Glaciation, and Glacial History of British Columbia During Fraser Glaciation

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ABSTRACT

In high latitude regions, continental and mountain glaciation are both some of the most significant land sculpting processes. As a result of glaciation, large glacial scour processes erode away glacier-derived sediments which get deposited in new areas affecting water drainage patterns while remolding the topography. Mixing glaciation with active tectonic uplift intensifies the magnitude of glacial erosion (Spotila in press). The history of the Wisconsinan Fraser glaciation physically altered the British Columbia province. Climate fluctuation has controlled the extent and retreat of glaciation thereby affecting the degree of erosion while sculpting valleys in British Columbia. Glaciers have dominated the British Columbia region recently during the Fraser Glaciation. During this glaciation, three phases dominated: 1) ice expansion phase, 2) ice maximum phase, and 3) late glacial phase. Each phase saw large glacial movement and glacial volumetric changes while controlling the surface geology of the region.

INTRODUCTION

The advance and retreat of continental and mountain glaciers are one of the more significant landscape-forming processes in high latitude regions. Glacial scour along with deposition of glacially-derived sediments can play a primary role in shaping the topography and drainage patterns of watersheds. The impacts of glaciation are particularly dramatic in regions that are tectonically active (Spotila in press). In central British Columbia, all major watersheds reflect the interaction of Pleistocene and Holocene glaciation with regional patterns of uplift (Clague 1984, Stumpf 2000). Glacial advance and retreat, driven by climate cycles, has sculpted the valleys and canyons of British Columbia and directly affected rates or erosion of the mountains (McCuaig 2002).

The Skeena River drainage system, located in north-central British Columbia, is one of the largest river drainages of the region. The hydrologic and geomorphic characteristics of the Skeena record regional glacial scour during the Last Glacial Maximum-the Fraser Glaciation (Stumpf 2000). During early phase ice flow, regional glaciers flowed roughly southeast across both the interior of the Skeena watershed and the mountainous western portions of the watershed (McCuaig 2002). During late-phase glaciation, the interior of the Skeena was sculpted by topographically-controlled southeast-oriented glacial flow. The western portions of the watershed traversed the Coast Mountains, which are undergoing rapid tectonic uplift (Stumpf 2000).

This paper reviews the current and past glacial and tectonic processes that have controlled the formation of the Skeena River and its watershed. The goal of this paper is to provide a general context for companion studies focused on habitat conditions for anadromous fishes within the basin. A brief overview of glaciers and glaciation is provided, along with an analysis of the history of the Fraser Glaciation in central British Columbia and its control on watershed topography.

AN OVERVIEW OF GLACIERS

Glaciers are slowly moving bodies of snow and ice that form as a result of the recrystallization of snow due to increased pressure. Snow falls on glaciers, then recrystallizes into firn, which recrystallizes into glacial ice with added pressure and an increase in density (Haeberli 1992). Recrystallization of snow mixed with ongoing precipitation rates in accumulation centers force layered large dense volumetric portions of ice within a glacier to move in a distinct direction (Oerlemans 2001). Figure one visually resembles an entire glacier.

As shown in figure one, a glacier can be subdivided into different facies at the surface and at depth (Menzies 2002). In order to properly categorize all the facies, there are two main physical distinctions which must be made: controls on the surface of a glacier versus controls of a glacier at depth. The surface of a glacier is controlled by precipitation, varying climatic conditions, and how much ice is exposed. The primary surface of a glacier is subdivided into the accumulation zone and the ablation zone with an equilibrium line oscillating between both zones due to near-surface and atmospheric temperatures. In figure one, note the zone of accumulation and the zone of ablation.

Depending which zone the equilibrium line is closer to will control the extent of glacial growth or decline.

