relations between the two components would influence inferences about the hydroclimatic system using indirect and proxy data sources from the biogeophysical system. The strategy was to test for the type and strength of linkages within the entire system over recent time scales - in particular, the post-1945 interval. The intention was to quantify the linkages for retrospective estimation of pre-instrument climate and hydrology using the longer response record.

A selection of biological and geophysical attributes was made in recognition of the fact that any single attribute would not provide incontrovertible evidence of changing hydroclimatology. This is because the actual temporal resolution of these elements varies from seasonal to approximately decennial and because the extent to which any attribute provides evidence of regional changes in hydrology was not initially known. It is clear from this study that inferences about hydroclimate are twice removed from the available biogeophysical evidence; one set of relations intervenes between the biogeophysical elements and hydrological controls, and a second between the system hydrology and climatological controls. Several of the relations are neither direct nor linear, and are complicated by patterns of response which may be continuous or discrete.

Recent Hydroclimatic Record

A significant difference in the seasonal frequency of objectively classified, daily synoptic patterns distinguishes periods with extreme positive departures in temperature and precipitation from ones with extreme negative departures. That is to say extreme climate is well defined in the context of synoptic scale circulation. A similar conclusion cannot be made about less extreme departures. Anomalies in thermal climate are not particularly well coordinated with ones in pluvial climate. The implication is that those environmental attributes which respond to precipitation variability (*e.g.* Douglas fir growth under extreme drought) will provide a different indication about climate than those influenced by temperature (*e.g.* subalpine fir near tree line).

The results presented in this thesis show that approximately 60% of the variance in Bella Coola River spring runoff is accounted for by seasonal temperature and precipitation indices. The remaining 40% unexplained variance is probably tied to three factors: 1) extreme climatic departures; 2) imperfectly measured climate and runoff indices; and 3) a random component. The later two factors are likely to be proportionally small in relation to the first factor which itself is related to conditions not indexed by seasonal climate parameters.

In contrast to spring runoff events, autumn floods in coastal basins of southwestern British Columbia appear to have very distinct synoptic controls. This manifests itself as similar secular variations in the frequency of one or two day runoff events within these coastal basins. However, flood magnitude is still very much dependent on storm track orientation, therefore, does not show the same spatial coherence. A major conclusion is that extreme weather produces key hydrological events (*e.g.* January 1968 flood) which are likely to have strong imprints in the geophysical environment.

Tree-Ring Response Record

The biological phenomena studied here yield a range of resolutions for interpretation of former hydrological environments (see figure 1.2) but tree-rings provide the most workable data. Douglas and subalpine fir trees provide annual resolution of growth variability; hence, under suitably stressed conditions they have potential for inferences of similar resolution about climate. However, the climatic information recoverable from these proxy data is at considerably lower than annual resolution. This is related to two effects: 1) an element of interannual persistence in tree growth which is not simply related to departures in any one climatic parameter and 2) an element of extreme responses in growth that represents confounding effects of climatic and hydrological conditions at specific sites (*e.g.* snow cover).

For these reasons working back from the tree-ring record to the hydroclimatological, hence climatic, antecedents becomes difficult. Linear modeling techniques, although widely used (*cf.* Hughes *et al.*, 1982), appear not to be capable of reflecting the combination of persistent and episodically extreme behavior. Even after attempts to remove the stochastic portion of the long-period variance and several trials at removing suspected non-climatic trends, explained variance generally remains low. Douglas fir chronologies account for 29% and subalpine fir for 45% of the variance in precipitation and temperature indices, respectively, over the calibration period (1933 to 1983). The implication is that reconstructions of annual climate from tree-rings in environments such as the Pacific Northwest are suspect. Although these difficulties preclude confident quantitative reconstruction of former temperature and precipitation values, there remains a broad concordance between tree growth and climate. In the study basin tree-ring chronologies are short and based only on living individuals. The humid maritime climate, limited treeline fluctuations and logging are not conducive to the preservation of older wood. Thus, extension of the record beyond 350 to 400 years B.P. does not appear possible.

Glacio-Lacustrine Response Record

Glacio-lacustrine environments provide a record of sediment flux with annual resolution, which possibly can be extended to seasonal resolution depending on the rate and processes of sedimentation. Preliminary reconnaissance of sediments in other glacial and non-glacial lakes in the region indicates that proximity to the glacial source, a high seasonal influx of suspended sediment and a deep receiving basin are minimum requirements for development of varve-like sediments (Desloges and Church, 1984). A composite record of 150 years was obtained from Ape Lake. Results from sub-bottom profiling indicate that a longer record may exist.

Lake-wide correlations in sedimentation variability demonstrate a similarity in depositional processes throughout the lake. However, inferences about the hydrological environment are complicated by the fact that different depositional processes may have operated at different times. The long period variance in the sedimentary record provides evidence of continuous processes such as glacier retreat and changing contributions of sediment from the glacier due to intermediate storage. A seasonally integrated response to snow and glacier melt produces a high frequency variance of less extreme departures. Finally, responses to events such as autumn rainstorms are in the form of extreme rates of sedimentation. There are few objective means by which these processes can be discriminated in the original sediment series.

Associations studied here indicate a positive linkage between sediment yield to Ape Lake and spring snow cover and ablation season temperatures. The correlation between snow cover and sedimentation rates is thought to be related to contributions from both glacial and non-glacial terrain surrounding the basin. The most striking feature of the record is anomalously high sedimentation rates in some years. These major units show considerable correspondence with high-magnitude runoff events recorded in the regional drainage network.

In general, glacier extent is directly related to sedimentation rate inasmuch as intervening storage capacity increases as glacier size decreases. This was a strong local effect in Ape Lake and has been noted for other glacio-lacustrine lakes in the Canadian Cordillera (*cf.* Leonard 1981). Implications for the whole of Bella Coola basin are less clear since it is the transport of clastic material from the glacier forefield which is most apt to have an impact on the downstream environment. More important might be the frequency and size of storage sites such as proglacial lakes and their integration into the high-order drainage network. In this regard, sediment source mapping is a necessary requirement for proper interpretation of down-valley lacustrine records.

Glacier Response Record

The conclusions given above are based on high resolution evidence that provide indications of annual variability. The glacier response record provides a low resolution record only. Contemporary data from southwestern British Columbia were used to estimate the magnitude of climate departures required to cause a significant response in Bella Coola glaciers. A climate-ELA model, calibrated using data from southwestern British Columbia, demonstrates that a 20% increase in winter precipitation or a 0.5 C° decrease in summer temperature (from the 1965-1984 mean index values for the coastal region) will produce a 100 m change in ELA. Glacier snouts in the region show sensitivity to changes in climate after a minimum of 8 to 12 years of persistently above average precipitation (7-10%) and respond as soon as 5 years after the effective change in climate (*i.e.* total response time for some glaciers is 13 to 17 years).

Actual glacier movement and the formation of sedimentary evidence which records these movements are responses to the persistence of climatic departures. Glacial moraines provide a record of changing environmental conditions at a much lower resolution. At best, decennial fluctuations in glacier front positions may be discerned if climatic changes are both persistent in time and consistent in direction, and the perturbations are sufficiently extreme in relation to the size of glacier being affected. More typically, glacier responses on the order of two or three decades can be identified using the moraine sequence at low gradient sites where recessional moraines have a high probability of preservation. The resolution of glacier moraine chronologies is further limited by the nature of techniques used to establish the glacier chronology. Dendrochronologies cannot be firmly established at all sites because of ecesis effects but the available evidence indicates that not more than 30 years is required for colonization. Large differences in lichen growth (*i.e.* maximum diameter) and soil development (*i.e.* depth of Bm horizon) and are evident between pre- and post-Little Ice Age surfaces. Corroboration of advance and retreat using tree damage from glacier overriding is desirable but was not observed at any of the sites.

Lichen and soil were used primarily as dating tools for establishing relative differences in the age of glacier moraines. At best, these proxy data discriminate between surfaces that vary in age by several decades or more (Birkeland, 1974; Innes, 1985b). As confirmatory techniques dating by lichen and/or soil can be a valuable means of assessing age differences provided that complicating factors such as elevation, aspect and slope stability are accounted for (Innes, 1985a). The comparatively youthful character of most moraines in the Bella Coola area precluded the development of a useful lichen curve.

Alluvial Response Record

Reaction wood in trees gives a record of disturbance which is due to discrete hydrologic events (*e.g.* floods) at specific sites on the floodplain and tributary alluvial fans. Younger specimens along the margins of tributary alluvial fans are the best source of information regarding the frequency of recent flood events. Conifers offer the best record due to distinct ring boundaries. Cottonwood, and in many instances alder, species which dominate newly constructed alluvial surfaces of the Bella Coola River for up to 100 years, are less desirable due to diffuse ring-boundaries. Even with attempts to enhance the recognition of growth irregularities (*e.g.* use of wood stains) these species appear to be less sensitive to inundations of flood waters. Only a very short and recent record of flood damage (50 - 60 years) could be obtained. Older trees on adjacent and higher surfaces yield evidence of several major floods during the late 18th and early 19th centuries. Basal scarring rather than reaction wood is the primary evidence for extreme events from these older trees. Flooding at several additional sites was thought to have occurred but a lack of consistent evidence from several regions of the floodplain may be related to isolated flood damage due to channel shifting and log jams.

Pollen variation in alluvial sediments from several depositional environments provided valuable information regarding the rates of floodplain sediment accumulation. Several factors complicated the exercise and these include: pollen sources, which were difficult to sort out because of local or regional effects based on both airborne and riverborne transport processes introducing pollen to the site; preservation of pollen in the coarser-grained alluvial sediments (*i.e.* medium sand or coarser), which is not well documented; and the possibility for the pollen evidence to be very site-specific, depending on the configuration of the channel and depositional site.

The resolution of alluvial sediments within the floodplain depends on the type of depositional environment. Vertical accretion deposits in back-channel areas and on distal floodplain surfaces offer the best environments for preservation. Sites containing sediments with the highest temporal resolution are connected to primary river flows on a seasonal basis by one or more overflow channels. Inundation occurs at moderate or high stages and low runoff years are usually not represented in the sediments. Macrofossils (autumn leaf litter) are evidence for the possibility of annual resolution at some sites.

Channel morphology in reaches which are transitional between a laterally stable and unstable habit provide the best locations for inferences about changing hydrology. Examination of floodplain age reveals that less than 30% of the floodplain is older than 280 years, leaving only a small number of sites available for long-term reconstruction. The last millenium is probably the maximum time period open for study in these laterally unstable coastal rivers. The chronology of flood units within the sediments must be established using absolute dating control. Charcoal is abundant in floodplain sediments of Bella Coola River but is a particularly poor material for dating alluvial sediments because of a recycling effect.

Overbank sedimentation is mostly related to the key flood events which are not exclusively linked to a specific set of mean climatic conditions. However, because the largest responses of the alluvial system, particularly morphological adjustments, occur as a result of a series of concentrated flood events, periods of high sedimentation can be supposed to represent intervals of increased flood frequency. This conclusion can be made because of the apparent vertical stability of the river and known morphologic histories of specific sites. In the absence of an understanding of the processes of floodplain development such conclusions can not be substantiated.

General Conclusion

True paleohydrological models are based on confident identification of the functional relationships between hydroclimatic and biogeophysical elements selected to extend the record. The major conclusion here is that events of several types are characteristically mixed in a response record. Proper analysis must include deconvolving the separate signals and then building each into the model for synthesis. In the case of drainage basin development in glacierized regions of coastal British Columbia the extreme events appear to be, at the least, equally important in determining adjustments in morphology and sediment transfer processes.

Except in very favorable circumstances, paleoenvironmental reconstruction methods do not have the resolution promised. Geophysical records of environmental change are the most complex elements studied here and generally yield lower resolution and less precise reconstructions when compared with biological sources. The evidence may be both discontinuous and incomplete. The distinct advantage is that the response records contain information related to events which may have been proportionally more significant in long-term adjustments of the sedimentary system. The regional significance of the reconstructions based on geophysical evidence is less clear because: 1) the response threshold can be site specific (*e.g.* elevation of an overflow channel) and 2) corroborating evidence of this nature is lacking from elsewhere in the Coast Mountains.

The implication for standard methods of paleo-environmental reconstruction is two-fold. First, and most important, is that analysis of a response record for long period variations using linear modeling techniques may provide only part, and in the case of environments such as coastal British Columbia, only a small part of the overall character of environmental change. Secondly, if an understanding is to be gained concerning earth surface responses to climate change then the integrated paleohydrological approach should be an attractive one.

(8.2) Recent Environments of the British Columbia Mid-Coast

Twentieth century climate for the central coast of British Columbia has exhibited variations roughly coincident with major hemispheric changes in global circulation. Meridional circulation in the early 20th century and post-1945 interval produced persistent departures in seasonal temperature and precipitation indices (see figure 3.1). Zonal circulation dominated the interval 1920 to 1945, resulting in greater year-to-year variability in precipitation. Post-1945 data indicate that sea surface temperature anomalies are strongly linked to variations in the position and frequency of blocking ridges of high pressure, hence, precipitation variability (see figure 3.2). Above average SSTs produce consistently above average precipitation along the north coast and below average precipitation along the south coast. Secular trends in temperature are more spatially homogeneous. A fall storm signal is evident throughout much of the south coast yielding a coherent pattern of runoff frequency, but actual runoff magnitudes are dependent on antecedent moisture and specific storm track orientation and so vary within the region. High magnitude autumn floods have a higher probability of occurrence during periods of meridional flow but are not restricted to this circulation regime (consider for example the floods of 1934 and 1936).

Tests for these trends in hydrological, biological and geophysical records during the period of instrument records reveal the following: 1) increased snowpack accumulations, higher spring runoff and the turning of glaciers followed persistent positive anomalies in precipitation after the mid-1960s - an unprecedented trend in the 20th century record of observed and inferred hydroclimates for the central coast; 2) growth anomalies in trees tend to cluster in intervals which correspond to the three periods defined on the basis of circulation type (pre-1920, 1921 to 1945 and post-1945); and 3) channel changes between 1900 and 1979 account for almost one quarter of the newly constructed valley-fill -an amount which is disproportionately large in relation to changes over the last 280 years. Similar studies have not been conducted on other floodplains of this size and in this region but it is likely that similar high turnover rates do occur.

Post-1960 floods mobilized sediment stored in the lower reaches of tributary streams and significantly widened channels on outlet alluvial fans. Although evidence is limited to only a few sites in river back channel areas that have existed throughout the 20th century, it appears that post-1960 floodplain sedimentation rates were the highest. The implication is that there was an increase in sediment transport and perhaps sediment yield during this period.

Little Ice Age Environments

Several lines of evidence point towards specific departures in temperature and precipitation during the Little Ice Age. Paleo-glacial evidence produces a wide range of estimates for the lowering of the climatic snowline during the Little Ice Age but the methodologically consistent technique of determining the lowering of the glacial equilibrium line altitude produces a realistic estimate of approximately 100 m. Using the ELA-climate model a maximum increase of 20% in winter precipitation above 1951-1980 normals is inferred for the Bella Coola basin. The glacial moraine chronology for Bella Coola basin, similar to that shown elsewhere in the southern Coast Mountains, demonstrates that glaciers reached Little Ice Age maximum positions during the middle of the 19th century. Retreat of most glaciers did not commence until the late 19th century.

Growth variations in Douglas fir and subalpine fir provide corroborating evidence for the timing and magnitude of climatic departures during the Little Ice Age advance of glaciers. It is unlikely that the tree ring records span the entire Little Ice Age interval but they do indicate a deterioration of climate during the late 18th and early 19th centuries which maintained glaciers in their advanced positions. Below average temperatures between 1695 and 1765 may have facilitated the turning of some glaciers. An inferred trend towards increasing annual precipitation and decreasing summer temperatures began around 1800 AD. Temperatures were average to below average between 1800 and 1855 but inferred precipitation was, on average, between 25 and 30% greater than the 1951-1980 mean. This led to maximum advances by the middle and late 19th century. Glacier recession, which commenced at some sites by the 1870s, was slowed by the return to cool and wet climates during the interval 1885 to 1900. Based on interpretations of the tree-ring record, the interval 1900-1950 was probably the warmest and driest in the last 330 years.

Evidence presented in chapter 6 indicates that the frequency of extreme departures in tree-growth may reflect characteristics of the general circulation. Inferences about atmospheric circulation are made by establishing periods in which persistent or high year-to-year variance in tree-growth occurred. Zonal circulation is inferred for 1710 to 1745, 1790 to 1800 and 1830 to 1870. The latter interval partly overlaps with high year-to-year fluctuations in sedimentation at Ape Lake (specifically 1836 to 1844 and 1862 to 1885). These show some agreement with circulation trends presented by Blasing and Fritts (1976). The frequency of flooding is thought to have been lower during those periods dominated by zonal circulation.

Tree damage on the Bella Coola River floodplain indicates that a major flood probably occurred in 1805/06. Douglas fir growth during these same two years was strongly positive suggesting well above average precipitation. Flood damage and anomalously high growth of Douglas fir were also noted for 1826 and 1885. This later event (1885) is also recorded by well above average sedimentation in Ape Lake. Other inferred flood events from the tree record occurred in 1785 and 1896, the latter of which is supported by observations of early settlers in the valley. However, these events do not correlate across all proxy data sources.

Accumulation of sediments in overbank deposits indicates that the frequency of overbank flows in the last century has probably exceeded those of any other time during the last millenium (see evidence from Brekke site - chapter 7). This conclusion remains tentative since the error associated with vertical and lateral changes in river position cannot be precisely determined.

(8.3) Future Directions

Spatial variations in the climatological response of coastal areas in British Columbia to changing circulation regimes require careful examination of climatological and hydrological records from a much larger area than that studied here. In particular, the differential trends in precipitation between north and south coastal areas must be documented from a wider regional analysis. Additionally, the implications of persistence in synoptic scale circulation must be sorted out more completely if biological and geophysical evidence is to be of any value in the extension of these records. This should involve careful evaluation of homogeneity amongst hydroclimate stations because of the strong west to east climatic gradient.

Modifications to the alluvial environment are indirectly related to changing hydrological conditions. It is the direct effect of sediment redistribution and ultimately sediment supply which influence river channel patterns and floodplain development. Clastic sediment sources and routing pathways over basin scales similar to that of the Bella Coola should be examined more closely (comparisons can be made with Homathko, Squamish, Lillooet and other rivers). In particular, the important sediment sources which supply debris to the high-order fluvial network need to be identified, and the potential residence times at specific storage sites should be determined. Sources appear to be very much more localized than conventional geomorphological theory would suggest. While sediment budgets of this type are usually only practical for very small basins, an integrated approach which utilizes a variety of remotely sensed data sources may yield results which are practical for planning and resource development.

High-resolution biogeophysical evidence, in particular tree-ring and lake sedimentation chronologies, should be sought elsewhere in coastal basins since it appears to provide a detailed, albeit potentially complex, record of environmental changes. Other tree species could be sampled and alternative modeling techniques employed. The Little Ice Age time frame is particularly appropriate for three reasons: 1) the most significant Holocene environmental changes appear to have occurred during this interval and 2) calibration and verification of growth or sedimentation models to changing hydroclimates can be more readily accomplished over these shorter time periods and 3) practical implications for contemporary planning exercises.

Non-linear modeling techniques which take advantage of the high-frequency information within the data series may provide a better separation of extreme events from departures much closer to the mean. In this context exceptional hydroclimatological events can be isolated in the record and their importance evaluated in terms of impact on the geophysical environment. At the same time this would help reduce the standard error on linear models and improve the estimation of responses in such parameters as seasonal or annual water yield. In essence, an elaboration of figure 1.1can be made.

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APPENDIX A

RADIOCARBON DATES OF SELECTED DEPOSITS FROM THE BELLA COOLA BASIN

(A.1) Radiocarbon Dating

Radiocarbon dating of organic materials collected from various deposits was undertaken to supplement a small number of previously published and unpublished ^{14}C dates from in and around the Bella Coola basin. The types of deposits include marine clays, floodplain sands and silts, deltaic sands and gravels, lake silts and glacial till. The dated materials were mostly wood or charcoal fragments. A summary of ^{14}C dates is given in table A.1.

Fluctuations in the concentration of atmospheric ^{14}C, particularly during the late Holocene Epoch, result in a discrepancy between radiocarbon dates and calendar years (Stuiver, 1982). For example, a radiocarbon age of 150 ± 30 years before present may represent any calendar date between 1680 and 1950 AD. During the last 1000 years the discrepancies are greatest for radiocarbon ages centered on 900 years B.P. and those materials younger than 370 radiocarbon years. Other sources of evidence must be used to resolve any inconsistencies surrounding the ^{14}C date. Ages reported in Table A.1 are those provided by the respective laboratories.

Table A.1. Radiocarbon Dates in and near Bella Coola River Basin.

Laboratory dating No.[1]	Radiocarbon years B.P	Location Lat. N	Long. W	Elevation (m)[2]	Dated Material	Dated Sediments	Reference
S - 2697	<100	$52^{\circ}23'$	$126^{\circ}38'$	35	wood(Thuja plicata?)	floodplain sands	this study
GSC - 4163	20 þ 60	$52^{\circ}10'$	$126^{\circ}07'$	1420	wood(silex)	silt, clay and sand	this study
GSC - 4030	460 þ 50	$52^{\circ}11'$	$126^{\circ}22'$	545	wood	glacial outwash	unpublished
GSC - 4191	480 þ 50	$52^{\circ}10'$	$126^{\circ}21'$	595	wood(Thuja plicata)	till	this study
GSC - 4028	770 þ 60	$52^{\circ}05'$	$126^{\circ}10'$	1395	wood	lake silts	unpublished
S - 2698	955 þ 120	$52^{\circ}24'$	$126^{\circ}31'$	61	charcoal	floodplain sands	this study
S - 2730	1725 þ 145	$52^{\circ}24'$	$126^{\circ}31'$	61	charcoal	floodplain sands	this study
Gak - 3721	3010 þ 100	$52^{\circ}09'$	$126^{\circ}55'$	1070	peat	peat over till	Retherford, 1972
GSC - 4155	2470 þ 50	$52^{\circ}02'$	$126^{\circ}04'$	1370	wood(Abies)	till	this study
GSC - 3964	9550 þ 80	$52^{\circ}29'$	$126^{\circ}33'$	61	wood(angiosperm)	marine clay	this study
GSC - 3980	9580 þ 80	$52^{\circ}23'$	$126^{\circ}36'$	-38	wood(Pinus contorta)	marine clay	this study

1. Laboratories: S - Saskatoon; GSC - Geological Survey of Canada;
   Gak - Gakushuin University
2. Measured with respect to mean sea level.

APPENDIX B

TEXTURAL AND MINERALOGICAL VARIATIONS OF BELLA COOLA VALLEY-FILL SEDIMENTS

(B.1) Introduction

Grain size, lithology and mineralogy of alluvial sediments were used to discriminate between principal sources of sediment for the Bella Coola River. Observations on the distribution of bedrock types indicated that several tributaries might yield sediments of distinct composition which would be traceable in sediments of the main valley. To characterize source area materials subsurface, bulk gravel samples were obtained at the mouths of 11 major tributaries and coarse sand samples were taken from an additional 5 tributaries representing, in total, 95% of the sediment source area for the floodplain. Sand samples were also collected from 11 sub-basins within some of the larger tributaries. Sampling locations are shown in figure 2.2 of the main text.

To characterize alluvial-fill materials bulk gravel samples were extracted from 17 sites along 52 km of the river channel and clast lithologies were identified for 100 grid-sampled, bar-surface particles at each site. Floodplain sediments were extracted from 106 auger boreholes positioned along 18 cross-valley transects. Coarse sands were sampled at the contact between underlying gravels (i.e., auger refusal depth) and the overlying floodplain material: this procedure was adopted as likely to yield the most consistent suite of samples from the boreholes with respect to depositional environment.

(B.2) Bulk Gravel Samples

A set of standardized criteria was adopted for characterizing grain sizes of bulk gravel samples taken from the main river and tributary alluvial fans. Samples from the main river channel were extracted from bank-attached lateral bars and medial bars spaced approximately three river-kilometers apart. Sites were also selected so that at least one sample was taken from positions upstream and downstream of major tributary confluences.

Reconnaissance of gravel bars revealed those locations characterized by the coarsest surface clasts. For medial bars this was always in the bar head area whereas for lateral bars it was either at the bar head or along the outer streamside margin of the bar extending downstream. A sampling area within the coarsest region of the bar was arbitrarily selected by visual inspection and then a sampling plot was located randomly within the area. A similar procedure was used for tributary alluvial fans but in this case the selection of a representative sampling site was restricted to areas near the toe of the fan on bar surfaces within or adjacent to the primary channel.

The texture of surface materials is subject to wide spatial variations due to differential removal of finer sediments during deposition. For this reason subsurface bulk samples were collected using a standardized procedure at each sampling location. A one meter plot was marked off and then surface clasts were removed to expose the underlying subsurface material. A surface layer, defined to a depth equal to the most deeply buried point on a clast exposed in this top layer, was then removed. Finally, 100 to 200 kilograms of the subsurface material was extracted, field sieved to sizes less than $-5.0\ \phi$ and then the sub $-5.0\ \phi$ fraction was split into smaller samples. One of the splits was returned to the laboratory to determine the size distribution for clasts (grains) smaller than $-5.0\ \phi$.