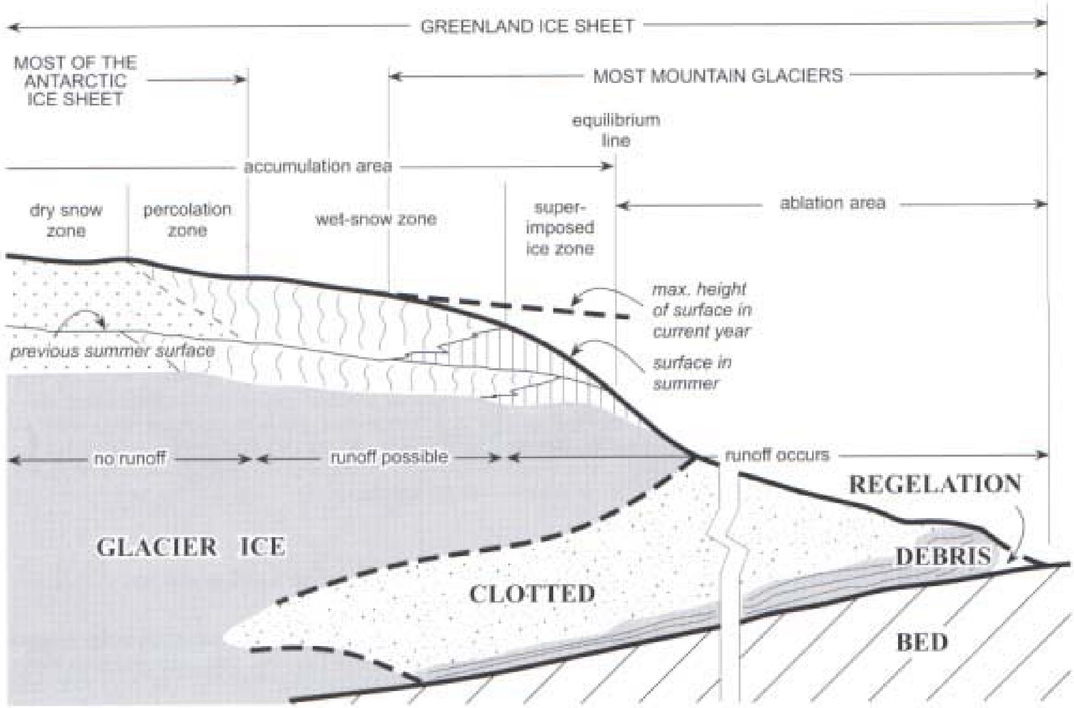

Figure 1. Ice facies distribution on theoretical ice mass (Menzies 2002, modified after Benson and Knight 1992)

The extent of melting in the ablation area hinges on the reflectivity of the material exposed on the glacial surface and the heat from the atmosphere. Fresh snow reflects 80-95%, older snow reflects about 50-80%, firn reflects about 30-60%, and glacial ice reflects about 15-40% (Haeberli 1992). In other words, the older and more compressed the exposed snow and ice, the easier melting occurs by speeding up the process of heat absorption. Heating from the atmosphere can also add to the amount of ablation. Depending on the actual temperature of the surrounding air, the amount of ablation will be greatly affected (Haeberli 1992). The amount of melting due to heating by the atmosphere decreases with higher elevations and higher latitudes, and increases with lower latitudes and lower altitudes. Often temperature alone can be used to estimate ablation rates. At depth, a further ice facies classification can be made based on ice crystalline structures and fabrics (Menzies 2002). Therefore, the controls on the surface of a glacier and the controls of a glacier at depth will control a glacier's facies distribution and control the extent of glaciation.

The temperature distribution within an ice mass is a function of: 1) near-surface and atmospheric temperatures, 2) vertical and horizontal components of velocity at any point within the ice mass, 3) rate of geothermal heat flux at the base of the ice mass, and 4) basal ice friction at the glacier bed (Hooke 1998). Together, these components greatly control the growth, movement, development and longevity of a glacier. Depending on the overall temperature distribution within a glacier, either accumulation zones or ablation zones, or neither will control the size and state of any given glacier (Menzies 2002).

Due to the weight, size, and power which glaciers possess, landscapes can be severely altered. In northern latitudes, small global climate shifts can greatly alter a glacier's surficial characteristics. As a result, widespread glaciation and deglaciation within the northern latitude has shown strong oscillating patterns back and forth through time. Figure 2 visually represents these oscillating patterns.