Results

The texture of the channel gravels shows no systematic downstream trend, maximum sizes on the main river bars approach $-7.5\ \phi$ (180 mm)

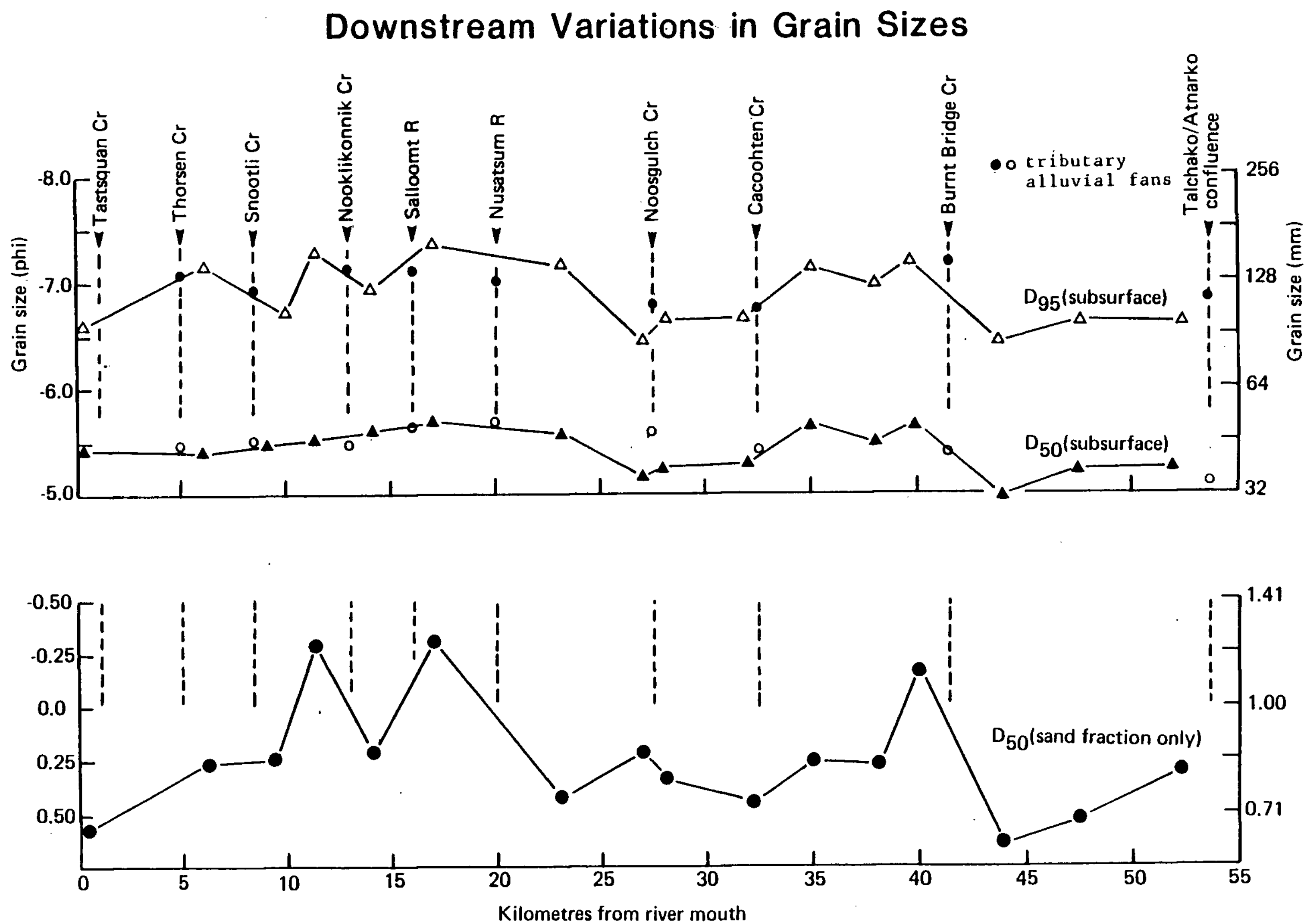

Figure B.1. Downstream variations in D_{95} and D_{50} grain size fractions for bulk gravel samples and sand fraction of subsurface river gravels. Vertical lines represent tributary junctions. Offset circles are respective grain sizes for samples taken on tributary alluvial fans just upstream from the confluence with Bella Coola River.

everywhere (Appendix figure B.1). A noticeable increase in D_{95} grain size is found in gravels immediately downstream of major tributaries such as Burnt Bridge and Noosgulch Creeks and Nusatsum River suggesting some contribution of coarser material from tributary alluvial fans. D_{50} grain sizes show less overall variance (Appendix figure B.1a) and there is a noticeable increase in the mean size of the sub 2 mm fraction downstream of the three main tributaries (Burnt Bridge, Nusatsum, Salloomt) (Appendix figure B.1b). The D_{95} grain sizes of sediments on tributary alluvial fans show little divergence from the sizes available in the main channel.

These results indicate several things. A lack of appreciable attrition of sediments over a distance of 80 km is probably due to a combination of the following factors: relatively resistant lithologies, short transport distances, or an influx of coarser material from tributaries. The increase in size of the D_{95} fraction of samples downstream of tributary fans provides some evidence for the importance these sources. This is further supported by the similarity in textures between alluvial fan and channel zone sediments. Slightly larger mean grain sizes in the lower 20 km of the river, with the exception of those sampled between km 35 and 40, may be due to either proportionately greater contributions from downstream tributaries or an influx of more resistant (plutonic) lithologies from these same sources. Results given below suggest that the latter hypothesis is more plausible.

(B.3) Mineralogy of Valley-fill Sediments

Megascopic identifications of quartz, feldspar, mafic and accessory minerals were made along with chlorite-rich groundmass particles for four groups of 100 grains in the 1.0 to 0.5 ϕ (0.50 - 0.71 mm) size fraction of main channel, tributary and floodplain sand samples. The coarse sand range was used because it is the largest size in which mostly non-aggregate minerals and identifiable rock fragments were visible (Appendix figure B2). Heavy minerals were not analyzed separately because of the selective sorting processes which are common in the deposition of alluvial sediments. Under these conditions heavy minerals may be preferentially deposited or winnowed in certain regions of the bar (*cf*. Ricks, 1985). Microscopic examination of prepared thin sections and examination of other size ranges confirmed the consistency of the results. Clast lithologies were classified using standard petrologic guidelines (Ehlers and Blatt, 1981).

The percentages of certain bedrock types (volcanic, plutonic and mixed) in each tributary were used as grouping criteria in a discriminant analysis of mineral proportions sampled at each tributary outlet (see Appendix table B1 for group assignments).[1] Of the 28 basins or subcatchments sampled all but four were assigned to an expected group (Appendix table B2). The four misclassified cases resulted from situations where sediments appeared to be preferentially derived from one bedrock type even in cases where other bedrock types may dominated a particular catchment. In this regard the analysis was particularly well suited for identifying those catchments which have a somewhat unique distribution of sediment sources in relation to the spatial variation of bedrock lithologies.

Results

The source areas (tributary samples) can be divided into three dichotomous groups that reflect the volcanic, plutonic and mixed bedrock geology (Appendix figure B3). Comparison of the valley alluvial samples

1 Mixed is used for basins with relatively equal proportions of plutonic and volcanic bedrock types. Actual proportions for this category are between 30/70 and 70/30 volcanic to plutonic bedrock types.

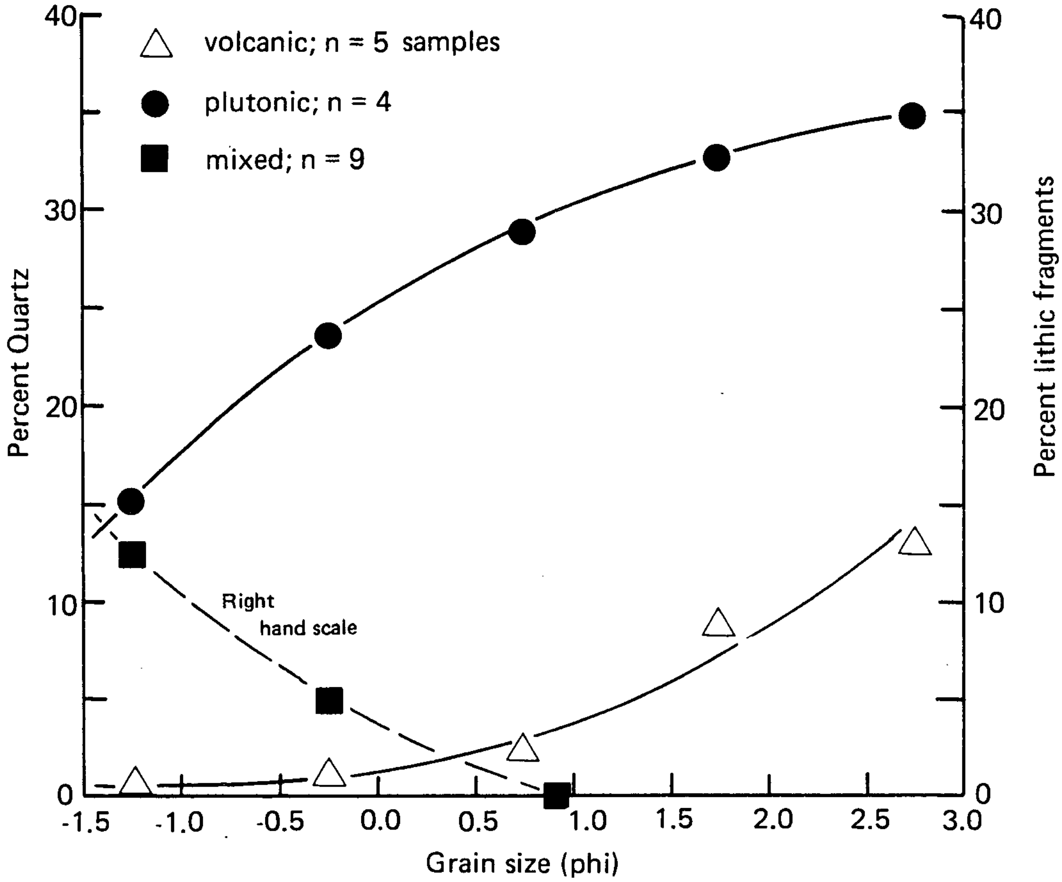

Figure B.2 Variations in the concentration of quartz and lithic fragments in sieve fractions of sand samples taken at the mouths of major tributaries to the Bella Coola River. With decreasing grain size quartz becomes a major constituent mineral in sediments derived from plutonic terrain and is also of increasing importance in sediments derived from volcanic terrain. The percentage of lithic fragments decreases to near zero for size fractions smaller than 1.0 phi in both groups.

Table B.1. Consituent minerals[1] of floodplain and channel zone sediments for the Bella Coola Valley

Site	Q	F	M	C	Gp(2)	Site	Q	F	M	C	Gp(2)	Site	Q	F	M	C	Gp(2)
27	7	38	13	42	1	67	20	43	10	27	2	108	5	32	18	45	1
28	9	34	17	40	2	68	6	35	32	27	1	109	6	41	27	26	1
29	5	35	13	47	1	69	20	40	20	20	2	110	14	76	6	4	2
30	5	34	15	46	1	70	6	50	20	24	3	111	9	31	55	5	1
31	6	36	16	42	1	71	6	56	15	23	3	112	2	43	29	27	1
32	8	40	8	44	1	72	5	49	19	27	3	113	9	41	11	39	1
33	3	31	18	48	1	73	6	52	12	30	3	114	5	33	18	44	1
34	3	36	20	41	1	74	12	48	18	22	2	115	8	39	20	33	1
35	2	32	14	52	1	75	30	60	8	2	2	116	5	40	21	34	1
36	3	25	25	47	1	76	5	49	12	34	3	117	13	43	10	34	2
37	2	34	20	44	1	77	7	38	20	35	1	118	8	40	19	32	1
38	2	30	24	44	1	78	9	34	17	40	1	119	6	36	30	28	1
39	2	25	17	55	1	79	3	38	21	38	1	120	7	37	17	38	1
40	3	30	13	54	1	80	4	37	16	43	1	121	7	41	18	34	3
41	2	34	17	47	1	81	5	38	21	36	1	122	27	58	9	6	2
42	2	36	21	41	1	82	9	42	15	34	3	123	26	57	8	9	2
43	3	50	16	31	3	83	4	37	14	45	1	124	4	35	18	43	1
44	5	55	17	23	3	84	6	46	10	38	3	125	9	41	19	31	3
44	2	38	14	46	1	85	9	34	14	43	1	126	5	40	18	37	1
45	2	33	15	50	1	86	4	46	20	30	3	127	9	50	13	28	3
46	0	38	18	34	2	87	4	41	12	43	1	128	6	40	14	40	1
47	3	29	29	39	1	88	6	35	15	44	1	129	8	36	16	40	1
48	8	77	8	7	3	89	5	41	19	35	1	130	7	41	26	26	1
49	4	75	10	1	2	90	9	42	14	38	1	131	5	39	16	41	1
50	6	41	16	27	2	91	6	27	15	52	1	132	4	41	14	41	1
51	7	33	17	43	1	92	4	26	16	55	1	133	4	30	45	21	1
52	4	39	9	48	1	93	9	47	15	29	3	134	5	28	20	47	1
53	7	36	17	40	1	94	5	50	16	29	3	135	2	35	21	42	1
54	7	55	9	9	2	95	5	46	19	30	3	136	9	30	16	45	1
55	9	46	15	30	3	96	6	62	7	25	3	137	3	40	17	40	1
56	1	45	9	45	3	97	11	41	18	30	2	138	5	34	21	40	1
57	5	41	19	35	2	98	18	41	11	30	2	139	7	37	15	41	1
58	7	61	21	11	3	99	12	36	20	32	2	140	5	39	17	39	1
59	0	35	32	3	2	100	6	36	15	43	1	141	9	35	14	42	1
60	9	37	23	31	1	101	8	35	22	35	1	142	5	43	19	33	1
61	5	51	17	18	2	102	7	35	19	39	1	143	6	49	14	31	3
62	7	45	16	32	3	103	6	31	24	39	1	144	5	44	15	36	1
63	1	50	11	28	2	104	13	34	21	32	2	145	8	50	15	27	3
64	0	25	25	0	2	105	6	62	12	20	3	146	5	34	13	48	1
65	7	40	18	35	1	106	9	35	20	36	1	147	6	34	16	44	1
66	1	29	16	44	2	107	8	45	16	31	3						

1. Q is quartz, F is feldspar, M is mafic minerals and C is chlorite rich minerals.
2. Assigned group is based on discriminant functions: (1) Volcanic affinity (2) plutonic affinity (3) mixed sources.

Table B.2. Alluvial sediment and bedrock types for major tributaries.

Catchment	Area (km^2)	Area (%)	% Volcanic Bedrock	% Plutonic Bedrock	Group[1] Expected	Group[1] Actual	Mineral Content Qtz	Feld	Mafic	Chl	Qtz/Chl
Tastsquan Cr.	29	0.6	0	100	2	2	25	53	21	1	25.0
Thorsen Cr.	88	1.7	69	31	3	2 *	23	61	1	1	23.0
Snooka Cr.	5	0.1	20	80	2	2	28	58	9	5	14.0
Noohalk Cr.	9	0.2	0	100	2	2	22	62	15	1	22.0
Snootli Cr.	39	0.8	6	93	2	2	34	49	16	0	34.0
Nooklikannik Cr.	42	0.8	35	65	3	2 *	28	45	13	15	3.0
Lenci Cr.	5	0.1	3	97	2	2	31	62	7	0	31.0
Salloomt Cr.	170	3.3	69	31	3	3	3	29	29	39	0.08
Tseapseahoolz Cr.	31	0.6	39	61	3	3	14	60	15	11	1.27
Noosgulch Cr.	148	2.9	96	4	1	3 *	1	64	21	14	0.07
Cacoohten Cr.	53	1.1	97	3	1	1	12	50	15	23	0.52
Noomst Cr.	95	1.9	100	0	1	1	1	6	15	78	0.01
Burnt Bridge Cr.	215	4.3	89	11	1	1	4	36	25	35	0.11
Atnarko R.	2,546	50.4	42	58	3	3	3	50	6	31	0.10
Young Cr.	241	4.7	73	27	1	1	2	31	46	21	0.10
Talchako R.	1,402	27.8	81	19	1	1	2	33	15	50	0.04
Tsini Tsini Cr.	47	0.9	100	0	1	1	1	26	22	51	0.02
Nordschow Cr.	97	1.9	100	0	1	1	1	18	17	64	0.02
Gyllensptez Cr.	139	2.8	98	2	1	1	2	33	15	50	0.04
Ape Cr.	106	2.1	70	30	1	3 *	3	33	16	48	0.06
Jacobsen Cr.	336	7.2	85	15	1	1	6	28	29	37	0.16
Nusatsum R.[2]	270	5.3	67	33	3	3	10	38	18	34	0.29
Subcatchments											
Lower Basin A	19.4	7	35	65	3	3	13	36	17	34	0.38
Lower Basin B	24.5	9	85	15	1	1	7	34	32	27	0.25
Middle W. Fork	19.3	7	72	28	1	1	8	72	17	3	2.60
West Fork	34.1	13	100	0	1	1	5	68	7	20	0.25
East Fork	111.0	41	65	35	3	3	29	61	7	3	9.70
Other	62.0	23	59	41	3	3	8	45	37	10	0.80

1. Expected grouping assignment based on proportion of bedrock types in each basin. Actual group is classification accorrding to mineral proportions :(1) volcanic affinity, (2) plutonic and (3) mixed.
2. Alluvial sands were sampled at 4 different locations on the surface of the outlet fan. No statistically significant difference in proportions of minerals was found.

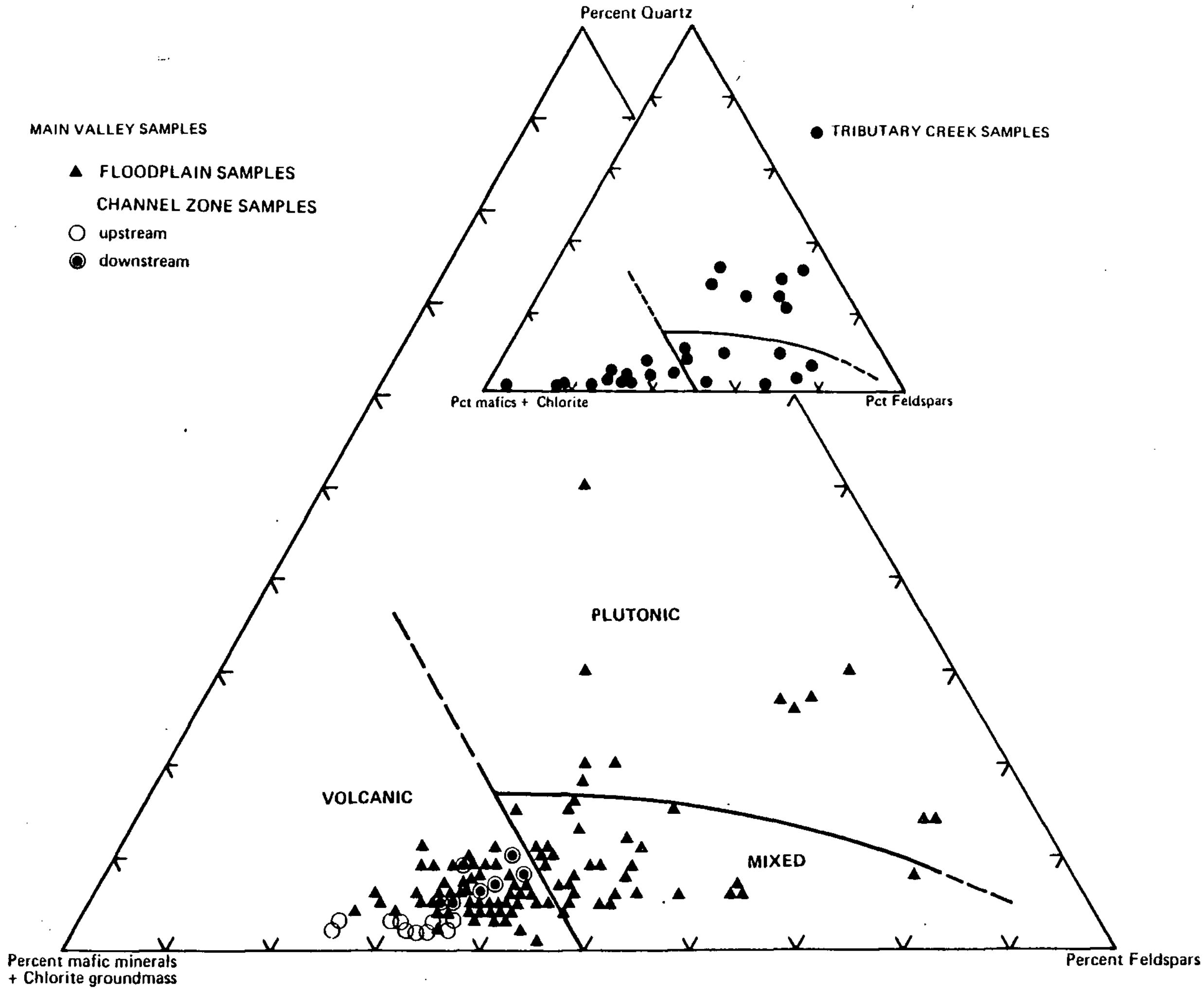

Figure B.3 Ternary diagram showing mineral composition of channel and floodplain sand samples. Boundaries separating the three sediment source groups are based on discriminant functions derived from grouped mineral data of tributary creek sand samples (inset). The discriminating functions correctly classify more than 80 percent of the tributary creek samples.

with these groups reveals the dominance of volcanic source areas. All 17 sand samples from the active channel were classified as having volcanic affinity (Appendix figure B3). A difference of means test conducted for mineral data, grouped on the basis of samples taken upstream and downstream of the Nusatsum confluence, shows a statistically significant difference ($t=2.69$, $\alpha = 0.10$) with higher quartz and feldspar and lower mafic content in the downstream group. In the channel gravels, 61% of the clasts are volcanic and 31% plutonic; a proportion comparable with the ratio of bedrock types in the basin. Unlike the sand-sized mineral data, this ratio does not change significantly when the sample is split into upstream and downstream subsets. However, the plutonic fraction from downstream samples exhibits an increasing percentage of quartz monzonite, a lithology found from the Nusatsum River west. Shale clasts are present in sampling sites immediately downstream from Nusatsum and Salloomt Rivers, the only major sources for this particular rock type.

Of the 106 floodplain sand samples, 51% have volcanic, 9% plutonic and 40% mixed affinity (see figure 3.6 of main text and Appendix figure B3). All sites with plutonic affinity are located downstream adjacent to alluvial fans, tributary creeks or rock walls comprised primarily of quartz monzonite. Samples of volcanic origin are located close to the active channel or active slough channels. Mixed sediments are mostly found in the downstream floodplain within a transition zone between single source sites or downstream from the Nusatsum alluvial fan.

These results suggest that channel zone sediments are derived mostly from upstream source areas. Sediments are transported directly through the channel system. Admixture from downstream plutonic source areas is detectable but not dominant in the channel zone gravels. With increasing lateral distance from the main channel of the lower Bella Coola River there are increasing contributions of fine-grained sediments from tributary alluvial fans.

APPENDIX C

RECENT PATTERNS OF SEDIMENTATION IN THE BELLA COOLA VALLEY

(C.1) Introduction

Results and data given in this appendix pertain to recent sedimentation patterns in several sloughs and back-channels of the Bella Coola River. A brief description of methods is followed by interpretations of mostly flood-related materials deposited in channels which have been abandoned over the last five decades.

(C.2) Sampling Design and Methods

Abandoned channels and backwater areas are common in unstable reaches of the Bella Coola River. Inspection of medium scale air photographs and large-scale maps revealed several potential sites where slough sediments might be used as indicators of recent sedimentation variability. The majority of the sites were below the Nusatsum confluence in very wide unstable regions of the floodplain. Vegetation patterns and examination of aerial photographs taken during the 1980 flood indicate that some of the abandoned channels are periodically active during which time sediments may be introduced. In total, 15 sites were investigated but only six provided sufficient variations and thickness of sediment to be meaningful. The following is a summary of the methods and results from these investigations

Cores from back channels, sloughs and ponds were extracted using a core tube constructed from PVC pipe (9 cm inside diameter). To facilitate sediment extraction the pipe was beveled on the cutting edge and fitted with tin core catchers. Sampling consisted of pushing the cutting edge of the pipe slowly into the softer surface sediments followed by pounding into more resistant basal sands. A medium weight sledge hammer was used for pounding. Coring was done using hip-waders in shallow water (< 1.0 m) and from an inflatable boat in deeper ponds. After extraction, excess water was drained, the tubes were sealed with rubber stoppers and then the cores were transported to the laboratory in an upright position. Cores were split, partially dried, logged and photographed prior to sampling. In order to assess compaction, penetration distances were measured and compared with the length of recovered core. In each case the two measurements agreed to within 10% suggesting that corer-induced compaction was not significant.

Initially, several potentially useful methods for establishing chronologic control of specific sedimentary sequences were investigated. These included radiometric dating (^{137}Cs, ^{14}C), organic matter content, grain size, mineralogy and pollen analysis. Evaluation of aerial photographs, vegetation damage and personal communication with long-term residents in the vicinity of several sampling sites provided additional data. Problems associated with the affinity of ^{137}Cs isotopes for only certain grain-sizes and potential for significant water circulation during the influx of the isotope precluded a feasible radiometric sampling scheme (*cf.* Alberts *et al.*, 1979).

Inspection of the cores revealed sediment textures ranging from fine silt to coarse sand, sometimes layered with macrofossils (whole leaves, twigs) and humified organic material. Sampling for pollen was restricted to the finer-grained sediments since the preservation and decreased vertical mobility of pollen grains are greatest in these materials. In addition to pollen, texture and organic matter contents (loss on ignition) were determined at several levels within each core.

Approximately 1 cm^3 of sediment was removed from each sampling horizon and prepared for pollen analysis using the procedures outlined in Faegri and Iversen (1975). Exotic *Lycopodium* spore tablets of known concentration were added to each sample and then treated with hydrochloric and hydrofluoric acids to remove base metals and silica respectively. Heavy liquid separation ($ZnBr_2$) was then used to separate residual mineral matter from the less dense pollen. The supernatent from the separation was washed, bleached, stained with safranin and permanently mounted on glass slides. Between 200 and 400 pollen grains were counted and identified for each sample and plotted as percentages. Pollen and spores were identified using a modern pollen reference collection together with a key to common pollen and spore types of the Pacific Northwest (R.J. Hebda, G.E. Rouse and M.A. North, in preparation).

Several problems exist with the preparation and interpretation of temporal changes of pollen in sediments. Although generally resistant to the application of acids during sample preparation, pollen grains may be preferentially destroyed through each successive stage (Barber, 1976). This is particularly true for pollen produced by some of the hardwood trees endemic to frequently inundated floodplains (Rouse, personal communication). Certain species produce more pollen than others; thus, although one species may not dominate the surrounding ecological community, it may become over represented in the sediments. This is particularly true in sediments of mixed texture in which the downward displacement of silt and clay-sized pollen grains can occur in coarser sediments and then become concentrated on the surface of finer-grained materials lower in the profile. Finally, pollen found in the sediments sampled in this study are transported by both air and water currents, therefore, both local and regional vegetation are apt to be represented.

Solomon *et al.* (1982) have demonstrated that pollen preserved in alluvial sequences from channel zone or floodplain environments generally represent the following sources: floodplain vegetation, waterborne pollen from upstream and regional vegetation in decreasing order of significance (almost to the exclusion of the final source). However, Brown (1985) has shown that waterborne pollen, albeit still representative of vegetation adjacent to stream sources, can constitute a significant proportion of pollen deposited in backwater and overbank areas of the floodplain. Although identification of all contemporary sources of pollen is beyond the scope of this project, some attempt has been made to determine the relative contribution of riverine pollen by establishing the concentration and type of pollen carried by the Bella Coola River.

Depth integrated water samples extracted for determination of suspended sediment concentrations near the main river gauge were also analyzed for pollen concentrations. Using a set of samples taken during three separate sampling days (all moderate summer flows < 200 m^3/s) the concentrations were approximately 100, 250 and 300 grains per liter. Almost all the pollen was arboreal in type dominated by conifers, *Pinus* and *Abies* in particular. Some *Alnus rubra* and *Monolete polypodiaceae* were noted although in very low concentrations. The overall measured concentrations were very low, between 0.5 and 3 orders of magnitude lower than those reported by Brown (1985), so it is possible that either pollen is concentrated in the surface water layer because of the high velocities throughout the profile or concentrations are low because of the time of year in which sampling was done (late June/early July when flowering of many perennials is complete). Brown (1985) has shown that pollen concentration exhibits many similarities to suspended sediments including increases with increasing discharge and hysteretic tendencies. Therefore, sediments deposited during floods are likely to contain significantly higher concentrations of pollen (one or two orders of magnitude) than those measured here in the main river channel.

Sediment sampling locations are shown in Appendix figure C.1 and descriptions of five of the six main sites are given here. See the main text in chapter 4 for the summary and interpretation of the sixth site.

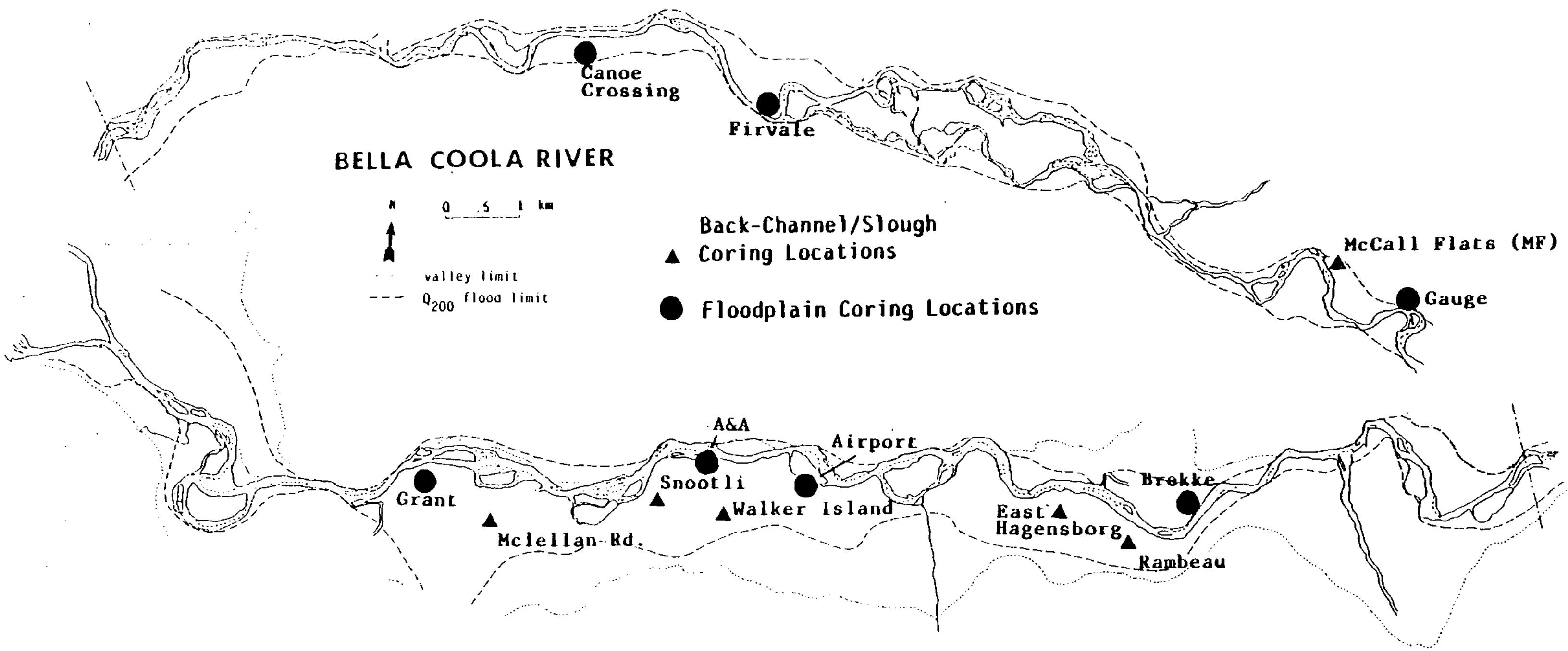

Figure C.1 Sampling locations of slough/backwater channel deposits (see chapter 3) and floodplain sediments (see chapter 7) in the Bella Coola River valley.

Rambeau Slough

Oblique photographs taken at the turn of the century show that this slough was a principal channel of Bella Coola River just below Nusatsum River. The slough may have been a subordinate active channel as late as 1934/36 but was definitely isolated by 1946 and at that time was flanked by an immature cover of scrub alder. Stratigraphy and pollen concentrations of the sediment are shown in Appendix figure C.2a.

Zone I sediments in Appendix figure C.2a show a dominance of alder, grasses and ferns. A trend towards decreasing alder pollen and increasing frequency of conifer pollen is found between zone I and the silty, organic-rich sediments of zone III. Concentrations are also lower towards the top of the core reflecting the decreasing importance of alder in the surrounding vegetation community. Evidence from air photographs, ground surveys and personal communication suggest that the sequence represents continued recovery (succession) of the surrounding vegetation after flooding in 1934/36. The two coarse units in zone II are likely related to the 1965/68 floods, since active flows were observed during both these events. Dyking completed after 1968 has eliminated any further significant flows and the slough now contains stagnante shallow water.

East Hagensborg Slough

This channel was part of the pre-1930 drainage and appears to have been cut off from primary flows in the 1934/36 floods but remained an active back channel until the late 1950s. The channel was abandoned completely by 1961 and has since contained standing water only. With the exception of the disappearance of pine pollen in the upper 30 cm of the sediment there is no strong trends towards changing frequencies or

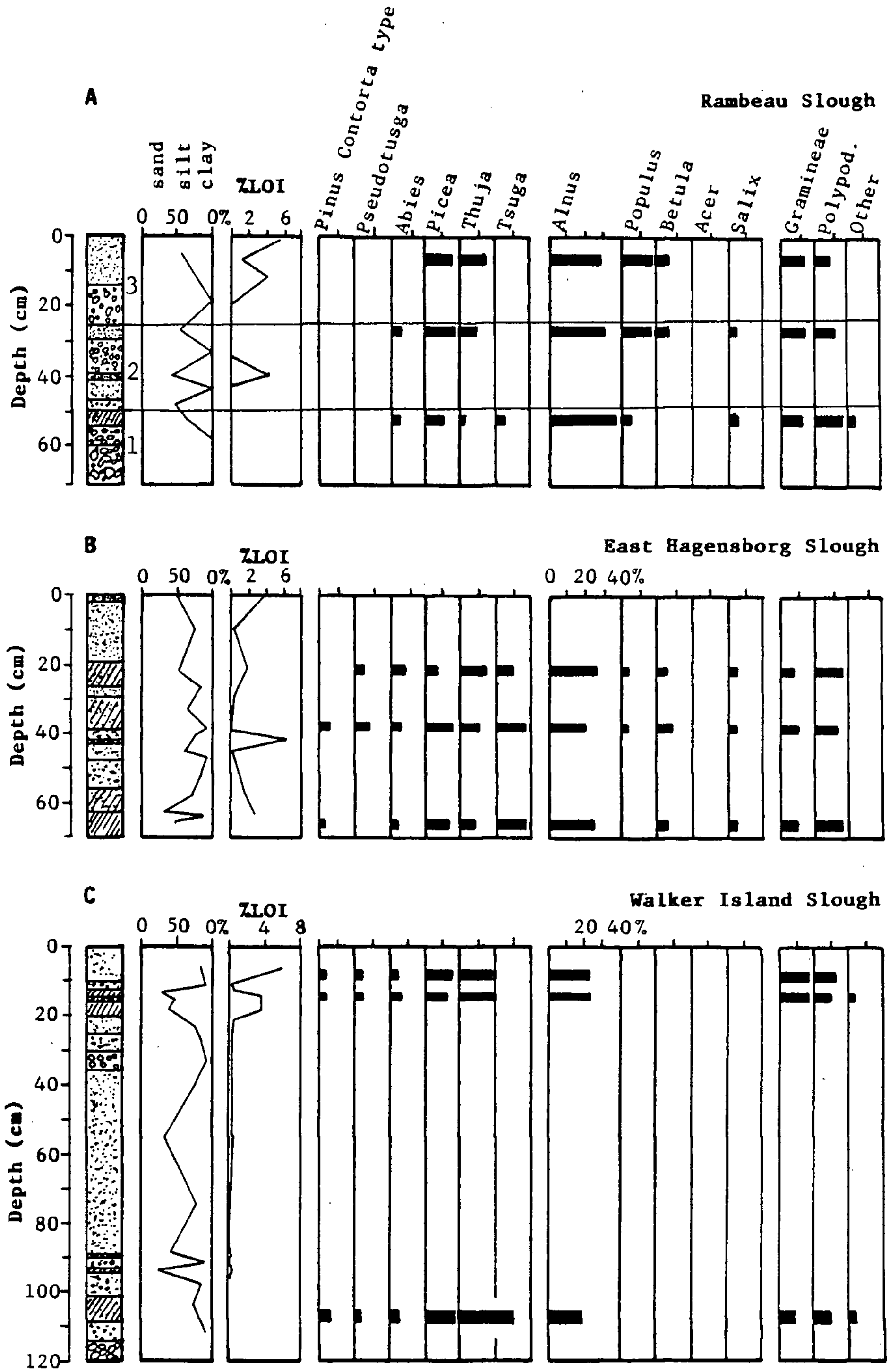

Figure C.2 (a-c) Stratigraphy, grain size variations, organic matter content and pollen concentrations of slough and backwater sediments for selected sites along Bella Coola River. See figure C.1 for sampling locations.

concentrations of pollen (Appendix figure C.2b). The absence of pine pollen may represent the final cutoff of flows from the main river and total isolation of the slough.

Sediments of the upper core are more massive whereas grain sizes and organic matter content show less variability. There is an overall trend towards decreasing grain-sizes and higher organic matter in the upper sedimentation units. Laminations in the lower core are suggestive of changing flow regimes while the upper core indicates more massive sedimentation processes. The combined evidence would suggest that the lower 30 cm of this sequence was deposited as backwater sediments prior to 1961 while the upper 40 cm has accumulated since the channel was completely abandoned in the early 1960s. Although the lower units may have been subject to erosional events it appears that the majority of this sediment was probably deposited prior to 1975 since by this time the main river had shifted northward significantly.

Walker Island

More than a meter of fine sands and silts have accumulated in this pond which is situated in a former active channel of Bella Coola River. The frequency of well exposed gravel bars and scarce vegetation in the 1946 photography of this site suggest the channel was only recently abandoned (1934/36). Although alders appear to have dominated at least one of the channel margins during the initial abandonment phase, mature forest vegetation, particularly hemlock, is best represented in the lower core (Appendix figure C.2c). Two mechanisms may explain this: (1) an influx of pollen from the mature forest which occupies stable surfaces north and west of the pond or (2) pollen contributions from the inundation of Bella Coola flood waters. The absence of coarser-grained flood units in the lower core and the concentrations and types of pollen in river waters given earlier suggest that the former mechanism is most likely.

Massive to partly laminated fine sands represent approximately two-thirds of the sedimentary sequence. Indications as to the origin of these sediments are difficult to detect because pollen recovery was poor. Overlying the fine sand is a unit of coarser sand, which may be related to the last observed flood through the channel in 1968. Slightly higher alder pollen counts in the upper core are probably associated with a significant change in the local vegetation due to commercial logging of hemlock, fir and spruce around 1974. The evidence in this core suggests that most of the sedimentation occurred sometime after 1946 and prior to 1969.

Snootli Slough

Approximately 60 cm of fine sands and silts have accumulated in this channel prior to the early 1970s. The lowest 20 cm of the sequence is characterized by finely laminated sediments dominated by conifers and ferns (Appendix figure C.3a). The laminations may have formed during seasonal inundation of the Bella Coola River since a well developed back channel was directly connected to the slough system until at least the 1965 flood. Higher in the core, alder and grasses are more important and sedimentation patterns less regular. These characteristics are indicative of more episodic inundations and increasing importance of the local vegetation in contributing pollen to the site.

Removal of cedar, hemlock and spruce between 1968 and 1971 from the surrounding floodplain does not appear to have had a substantial impact on the pollen rain at this site, although it may have contributed to the declining importance of conifer pollen in the upper portions of the core. The mineralogy of the sediments lends some supporting evidence to the

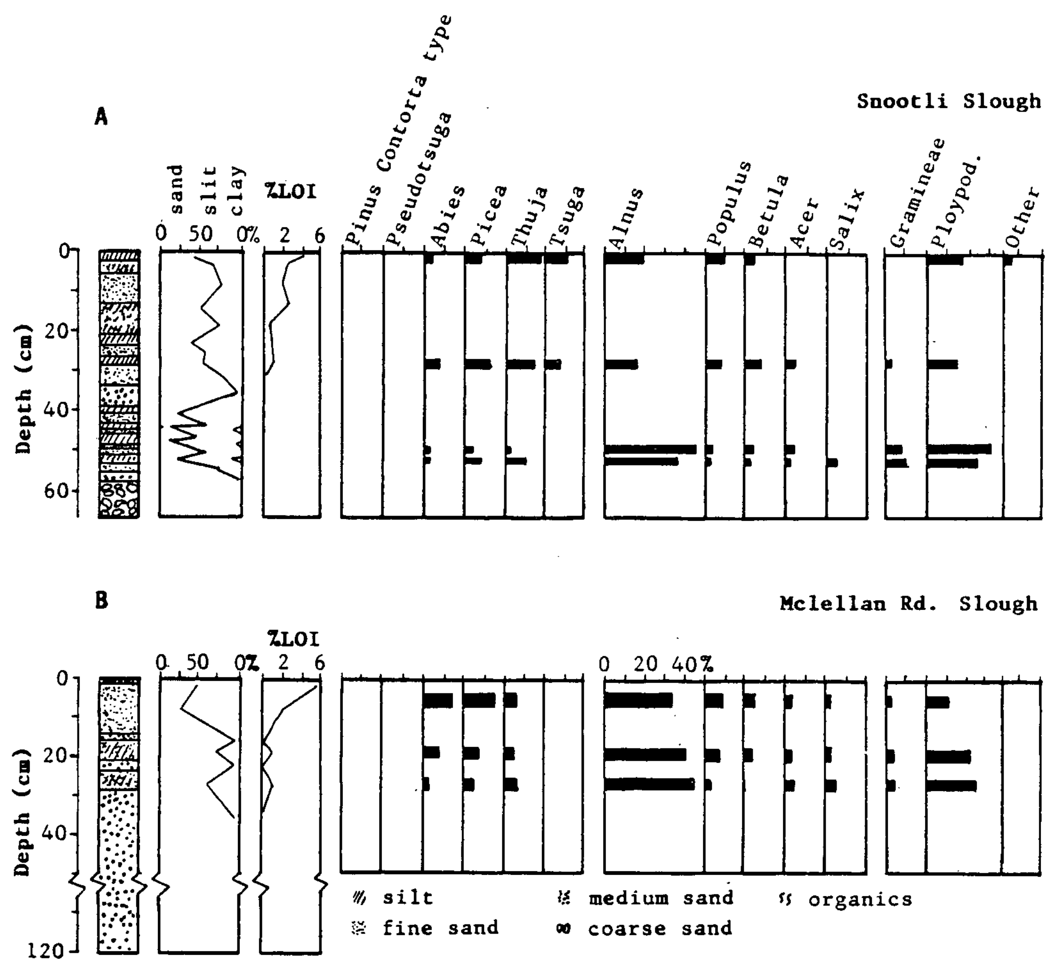

Figure C.3 (a and b) Stratigraphy, grain size variations, organic matter content and pollen concentrations of slough and backwater sediments for selected sites along Bella Coola River. See figure C.1 for sampling locations.

declining significance of Bella Coola waters as a source of sediment and pollen at this site. Constituent minerals were identified in the medium sand fraction at several levels in the core and classified using the discriminating criteria of Appendix B. The lowest 20 cm of the core was classified as having volcanic affinity, or an upstream source, whereas the upper 40 cm were characterized by increasing proportions of quartz, feldspars and biotite. This latter composition is indicative of the plutonic terrain underlying Snootli Creek which enters the floodplain above this site as a series of distributary channels. Flooding from this tributary appears to have had an increasing influence on the nature of sedimentation in the slough channel over recent decades.

Mclellan Road

A series of slough channels upstream of the Thorsen Creek confluence delineate former channel positions of the Bella Coola River. Currently the main stem of the river is on the extreme north side of the valley and during flooding a substantial portion of the floodwaters re-occupy a wide, gravel-bedded channel located between this site and the main river. Hence, inundation of floodwaters to these more southerly sloughs has been less frequent. More than 90 cm of massive to horizontally stratified sands overlie former channel gravels and are thought to have originated from infilling during the late 19th and early 20th centuries (Appendix figure C.3b). Finer-grained materials accumulated in standing water as an early successional forest of alder dominated the surrounding terrain. Subsequent flows have been infrequent and sedimentation rates very low, although it is difficult to determine the interval when inundation became less frequent.

The evidence presented above suggests that there is a high spatial variability in the character and rate of infilling of recently abandoned channels and sloughs of Bella Coola River. This is related partly to real sedimentation differences and partly to shifts in the primary position and degree of anabranching of the main channel. Former channel gravels at all sites are overlain by massive to moderately stratified sands and silts punctuated with organic-rich debris which probably separates flood sediments from over- or underlying backwater deposits and standing water sedimentation. Rates of sedimentation are dependent on the horizontal and vertical displacement of the site from the contemporary river and the frequency of flooding capable of mobilizing sediments in these subordinate channels. River training works, principally dyking, have eliminated further sedimentation at some sites and commercial logging activity has altered the characteristics of the local pollen-rain, thus providing some chronostratigraphic markers. Regional pollen, transported by riverine processes, appear to be incorporated into the sediments in backwater areas of the river but then declines in abundance with the decreasing influence of main river flows.

Although difficult to separate causes, some of the increased sedimentation noted at these sites is likely related to an increased frequency of flooding since the major floods of 1934/36. The implication is that increased slough activity and redistribution of sediment is an important component to changing environmental conditions over the last several decades in the Bella Coola valley. River training works have reversed this trend since 1968 particularly in the lower reaches of the river and may be important in the throughput of sediment.

APPENDIX D

APE LAKE SEDIMENTS:
SAMPLING, MEASUREMENT AND SUMMARY OF RESULTS

(D.1) Introduction

Ape Lake sediments were sampled during the summer and autumn of 1984 and summer of 1985. The first section of this appendix is devoted to a discussion of field methods and laboratory procedures used in the collection and analysis of these sediments. A summary of isotope dating methods considered for the analysis of these sediments is given in the second section and a summary of the sediment data for Ape Lake is provided in the third section.

(D.2) Field and Laboratory Procedures

Sediments from the bottom of Ape Lake were extracted using a portable percussion corer (Gilbert and Glew, 1985) and an Ekman grab sampler. An inflatable Zodiac boat provided a reasonably stable platform from which to operate. The Ekman grab sampler was capable of recovering surface sediments of thicknesses between 5-7 cm whereas the percussion corer yielded cores up to 73 cm long.[2] This model of percussion corer has advantages over other coring devices (*e.g.*, gravity, piston) because of its portability, comparatively light weight and capability to be operated in relatively deep lake settings, *i.e.*, up to 100 m. Samples from the Ekman were extracted at each sampling station by using tin cans pushed slowly into the sediment contained in the sampler. Free standing water was allowed to evaporate from the open ended tins before and after transport to the laboratory until the sediment was firm enough to be handled. Core barrels recovered from the lake bottom were drained, sealed with rubber stoppers and then transported in an upright position to the laboratory. Each core was allowed to

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2. The barrel of the percussion corer consists of 7.37 cm (outside diameter) PVC plastic pipe fitted with a flexible core catcher. The total usable length of the 3.05 m core barrel is approximately 1.5 m.

partially dry within the core barrel until firm enough to be split, which was done by etching the plastic barrel on two sides and then slicing with thin wire.

All sediment surfaces were shaved with a sharp knife, photographed and then prints were produced at scales commensurate with the character of laminations and structures visible in each sample, usually 1:1. Width measurements and observations on texture and structure were made from the percussion cores directly using a calibrated micrometer scale attached to a sliding binocular microscope. Measurements for the larger number of Ekman samples were taken from the photographs directly. All subsequent cross-correlations between grab samples and sediment cores were conducted using the photographs only.

Compaction and Distortion of Sediments

Interpretations of sedimentation rates from badly disturbed or compacted cores are often unreliable. Compaction and distortion may result from sampler design problems, sampling procedure or from the sediment characteristics themselves (Gilbert, 1975). Devices such as the Ekman grab sampler are designed to minimize disturbances to the upper layers of sediment and examinations of Ekman samples from Ape Lake show very little evidence of differential compaction or boundary disturbances among samples. The relatively large dimensions (15.2 x 15.2 x 22.9 cm) and limited penetration of the Ekman preclude any significant compaction of sediment in these samples (Blomqvist, 1985).

It was anticipated that the slow, impulsive movement of the percussion corer would reduce sediment distortion which commonly is found in gravity core samples (Wright, 1987). Distortion ranged from very slight downwarping along the outside edge of the sediment to extensive downwarping and homogenization in limited sections of a few cores. The greatest disturbances were in the upper 2-3 cm of each core where the more recently deposited sediments may have become re-suspended during penetration due to high water contents in layers adjacent to the sediment-water interface.[3] It is probable that the limited distortion occurred during removal of the core barrel from the lake bottom as a result of the relatively high water contents within the sediments. No evidence for bioturbation or dissolution of gases was found in any of the samples.