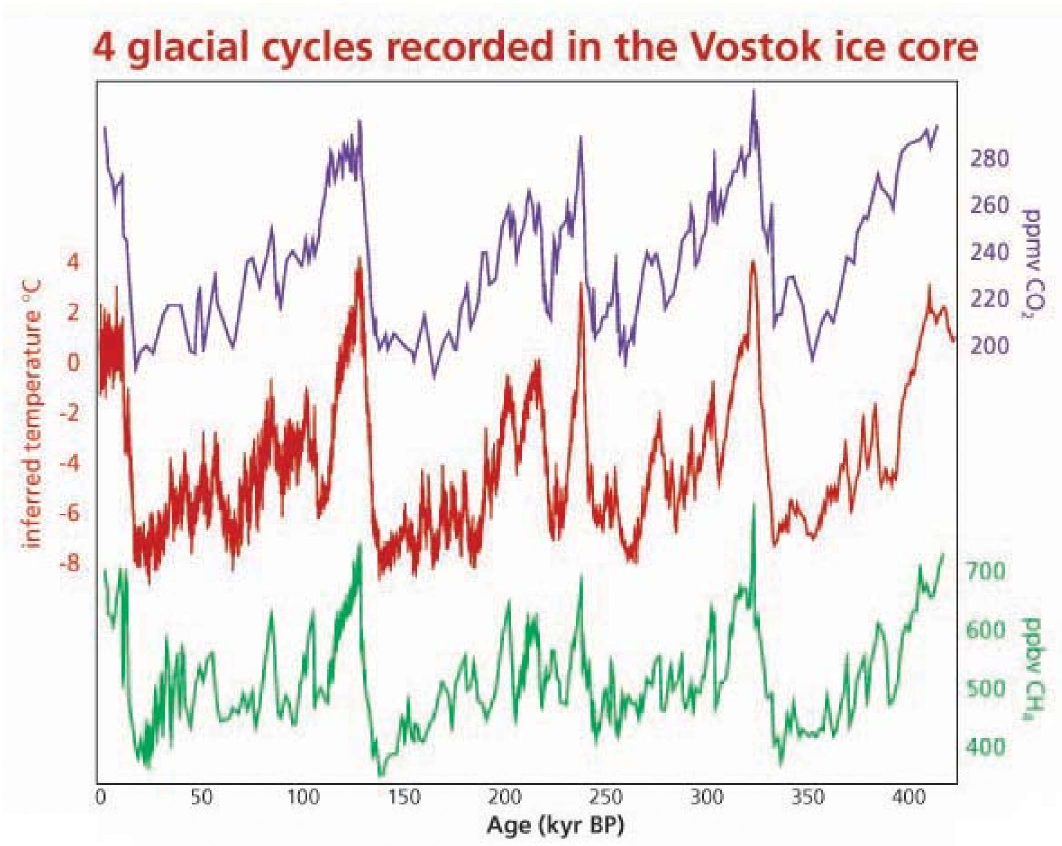

Figure 2. Visually represents climate fluctuations over the last 400,000 years. These climate fluctuations highly control high latitude glaciation. (Petit 1999)

GLACIAL MOVEMENT

With time, what is lost in the ablation area is offset by the movement of snow, ice, and sediments downhill from the accumulation zone. The amount of this material which moves approximately equals the amount of material which falls on a glacier. The movement of a glacier begins in the accumulation area. Movement consists of layers of snow, firn and glacial ice moving over each other with increasing speed as they approach the equilibrium line (Haeberli 1992). Snow recrystallizes into spherical particles called firn, which then can recrystallize and compress into glacial ice with a further increase in pressure (Oerlemans 2001). Acceleration of the ice stretches and extends the glacier downhill, producing crevasses and resulting in the firn being forced deeper into the glacier (Haeberli 1992). Figure 3 shows a conceptualized schematic of glacial movement, while the photo in figure 4 shows the clear delineation between the ablation and accumulation zones.