Compaction of sediments may be induced by the sampling procedure, sampling device, drying or simply by normal post-depositional loading on successive layers. While these four sources of distortion may be difficult to distinguish, for a limited length of core (less than 1 m) diagenetic changes are likely to be least important. Blomqvist (1985) shows that depth of penetration and core tube diameter are the most important factors governing core-shortening. For the core diameter used here (7.37 cm) and a maximum penetration of 1 m as much as 7.5 cm of shortening might have occurred given the relationship developed by Blomqvist. In order to assess the degree of compaction in sediment cores two methods were selected; (i) the measurement of dry bulk density down one percussion core and (ii) a comparison of correlated laminae thicknesses between Ekman samples and percussion cores taken in proximity to each other. The former method is a test for a significant, corer-induced, density gradient indicative of non-linear compaction (*cf*. Leonard, 1981; Emery and Huselmann, 1965) and the latter is an alternative but direct check of differential compaction in surface layers.

3. The mean water content of sediments taken from the grab samples was approximately 65%, but values as high as 85% were measured in some of the sediments.

For a freshly opened core 1 cm^3 plugs of sediment were sampled using a calibrated 2 cm^3 syringe. Pre-weighed samples were oven dried for 24 hours at 80°C, an additional 24 hours at 120°C and then re-weighed. Results, shown in figure D.1a, indicate that some compaction may have occurred in the first few centimeters but dry bulk density values are relatively consistent throughout the upper 20 cm. Higher dry bulk density values towards the base of the core are likely related to textural changes and not compaction. Thicknesses of cross-correlated laminae between Ekman samples (undisturbed) and core samples are plotted in figure D.1b. Values which plot below the 1:1 line may be due to compaction or to real differences in sedimentation rates between the two sampling sites while those which plot above the line are likely due to differential sedimentation rates only (assuming the Ekmans represent undisturbed samples). Both methods show that compaction below 2 cm in the percussion cores is generally less than 10%.

(D.3) Periodicity of Sedimentation in Ape Lake

Since the prediction was made that annually-laminated sediments might occur in lacustrine environments (Wesenberg-Lund, 1901), a variety of methods have been employed to determine the relative and absolute ages of rhythmic lake bottom sediments. These include: the presence of light, coarse laminae alternating with thinner, darker and finer sediments exhibiting a high spatial coherence (Antevs, 1925; Gilbert, 1975); actual measurements of seasonal sedimentation rates using sediment traps (Smith *et al*., 1982; Gilbert and Shaw, 1981) or repeated bathymetric surveys; the use of pollen, diatoms or other biological indicators which fluctuate over a known time interval (Boyko-Diakanow, 1979; Brush *et al*., 1982; Renberg, 1982, 1984); the frequency of couplets between horizons of known age

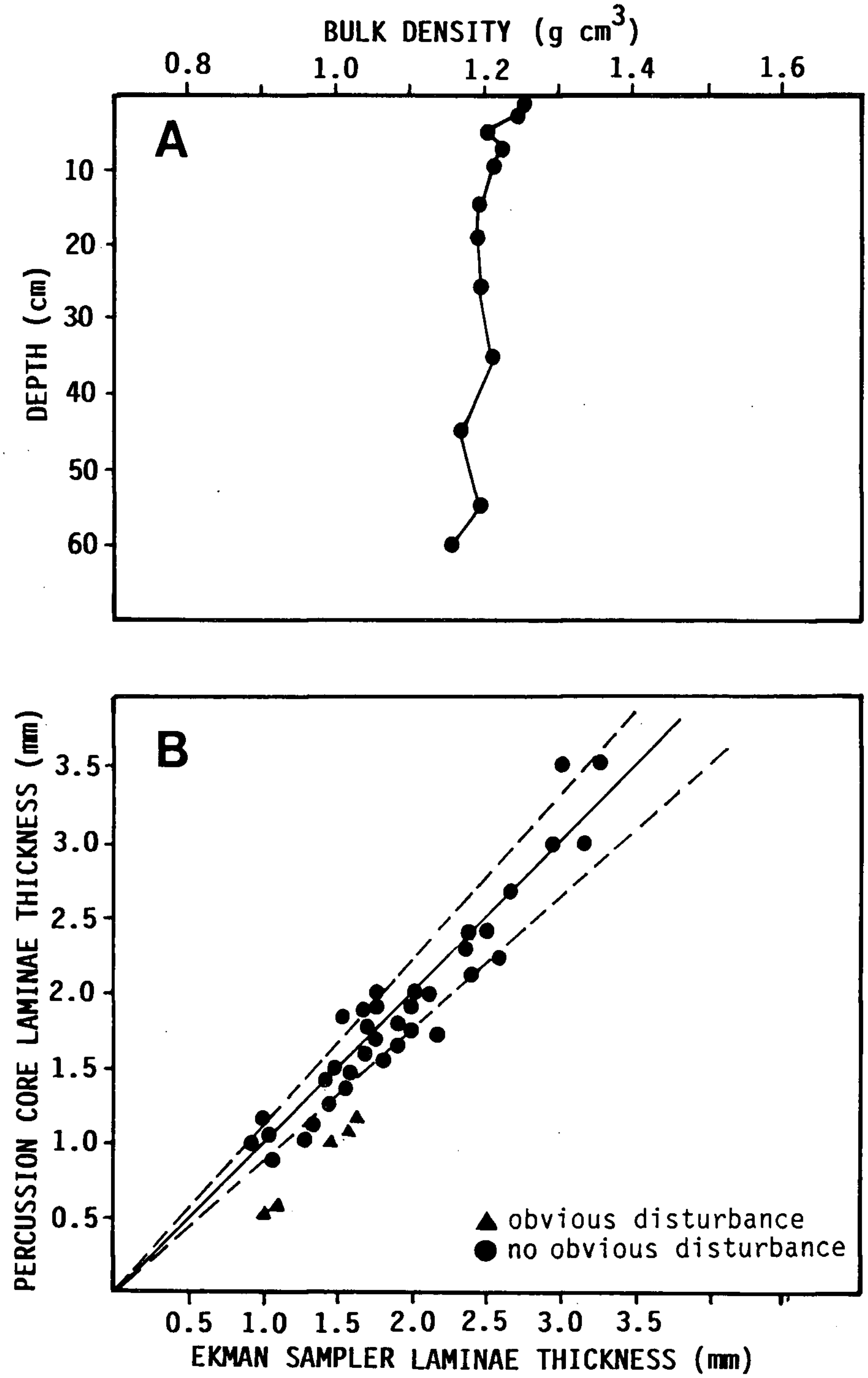

Figure D.1 Estimates of differential sediment compaction in Ape Lake sediments. (A) variations in bulk density measured down core 84C2. (B) comparison of selected laminae thicknesses between cross-correlated units of percussion core and Ekman core samples.

(Ostrem, 1975; Lambert and Hsu, 1979; Ludlum, 1981; Perkin and Sims, 1983); and the use of radionuclides with known flux rates and decay profiles (Antevs, 1957; Ashley and Moritz, 1979; Dominik *et al.*, 1981; Leonard, 1986).

Organic matter contents in all samples from Ape Lake were extremely low (<1%) precluding the possibility of using ^{14}C as an absolute dating tool. Even when present, dating organic carbon in lake sediments using the ^{14}C technique can be misleading. The lake may act as a reservoir for older carbon leading to anomalously old dates (Olsson, 1979) or conversely an influx of modern humus from the surrounding watershed each season can lead to dilution of the old ^{14}C (Karlen, 1981). Lead 210 and the ratio of $^{228}Th/^{232}Th$ (activity ratio) have also been used to establish absolute dates (Appleby and Oldfield, 1978; Bertine *et al.*, 1978). Since ^{228}Th has a comparatively short half-life (1.9 years) it is not a useful method when sedimentation rates are low (Dominik *et al.*, 1981). Additionally, an assumption of the technique is that the lithogenic isotope fraction is in equilibrium and that excess ^{228}Th absorbed onto particles in the water column is lost by decay only and not through exchanges with the lithogenic fraction (Kodie *et al.*, 1973).

Lead 210, a naturally occurring radioisotope, has also been used for establishing dating control of sediment sequences over a period of up to 200 years. However, the technique does not allow for a high-resolution discrimination of unsupported ^{210}Pb activity thus precise calendar dates for a given horizon are not possible (Appleby and Oldfield, 1978). Limitations for these other techniques make cesium-137 a particularly attractive isotope for the dating of recent sediments.

Cesium 137 is not a naturally occurring isotope but was introduced as a result of atmospheric testing of nuclear devices in the 1950s and 1960s.

Significant fallout of the isotope followed testing in 1958-59 and 1962-64 with comparatively smaller amounts introduced over the period 1954-1972. Prior to 1954 and after 1972 atmospheric concentrations were negligible. Measurements of ^{137}Cs concentrations have been made for lacustrine sediments (Ritchie *et al.*, 1973; Pennington *et al.*, 1973; Ashley and Moritz, 1979; Leonard, 1986) and fine-grained solids in surrounding watersheds (Low and Evardson, 1970; McHendry *et al*, 1973; Ritchie *et al.*, 1973; Wise, 1980).

Physical, biological and chemical characteristics of the environment may control the distribution of cesium within a particular catchment. For example, extensive bioturbation can homogenize lake sediments and the associated cesium profile. As with ^{210}Pb dating, an assumption of this technique is the immobility of ^{137}Cs within the sediment column. Dominik *et al.*, (1981) found high concentrations of cesium in the younger surface layer only and suggested that upward molecular diffusion is possible if there is groundwater recharge into the lake bottom. Seasonal cycling of ^{137}Cs has also been reported (Alberts *et al.*, 1979). Dissolution of ferric oxides resulting in the uptake of cesium at the lake bottom may occur during thermal stratification when the hypolimnion is depleted of oxygen. Following lake overturn oxygen is enriched at the lake bottom and the iron oxides are reprecipitated along with cesium.

Results

Analyses of samples from Ekman core 84E2 were conducted by Chemex Ltd using a gamma spectroscopy [Ge(li)] detection system housed at TRIUMF on the UBC campus. Because of the large sample sizes (>9 g) no pretreatment of the material was required. Samples were counted for *ca*. 14 hours using standard 4350B (river sediment) of a similar weight for reference. The 95%

detection limit is approximately 0.1 pCi/g. Concentrations of ^{137}Cs in the Ekman samples are given in table D.1. and plotted in figure 5.6 of the main text. Values range from 0.11 to 19.32 pCi/g the latter of which is amongst the highest recorded for sediments sampled in western British Columbia (G. Leget, personal communication, May, 1986). See chapter V for a discussion and summary of these results.

Based on the cesium results calendar dates were assigned to each lamina in both the Ekman and percussion core samples. A summary of laminae thicknesses and the standardized chronology for Ekman samples is given in table D.2. Table D.3 contains percussion core data and the standardized sedimentation chronology is shown in tables D.4.

Table D.1 Cesium Results for Core 84E2

Depth[1]	Couplet Numbers	Weight (g)	137Cs (pci/g)
7	3-5	8.70	5.91
13	6-9	9.15	5.82
21	10-13	9.12	8.28
25	14-16	9.80	8.34
28	17-18	9.47	15.68
33	19-21	9.80	19.32
41	21-25	9.92	7.87
48	26-27	9.93	2.90
53	28-30	9.87	1.91
58	31-33	9.80	0.17

Depth to bottom of lowest laminae.

Table D.2. Sedimentation indices derived from Ekman Cores

	Individual				Composite(mm)
YR	84E1	84E2	84E7	84E10	Index[1]
1984	0.00	0.59	2.33	0.81	1.16
1983	1.11	0.74	1.21	0.90	0.94
1982	2.72	1.48	1.52	2.53	1.96
1981	1.01	0.81	1.42	1.75	1.18
1980	1.41	0.81	1.31	1.23	1.12
1979	0.91	0.74	1.01	0.81	0.83
1978	0.60	0.44	0.61	0.78	0.58
1977	0.70	0.71	0.81	0.76	0.72
1976	2.42	1.51	2.43	2.17	2.02
1975	1.01	0.77	1.31	0.76	0.91
1974	1.01	0.74	0.91	1.05	0.89
1973	0.81	0.56	1.01	0.81	0.75
1972	0.81	0.44	0.91	0.54	0.631.
1971	1.21	0.81	0.81	0.81	0.87
1970	0.81	1.11	0.50	0.72	0.78
1969	1.21	1.04	1.11	1.62	1.20
1968	1.21	0.96	1.62	1.08	1.15
1967	1.41	1.41	1.11	1.35	1.29
1966	0.70	0.59	0.81	0.54	0.63
1965	1.11	0.74	1.01	0.90	0.89
1964	0.81	0.81	0.91	0.90	0.83
1963	1.41	0.89	1.21	0.90	1.05
1962	1.05	1.11	0.91	0.81	0.95
1961	0.77	0.59	1.01	0.63	0.71
1960	1.41	1.33	1.21	1.26	1.05
1959	2.22	1.78	2.02	1.53	1.52
1958	1.46	1.33	1.52	1.05	1.08
1957	0.70	0.74	0.71	0.58	0.55
1956	1.01	1.04	0.61	0.90	0.71
1955	0.60	0.81	0.61	0.81	0.57
1954	1.21	1.41	0.61	0.99	0.84
1953	2.22	2.44	1.92	1.98	1.72
1952	0.00	1.85	0.00	1.71	1.45
1951	0.00	0.81	0.00	0.99	0.74
1950	0.00	1.11	0.00	0.00	0.87
1949	0.00	1.18	0.00	0.00	0.93
1948	0.00	0.74	0.00	0.00	0.58

1. Sedimentation indices were calculated using period means for 1948-1960 and 1961-1984. An unweighted average of all four Ekmans was used to produce a composite index. Indivdual yearly values were divided by the respective period means.

Table D.3. Winter (w) and summer (s) lamina thickness for Ape Lake percussion cores (mm)

	84C5		84C2		84C4		84C3	
YEAR	S	W	S	W	S	W	S	W
1965	1.2	0.6	1.5	0.7	1.1	0.6	0.0	0.0
1964	1.1	0.6	2.0	0.6	1.3	0.5	0.0	0.0
1963	1.9	0.4	1.9	0.8	1.3	0.6	0.0	0.0
1962	1.0	0.9	1.6	1.0	1.4	0.8	0.0	0.0
1961	0.8	0.6	1.0	0.5	0.5	0.4	0.0	0.0
1960	2.0	0.6	2.0	0.9	1.7	0.7	0.0	0.0
1959	3.0	1.8	3.0	0.8	2.7	1.6	0.0	0.0
1958	2.0	1.0	2.0	0.5	1.6	0.5	0.0	0.0
1957	1.2	0.4	1.0	0.8	0.5	1.0	0.0	0.0
1956	1.1	0.5	1.7	0.8	1.0	0.7	0.0	0.0
1955	1.8	0.8	1.8	0.4	0.9	0.7	0.0	0.0
1954	2.9	0.9	2.4	0.9	1.9	0.7	0.0	0.0
1953	4.0	1.2	3.9	1.1	3.7	1.3	0.0	0.0
1952	3.2	1.0	4.8	1.2	3.6	1.5	0.0	0.0
1951	1.1	0.8	1.2	0.9	1.6	1.1	2.2	1.2
1950	1.4	1.0	1.6	1.2	1.4	1.0	2.0	0.8
1949	2.0	0.8	2.0	0.8	1.9	1.0	0.6	0.8
1948	3.8	1.0	4.0	0.4	3.2	0.7	2.0	0.8
1947	2.0	0.2	3.0	1.1	1.4	1.0	1.0	0.8
1946	2.8	0.6	2.0	1.0	2.4	0.4	1.9	1.0
1945	5.8	1.2	5.0	1.7	6.0	1.0	3.0	0.8
1944	7.1	0.8	6.1	1.1	8.0	0.5	5.5	1.1
1943	3.0	0.9	2.7	0.6	2.2	0.5	1.9	0.8
1942	2.4	0.8	2.0	0.6	2.0	0.5	2.2	0.8
1941	3.9	0.7	3.9	0.3	3.5	0.4	2.9	0.9
1940	7.6	1.6	8.0	0.9	9.3	1.0	7.4	1.0
1939	1.8	0.8	2.0	0.3	2.3	0.5	2.0	0.7
1938	1.6	0.8	1.8	0.9	1.8	0.6	1.7	0.8
1937	1.6	0.4	1.9	0.6	2.3	0.5	1.3	0.8
1936	3.9	0.8	5.4	0.7	7.0	1.0	3.6	0.8
1935	1.0	0.2	1.0	0.2	1.8	0.5	0.8	1.9
1934	8.0	2.1	7.6	0.8	7.9	1.8	5.8	1.9
1933	1.2	0.4	1.0	0.7	2.0	0.5	0.8	0.8
1932	2.2	0.8	2.8	0.9	2.7	0.7	1.9	0.8
1931	1.2	0.4	1.0	0.3	1.4	0.7	0.8	0.4
1930	1.0	0.3	0.9	0.2	2.4	0.8	0.8	0.3
1929	0.8	0.2	1.6	0.2	1.8	0.4	0.8	0.3
1928	1.0	0.3	1.0	0.2	4.3	1.0	1.4	0.3
1927	0.9	0.2	1.9	0.2	2.4	0.7	0.9	0.3
1926	1.0	0.2	1.0	0.2	2.7	0.5	0.7	0.3
1925	1.2	0.2	1.0	0.2	1.7	0.6	0.8	0.3
1924	11.0	1 0	9.0	1.0	16.0	1.3	7.0	1.0
1923	2.0	0.2	2.0	0.2	3.0	0.7	1.7	0.5
1922	2.2	0.4	2.2	0.4	3.5	0.7	1.1	0.3
1921	0.8	0.2	0.8	0.2	1.5	0.4	0.9	0.2
1920	0.8	0.3	0.8	0.3	2.2	0.3	0.6	0.2
1919	1.2	0.2	1.2	0.2	5.8	0.6	1.5	0.3
1918	5.1	1.0	5.0	1.0	7.3	4.1	3.0	0.3
1917	7.0	0.8	7.0	0.8	12.8	2.6	4.0	0.5
1916	1.8	1.0	1.8	1.0	7.4	1.7	1.3	0.5
1915	1.8	0.9	1.8	0.9	5.2	0.5	1.0	0.1
1914	2.6	0.8	2.6	0.8	3.3	0.5	1.0	0.1
1913	1.9	0.3	1.9	0.3	3.2	0.7	1.7	0.2
1912	4.1	0.2	4.1	0.2	4.0	0.4	0.8	0.1
1911	1.1	0.2	1.1	0.2	2.7	0.5	1.0	0.2
1910	1.5	0.2	1.5	0.2	2.6	0.5	1.0	0.1
1909	1.8	0.2	1.8	0.2	2.8	0.6	0.7	0.1
1908	1.8	0.2	1.8	0.2	1.8	0.4	0.6	0.1
1907	3.0	0.1	3.0	0.1	3.0	0.3	1.8	0.1
1906	6.8	0.2	6.8	0.2	3.5	0.4	1.9	0.2
1905	5.0	0.6	5.0	0.6	5.9	0.5	1.0	0.2
1904	9.0	1.0	9.0	1.0	8.0	0.5	2.6	0.2
1903	7.1	0.8	7.1	0.8	0.0	0.0	2.2	0.2
1902	4.0	0.5	4.0	0.5	0.0	0.0	3.1	0.2
1901	2.8	0.4	2.8	0.4	0.0	0.0	2.0	0.2
1900	5.8	0.2	5.8	0.2	0.0	0.0	3.0	0.2
1899	3.8	0.2	3.8	0.2	0.0	0.0	1.4	0.2
1898	6.8	0.2	6.8	0.2	0.0	0.0	3.0	0.2
1897	5.0	0.4	5.0	0.4	0.0	0.0	2.0	0.2
1896	10.0	0.4	10.0	0.4	0.0	0.0	4.6	0.2
1895	4.8	0.8	4.8	0.8	0.0	0.0	1.9	0.3
1894	2.0	1.2	2.0	1.2	0.0	0.0	2.2	0.4
1893	4.0	0.2	4.0	0.2	0.0	0.0	4.5	0.3
1892	2.9	0.2	2.9	0.2	0.0	0.0	2.8	0.3
1891	1.3	0.2	1.3	0.2	0.0	0.0	2.1	0.3
1890	4.0	0.2	4.0	0.1	0.0	0.0	7.0	0.3
1889	5.6	0.2	5.6	0.2	0.0	0.0	6.0	1.0
1888	5.4	0.8	5.4	0.8	0.0	0.0	4.9	0.3
1887	7.8	1.2	7.8	1.2	0.0	0.0	5.2	0.5
1886	8.0	1.1	8.0	1.1	0.0	0.0	2.9	0.8
1885	14.1	1.0	13.0	1.0	0.0	0.0	7.6	0.6
1884	3.0	1.2	3.0	1.2	0.0	0.0	2.0	0.4
1883	4.7	0.4	4.7	0.4	0.0	0.0	4.0	0.3
1882	2.8	0.4	2.8	0.4	0.0	0.0	3.5	0.4
1881	0.8	0.2	0.9	0.1	0.0	0.0	1.0	1.0
1880	1.8	0.2	1.8	0.3	0.0	0.0	2.3	0.9
1879	3.0	0.3	2.1	0.6	0.0	0.0	3.3	0.8
1878	4.0	0.3	2.2	0.2	0.0	0.0	3.2	0.5
1877	7.5	0.8	6.1	1.1	0.0	0.0	7.9	0.4
1876	3.6	0.8	2.8	0.3	0.0	0.0	1.9	1.0
1875	6.7	0.9	4.5	0.7	0.0	0.0	4.8	0.9
1874	4.9	1.2	4.1	0.7	0.0	0.0	4.0	0.9
1873	5.0	0.6	3.9	0.6	0.0	0.0	2.3	0.8
1872	11.0	1.0	6.1	1.1	0.0	0.0	4.7	1.2
1871	12.5	2.8	7.0	1.2	0.0	0.0	6.8	0.6
1870	6.0	1.0	4.0	0.6	0.0	0.0	4.0	0.8
1869	8.9	1.2	9.1	2.2	0.0	0.0	6.6	0.9
1868	3.8	0.8	4.0	0.7	0.0	0.0	4.7	0.5
1867	3.1	0.6	3.5	0.6	0.0	0.0	2.5	0.8
1866	3.1	0.4	4.2	0.7	0.0	0.0	3.0	1.0
1865	10.0	1.0	10.0	0.8	0.0	0.0	7.0	0.9
1864	0.0	0.0	3.5	0.8	0.0	0.0	0.0	0.0
1863	0.0	0.0	11.0	0.6	0.0	0.0	0.0	0.0
1862	0.0	0.0	4.0	0.7	0.0	0.0	0.0	0.0
1861	0.0	0.0	3.2	0.7	0.0	0.0	0.0	0.0
1860	0.0	0.0	5.2	0.7	0.0	0.0	0.0	0.0
1859	0.0	0.0	2.0	0.3	0.0	0.0	0.0	0.0
1858	0.0	0.0	5.2	0.9	0.0	0.0	0.0	0.0
1857	0.0	0.0	4.0	0.8	0.0	0.0	0.0	0.0
1856	0.0	0.0	4.6	0.9	0.0	0.0	0.0	0.0
1855	0.0	0.0	6.1	2.0	0.0	0.0	0.0	0.0
1854	0.0	0.0	4.8	0.7	0.0	0.0	0.0	0.0
1853	0.0	0.0	2.9	0.5	0.0	0.0	0.0	0.0
1852	0.0	0.0	4.5	0.7	0.0	0.0	0.0	0.0
1851	0.0	0.0	3.8	0.6	0.0	0.0	0.0	0.0
1850	0.0	0.0	5.1	0.8	0.0	0.0	0.0	0.0
1849	0.0	0.0	4.9	0.8	0.0	0.0	0.0	0.0
1848	0.0	0.0	5.8	0.9	0.0	0.0	0.0	0.0
1847	0.0	0.0	5.8	0.9	0.0	0.0	0.0	0.0
1846	0.0	0.0	6.0	1.8	0.0	0.0	0.0	0.0
1845	0.0	0.0	3.5	0.6	0.0	0.0	0.0	0.0
1844	0.0	0.0	4.0	0.7	0.0	0.0	0.0	0.0
1843	0.0	0.0	12.0	3.0	0.0	0.0	0.0	0.0
1842	0.0	0.0	7.8	0.9	0.0	0.0	0.0	0.0
1841	0.0	0.0	5.2	0.7	0.0	0.0	0.0	0.0
1840	0.0	0.0	4.8	0.8	0.0	0.0	0.0	0.0
1839	0.0	0.0	8.9	2.4	0.0	0.0	0.0	0.0
1838	0.0	0.0	3.0	0.5	0.0	0.0	0.0	0.0
1837	0.0	0.0	2.5	0.3	0.0	0.0	0.0	0.0
1836	0.0	0.0	7.0	2.1	0.0	0.0	0.0	0.0

0.0 - no data

Table D.4. Composite Chronology for Ape Lake Cores

1965	0.30	1911	0.42	1853	0.67
1964	0.34	1910	0.44	1852	1.04
1963	0.40	1909	0.44	1851	0.90
1962	0.31	1908	0.35	1850	1.18
1961	0.18	1907	0.68	1849	1.13
1960	0.45	1906	1.04	1848	1.34
1959	0.68	1905	0.97	1847	1.34
1958	0.44	1904	1.64	1846	1.39
1957	0.21	1903	1.23	1845	0.81
1956	0.30	1902	1.02	1844	0.93
1955	0.35	1901	0.68	1843	2.78
1954	0.56	1900	1.22	1842	1.81
1953	0.91	1899	0.70	1841	1.20
1952	0.91	1898	1.34	1840	1.11
1951	0.42	1897	0.95	1839	2.06
1950	0.43	1896	2.00	1838	0.69
1949	0.40	1895	0.91	1837	0.58
1948	0.82	1894	0.63	1836	1.62
1947	0.46	1893	1.27		
1946	0.59	1892	0.84		
1945	1.25	1891	0.52		
1944	1.73	1890	1.71		
1943	0.63	1889	1.72		
1942	0.57	1888	1.50		
1941	0.92	1887	1.84		
1940	2.11	1886	1.46		
1939	0.53	1885	3.01		
1938	0.45	1884	0.71		
1937	0.45	1883	1.26		
1936	1.27	1882	0.95		
1935	0.29	1881	0.25		
1934	1.89	1880	0.55		
1933	0.32	1879	0.79		
1932	0.62	1878	0.86		
1931	0.28	1877	1.99		
1930	0.32	1876	0.72		
1929	0.32	1875	1.44		
1928	0.49	1874	1.17		
1927	0.38	1873	0.97		
1926	0.34	1872	1.89		
1925	0.30	1871	2.33		
1924	2.73	1870	1.25		
1923	0.59	1869	2.18		
1922	0.58	1868	1.16		
1921	0.29	1867	0.81		
1920	0.31	1866	0.92		
1919	0.72	1865	2.39		
1918	1.33	1864	0.81		
1917	2.02	1863	2.55		
1916	0.87	1862	0.93		
1915	0.67	1861	0.74		
1914	0.58	1860	1.20		
1913	0.60	1859	0.46		
1912	0.73	1858	1.20		
		1857	0.93		
		1856	1.06		
		1855	1.41		
		1854	1.11		

APPENDIX E

TREE-RING SITE CHRONOLOGIES IN AND NEAR BELLA COOLA RIVER BASIN

(E.1) Introduction

Tree-ring data are best acquired under the auspices of a well-developed sampling scheme oriented towards maximizing the growth signal of interest. The following discussion is a brief summary of sampling techniques, cross-dating procedures and steps involved in the development of the tree-ring chronologies used in this thesis. Chronology data are listed in a series of standardized tables with several summary statistics.

(E.2) Tree-Ring Data Acquisition

Site and specimen selection criteria were developed to minimize the effect of disturbance factors such as insect damage, slope instability and physical effects related to the surrounding vegetation. General accessibility was also an important consideration for site selection. Indicators of physical stress such as scar damage or heart-rot were used to discriminate between sites and specimens. A sample constituted a minimum of two cores per tree (free of these disturbance effects) and a minimum of 10 trees per site. Typically, more cores were sampled because of the difficulty in locating the center of the tree.

Increment corers with 5 mm inside diameters were used to provide as large a sample as possible. Increment cores were then stored and transported in plastic straws. Storage was for brief periods only to avoid the growth of mold and drying of the cores. Swede and chain saws provided the best means for collection of tree disks which were taken at several levels from the felled tree.

Increment cores were individually mounted in grooved core-holders, wetted, shaved with a sharp blade and then dusted with zinc oxide to enhance visibility of cell structure. The smaller tree disks were sanded with several grits of sandpaper to smooth the cut surface whereas the larger disks (diameters greater than 20 cm) were first cut into six or eight radial sections and then sanded.

Several techniques for cross-dating or assigning calendar dates to each ring from every core at a site have been proposed (Stokes and Smiley, 1968; Fritts, 1976; Holmes, 1985). Improperly cross-dated samples when combined in a site chronology produce meaningless results. Cross-dating was done here by using a low magnification binocular microscope to count ring positions between the inner bark and pith. During counting, sequences of anomalously narrow and wide rings were used as markers to ensure accurate cross-dating between specimens at a site. Microscopic or potential missing rings were noted for each core (particularly Douglas fir) and confirmed after the completion of counting of all site samples. Missing or microscopic rings were then replaced with a mean value based on ± 10 rings surrounding the suspect datum. Cross-dating for sites exhibiting more complacent growth was facilitated by computer assisted techniques given in Holmes (1985).

After cross-dating ring-width data were acquired using a commercially-made ring measuring machine available in the Faculty of Forestry at UBC. Cross-hairs in a binocular microscope with 10X and 30X magnification are used to identify ring boundaries on an electrically mobile stage with the ability to compensate for core angles up to 25 degrees from vertical. Width measurements are accurate to within ± 0.01 mm. Digitally produced data were transferred to micro and mainframe computers for analysis.

Removal of growth trends, standardization procedures and chronology development were done using software developed for the desktop computer (Cho and Desloges, 1986). The procedures available in this software were modeled closely on those given in Fritts (1976) and Hughes *et al.* (1982).

Software developed by Cook and Holmes (1985) provided a means for assessing the degree of autocorrelation within the standardized time series of ring-width values. When first-order serial correlation was significant the autoregressive integrated moving average or ARIMA model (ARSTAN chronology) was used in place of the regular standardized chronology. In most cases these processed chronologies diverged minimally from the regular chronologies and represented reductions in variance due to serial correlation of only 3 to 10%. This is most likely due to the very "noisy" nature of the original ringwidth values for both Douglas and alpine firs. Tree-ring chronologies developed for this study as well as previously published data sets are summarized in the following tables.

SITE: Burnt Bridge Slope
LONGITUDE: 126 05' LATITUDE: 52 25'
ELEVATION: 155 m
SPECIES: Pseudotsuga menziesii
COLLECTED BY: J. Desloges/J. Wieniger
CHRONOLOGY TYPE: "B & C"
YEARS: 1705 to 1983

TREE RING INDICES

DATE	0	1	2	3	4	5	6	7	8	9
1700						1.52	1.20	1.36	1.99	1.26
1710	2.44	0.76	0.89	1.22	1.53	1.34	1.04	1.10	1.29	1.09
1720	0.95	0.96	0.78	0.99	0.60	1.20	0.93	0.60	0.89	0.53
1730	0.63	0.53	0.31	0.60	0.53	0.75	0.92	1.01	0.87	1.12
1740	0.71	0.69	0.65	0.73	0.79	0.77	0.66	1.17	0.78	0.70
1750	0.62	0.39	0.66	0.85	0.53	0.67	0.98	0.84	1.20	0.78
1760	0.79	0.93	0.96	0.79	1.21	1.24	1.36	1.17	1.18	1.41
1770	1.04	0.73	1.82	1.26	1.25	1.41	1.38	1.65	1.38	1.05
1780	1.09	1.08	1.55	1.26	1.28	0.88	1.02	1.19	0.99	0.98
1790	0.97	1.08	1.28	1.26	1.26	0.70	0.73	0.51	0.45	0.59
1800	0.70	0.59	0.71	0.75	0.88	0.85	1.50	1.67	1.33	1.36
1810	1.39	1.10	0.82	0.81	0.93	0.88	0.97	0.95	0.79	1.02
1820	0.89	0.75	0.81	0.90	0.83	0.90	0.57	0.74	1.28	0.99
1830	1.56	1.45	1.12	1.09	1.12	0.80	1.49	2.07	1.61	2.11
1840	1.80	1.94	1.47	1.04	1.38	1.16	0.96	1.49	1.34	1.94
1850	1.28	0.68	0.87	1.48	1.18	0.98	1.18	0.99	0.82	0.81
1860	0.75	0.60	0.73	0.88	0.43	0.63	0.68	0.63	0.61	0.45
1870	0.68	0.50	0.71	0.56	0.74	0.81	0.79	0.81	0.61	0.74
1880	0.68	0.57	0.66	0.82	1.05	0.74	1.17	1.89	1.15	1.01
1890	0.99	0.64	0.71	1.37	0.87	0.81	1.26	0.85	0.97	1.22
1900	1.62	1.02	0.85	0.85	0.95	0.74	0.98	1.11	0.92	2.06
1910	1.79	1.18	0.76	1.21	1.01	0.98	0.88	1.01	0.71	0.69
1920	0.88	0.64	0.74	0.79	1.07	0.84	0.81	0.81	0.74	0.98
1930	1.44	0.80	0.71	0.93	0.92	1.23	0.97	1.13	0.81	1.09
1940	0.69	0.97	0.92	1.89	1.30	0.94	1.09	1.04	0.75	0.89
1950	1.74	1.04	0.93	0.80	1.15	1.08	0.53	0.79	0.92	0.57
1960	0.94	0.66	0.51	0.72	1.26	1.27	1.06	1.95	1.53	2.39
1970	1.28	0.89	1.19	1.02	0.77	0.62	0.65	0.75	0.44	0.61
1980	0.90	0.90	0.67	0.89						

NUMBER OF RADII

DATE	0	1	2	3	4	5	6	7	8	9
1700						1	1	1	1	1
1710	2	3	3	3	3	4	4	4	4	4
1720	4	4	4	4	4	4	4	4	4	4
1730	4	4	4	5	5	5	5	5	6	6
1740	6	6	6	6	6	6	6	6	6	6
1750	7	7	8	9	10	10	10	10	10	10
1760	14	14	14	14	14	15	15	15	15	15
1770	15	15	15	15	15	15	15	15	15	15
1780	15	15	15	15	15	15	15	15	15	15
1790	15	15	15	15	15	15	15	15	15	15
1800	15	15	15	15	15	15	15	15	15	15
1810	15	15	15	15	15	15	15	15	15	15
1820	15	15	15	15	15	15	15	15	15	15
1830	15	15	15	15	15	15	15	15	15	15
1840	15	15	15	15	15	15	15	15	15	15
1850	15	15	15	15	15	15	15	15	15	15
1860	15	15	15	15	15	15	15	15	15	15
1870	15	15	15	15	15	15	15	15	15	15
1880	15	15	15	15	15	15	15	15	15	15
1890	15	15	15	15	15	15	15	15	15	15
1900	15	15	15	15	15	15	15	15	15	15
1910	15	15	15	15	15	15	15	15	15	15
1920	15	15	15	15	15	15	15	15	15	15
1930	15	15	15	15	15	15	15	15	15	15
1940	15	15	15	15	15	15	15	15	15	15
1950	15	15	15	15	15	15	15	15	15	15
1960	15	15	15	15	15	15	15	15	15	15
1970	15	15	15	15	15	15	15	15	15	15
1980	15	15	15	15						

STATISTICS:

Time span	= 279 years.	Skewness	= 1.146
Mean	= 0.993	Kurtosis	= 4.601
Median	= 0.931	Mean Sensitivity	= 0.243
Std. Dev.	= 0.351	Auto Corr. Coeff.	= 0.535

SITE: Kamloops
LONGITUDE: 120 33′ LATITUDE: 50 45′
ELEVATION: 820 m
SPECIES: *Pseudotsuga menziesii*
COLLECTED BY: H.C. Fritts
CHRONOLOGY TYPE: "B & C"
YEARS: 1420 to 1965

TREE RING INDICES

DATE	0	1	2	3	4	5	6	7	8	9
1420	1.16	1.74	1.30	0.54	1.12	0.67	0.48	1.33	0.50	1.38
1430	1.40	1.66	1.21	0.59	0.59	0.78	0.93	0.69	0.90	1.10
1440	0.89	0.94	1.03	0.85	1.23	0.87	1.32	1.20	0.89	1.07
1450	1.07	1.06	1.12	0.71	0.56	0.82	0.98	0.65	0.60	1.26
1460	1.19	1.05	1.54	0.82	1.16	1.24	0.85	0.56	0.83	0.75
1470	1.15	1.74	0.64	1.30	1.40	0.73	0.66	1.03	0.99	0.91
1480	0.98	1.35	1.52	1.19	1.55	1.07	1.42	1.51	1.89	0.55
1490	1.09	1.08	0.75	0.86	0.64	1.17	1.03	0.78	0.89	0.72
1500	1.46	0.50	0.88	0.01	1.37	1.21	0.79	1.07	0.81	1.05
1510	1.11	1.21	1.36	0.89	0.90	0.54	1.11	1.27	1.16	1.33
1520	0.61	0.81	1.09	0.76	1.05	0.99	0.89	0.77	1.08	0.52
1530	0.71	0.74	1.29	1.34	0.61	0.99	1.84	1.38	0.99	1.21
1540	0.68	0.94	0.80	0.67	0.76	0.49	0.48	0.97	0.94	0.96
1550	1.20	0.91	1.32	0.83	0.80	0.50	0.79	0.80	0.58	0.81
1560	0.81	0.89	0.69	0.65	1.17	0.86	1.41	0.79	0.72	1.22
1570	0.79	0.51	1.47	1.21	1.05	1.14	1.07	1.15	1.16	1.30
1580	1.33	0.89	0.88	0.64	0.91	0.77	1.01	1.33	0.91	1.09
1590	1.34	1.29	1.47	0.97	0.56	1.02	1.21	1.64	1.25	1.32
1600	0.75	1.55	1.34	1.14	1.16	0.84	1.07	1.25	0.93	1.03
1610	0.89	1.26	1.43	1.46	1.42	0.82	0.87	0.94	0.51	0.76
1620	0.67	1.08	1.14	0.43	1.17	1.29	0.61	1.16	1.17	0.91
1630	1.10	1.46	1.45	1.48	0.72	1.31	1.38	0.78	0.57	0.30
1640	0.51	0.72	0.74	0.31	0.78	0.70	0.74	0.41	0.82	0.45
1650	0.60	0.77	1.03	0.54	1.15	0.70	1.04	0.75	1.09	0.91
1660	0.01	0.69	1.25	1.09	0.55	0.72	1.05	1.49	1.63	1.23
1670	1.18	0.88	1.76	1.01	0.99	1.08	1.15	1.48	0.84	0.68
1680	0.76	1.07	0.99	0.31	0.91	0.48	0.80	1.13	0.87	1.14
1690	0.82	0.87	1.41	1.32	1.33	0.71	1.03	1.49	1.59	1.28
1700	1.94	1.32	2.10	1.03	0.96	0.39	0.55	1.15	0.76	0.88
1710	0.72	0.04	0.82	1.35	1.12	0.62	1.08	0.78	0.54	0.55
1720	0.70	0.57	0.98	0.38	1.46	1.24	0.59	1.45	1.03	0.67
1730	0.73	0.78	0.89	1.45	1.03	0.61	0.52	0.01	1.55	0.69
1740	1.42	0.65	1.12	1.51	0.37	1.09	0.79	0.77	0.65	0.85
1750	1.23	0.74	1.24	1.37	0.01	0.93	0.90	1.06	0.82	0.57
1760	0.70	1.09	1.73	1.04	1.06	0.74	1.37	1.59	0.80	1.30
1770	0.90	0.29	0.80	0.74	1.10	1.30	0.86	0.33	1.22	0.86
1780	1.02	1.11	1.26	0.61	0.55	1.22	1.23	0.81	0.89	1.66
1790	0.97	0.69	1.70	1.04	0.77	0.86	0.94	0.67	0.93	0.69
1800	0.31	1.01	1.22	1.02	0.78	0.30	1.01	1.14	0.73	0.76
1810	0.82	1.13	1.25	0.91	1.16	1.40	1.31	1.16	1.04	1.88
1820	1.56	1.36	1.54	1.12	0.68	1.43	1.55	0.87	1.31	1.09
1830	1.73	1.31	1.57	0.82	0.62	1.13	2.02	0.60	1.48	1.56
1840	1.11	0.92	0.58	0.35	0.70	1.02	1.13	0.70	0.58	0.98
1850	0.99	0.95	0.01	0.86	0.89	0.88	0.97	1.07	0.72	0.85
1860	1.14	1.49	1.19	1.21	0.46	0.92	1.20	1.21	1.10	0.60
1870	1.16	0.93	0.76	1.06	0.79	1.05	1.41	1.25	1.18	1.46
1880	1.70	1.71	1.90	1.30	1.21	1.44	0.64	0.93	1.44	1.02
1890	0.92	0.30	1.06	1.09	0.91	0.76	0.77	0.68	1.11	1.13
1900	1.33	1.52	1.45	0.83	1.52	0.46	1.37	1.25	1.33	0.61
1910	0.67	1.02	1.25	1.39	1.22	0.79	1.53	1.83	0.87	0.57
1920	0.53	0.76	0.38	0.58	0.51	0.58	0.54	0.70	0.85	0.75
1930	0.59	0.43	0.62	0.68	0.83	0.70	1.10	0.93	0.84	1.08
1940	0.97	0.38	1.42	1.24	0.98	0.98	1.26	0.89	1.11	1.12
1950	1.23	0.93	1.18	0.92	1.17	1.22	1.20	0.99	0.75	0.64
1960	1.21	0.80	1.03	0.64	0.60	0.79				

NUMBER OF RADII

DATE	0	1	2	3	4	5	6	7	8	9
1420	1	1	1	1	1	1	1	1	1	1
1430	1	1	1	1	1	1	1	1	1	1
1440	1	1	1	1	1	1	1	1	1	1
1450	1	1	1	1	1	1	1	1	1	1
1460	1	1	1	1	1	1	1	1	1	1
1470	2	2	2	2	2	2	2	2	2	2
1480	2	2	2	2	2	2	2	2	2	2
1490	2	2	2	2	2	2	2	2	2	2
1500	5	5	5	5	5	6	6	6	6	6
1510	6	6	6	6	6	6	6	6	6	6
1520	6	6	6	6	6	6	6	6	6	6
1530	6	6	6	6	6	6	6	6	6	6
1540	6	6	6	6	6	6	6	6	6	6
1550	7	7	7	7	7	7	7	7	7	7
1560	7	7	7	7	7	7	7	7	7	7
1570	7	7	7	7	7	7	7	7	7	7
1580	7	7	7	7	7	7	7	7	7	7
1590	7	7	7	7	7	7	7	7	7	7
1600	7	7	7	7	7	7	7	7	7	7
1610	7	7	7	7	7	7	7	7	7	7
1620	6	6	6	6	6	6	6	6	6	6
1630	6	6	6	6	6	6	6	6	6	6
1640	6	6	6	6	6	6	6	6	6	6
1650	6	6	6	6	6	6	6	6	6	6
1660	8	8	8	8	8	8	8	8	8	8
1670	8	8	8	8	8	8	8	8	8	8
1680	11	11	11	11	11	11	11	11	11	11
1690	14	14	14	14	14	14	14	14	14	14
1700	14	14	14	14	14	14	14	14	14	14
1710	14	14	14	14	14	14	14	14	14	14
1720	14	14	14	14	14	14	14	14	14	14
1730	15	15	15	15	15	15	15	15	15	15
1740	15	15	15	15	15	15	15	15	15	15
1750	16	16	16	16	16	16	16	16	16	16
1760	19	19	19	19	19	19	19	19	19	19
1770	19	19	19	19	19	19	19	19	19	19
1780	19	19	19	19	19	19	19	19	19	19
1790	19	19	19	19	19	19	19	19	19	19
1800	19	19	19	19	19	19	19	19	19	19
1810	20	20	20	20	20	20	20	20	20	20
1820	20	20	20	20	20	20	20	20	20	20
1830	20	20	20	20	20	20	20	20	20	20
1840	20	20	20	20	20	20	20	20	20	20
1850	20	20	20	20	20	20	20	20	20	20
1860	20	20	20	20	20	20	20	20	20	20
1870	20	20	20	20	20	20	20	20	20	20
1880	20	20	20	20	20	20	20	20	20	20
1890	20	20	20	20	20	20	20	20	20	20
1900	20	20	20	20	20	20	20	20	20	20
1910	20	20	20	20	20	20	20	20	20	20
1920	20	20	20	20	20	20	20	20	20	20
1930	20	20	20	20	20	20	20	20	20	20
1940	20	20	20	20	20	20	20	20	20	20
1950	20	20	20	20	20	20	20	20	20	20
1960	20	20	20	20	20	20				

STATISTICS:

Time span	= 546 years.	Skewness	= 0.135
Mean	= 0.986 mm.	Kurtosis	= 3.138
Median	= 0.980 mm.	Mean Sensitivity	= 0.370
Std. Dev.	= 0.346 mm.	Auto Corr. Coeff.	= 0.239

SITE: Pavilion
LONGITUDE: 121 41' LATITUDE: 50 50'
ELEVATION: 1160 m
SPECIES: Pseudotusga menziesii
COLLECTED BY: H.C. Fritts
CHRONOLOGY TYPE: "B & C"
YEARS: 1480 to 1965

TREE RING INDICES / NUMBER OF RADII

DATE	0	1	2	3	4	5	6	7	8	9	0	1	2	3	4	5	6	7	8	9
1480	0.71	1.26	1.27	0.92	1.33	0.72	1.60	1.29	1.31	0.92	1	1	1	1	1	1	1	1	1	1
1490	1.02	1.07	1.05	0.53	0.84	1.22	0.95	0.88	1.13	0.69	1	1	1	1	1	1	1	1	1	1
1500	1.20	0.91	0.74	0.94	1.12	1.20	1.10	0.71	1.12	1.58	1	1	1	1	1	1	1	1	1	1
1510	1.06	0.97	1.11	0.52	0.79	0.52	1.01	1.16	0.94	1.68	1	1	1	1	1	1	1	1	1	1
1520	0.61	1.07	0.97	0.61	0.93	1.12	0.48	0.86	0.72	0.71	1	1	1	1	1	1	1	1	1	1
1530	0.63	0.75	0.87	0.85	0.39	0.93	1.64	0.82	0.86	0.88	1	1	1	1	1	1	1	1	1	1
1540	0.41	0.57	1.10	0.69	0.77	0.64	0.61	0.84	0.84	1.04	1	1	1	1	1	1	1	1	1	1
1550	1.72	1.32	1.34	0.62	0.66	1.11	1.59	1.03	0.66	0.76	1	1	1	1	1	1	1	1	1	1
1560	0.48	0.40	0.37	0.53	1.01	0.72	1.15	1.05	0.72	1.35	1	1	1	1	1	1	1	1	1	1
1570	0.77	0.88	1.68	2.16	2.33	1.44	1.08	1.66	1.93	1.72	1	1	1	1	1	1	1	1	1	1
1580	1.55	1.51	0.70	0.59	1.12	1.40	0.92	1.60	1.12	1.24	1	1	1	1	1	1	1	1	1	1
1590	1.26	1.31	1.31	0.67	0.72	0.79	0.91	1.11	1.39	1.55	3	3	3	3	3	3	3	3	3	3
1600	0.78	1.84	1.54	1.11	1.19	1.28	1.08	1.40	1.61	1.36	4	4	4	4	4	4	4	4	4	4
1610	1.17	1.27	1.07	1.09	1.28	0.98	0.77	1.22	0.63	1.32	4	4	4	4	4	4	4	4	4	4
1620	0.40	0.39	0.53	0.45	0.92	1.02	0.51	0.71	0.64	0.58	4	4	4	4	4	4	4	4	4	4
1630	1.11	1.25	1.23	1.27	0.94	0.91	1.51	1.02	0.64	0.35	5	5	5	5	5	5	5	5	5	5
1640	1.23	1.08	1.12	0.32	0.90	0.53	0.73	0.54	1.05	0.41	6	6	6	6	6	6	6	6	6	6
1650	0.45	0.59	0.89	0.73	1.53	0.70	1.05	0.47	1.04	0.71	6	6	6	6	6	6	6	6	6	6
1660	0.88	0.52	1.20	0.59	0.38	0.94	1.24	1.12	1.08	0.49	6	6	6	6	6	6	6	6	6	6
1670	1.09	0.86	1.65	0.64	0.25	1.32	1.19	1.10	1.06	0.93	6	6	6	6	6	6	6	6	6	6
1680	0.83	1.19	1.27	0.34	0.93	0.77	0.68	0.78	0.62	0.94	6	6	6	6	6	6	6	6	6	6
1690	0.89	1.08	0.44	0.53	1.33	1.41	1.14	2.17	2.23	1.64	6	6	6	6	6	6	6	6	6	6
1700	2.37	1.16	1.70	0.93	1.36	0.82	0.33	0.82	0.75	1.06	7	7	7	7	7	7	7	7	7	7
1710	0.86	1.02	0.72	1.36	0.95	0.95	0.94	0.37	0.81	0.65	7	7	7	7	7	7	7	7	7	7
1720	0.35	0.64	1.04	0.10	1.08	0.66	0.56	0.94	1.01	1.33	7	7	7	7	7	7	7	7	7	7
1730	0.75	0.85	1.27	2.12	0.84	1.29	0.60	1.21	1.32	0.30	7	7	7	7	7	7	7	7	7	7
1740	1.35	0.26	1.25	1.53	0.47	1.82	0.73	1.45	0.63	1.27	7	7	7	7	7	7	7	7	7	7
1750	1.13	0.46	1.56	1.31	1.40	0.87	0.80	1.14	0.71	0.21	11	11	11	11	11	11	11	11	11	11
1760	0.64	1.05	1.56	1.33	0.70	1.07	1.31	1.18	0.87	1.34	11	11	11	11	11	11	11	11	11	11
1770	0.72	1.12	0.43	1.24	0.96	1.31	0.88	0.52	0.90	1.17	13	13	13	13	13	13	13	13	13	13
1780	0.78	1.53	1.04	1.09	0.79	0.65	0.98	1.03	0.74	1.70	13	13	13	13	13	13	13	13	13	13
1790	0.91	1.14	0.82	0.77	0.33	0.98	0.94	0.66	0.84	0.80	13	13	13	13	13	13	13	13	13	13
1800	0.52	1.30	1.45	1.39	0.80	2.08	2.09	1.89	1.24	0.56	13	13	13	13	13	13	13	13	13	13
1810	1.01	1.12	1.16	1.47	0.93	0.91	0.91	1.13	1.66	2.02	13	13	13	13	13	13	13	13	13	13
1820	1.53	1.41	1.69	1.00	0.75	0.80	1.66	1.17	1.29	1.84	16	16	16	16	16	16	16	16	16	16
1830	1.63	0.83	1.13	0.60	0.92	0.60	1.39	0.62	1.53	1.32	18	18	18	18	18	18	18	18	18	19
1840	1.19	1.51	0.58	0.44	0.68	1.62	1.58	0.83	0.83	1.07	19	19	19	19	19	19	19	19	19	19
1850	0.97	0.80	1.14	0.96	0.92	1.12	1.30	0.83	0.54	0.56	20	20	20	20	20	20	20	20	20	20
1860	1.33	1.91	1.11	0.75	0.69	0.81	1.06	0.80	0.94	0.25	20	20	20	20	20	20	20	20	20	20
1870	0.98	0.58	0.96	1.08	0.86	0.71	0.88	1.14	1.20	1209	20	20	20	20	20	20	20	20	20	20
1880	1.28	1.88	1.52	1.13	0.61	1.52	0.82	1.33	1.31	1.11	20	20	20	20	20	20	20	20	20	20
1890	1.44	0.92	0.83	1.34	1.17	0.96	0.63	0.69	1.05	0.99	20	20	20	20	20	20	20	20	20	20
1900	0.90	1.38	0.51	0.71	1.08	0.60	0.88	0.78	0.94	0.66	20	20	20	20	20	20	20	20	20	20
1910	0.65	0.45	0.84	0.93	1.25	1.05	1.61	0.90	0.78	0.52	20	20	20	20	20	20	20	20	20	20
1920	0.47	0.84	0.66	0.36	0.75	0.72	0.38	0.81	0.95	0.61	20	20	20	20	20	20	20	20	20	20
1930	0.81	0.32	0.79	0.57	0.86	0.96	0.98	1.28	0.76	0.75	20	20	20	20	20	20	20	20	20	20
1940	0.99	0.46	1.70	0.92	1.34	1.48	0.81	0.66	1.13	0.88	20	20	20	20	20	20	20	20	20	20
1950	1.15	1.83	0.99	1.16	1.35	1.31	1.08	1.64	1.09	0.52	20	20	20	20	20	20	20	20	20	20
1960	1.15	0.85	1.14	0.93	1.46	1.51					20	20	20	20	20	20				