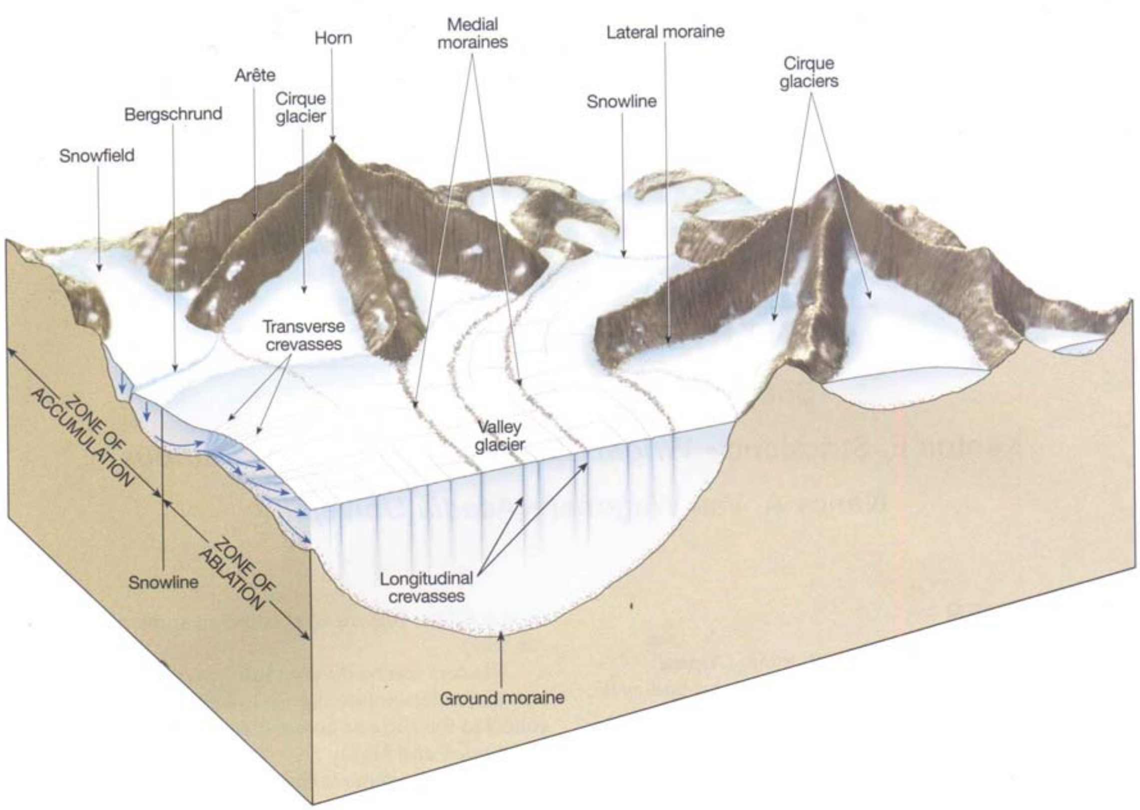

Figure 3. Visually illustrates active mountain glaciation in a hypothetical region. Note cutaway views of glacial ice, showing flow lines and direction (Blue lines and arrows), (American Geological Institute and the National Association of Geoscience Teachers 2003).

Figure 4. The two zones of a glacier are revealed by the differing visibility of crevasses in this aerial photograph of Washington's South Cascade Glacier. Above in the accumulation zone, lower temperatures preserve most of the precipitation thereby limiting the number of crevasses. In the ablation zone, crevasses are clearly evident where melting has occurred (Bailey 1982).

Once the firn and ice reach the ablation zone, the glacier begins to slow down and the force of deceleration causes compression in the ice sheet. Because ice crystals are incompressible, the ice sheet is forced to spread sideways in both directions towards both margins of the glacier. This movement tends to create a thin claw-shaped glacial tongue for mountain glaciation (Benn 2001). Generally, glaciers that move down steep slopes tend to be thin, while glaciers that move down gradual slopes tend to be thick. As the ice decelerates and compression forces the ice to thin, ice in the ablation zone is also forced up to the surface. This newly exposed ice on the surface is very old and has not been in contact with Earth's atmosphere for a long time. The percentages of elements and sediments that exist in this newly exposed ice accurately portray the climates of the past throughout the years when the glacier(s) existed (Benn 2001).

Glacial movement forces massive bodies of ice over bedrock and underlying sediment thereby physically modifying surrounding landscapes. With this movement, glacial scour along with the erosion, transportation, and deposition of glacially-derived sediments plays a major role in altering surrounding topography and drainage patterns of watersheds (Menzies 2002). Early in a glacial cycle, valleys and lowland topography will fill with ice. Mild glacial movement will dominate this lowland topography. Eventually, with increased accumulation, glacial advancement and depth will further increase. The power and pressure behind large glaciers increases exponentially causing larger volumes of glacial scour and glacially-derived sediments to flow with glacial movement. With this glacial mass, erosional features such as cirques, aretes, and glacial troughs will begin to form, while depositional features of glaciated regions like moraines and drumlins will form depositing what was once eroded (Busch 2003). These newly formed erosional and depositional features resemble different components of newly reshaped landscape. With time, additional depositions of glacial-derived sediments will occur further resculpting the landscape. In turn, river drainage patterns will be heavily impacted and altered due to the heavily altered landscape.

The overlying topography of a region act like barriers for glacial movement. Glaciation at first will be confined to lower topographic planes and localized directions. As a glacier or set of glaciers increases in mass, the height and flow of the glacier will change. With increased depth in the ice sheets, topography is less of a barrier, enabling ice to flow in a wide array of directions, even over existing terrain. With varying topography, a combination of different geological regions, and a distribution of differing topographic barriers, glacial movement can occur in a complex variety of directions through time.

Formations Caused by Glaciation

As glaciation progresses, subglacial bedforms and ice-marginal morraines will accrete along and around active glacial movement (Benn 1998). Subglacial bedforms are aligned parallel to moving glaciers made up of accumulations of glacial-derived sediments (Benn 1998). Drumlins, flutings and megaflutings or Rogen moraines are prime examples of subglacial bedforms (Benn 1998). Ice-marginal morraines are depositional features formed near the edges of glaciers. Large unsorted blocks of sediment creating a wide variety of morraines result from glacial flow and will be able to be viewed after a glacier fully recedes and melts (Benn 1998). Through time, these glacial structures become an integrated part of a new surficial geological region.

THE INTERCONNECTION BETWEEN GLACIAL EROSION AND CRUSTAL DEFORMATION

When glacial erosion and tectonic uplift interact, higher rates of erosion often occur. In regions with active tectonic uplift, climate and geographic setting control erosional rates and landscape alteration (Spotila in press). In mountain belts dominated by glaciation, erosion rates are controlled by the extent or magnitude of crustal deformation, climate flux and precipitation. The greater erosional power of ice movement generally contributes to a greater volume of erosion than uplift alone (Spotila in press). However, when climatic conditions are conducive to glaciation in a region with active uplift, erosion has the potential to keep up with the rate of uplift, limiting the height of topography. In this scenario, controls or limitations to the height limit of topography can be met with large rates of erosion (Spotila in press). Denudation, the sum of processes that result in wearing away or the progressive lowering of Earth's surface by weathering, mass wasting, and transportation, is the complete set of methods used with the interconnection of glaciation and crustal deformation for accurately portraying glacial erosional rates (Bates and Jackson 1984). Exhumation, the process of ice uncovering and exposing sediments at depth further increases the speed of tectonic uplift over time (Spotila in press). As a glacier recedes, exhumation coupled with erosion releases pressure on the crust, allowing the surface to rebound and uplift. Once the ice melts, erosion rates decrease as climate and topography become the primary controls (figure 5).

The power and shear force behind glaciation greatly affect the physical characteristics of surrounding landscapes. The study produced by Spotila (in press) concluded that glacially-influenced denudation rates measured over a 100k year time scale are significantly less than rates at a 100 year time scale. Short time scale erosion rates may be the result of a short glacial cycle (Koppes and Hallet 2002). With large denudation rates over short periods of time, tectonic uplift coupled with glacial activity can further increase the magnitude of erosion and landscape alteration in a region.