STATISTICS:

Time span	= 486 years.	Skewness	= 0.574
Mean	= 1.005 mm.	Kurtosis	= 3.431
Median	= 0.960 mm.	Mean Sensitivity	= 0.387
Std. Dev.	= 0.388 mm.	Auto Corr. Coeff.	= 0.292

SITE: Willams Lake/Alkali Lake/Trenquille
LONGITUDE: various LATITUDE: various
ELEVATION: 600-800 m
SPECIES: Pseudotsuga menziesii/Pinus ponderosa
COLLECTED BY: E. Schulman
CHRONOLOGY TYPE: "B & C"
YEARS: 1420 to 1944

TREE RING INDICES

DATE	0	1	2	3	4	5	6	7	8	9
1420	1.16	1.74	1.30	0.54	1.12	0.67	0.48	1.33	0.50	1.38
1430	1.40	1.66	1.21	0.59	0.59	0.78	0.93	0.69	0.90	1.10
1440	0.89	0.94	1.03	0.85	1.23	0.87	1.32	1.20	0.89	1.07
1450	1.07	1.06	1.12	0.71	0.56	0.82	0.98	0.65	0.60	1.26
1460	1.19	1.05	1.54	0.82	1.16	1.24	0.85	0.56	0.83	0.75
1470	1.15	1.74	0.64	1.30	1.40	0.73	0.66	1.03	0.99	0.91
1480	0.98	1.35	1.52	1.19	1.55	1.07	1.42	1.51	1.89	0.55
1490	1.09	1.08	0.75	0.86	0.64	1.17	1.03	0.78	0.89	0.72
1500	1.46	0.50	0.88	1.00	1.37	1.27	0.76	0.98	0.78	0.91
1510	1.04	1.16	1.42	0.86	0.88	0.46	0.97	1.18	1.04	1.25
1520	0.51	0.72	0.98	0.78	1.03	1.08	0.88	0.78	1.07	0.49
1530	0.62	0.61	1.20	1.31	0.59	0.97	1.75	1.42	1.00	1.20
1540	0.66	0.96	0.75	0.66	0.70	0.44	0.50	1.03	1.00	1.01
1550	1.21	0.95	1.35	0.88	0.86	0.53	0.86	0.79	0.55	0.77
1560	0.81	0.95	0.78	0.73	1.14	0.81	1.34	0.82	0.76	1.30
1570	0.90	0.59	1.54	1.31	1.14	1.31	1.07	1.18	1.25	1.42
1580	1.46	0.90	0.85	0.62	0.87	0.67	0.93	1.34	0.81	1.06
1590	1.22	1.25	1.37	0.98	0.55	1.00	1.23	1.71	1.20	1.34
1600	0.75	1.36	1.25	1.17	1.25	0.83	0.99	1.11	0.87	0.96
1610	0.81	1.04	1.42	1.53	1.41	0.82	0.88	0.95	0.36	0.64
1620	0.32	0.43	0.51	0.10	0.52	0.67	0.24	0.68	0.71	0.66
1630	0.82	1.67	1.56	1.40	0.75	1.14	1.08	1.07	0.83	0.54
1640	0.99	1.23	0.99	0.71	0.77	1.05	0.83	0.48	0.99	0.65
1650	0.62	0.53	1.05	0.89	1.40	1.02	1.20	0.69	0.95	1.07
1660	1.10	0.89	1.34	0.87	0.59	0.99	1.71	1.52	1.25	0.89
1670	1.26	0.95	1.83	1.17	0.74	1.19	1.14	0.96	0.75	0.75
1680	0.89	1.20	0.86	0.26	0.59	0.50	0.72	1.01	0.92	0.93
1690	0.93	0.78	0.83	0.78	1.11	1.07	1.23	1.46	1.58	0.80
1700	1.64	1.16	1.70	1.11	1.04	0.77	0.98	1.08	0.84	0.93
1710	0.96	1.38	1.20	1.74	1.17	1.08	1.28	0.62	0.89	0.77
1720	0.93	0.90	1.39	0.58	0.97	1.11	1.06	1.47	1.10	1.10
1730	0.75	0.87	1.19	1.43	0.72	0.77	0.71	0.89	1.33	1.02
1740	1.36	0.72	1.23	1.31	0.41	1.21	0.91	0.84	0.41	0.63
1750	0.97	0.71	1.36	1.47	1.50	1.50	1.02	1.15	0.98	0.78
1760	0.69	1.23	1.68	1.14	1.60	1.22	1.64	1.69	0.87	1.24
1770	0.77	0.39	0.77	0.67	1.08	1.05	1.05	0.27	1.01	0.90
1780	0.78	1.10	1.16	0.74	0.76	1.08	1.10	1.24	0.99	1.84
1790	0.96	0.68	1.49	0.69	0.69	0.89	0.94	0.52	0.80	0.75
1800	0.36	0.83	1.19	1.12	0.91	0.75	0.75	1.15	0.88	0.81
1810	0.73	1.19	1.18	0.67	1.04	1.39	1.14	1.14	0.82	1.44
1820	1.38	1.36	1.51	1.08	0.72	0.90	1.50	0.93	1.44	1.35
1830	1.05	1.04	1.51	0.54	0.63	0.95	1.54	0.41	1.26	1.10
1840	0.87	0.62	0.46	0.27	0.69	0.93	1.01	0.74	0.77	0.85
1850	1.00	0.90	0.92	0.75	1.10	0.93	0.95	0.99	0.53	0.73
1860	1.16	1.27	1.27	1.34	0.43	0.96	1.04	1.50	1.38	0.18
1870	1.01	0.76	0.88	1.00	0.91	1.20	0.99	1.22	0.87	1.40
1880	1.64	1.68	1.50	1.13	1.18	1.30	0.51	1.05	1.19	1.04
1890	0.94	0.54	0.79	1.08	0.90	0.82	0.67	0.73	1.19	1.06
1900	1.18	1.43	1.15	0.98	1.24	0.47	1.08	1.21	1.21	0.59
1910	0.88	0.86	1.10	1.10	1.42	0.91	0.91	1.39	0.96	0.69
1920	0.91	0.72	0.61	0.67	0.59	0.82	0.61	0.67	1.14	0.88
1930	0.77	0.35	0.75	0.98	0.96	1.07	1.04	1.10	0.93	1.50
1940	1.02	0.77	1.42	1.54	1.30					

NUMBER OF RADII
(approximate)

DATE	0	1	2	3	4	5	6	7	8	9
1420	1	1	1	1	1	1	1	1	1	1
1430	1	1	1	1	1	1	1	1	1	1
1440	1	1	1	1	1	1	1	1	1	1
1450	1	1	1	1	1	1	1	1	1	1
1460	1	1	1	2	2	2	2	2	2	2
1470	2	2	2	2	2	2	2	2	2	2
1480	2	2	2	2	2	2	2	2	2	2
1490	2	2	2	2	2	2	2	2	2	2
1500	10	10	10	10	10	10	10	10	10	10
1510	10	10	10	10	10	10	10	10	10	10
1520	10	10	10	10	10	10	10	10	10	10
1530	10	10	10	10	10	10	10	10	10	10
1540	10	10	10	10	10	10	10	10	10	10
1550	10	10	10	10	10	10	10	10	10	10
1560	10	10	10	10	10	10	10	10	10	10
1570	10	10	10	10	10	10	10	10	10	10
1580	10	10	10	10	10	10	10	10	10	10
1590	10	10	10	10	10	10	10	10	10	10
1600	10	10	10	10	10	10	10	10	10	10
1610	10	10	10	10	10	10	10	10	10	10
1620	10	10	10	10	10	10	10	10	10	10
1630	10	23	23	23	23	23	23	23	23	23
1640	23	23	23	23	23	23	23	23	23	23
1650	23	23	23	23	23	23	23	23	23	23
1660	23	23	23	23	23	23	23	23	23	23
1670	23	23	23	23	23	23	23	23	23	23
1680	23	23	23	23	23	23	23	23	23	23
1690	23	23	23	23	23	23	23	23	23	23
1700	23	23	23	23	23	23	23	23	23	23
1710	23	23	23	23	23	23	23	23	23	23
1720	23	23	23	23	23	23	23	23	23	23
1730	23	23	23	23	23	23	23	23	23	23
1740	23	23	23	23	23	23	23	23	23	23
1750	23	23	23	23	23	23	23	23	23	23
1760	23	23	23	23	23	23	23	23	23	23
1770	23	23	23	23	23	23	23	23	23	23
1780	23	23	23	23	23	23	23	23	23	23
1790	23	23	23	23	23	23	23	23	23	23
1800	44	44	44	44	44	44	44	44	44	44
1810	44	44	44	44	44	44	44	44	44	44
1820	44	44	44	44	44	44	44	44	44	44
1830	44	44	44	44	44	44	44	44	44	44
1840	44	44	44	44	44	44	44	44	44	44
1850	44	44	44	44	44	44	44	44	44	44
1860	44	44	44	44	44	44	44	44	44	44
1870	44	44	44	44	44	44	44	44	44	44
1880	44	44	44	44	44	44	44	44	44	44
1890	44	44	44	44	44	44	44	44	44	44
1900	44	44	44	44	44	44	44	44	44	44
1910	44	44	44	44	44	44	44	44	44	44
1920	44	44	44	44	44	44	44	44	44	44
1930	44	44	44	44	44	44	44	44	44	44
1940	44	44	44	44	44					

STATISTICS:

Time span	= 525 years.	Skewness	= 0.188
Mean	= 0.992 mm.	Kurtosis	= 2.838
Median	= 0.980 mm.	Mean Sensitivity	= 0.315
Std. Dev.	= 0.315 mm.	Auto Corr. Coeff.	= 0.289

SITE: Bull Canyon
LONGITUDE: 123 15' LATITUDE: 52 06'
ELEVATION: 750 m
SPECIES: Pseudotsuga menziesii
COLLECTED BY: J. Desloges/L. Davies
CHRONOLOGY TYPE: "B & C"
YEARS: 1650 to 1983

TREE RING INDICES

DATE	0	1	2	3	4	5	6	7	8	9
1650	1.27	0.90	1.28	1.08	0.81	1.41	0.98	1.49	1.81	1.28
1660	1.49	1.02	0.66	1.03	0.89	0.35	0.93	0.79	0.61	0.66
1670	0.71	0.79	0.94	1.58	1.12	0.47	1.54	1.90	1.51	1.15
1680	0.92	0.88	1.79	1.48	0.33	0.37	0.61	0.83	1.06	0.64
1690	0.76	0.55	1.04	1.96	0.75	1.12	1.07	1.48	1.39	0.62
1700	1.27	0.70	1.13	0.66	0.85	0.57	0.83	0.74	0.40	0.82
1710	0.86	1.41	1.26	1.85	1.02	1.01	1.15	0.76	1.19	0.90
1720	0.71	1.10	1.28	0.84	0.51	0.85	0.82	1.18	1.12	1.41
1730	0.80	0.99	1.32	0.94	0.73	1.04	1.39	1.35	1.47	0.69
1740	1.18	0.98	1.21	1.27	0.27	0.91	0.47	0.58	0.64	0.45
1750	0.88	1.26	1.43	1.31	1.53	1.29	1.21	0.89	1.42	0.67
1760	0.56	1.15	1.23	1.34	1.32	1.31	1.34	1.83	0.86	1.00
1770	0.71	0.37	0.50	0.78	0.56	0.77	0.67	0.40	0.68	0.99
1780	1.10	0.99	1.05	1.28	1.54	1.19	1.51	1.44	1.21	1.78
1790	1.02	0.46	1.13	0.48	0.64	0.92	0.59	0.60	0.68	0.92
1800	0.59	0.93	1.01	1.01	1.14	1.27	1.32	1.69	1.43	1.18
1810	0.44	1.40	0.93	0.51	0.72	1.26	0.98	1.24	0.75	1.02
1820	1.23	1.27	1.35	1.40	1.01	0.74	1.12	1.10	1.31	2.09
1830	1.95	1.29	1.28	0.41	0.86	0.46	1.08	1.08	1.66	1.12
1840	1.04	0.88	0.54	0.64	1.01	1.10	1.18	0.67	0.99	0.75
1850	1.11	0.91	0.73	0.63	0.97	0.71	0.84	0.55	0.52	0.51
1860	0.72	1.18	1.18	1.32	0.53	0.67	0.82	0.61	0.88	0.28
1870	0.60	0.26	0.57	0.51	0.75	1.10	1.31	1.12	0.92	1.30
1880	1.34	1.75	1.61	1.20	1.06	0.97	1.44	1.45	1.76	1.53
1890	1.64	1.18	0.97	1.31	1.45	1.21	1.08	1.15	1.44	1.17
1900	1.15	1.36	0.96	0.78	1.13	0.90	0.87	0.82	1.05	0.76
1910	0.65	0.67	0.92	1.09	1.05	0.89	0.53	0.46	0.83	0.68
1920	0.72	0.61	0.57	0.73	0.58	0.66	0.60	0.57	0.73	0.61
1930	0.59	0.30	0.61	1.08	0.69	0.91	0.41	0.82	0.63	0.92
1940	0.85	0.43	1.14	1.03	1.19	1.21	0.51	1.47	1.38	1.20
1950	1.44	1.21	1.68	0.84	0.96	1.78	1.03	1.66	1.55	0.44
1960	1.55	0.91	1.11	1.26	1.33	1.10	1.18	1.07	1.14	1.09
1970	0.64	0.93	0.90	0.73	0.75	0.74	1.14	1.19	0.78	0.76
1980	0.69	0.71	0.54	0.77						

NUMBER OF RADII

DATE	0	1	2	3	4	5	6	7	8	9
1650	1	1	1	1	1	1	1	1	1	1
1660	1	1	1	1	1	1	1	1	1	1
1670	1	1	1	1	2	2	2	2	2	2
1680	2	2	2	2	2	2	2	2	2	2
1690	2	2	2	2	2	2	2	2	2	2
1700	2	2	2	2	2	2	2	2	2	2
1710	2	2	2	2	2	3	3	3	3	3
1720	3	4	4	4	4	4	4	4	4	4
1730	4	4	4	4	4	8	8	8	8	8
1740	8	8	8	10	10	10	10	10	10	10
1750	14	14	14	14	14	14	14	14	14	14
1760	14	14	14	14	14	14	14	14	14	14
1770	14	14	14	14	14	14	14	16	16	16
1780	16	16	16	16	16	16	16	16	16	16
1790	16	16	16	16	16	16	16	16	16	16
1800	16	16	16	16	16	16	16	16	16	16
1810	16	16	16	16	16	16	16	16	16	16
1820	16	16	16	16	16	16	16	16	16	16
1830	16	16	16	16	16	16	16	16	16	16
1840	16	16	16	16	16	16	16	16	16	16
1850	16	16	16	16	16	16	16	16	16	16
1860	16	16	16	16	16	16	16	16	16	16
1870	16	16	16	16	16	16	16	16	16	16
1880	16	16	16	16	16	16	16	16	16	16
1890	16	16	16	16	16	16	16	16	16	16
1900	16	16	16	16	16	16	16	16	16	16
1910	16	16	16	16	16	16	16	16	16	16
1920	16	16	16	16	16	16	16	16	16	16
1930	16	16	16	16	16	16	16	16	16	16
1940	16	16	16	16	16	16	16	16	16	16
1950	16	16	16	16	16	16	16	16	16	16
1960	16	16	16	16	16	16	16	16	16	16
1970	16	16	16	16	16	16	16	16	16	16
1980	16	16	16	16						

STATISTICS:

Time span	= 334 years.	Skewness	= 0.314
Mean	= 0.990	Kurtosis	= 2.688
Median	= 0.984	Mean Sensitivity	= 0.290
Std. Dev.	= 0.355	Auto Corr. Coeff.	= 0.437

SITE: Burnt Bridge Terrace
LONGITUDE: 126 05' LATITUDE: 52 25'
ELEVATION: 150 m
SPECIES: Pseudotusga menziesii
COLLECTED BY: J. Desloges/T. Millard
CHRONOLOGY TYPE: "B & C"
YEARS: 1704 to 1983

	TREE RING INDICES										NUMBER OF RADII									
DATE	0	1	2	3	4	5	6	7	8	9	0	1	2	3	4	5	6	7	8	9
1700					2.38	1.63	1.48	1.45	1.87	1.88					1	1	1	1	1	1
1710	1.47	1.13	1.10	1.58	2.03	1.04	0.96	1.81	1.79	1.69	1	1	1	1	1	1	2	2	2	2
1720	1.56	1.14	1.64	1.73	0.98	1.26	1.05	1.02	1.30	0.90	2	2	2	2	2	2	2	2	3	3
1730	1.06	1.23	1.06	1.78	0.98	1.13	1.28	1.05	0.81	0.96	3	3	3	3	3	4	4	4	4	5
1740	1.22	1.16	1.33	1.16	1.18	0.94	0.90	1.09	1.07	0.79	5	5	5	5	5	5	5	5	5	5
1750	0.88	1.00	1.01	1.27	1.10	1.48	1.25	0.92	1.16	1.08	5	5	5	6	6	6	6	6	6	6
1760	1.21	1.10	1.22	1.53	1.47	1.08	0.66	0.43	0.42	0.24	6	6	6	6	6	6	6	6	6	6
1770	0.31	0.30	0.32	0.41	0.44	0.37	0.42	0.82	0.88	0.88	6	6	6	6	6	6	6	6	6	6
1780	1.01	1.36	0.87	0.67	0.92	0.86	0.68	0.80	0.88	1.00	6	6	8	8	8	8	8	8	8	8
1790	0.60	0.63	0.51	0.50	0.78	0.69	0.66	0.60	0.54	0.52	8	8	8	8	8	8	8	8	12	12
1800	0.73	0.78	0.84	0.85	1.05	1.20	1.21	1.27	1.66	1.65	12	12	12	12	12	12	12	12	12	12
1810	1.73	1.59	1.36	1.18	1.41	1.52	1.53	1.39	1.19	1.05	12	12	12	12	12	12	12	12	12	12
1820	1.26	1.14	1.21	1.41	1.28	1.45	0.99	0.70	0.94	1.20	12	12	12	12	12	12	12	12	12	12
1830	1.12	1.21	1.29	1.20	1.40	1.43	1.28	1.25	1.59	1.39	12	12	12	12	12	12	12	12	12	12
1840	0.93	0.95	1.11	0.69	0.78	1.05	1.06	1.38	1.10	1.14	12	12	12	12	12	12	12	12	12	12
1850	1.06	0.73	0.89	0.98	1.07	1.26	1.18	1.22	1.23	1.02	12	12	12	12	12	12	12	12	12	12
1860	0.99	1.07	0.88	0.74	0.63	0.48	0.46	0.46	0.51	0.51	12	12	12	12	12	12	12	12	12	12
1870	0.42	0.56	0.36	0.31	0.59	0.59	0.37	0.46	0.40	0.41	12	12	12	12	12	12	12	12	12	12
1880	0.45	0.65	0.55	0.57	0.98	1.12	0.66	0.83	0.88	1.18	12	12	12	12	12	12	12	12	12	12
1890	0.79	0.97	0.95	0.88	0.79	0.71	1.22	0.82	0.96	0.95	12	12	12	12	12	12	12	12	12	12
1900	0.92	0.83	0.72	0.74	1.07	1.09	1.09	1.27	1.84	2.12	12	12	12	12	12	12	12	12	12	12
1910	1.83	0.86	1.16	0.96	1.18	1.27	0.98	1.24	1.30	1.48	12	12	12	12	12	12	12	12	12	12
1920	1.63	1.02	1.23	1.42	1.45	0.67	0.89	1.21	1.10	0.80	12	12	12	12	12	12	12	12	12	12
1930	1.22	0.93	0.83	1.27	1.39	0.89	0.40	0.47	0.65	0.77	12	12	12	12	12	12	12	12	12	12
1940	0.85	0.92	0.92	1.20	1.12	0.79	1.13	1.17	0.98	1.04	12	12	12	12	12	12	12	12	12	12
1950	1.35	0.83	1.45	1.39	1.14	1.11	0.88	0.88	1.14	0.97	12	12	12	12	12	12	12	12	12	12
1960	1.19	0.99	0.94	1.21	1.24	0.66	0.91	0.56	0.79	0.92	12	12	12	12	12	12	12	12	12	12
1970	1.07	0.82	0.85	0.78	0.85	0.92	1.16	1.45	0.87	1.35	12	12	12	12	12	12	12	12	12	12
1980	1.88	1.63	0.74	0.85							12	12	12	12						

STATISTICS:

Time span	=	280 years.	Skewness	=	0.211
Mean	=	1.000	Kurtosis	=	3.099
Median	=	1.028	Mean Sensitivity	=	0.336
Std. Dev.	=	0.346	Auto Corr. Coeff.	=	0.704

SITE: Stuie (Tweedsmuir Lodge)
LONGITUDE: 126 03' LATITUDE: 52 25'
ELEVATION: 140 m
SPECIES: Pseudotsuga menziesii
COLLECTED BY: J. Desloges/P. Desloges
CHRONOLOGY TYPE: "B & C"
YEARS: 1579 to 1983

TREE RING INDICES

DATE	0	1	2	3	4	5	6	7	8	9
1570										0.93
1580	1.34	1.39	1.55	1.23	0.98	1.05	0.91	1.17	1.10	1.00
1590	0.85	0.73	0.74	0.81	0.78	0.76	0.69	0.72	0.78	0.80
1600	0.95	0.95	0.85	0.80	0.84	0.88	0.87	0.85	0.98	0.94
1610	0.79	0.83	0.78	0.73	0.83	0.81	0.76	0.88	0.94	0.76
1620	0.72	0.70	0.74	0.64	0.67	0.65	0.60	0.61	0.67	0.71
1630	0.67	0.75	0.81	0.72	0.80	0.66	0.75	0.83	0.83	0.78
1640	0.79	0.70	0.76	0.83	0.98	0.87	0.85	0.90	0.90	0.91
1650	1.11	1.19	1.35	1.30	1.18	0.93	0.96	0.84	0.78	0.83
1660	0.72	0.74	0.60	0.46	0.49	0.50	0.56	0.66	0.82	0.80
1670	0.93	0.78	0.87	0.76	0.72	1.10	1.10	0.96	0.99	0.83
1680	0.81	1.11	1.14	1.36	1.09	1.34	1.72	1.46	1.65	1.35
1690	1.16	1.03	0.95	1.15	0.97	0.73	0.88	0.74	1.49	1.20
1700	1.26	0.91	1.13	1.12	1.21	0.93	0.79	0.88	1.05	1.11
1710	1.34	0.94	0.88	1.02	1.22	1.02	1.01	1.16	1.42	1.62
1720	1.44	1.36	1.39	1.42	1.43	1.53	1.19	1.18	1.18	1.05
1730	1.23	1.29	1.12	1.27	1.05	1.21	1.23	1.16	1.16	1.35
1740	1.14	1.24	1.29	1.13	1.11	0.99	0.90	1.09	1.07	0.95
1750	0.94	0.76	0.79	0.97	0.93	1.01	1.07	1.12	1.24	1.21
1760	1.45	1.43	1.45	1.50	1.63	1.31	1.21	0.81	0.99	0.99
1770	0.90	0.64	1.07	0.87	0.89	0.91	0.86	1.04	0.89	0.90
1780	0.98	1.13	1.28	1.07	1.06	0.96	1.04	1.05	1.03	0.68
1790	0.65	0.75	0.86	1.04	1.18	0.65	0.66	0.40	0.61	0.54
1800	0.85	0.98	1.01	0.94	1.06	1.00	1.03	1.28	1.17	1.15
1810	1.06	1.02	1.00	1.07	1.18	1.17	1.32	1.18	1.30	1.35
1820	1.27	1.11	1.33	1.37	1.37	1.22	0.93	1.06	1.00	1.26
1830	1.13	1.23	1.21	1.03	1.14	1.08	1.03	1.12	0.96	1.01
1840	0.78	0.87	0.74	0.68	0.95	0.86	1.08	0.92	0.97	0.95
1850	0.87	0.69	0.85	1.02	1.00	1.05	1.02	1.00	0.95	1.05
1860	0.94	0.98	1.06	1.07	1.17	1.01	1.17	1.01	0.97	0.85
1870	0.83	0.82	0.76	0.71	0.75	0.97	0.54	0.53	0.43	0.63
1880	0.60	0.66	0.67	0.66	0.72	0.71	1.02	0.88	0.82	0.93
1890	0.91	0.93	0.89	0.88	0.83	0.81	1.09	0.88	1.08	0.97
1900	1.03	0.94	1.06	1.02	1.03	1.02	1.01	1.08	1.06	1.47
1910	1.22	0.96	0.94	1.06	1.19	1.22	1.23	1.12	1.14	1.10
1920	1.08	0.85	1.01	0.94	1.25	1.00	0.95	0.99	0.97	0.79
1930	1.17	0.92	0.78	0.87	1.02	0.81	0.78	0.82	0.88	1.04
1940	1.14	1.01	1.10	1.28	1.28	1.08	1.42	1.31	1.21	1.26
1950	1.48	0.90	1.02	1.01	1.14	1.22	0.84	0.84	1.01	0.92
1960	1.22	1.23	1.05	1.18	1.18	0.95	1.06	0.96	0.90	0.97
1970	1.13	0.95	0.96	0.92	0.78	0.73	0.80	0.87	0.62	0.76
1980	0.94	0.92	0.71	0.72						

NUMBER OF RADII

DATE	0	1	2	3	4	5	6	7	8	9
1570										1
1580	1	1	1	1	1	1	1	1	1	1
1590	1	2	2	2	2	2	2	2	2	2
1600	2	2	2	2	2	2	2	2	2	2
1610	2	2	2	2	2	2	2	2	2	2
1620	3	3	3	3	3	3	3	3	3	3
1630	4	4	4	4	4	4	4	4	4	4
1640	4	4	4	6	6	6	6	6	6	6
1650	6	6	6	6	6	6	6	6	6	6
1660	6	6	6	6	6	6	6	6	6	6
1670	7	7	7	7	7	7	7	7	7	7
1680	7	8	8	8	8	8	8	8	8	8
1690	8	8	8	8	8	8	8	8	8	8
1700	10	10	10	10	10	10	10	10	10	10
1710	10	10	10	10	10	10	10	10	10	10
1720	11	11	11	11	11	11	11	11	12	12
1730	12	13	13	13	13	13	13	13	13	13
1740	13	13	13	13	14	14	14	14	14	14
1750	15	15	15	15	15	15	15	15	15	15
1760	15	15	15	15	15	15	15	15	15	15
1770	15	15	15	15	15	15	15	15	15	15
1780	15	15	15	15	15	15	15	15	15	15
1790	18	18	18	18	18	18	18	18	18	18
1800	18	18	18	18	18	18	18	18	18	18
1810	18	18	18	18	18	18	18	18	18	18
1820	18	18	18	18	18	18	18	18	18	18
1830	18	18	18	18	18	18	18	18	18	18
1840	18	18	18	18	18	18	18	18	18	18
1850	18	18	18	18	18	18	18	18	18	18
1860	18	18	18	18	18	18	18	18	18	18
1870	18	18	18	18	18	18	18	18	18	18
1880	18	18	18	18	18	18	18	18	18	18
1890	18	18	18	18	18	18	18	18	18	18
1900	18	18	18	18	18	18	18	18	18	18
1910	18	18	18	18	18	18	18	18	18	18
1920	18	18	18	18	18	18	18	18	18	18
1930	18	18	18	18	18	18	18	18	18	18
1940	18	18	18	18	18	18	18	18	18	18
1950	18	18	18	18	18	18	18	18	18	18
1960	18	18	18	18	18	18	18	18	18	18
1970	18	18	18	18	18	18	18	18	18	18
1980	18	18	18	18						

STATISTICS:

Time span	= 405 years.	Skewness	= 0.317
Mean	= 0.988 mm.	Kurtosis	= 3.069
Median	= 0.970 mm.	Mean Sensitivity	= 0.318
Std. Dev.	= 0.226 mm.	Auto Corr. Coeff.	= 0.661

SITE: East Nusatsum
LONGITUDE: 126 17' LATITUDE: 52 13'
SPECIES: Abies Lasiocarpa
COLLECTED BY: J. Desloges/J. Wieniger
CHRONOLOGY TYPE: "B & C"
YEARS: 1711 to 1983

TREE RING INDICES

DATE	0	1	2	3	4	5	6	7	8	9
1710		1.49	1.42	1.83	1.14	1.21	1.31	0.82	1.34	0.44
1720	0.79	0.99	0.78	0.88	1.05	1.04	0.94	1.27	0.90	0.96
1730	0.60	0.86	1.22	1.17	1.37	1.23	1.30	1.03	1.08	1.10
1740	1.02	0.70	0.76	0.85	0.78	0.89	0.82	0.98	0.85	0.99
1750	0.91	0.86	0.73	0.64	0.82	0.97	0.88	0.67	0.81	0.85
1760	0.89	0.78	0.94	1.08	0.93	1.10	1.29	1.09	1.01	0.93
1770	0.91	1.07	1.05	1.03	1.13	1.09	1.06	1.03	0.85	1.07
1780	0.79	0.96	0.96	0.87	1.08	1.03	0.92	1.00	1.04	1.13
1790	0.91	1.02	0.97	1.10	1.17	0.96	1.10	1.11	0.95	1.01
1800	0.98	0.98	1.20	0.88	1.16	0.99	0.96	0.88	1.06	0.84
1810	0.71	0.79	0.87	0.97	1.12	1.11	0.97	1.13	0.99	0.86
1820	0.93	1.16	1.21	1.02	1.29	1.07	1.27	0.98	1.10	1.18
1830	1.01	1.14	1.07	1.29	1.34	1.28	1.28	1.00	1.22	0.96
1840	1.04	1.08	1.21	1.11	0.92	0.97	1.11	1.22	1.09	0.94
1850	1.06	0.99	0.97	0.97	0.96	1.04	1.02	0.93	0.93	0.90
1860	0.92	0.81	0.86	1.02	1.06	1.32	1.13	1.02	1.06	1.01
1870	1.03	0.93	0.86	1.13	1.25	0.97	1.29	0.83	0.95	0.84
1880	0.84	0.83	1.07	0.95	0.74	1.04	1.19	1.09	1.12	0.99
1890	0.67	0.87	0.85	0.79	0.81	0.86	0.78	0.71	0.73	0.57
1900	0.75	0.64	0.55	0.68	0.68	0.70	0.67	0.72	0.79	0.76
1910	0.84	0.83	0.72	0.76	0.74	0.77	0.80	0.91	0.82	0.78
1920	0.78	0.83	0.94	0.85	0.77	0.77	0.88	0.87	0.91	0.85
1930	0.91	0.82	0.77	0.90	0.99	1.38	1.39	1.39	1.24	1.01
1940	0.88	0.97	1.11	1.31	1.48	1.23	1.33	1.10	1.61	1.30
1950	1.50	1.34	1.24	1.35	1.01	1.23	1.36	1.14	1.13	1.02
1960	0.83	1.04	0.90	1.20	0.97	1.01	1.12	0.80	0.93	0.79
1970	0.79	0.86	0.58	0.77	0.68	0.76	0.78	0.78	1.19	0.79
1980	0.87	0.92	0.88	0.99						

NUMBER OF RADII

0	1	2	3	4	5	6	7	8	9
	1	1	1	1	1	1	1	1	1
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1	1	1	3	4	4	4	4	4	4
4	4	4	4	4	4	4	4	4	4
4	4	4	4	4	4	4	4	4	4
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10	10	10	10	10	10	10	10	10	10
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10	10	10	10	10	10	10	10	10	10
10	10	10	10	10	10	10	10	10	10
10	10	10	10	10	10	10	10	10	10
10	10	10	10						

STATISTICS:

Time span	=	273 years.	Skewness	=	0.476
Mean	=	0.980	Kurtosis	=	3.217
Median	=	0.970	Mean Sensitivity	=	0.230
Std. Dev.	=	0.193	Auto Corr. Coeff.	=	0.591

SITE: Noohalk Lakes
LONGITUDE: 126 38' LATITUDE: 52 20'
ELEVATION: 1440 m
SPECIES: Abies Lasiocarpa
COLLECTED BY: J. Desloges/J. Wieniger
CHRONOLOGY TYPE: "B & C"
YEARS: 1871 to 1983

TREE RING INDICES

DATE	0	1	2	3	4	5	6	7	8	9
1870		0.51	0.49	0.36	0.60	0.57	0.44	0.47	0.33	0.40
1880	0.61	0.67	0.66	0.45	0.53	0.57	0.80	0.68	0.92	0.80
1890	1.42	1.43	1.39	1.17	1.07	0.83	1.09	0.92	1.15	1.05
1900	1.31	0.98	0.94	1.49	1.78	1.62	1.71	1.53	1.27	1.29
1910	1.02	1.03	0.86	0.61	0.72	0.86	1.61	1.36	1.66	1.74
1920	1.23	0.91	0.57	0.98	1.17	1.01	0.89	1.40	0.74	1.05
1930	0.79	1.09	1.20	0.90	0.77	0.98	0.93	0.68	0.64	0.89
1940	0.63	0.96	1.40	1.19	0.89	1.30	1.12	0.93	0.90	1.07
1950	0.91	0.98	0.68	0.48	0.66	0.55	0.64	0.56	0.63	1.03
1960	0.76	0.83	1.32	0.65	1.24	0.91	1.17	1.13	1.22	1.20
1970	0.83	1.06	0.70	0.56	0.83	0.50	0.75	0.48	1.33	1.00
1980	1.22	1.47	1.51	1.27						

NUMBER OF RADII

0	1	2	3	4	5	6	7	8	9
	1	1	1	1	1	1	1	1	1
1	1	3	3	3	5	6	6	6	8
8	8	8	8	8	8	8	8	8	8
8	8	8	8	8	8	8	8	8	8
12	12	12	12	12	12	12	12	12	12
12	14	14	14	14	14	14	18	18	18
18	18	18	18	18	18	18	18	18	18
18	18	18	18	18	18	18	18	18	18
18	18	18	18	18	18	18	18	18	18
18	18	18	18	18	18	18	18	18	18
18	18	18	18	18	18	18	18	18	18
18	18	18	18						

STATISTICS:

Time span	=	113 years.	Skewness	=	0.351
Mean	=	1.011	Kurtosis	=	2.478
Median	=	0.919	Mean Sensitivity	=	0.234
Std. Dev.	=	0.320	Auto Corr. Coeff.	=	0.665

SITE: Ape Lake
LONGITUDE: 126 10' LATITUDE: 52 05'
ELEVATION: 1400 m
SPECIES: Abies Lasiocarpa
COLLECTED BY: J. Desloges/T. Millard
CHRONOLOGY TYPE: "B & C"
YEARS: 1740 to 1983

TREE RING INDICES

DATE	0	1	2	3	4	5	6	7	8	9
1740	0.99	0.88	1.00	0.81	0.69	0.73	0.67	0.92	1.08	0.93
1750	1.07	0.91	0.91	0.91	0.76	1.04	1.06	0.87	0.87	0.87
1760	0.71	0.80	0.76	0.91	0.96	0.98	1.06	1.07	0.94	1.11
1770	1.04	1.16	1.14	1.40	1.37	1.18	1.36	1.49	1.19	1.09
1780	1.17	1.14	1.01	1.25	1.16	1.19	1.08	0.99	1.15	1.07
1790	1.06	1.19	1.23	1.19	1.16	1.18	1.16	1.10	1.21	1.08
1800	1.18	1.17	1.05	1.14	1.21	1.16	0.95	1.02	1.03	1.04
1810	0.88	0.96	0.94	0.96	0.99	0.83	1.13	1.07	0.99	0.90
1820	1.02	0.97	0.99	0.96	0.93	0.95	0.95	0.87	0.88	0.86
1830	0.90	0.85	0.89	0.96	0.93	0.92	0.91	1.01	0.88	1.07
1840	0.96	0.96	1.06	0.98	0.99	0.92	0.99	0.98	1.04	0.98
1850	0.96	1.03	1.00	0.90	0.97	0.96	0.86	0.90	0.91	0.93
1860	0.87	0.88	0.93	1.06	0.83	1.00	0.86	0.82	1.06	0.97
1870	0.85	0.90	0.65	0.93	1.01	1.02	0.80	0.85	0.79	0.87
1880	0.82	0.81	0.78	0.81	0.85	0.78	0.75	0.80	0.83	0.76
1890	0.88	1.09	1.01	0.83	0.94	0.92	0.87	0.84	0.86	0.86
1900	0.98	1.00	1.04	1.18	1.13	0.98	1.10	1.14	1.06	1.03
1910	1.07	1.01	0.85	0.95	1.10	1.11	0.85	1.12	1.26	1.24
1920	1.19	1.02	1.11	0.94	1.05	0.67	1.09	1.05	1.03	1.15
1930	1.20	1.02	1.05	1.13	1.08	1.13	1.13	0.87	1.24	0.92
1940	1.07	1.41	1.16	0.75	1.09	1.05	1.07	0.97	1.18	1.16
1950	1.19	1.07	0.95	1.01	0.87	0.94	0.70	0.68	1.14	0.83
1960	0.88	1.23	0.93	1.06	1.10	1.21	1.08	1.15	1.21	1.20
1970	1.24	1.14	0.94	0.94	0.84	0.98	0.59	1.10	1.11	0.93
1980	1.20	1.48	0.97	0.94						

NUMBER OF RADII

0	1	2	3	4	5	6	7	8	9
1	1	1	1	1	1	1	1	1	1
1	1	1	1	1	1	1	1	1	1
2	2	2	2	2	2	2	2	2	2
2	3	3	4	4	4	4	4	4	4
4	4	5	5	6	6	6	6	6	6
8	8	8	8	8	8	8	8	8	8
8	8	8	8	8	8	8	8	8	8
8	8	9	9	9	9	9	9	9	9
10	10	10	10	10	10	10	10	10	10
10	11	11	11	12	12	12	12	12	12
12	12	14	14	14	14	14	14	14	14
14	14	14	14	14	14	14	14	16	16
18	18	18	18	18	18	18	18	18	18
18	18	18	18	18	18	18	18	18	18
18	18	18	18	18	19	19	19	19	19
20	20	20	20	20	20	20	20	20	20
20	20	20	20	20	20	20	20	20	20
20	20	20	20	20	20	20	20	20	20
20	20	20	20	20	20	20	20	20	20
20	20	20	20	20	20	20	20	20	20
20	20	20	20	20	20	20	20	20	20
20	20	20	20	20	20	20	20	20	20
20	20	20	20	20	20	20	20	20	20
20	20	20	20	20	20	20	20	20	20
20	20	20	20						

STATISTICS:

Time span	= 244 years.	Skewness	= 0.221
Mean	= 1.000	Kurtosis	= 3.348
Median	= 0.992	Mean Sensitivity	= 0.309
Std. Dev.	= 0.151	Auto Corr. Coeff.	= 0.527

SITE: Rainbow Range
LONGITUDE: 52 33' LATITUDE: 125 49'
ELEVATION: 1665 m
SPECIES: Abies Lasiocarpa
COLLECTED BY: J. Desloges/T. Millard
CHRONOLOGY TYPE: "B & C"
YEARS: 1725 to 1984

TREE RING INDICES

DATE	0	1	2	3	4	5	6	7	8	9
1720						1.20	0.98	1.28	1.46	1.13
1730	0.80	1.02	1.18	0.87	1.07	0.70	0.66	1.10	0.92	0.99
1740	0.72	0.61	0.52	0.73	0.92	1.33	1.09	0.98	0.99	1.11
1750	0.76	0.59	1.23	1.21	0.81	1.01	0.79	0.99	1.09	0.96
1760	1.20	1.16	1.19	1.18	0.96	0.91	0.97	0.85	1.16	1.03
1770	0.93	1.10	1.32	1.10	0.74	0.76	1.03	1.09	1.05	1.06
1780	1.00	0.76	1.00	1.07	1.01	1.01	1.09	0.99	1.29	1.21
1790	0.97	0.91	0.99	1.07	1.07	1.42	1.26	1.28	1.13	1.21
1800	0.99	1.17	1.22	1.32	1.11	1.13	1.06	0.97	1.00	0.90
1810	1.01	0.99	1.03	0.84	0.69	0.94	0.79	0.72	0.65	0.76
1820	0.73	0.89	0.83	0.80	0.88	0.76	1.02	0.93	1.11	1.19
1830	1.22	0.93	1.09	1.43	1.22	0.85	1.05	1.03	1.31	1.05
1840	0.98	1.01	1.04	1.02	1.04	0.87	0.86	1.19	0.97	1.16
1850	1.02	1.14	0.84	1.16	1.20	0.91	0.94	0.97	0.95	0.68
1860	0.72	0.65	0.94	0.57	0.99	0.80	0.80	0.84	0.95	0.82
1870	0.71	0.64	0.76	0.74	1.03	0.57	0.60	0.57	0.84	0.88
1880	1.01	1.15	0.92	0.87	0.80	0.81	1.03	1.04	1.02	1.00
1890	1.32	1.12	1.11	1.27	1.08	1.05	1.09	1.03	0.90	0.97
1900	1.17	1.09	1.11	1.28	1.20	1.10	1.13	1.22	1.17	1.13
1910	1.25	1.06	1.25	1.31	1.32	1.07	1.20	1.62	1.29	1.38
1920	1.39	1.38	1.23	0.99	0.97	1.22	1.14	0.99	1.07	1.25
1930	1.14	1.01	1.29	0.98	1.14	0.99	1.08	1.08	1.01	1.07
1940	1.06	1.06	1.08	1.21	1.09	1.09	0.98	1.11	1.10	1.24
1950	1.08	1.06	1.05	1.02	1.15	0.82	0.91	1.18	1.07	0.96
1960	1.10	0.73	0.84	0.79	1.14	1.00	1.02	0.86	0.83	0.85
1970	0.78	0.66	0.73	0.63	0.62	0.66	0.73	0.70	0.63	0.63
1980	0.70	0.73	0.68	0.84	1.20					

NUMBER OF RADII

0	1	2	3	4	5	6	7	8	9
					1	1	1	1	1
3	3	3	3	3	3	3	3	3	3
3	3	3	3	3	3	3	3	3	3
3	3	3	3	3	3	3	3	3	3
3	3	3	4	4	4	4	4	4	4
4	4	5	5	5	5	5	5	7	7
7	7	7	7	7	7	7	7	7	7
7	7	7	7	7	7	7	7	7	8
8	8	8	9	9	9	9	9	9	9
10	10	10	10	10	10	10	10	10	10
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10	10	10	10	10	10	10	10	11	11
11	11	11	11	11	11	11	11	11	11
12	12	12	12	12	12	12	13	13	13
13	15	15	15	15	15	15	15	15	15
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15	15	15	15	15	15	15	15	15	15
15	15	15	15	15	15	15	15	15	15
15	15	15	15	15	15	15	15	15	15
15	15	15	15	15					

STATISTICS:

Time span	=	260 years.	Skewness	=	-0.122
Mean	=	0.999	Kurtosis	=	2.763
Median	=	1.016	Mean Sensitivity	=	0.122
Std. Dev.	=	0.198	Auto Corr. Coeff.	=	0.588

APPENDIX F

APPLICATION OF PALEOHYDROLOGICAL TECHNIQUES FOR RECONSTRUCTING FORMER CONFIGURATIONS OF GLACIERS IN THE THE BELLA COOLA RIVER BASIN

(F.1) Accumulation Area Ratios

The altitude of the equilibrium line on a glacier (ELA) can be estimated by assuming that some constant proportion of the glacier area lies above the equilibrium line (accumulation area). Since the ELA fluctuates from year to year the proportion of area in the accumulation zone will also vary. To facilitate comparisons and to ascribe proper climatic interpretations, measurements of the ELA must correspond to steady-state conditions of the glacial system (*i.e.* when accumulation equals ablation or zero net mass balance). Therefore, the accumulation area ratio can be estimated by using modern measurements of glacier mass balance on selected glaciers and determining the altitude of the ELA (hence AAR) during years of zero net mass balance.

Steady-state ELAs on selected Canadian glaciers reveals that AARs vary between 0.23 to 0.74 although most are between 0.50 and 0.65 (table F.1). The large range is representative of the various glacier hypsometries and long-term net mass balance trends at each site. For glaciers of "normal" or regular hypsometry in table F.1 an AAR value of between 0.55 and 0.60 is considered to be representative. This compares reasonably well with those values used by Leonard (1984) and Sutherland (1984). Glaciers which are more irregular will have lower or higher values depending on the actual configuration of the accumulation and ablation areas.

Two groups of glaciers were selected in the eastern and southern portions of the Bella Coola catchment in order to estimate contemporary ELAs. The first group of six glaciers, representing small ice masses (< 5.0 km^2) with fairly regular hypsometries and northerly or easterly aspects (see figure F.1), yielded an estimate of 1836 ± 206 m (using an AAR of 0.60). The fairly large standard deviation demonstrates the variability in morphology of cirque basins, differences in glacier hypsometry, measurement

Table F.1. Accumulation Area Ratios for Selected Canadian Glaciers.[1]

Glacier	Area (km^2)	Mean ELA (m)		Steady State AAR	Average Net Mass Balance	Period of Record
Alexander	5.83	1725	160	0.66	negative	1979-84
Andrei	92.15	1514	108	0.64	negative	1978-84
Bench	10.51	1860	34	0.58	negative	1981-84
Bridge	88.10	2198	126	0.61	negative	1977-84
Decade	8.65	1113	358	0.74	negative	1966-70
Helm	2.44	2103	89	0.23	negative	1977-85
Peyto	13.40	2680	73	0.50	negative	1966-84
Place	3.96	2189	150	0.44	negative	1965-84
Ram	1.80	1838	109	0.46	negative	1966-75
Sentinel	1.77	1818	110	0.72	positive	1966-84
Sykora	25.35	2197	127	0.55	negative	1974-84
Tiedemann	62.69	1891	195	0.60	negative	1981-84
Woolsey	3.87	2266	146	0.50	negative	1966-75
Yuri	3.58	1829	107	0.56	negative	1978-84
Zavisha	6.49	2281	82	0.64	negative	1977-81

1. Basic data given in Mokievsky-Zubok et al. (1985)

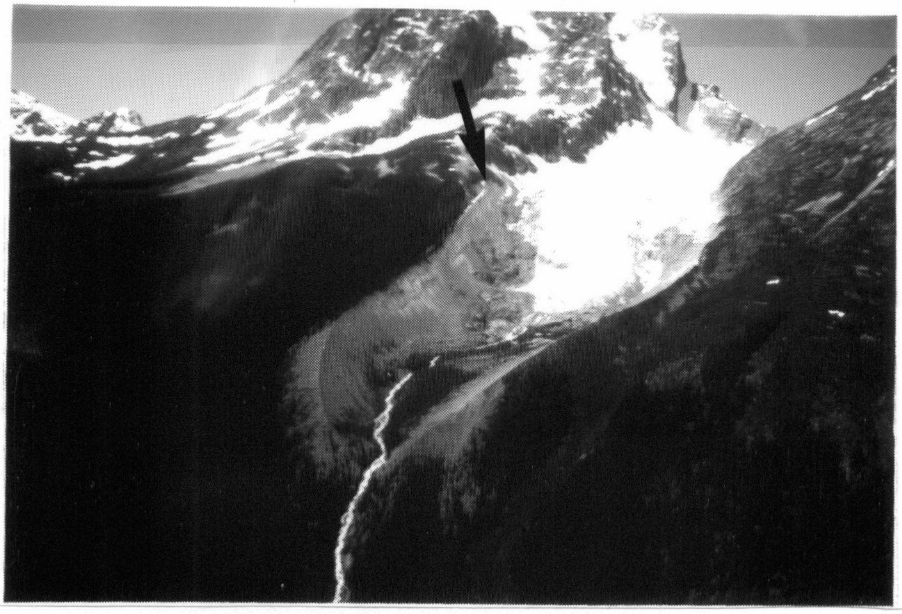

Figure F.1 Typical hanging glacial valley and well-defined Little Ice Age moraines on the west flank of Talchako River. Lateral moraines are used to contour the former configuration of the glacier in order to determine the height of the paleo ELA. Highest point on the lateral moraines here (arrow in figure) and at other sites is also used as an estimate of climatic snowline.

error associated with planimetry and the contouring error (± 100 feet) in the 1:50,000 topographic maps.[4] ELAs on a second group of larger glaciers between 5.0 and 25.0 km^2 (n = 5) (see figure 2.8b in main text) were determined by adopting a customized value for AAR derived after comparing hypsometries of each site with those glaciers associated with a known steady-state AAR (table F.1). The results show an equally high variance and a higher overall mean for contemporary ELA ($\bar{X}$ = 1924 ± 231). The difficulties in identifying ice-divides for some of the larger glaciers (*e.g.* Fyles, Ape) makes estimates from the latter group less precise.

(F.2) Reconstructing Former Climatic Snowlines

The reconstruction of former climatic snowlines can be accomplished using a variety of methods. In addition to the glaciation threshold, three other methods were used in this study. They include cirque floor altitudes, maximum elevation of lateral moraine deposits and the ELA of glaciers during their former advanced position.

Cirque floor altitudes are thought to represent the base level of glacial erosion during maximum ice advance. The two major difficulties of this method are identifying a break in slope in the cirque head wall which defines the top of the cirque floor and selection of sites which represent surfaces formed during the same glacial maximum. Several cirques in the Bella Coola basin were ice-free during the recent glacial maximum and are relict features of the Fraser ice-advance. All cirques with characteristics indicative of recent ice occupation and subsequent withdrawal (*e.g.* fresh trim line, little or no soil development, lichen free debris) in the

4. All glacier boundaries marked on the 1:50,000 topographic sheets were verified by examination of larger-scale air photographs.

eastern basin were sampled (n = 25). A mean estimate of the cirque floor elevations is 1610 ± 110 m.

The maximum altitude of lateral moraine deposits coincides with the boundary between the erosion and ablation zones of the glacier system and has been shown to be an approximation of the climatic snowline (Meierding, 1982). Lateral moraine deposits formed during the glacial maximum were mapped using vertical and oblique photographs in addition to field-based measurements using a Paulin altimeter (see figure F.1). The sample data were stratified into those deposits found in north facing basins of the western Bella Coola region and into north facing and south facing basins of the eastern catchment. A total of 36 sites were used six of which were verified with field-based measurements. The maximum altitude of lateral moraine deposits varies from 1320 ± 150 m in the west to 1980 ± 100 m in south-facing basins of the eastern catchment.

Estimates of former steady-state ELAs were made for the same set of glaciers used to estimate contemporary heights. Paleo-glacier surfaces were contoured using evidence of former ice extent at each site and the assumption that the surface profile of the ice and accumulation area ratios were similar for both sampling periods (see figure F.1). Results from the analyses and a discussion of their significance is given in the main text.

(F.3) Glacier Chronologies for the Bella Coola Basin

Evidence for the timing of glacier fluctuations in the Bella Coola basin was examined in order to document responses to changing climate over the last several centuries. In total, 16 glaciers (18 sites) were investigated ranging in size from small cirque glaciers (<2.0 km^2) to major valley glaciers (>35 km^2). Site selection criteria included the presence of vegetated, well-defined lateral or terminal moraines, the potential for stratigraphic exposures in the moraine and outwash environments and suitable access by air or on foot (figure F.2). Dating methods were based primarily on dendrochronology with lichenometry, soil development and ^{14}C dating used where possible to verify and supplement tree-ring derived ages.

Methods

A minimum limiting age of glacial deposits can be inferred from the age of the oldest tree growing on the surface of the deposit. The major assumptions of this dating method are that the sampled trees are part of the original colonizing vegetation and that the interval between exposure of the deposit and first germination (ecesis period) can be reasonably estimated (Lawrence, 1950; Sigafoos and Hendricks, 1969). Additional application problems include locating reliably the oldest, first-growth vegetation on the deposit and then sampling accurately the first growth-ring of the specimen(s).

In most instances the largest tree is the oldest but where the substrate is poorly developed and micro-climate most severe, smaller specimens may in fact be older. Sampling several of the largest and most stressed trees ensures a high probability of determining the maximum age of the vegetation. Increment cores and disks extracted from as close to the base of the tree as possible provide the best estimates of true age.

Stem analysis was conducted at several sites by intensively sampling trees at one or two meter intervals to determine the dating error associated with samples taken higher in the stem. Two relationships were uncovered. At middle and low elevation sites dominated by western hemlock, Sitka spruce and Douglas fir, annual changes in tree height where as much as 39 cm but an average of 25 cm or less was more common; thus, a correction of approximately 0.04 years/cm was applied to ages determined at

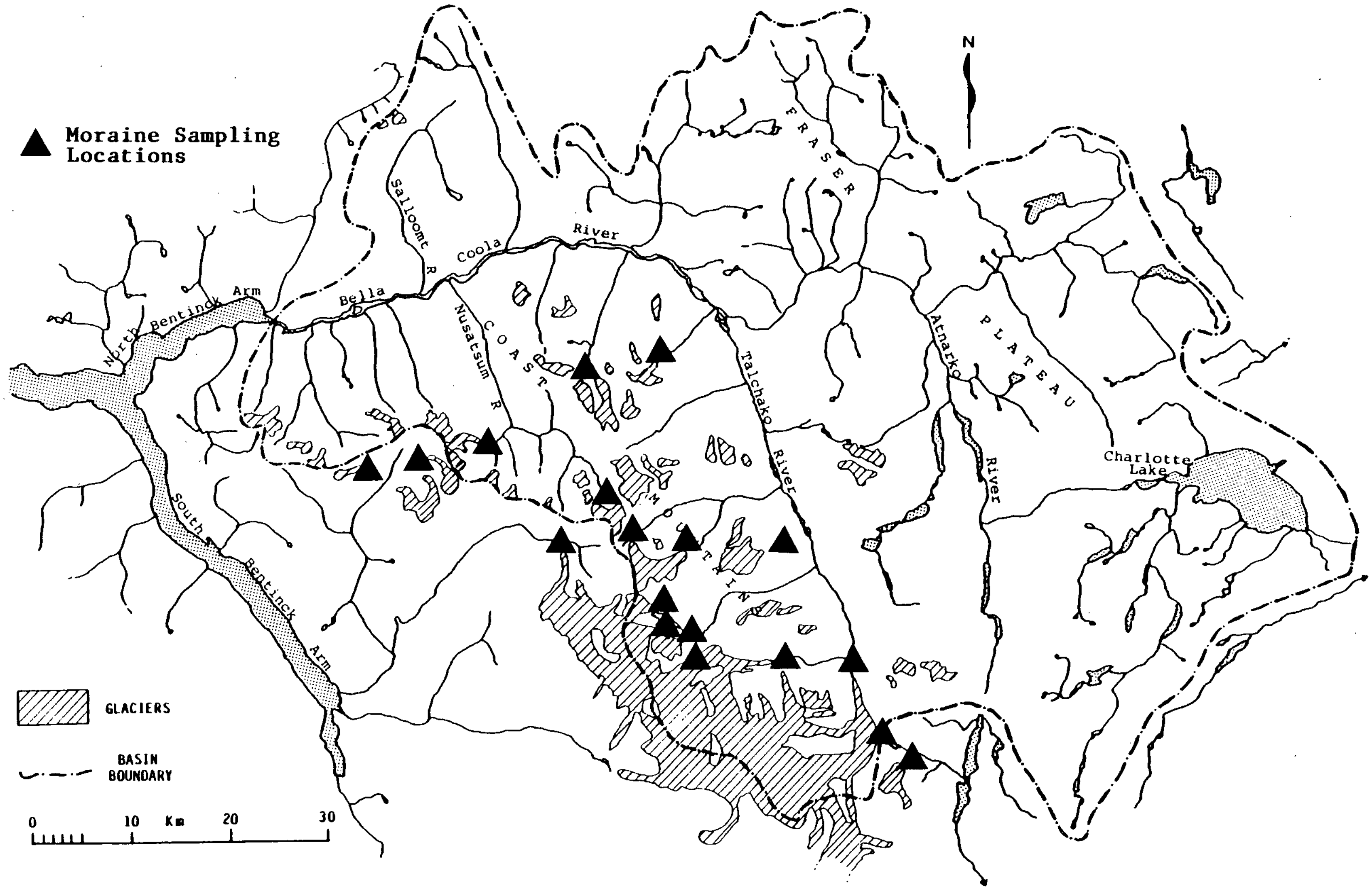

Figure F.2 Sampling locations of glaciers used to develop a recent chronology of glacier moraines in the Bella Coola River basin. See table F.1 for a summary of results.

a given sampling height above the base. At higher elevations where Alpine fir and Lodgepole pine dominate, and annual changes in height are much smaller, the estimated correction factor was 8 times or 0.32 years/cm. Thus an age of 100 years determined 25 cm above the base would be corrected to 108 years.

The ecesis interval, a major source of error, is influenced by a number of factors but local site-scale conditions such as texture of the substrate, distance from seed source and micro-climate are probably the most important (Brink, 1964). Ecesis estimates, varying from a few years to several decades, have been made for southwestern B.C. (Mathews, 1951; Ricker and Tupper, 1979), the Cascade and Olympic Mountains of Washington State (Heusser, 1957; Sigafoos and Hendricks, 1961, 1972; Heikkinen, 1984), the Coast Mountains of Alaska (Lawrence, 1950) and several sites in the Canadian Rockies (Luckman, 1986). A recent investigation by Ryder (unpublished) suggests that local variability exceeds differences in any of the regionally-based estimates. For this reason a few sites were selected to determine the interval of colonization by sampling vegetation on recently exposed glacial materials. The interval since deglaciation for each sampling location was estimated using multiple air photograph coverage.

Surveys for vegetation on recently exposed glacial deposits were conducted at four sites (Fyles, Deer, Atavist and Borealis Glaciers). Field measurements show that a limited number of widely scattered seedlings have colonized deposits exposed for intervals as short as 10 years (*e.g.* Fyles Glacier northwest margin). However, the ecesis interval is more typically 15-20 years. In almost every case seedlings have become established in pockets of finer-grained sandy gravel. In several locations low tangled growth indicates that wind shear across the moraine surface is an important limiting factor for establishment and development of arboreal vegetation. Seedlings were notably absent from cobble/boulder moraines except on benches or elevated, well-drained, depressions where fine-grained material had accumulated. Thus for areas in which ecesis estimates could not be made directly a value of 20 years was assumed if pockets of fine-grained material on the moraine were present, otherwise 30 years was used.

Combined errors involving tree-ring dating of glacial deposits are probably on the order of a decade for many of the features sampled in this study. It is only where the substrate is unusually coarse and the climate correspondingly severe that the error may be significantly larger (Luckman, 1986).

Lichenometry has been commonly used for establishing ages of surfaces for which other relative or absolute dating techniques are not applicable. Although no "absolute" calibration curve of regional significance has yet been established for the southern Coast Mountains, lichen measurements were made to verify the relative ages of deposits established using the dendrochronological techniques. Like dendro methods the assumptions of immediate colonization and accountability of growth factors other than time must also be made (Innes, 1985a, 1985b). Because several sites were visited only briefly and others were at sufficiently low elevation to be covered continuously by heath and moss rather than lichen, measurements were taken from eight sites only.

Soil development on major moraines at elevations greater than 1200-1300 m was rare and in most cases non-existent. For the remaining few sites the depth of soil development, degree of horizonation and surface organic matter contents were used to distinguish deposits of different relative ages (*cf*. Birkeland, 1974)

Buried organic material (soil, roots, logs) was also sampled at several locations to determine the general chronology for glacial outwash and glacial moraine sedimentary sequences. Stuiver (1982), Porter (1981) and others have shown that variations in ^{14}C content of the atmosphere over the last millennia significantly influence the relationship between radiocarbon years and calendar years. Hence, where ambiguities exist other lines of evidence are required. A summary of estimated moraine ages from all 17 sites is given in table F.2. Details for six of the most informative sites are discussed below.

(F.4) Description of Glacier Chronologies at Selected Sites

East Nusatsum Valley

Ice from three tributary sub-basins covering an area of approximately 20 km^2 coalesced to form a set of well-defined terminal and recessional moraines in the east Nusatsum watershed. The largest and most continuous moraine marks the terminus of these glaciers which were fully advanced as late as 1867 (figure F.3a). Recessional moraines dated at 1880, 1889, 1898, 1912 and 1934 suggest brief pauses in retreat during the late 19th and early 20th century. Soil development on each moraine is limited to a thin LF horizon over a shallow Ah/C profile. Although depths of the Ah horizon vary only from 5 to 10 cm on the outermost moraine to less than 2 cm on the younger 20th century moraines, LF horizons are substantially thicker on the 19th century moraines (15 cm versus 2-3 cm). The most rapid ice recession appears to have occurred between 1912 and 1934, averaging approximately 23 m/year, while the slowest rates were between 1867 and 1880 (8 m/year) and 1898 and 1912 (7 m/year).

Table F.2. Tree-ring derived ages of Late Neoglacial terminal moraines in study area.

Site	Elevation (m)	Aspect (true)	Soil Development	Lichen Cover[1]	Sampled Age	Tree-ring derived ages[2]		
						Corr Factor	Ecesis Int.	Final Age
East Nusatsum	1173	335°	Terminal - Ah/C Ah ≈ 5cm	mostly mosses and LFH	91	8	20	119 (1865)
Noomst	1036	355°	Terminal - Ah/C Ah ≈ 10cm	mostly mosses and LFH	83	4	20	107 (1877)
Deer Lake	1435	45°	none	RG (35)	74	12	25	111 (1874)
Ape Glacier	1430	15°	none	RG (15)	58	3	30	91 (1894)
Fyles Glacier	1402	45°	none	RG (20)	80	5	10	85 (1900)
Tsini Tsini	1183	50°	none	RG (20)	86	5	20	111 (1875)
E. Skunk Gl.	1285	50°	Lateral Ah/C Ah ≈ 10-15cm	mostly mosses and LFH	99	7	20	126 (1860)
Talchako Terminal	490	350°	Terminal - Ah/C Ah ≈ 10-15cm	mostly mosses and LFH	78	5	20	103 (1883)
Talchako Lateral	1450	0°	none	RG (40)	125	10	30	165 (1821)
Horseshoe Gl.	1265	$20\text{-}90^{\circ}$	Terminal - Ah/C Ah ≈ 2-3cm	-	33	5	20	58 (1928)
Jacobsen Terminal	1060	70°	none	A (105)	64	3	20	87 (1899)
Jacobsen Lateral	1280-1585	60°	none	-	87	10	30	127 (1859)
East Smitley	1300	290°	none	RG (25)	30	5	20	55 (1931)
Big Snow	1024	70°	Lateral - Ah/C Ah ≈ 5cm	-	63	10	20	93 (1893)
East Saugstad	975	110°	Lateral - Ah/C Ah ≈ 5-7cm	-	59	12	30	101 (1885)
Borealis	1435	35°	Terminal - Ah/C Ah ≈ 2-3cm	RG (20)	80	7	30	117 (1869)
Atavist	1340	355°	none	RG (12)	60	5	20	90 (1896)
Purgatory	595	25°	Terminal - Ah/Bm/C Ah ≈ 15cm Bm ≈ 5cm	-	110+	3	20	133 (1854)

1. Lichen cover refers to most dominant species. RG (Rizocarbon Geographicum); A (Alectoria). Value in brackets is largest measured diameter in millimeters.
2. Terminal moraines only. Correction factor is for sampling height. See text for discussion.

Figure F.3 (A) terminal and recessional moraine sequence (arrow shows terminal moraine) in the upland valley plain of the east Nusatsum watershed. (B) outer left lateral moraine of the Tsini Tsini Glacier. Note the lack of soil development and limited tree growth.

Tsini Tsini Glacier

A series of three moraines have formed on the left flank of the Tsini Tsini Glacier below Mt. Nyland (figure F.3b). The glacier snout terminated in a narrow valley bottom during the glacial maximum thus a terminal moraine was not formed and only limited exposures of moraine can be found on the right lateral margin due to the presence of bedrock. Left lateral moraines exist as comparatively subdued but distinct ridges which separate into a sequence of recession moraines in the down-valley direction.

Soil development was absent from all surfaces since most of the till was angular and blocky but lichens (mostly *Rhizocarbon geographicum*) were visible on the crest of the most distal lateral moraine only. Dendro dates suggest ages of 1875 for the outermost moraine and 1906 and 1936 for the two inner moraines, respectively.

Borealis Glacier

A single well-defined lobe of the Borealis Glacier terminated on a low-gradient upland valley forming a series of semi-circular terminal and recessional moraines (figure F.4a). A thin Ah/C soil (< 10 cm) has developed on the outermost moraine which yielded a dendro age of 1869. A series of four recessional moraines, covering a distance of approximately 300 m, delimit brief pauses in the retreat of Borealis Glacier with only the largest indicating a more lengthy pause (moraine B in figure F.4a). Approximately 5 trees were sampled from the sparsely vegetated surfaces of each recessional moraine but no significant differences in age are apparent (all trees were between 18 and 35 years old). The blocky moraine material and comparatively high elevation of the site (> 1500 m) may have precluded rapid colonization and yielded the fairly homogeneous age structure of the

Figure F.4 (A) moraine field of the Borealis Glacier. The largest and most distal moraine in the photograph obscures the subdued and forested terminal moraine. Lichen diameters show systematic decreases away from the terminal moraine. However, with the exception of the terminal moraine, the age structure for trees is relatively homogeneous. (B) recessional moraine sequence along the northwest margin of Fyles Glacier. Although vegetation cover is scarce, the frequency of lichen and seedlings decreases systematically towards the ice margin.

recessional moraine field or, alternatively, retreat was fairly rapid and the moraines represent intervals of less than a few years.

Rhizocarbon geographicum measurements yield significantly different maximum thallus diameters between the more distal B and C moraines (12 - 20 mm) and the two younger inner moraines (6 - 12 mm). Larger lichens were measured along distal lateral moraines on the south flank of the moraine field. Maximum diameters here approached 50 mm and are probably characteristic of this older terminal moraine. These differences in lichen diameter suggest that age differences between the two groups of moraines (*i.e.* distal and proximal groups) are probably much greater than what would be inferred from the dendro dates. However, since absolute dating was not possible it assumed that the outer Borealis moraine formed in the middle to late 19th century and the younger inner moraines are the product of early 20th century events.

Fyles Glacier

Ice emanating from the northwest corner of the Monarch icefield forms a broad, semi-concentric glacier which flows cross-valley to form the drainage divide between the Talchako and Noeick River valleys. Terminal moraines are formed on three sides of the glacier although the best preserved sequence of moraines is along the northeast margin of the ice. (figure F.4b). Trees sampled at several positions along the terminal moraine indicate that Fyles Glacier began to retreat after 1892 but most of this probably occurred as vertical thinning rather than horizontal movement. Two well-developed recessional moraines are assigned dates of 1924 and 1936. Horizontal retreat along the eastern edge was less than 15 m/year until at least the early 1940s after which recessional rates exceeded 50 m/year.

Lichens are absent from granitic tills younger than 25-30 years suggesting a colonization period of at least this length. *Rhizocarbon geographicum* and *Umbilicaria hyperborea* are the dominant species on all three sets of moraines on the eastern edge of the glacier. Maximum thalli diameters are 35% greater (45 mm) on the outermost moraine compared with two younger inner moraines where maximum lichen diameters are not significantly different (35 mm and 36 mm respectively).

Deer Lake Glacier

A set of two well-defined moraines marks the limit of advance for a tongue of ice flowing down off the northeast flank of Mt. Jacobsen between Jacobsen and Fyles glaciers (figure F.5a). There is no soil development on either surface but dendro dates indicate the outer moraine is somewhat older, forming sometime before 1869. The recessional moraine is dated at approximately 1900 years based on four trees sampled along the western margin. Maximum thalli diameters (mostly *Rhizocarbon geographicum*) on the outer moraine are again larger by 35-40% (45 mm) compared with measurements from the inner moraine. Recession from the 1900 ice front has yielded a very smooth till surface suggesting retreat was fairly rapid following the brief pause at the turn of the century.

Jacobsen Glacier

Jacobsen Glacier is a 9 km long outlet valley glacier emanating from the north side of the Monarch icefield. The left lateral moraine is unconfined and curves sharply in the down valley direction, thus several small bulges formed by ice moving perpendicular to the dominant flow direction, have created a series of terminal and recessional moraines (figure F.5b). In addition to these features, a well-defined terminal

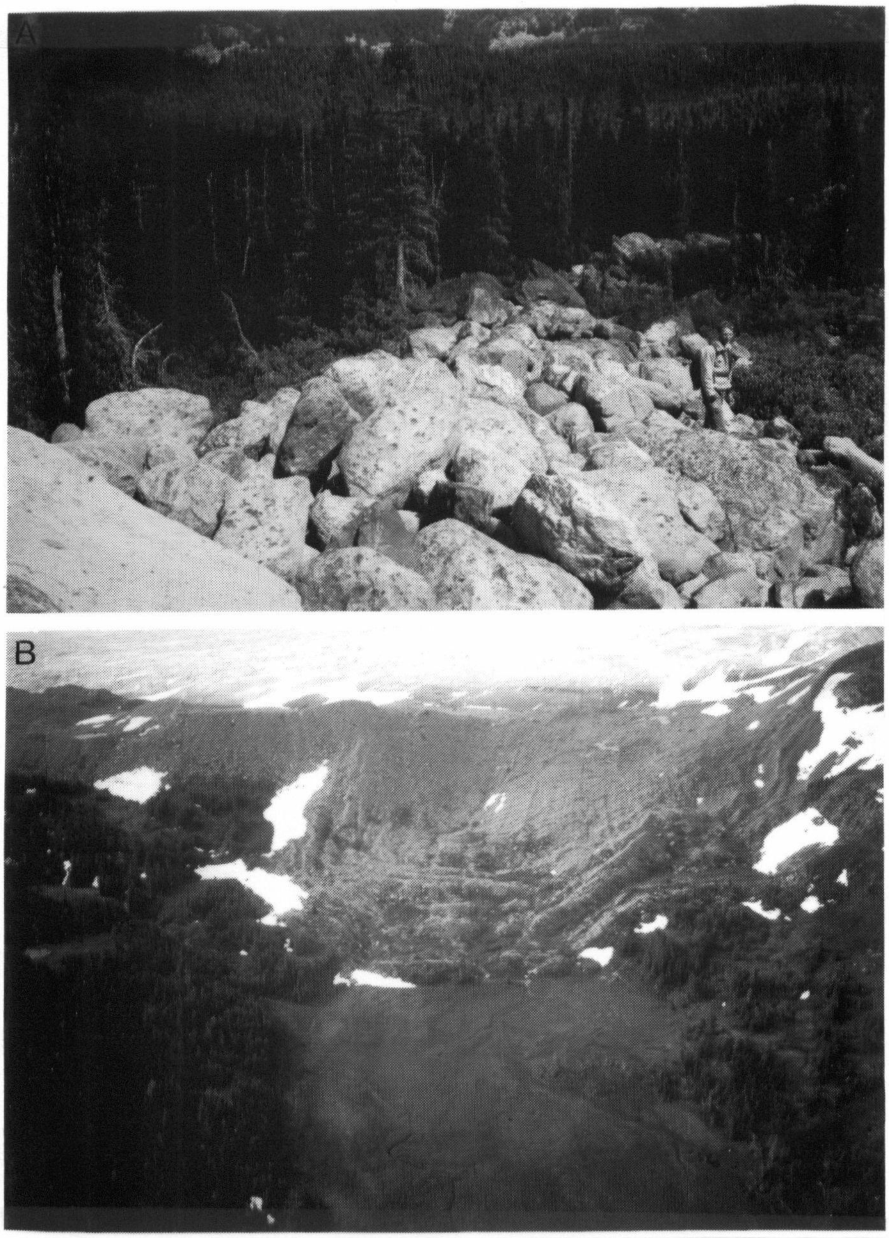

Figure F.5 (A) outer boulder moraine of the Deer Lake Glacier. Note the partly rounded morphology of the morainic debris and subalpine fir growing on the surface. (B) lateral moraine sequence formed by an ice bulge along the unconfined left lateral margin of the Jacobsen Glacier.

moraine was formed. Dendro dates on the terminal moraines indicate that Jacobsen glacier was in a fully-advanced position as late as 1915 with only one major recessional moraine formed prior to 1940. The broad, low-gradient glacier morphology, large accumulation area, northerly aspect and moderately high elevation (1000 m) of the glacier snout may have ensured a comparatively slower retreat and formation of moraines younger than those at other sites within the catchment.