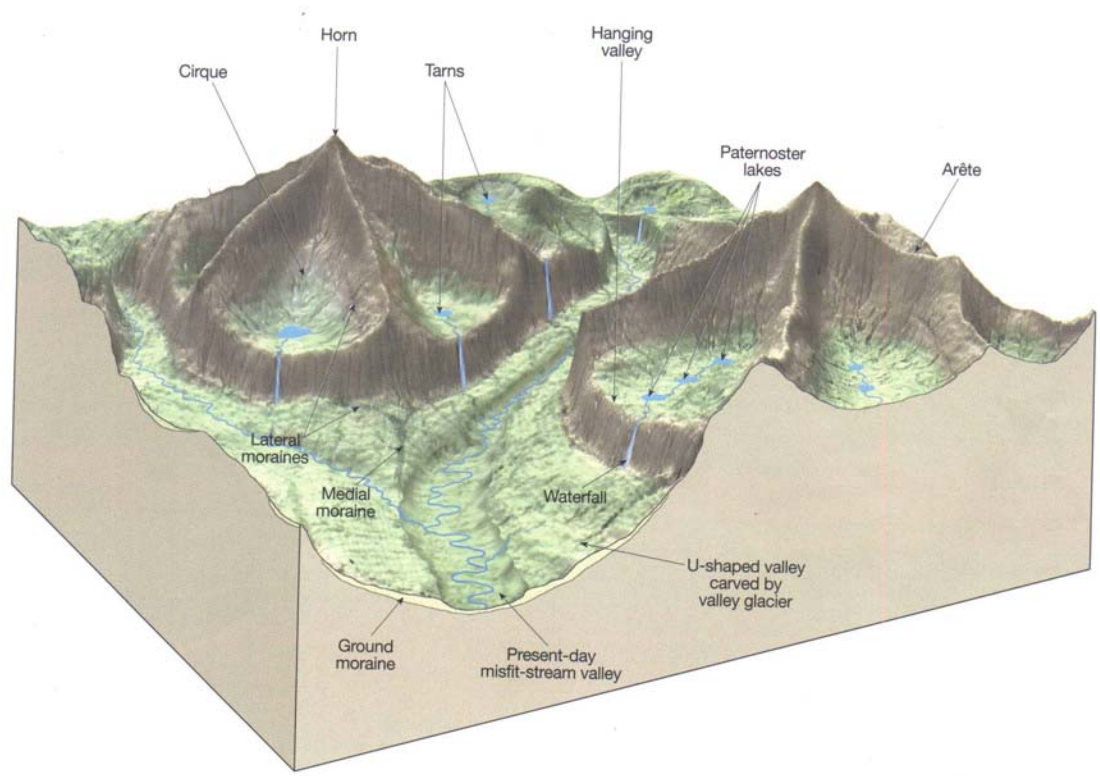

Figure 5. Erosional features remain after the total ablation of glacial ice. This is of same theoretical region as figure 2 (American Geological Institute and the National Association of Geoscience Teachers 2003).

HISTORY OF FRASER GLACIATION WITHIN BRITISH COLUMBIA

The last major glaciation event that covered British Columbia was called the Wisconsinan Fraser glaciation. There were three main glacial movement phases associated with this glaciation: 1) ice expansion phase, 2) maximum phase, 3) late glacial phase (Stumpf 2000). Subdividing the Wisconsinan Fraser glaciation in this way helps describe the dominant trends throughout that glacial period. Climatic alteration in association with the Fraser glaciation began as early as 29,000 years ago (Stumpf 2000). Extensive ice advance from mountains into valleys and fjords did not occur until 25,000 years ago. During this time, the ice expansion phase dominated glacial movement. The Glacial maximum occurred between 14,000-16,000 years ago, marking the glacial movement of the maximum phase (Stumpf 2000). Between 13,000-9,000 years ago, the late glacial phase occurred, and glaciers retreated back to early glacial positions prior to glacial maximum.

During the late Quaternary, colder temperatures and increased precipitation within the Pacific caused a large glacial advance (Stumpf 2000). Through the ice expansion phase period, glaciers grew in volume and in height. Slowly glaciers became interconnected with one another as the overall height of individual glaciers and the ice sheet grew. Due to variations in topographic features and relief, different origins of glacial accumulation forced ice to flow in various directions across the region. As ice built up, these different directional ice flows coalesced into multiphase directional ice flows. Elevation, topographic features and height above sea level controlled glaciation initially, confining glaciers to valleys with lower topographic points. Through time, as the ice sheets grew, topography no longer controlled ice flow. With varying phases of ice flow, the Wisconsinan Fraser glaciation could be described in multiphase flows.

Over the Babine Lake valley and southern Skeena Mountains, ice flow indicators reveal that ice flow followed along two main paths. Earlier during the Fraser glaciation, ice moved southeast over the province. In the Skeena Mountains, ice flow indicators show a second westerly ice flow direction at elevations around 2,200 meters and higher (Stumpf 2000). Today’s topography reflects these ice flows. Babine Lake is linear in shape and trends southeast and the Skeena river flows west to the Pacific Ocean.

In the Northern Skeena and Omineca Mountains, glaciers expanded outwards from alpine ice centers downslope along valleys during the early part of the Fraser glaciation (Stumpf 2000). These ice flow patterns were complex and were thought to vary a bit due to topography and local shifts and reversals of altering ice accumulation centers (Stumpf 2000). During the maximum extent of glaciation, ice flows paralleled mountain ridges within the Skeena and Omineca mountains, and generally trended eastward.

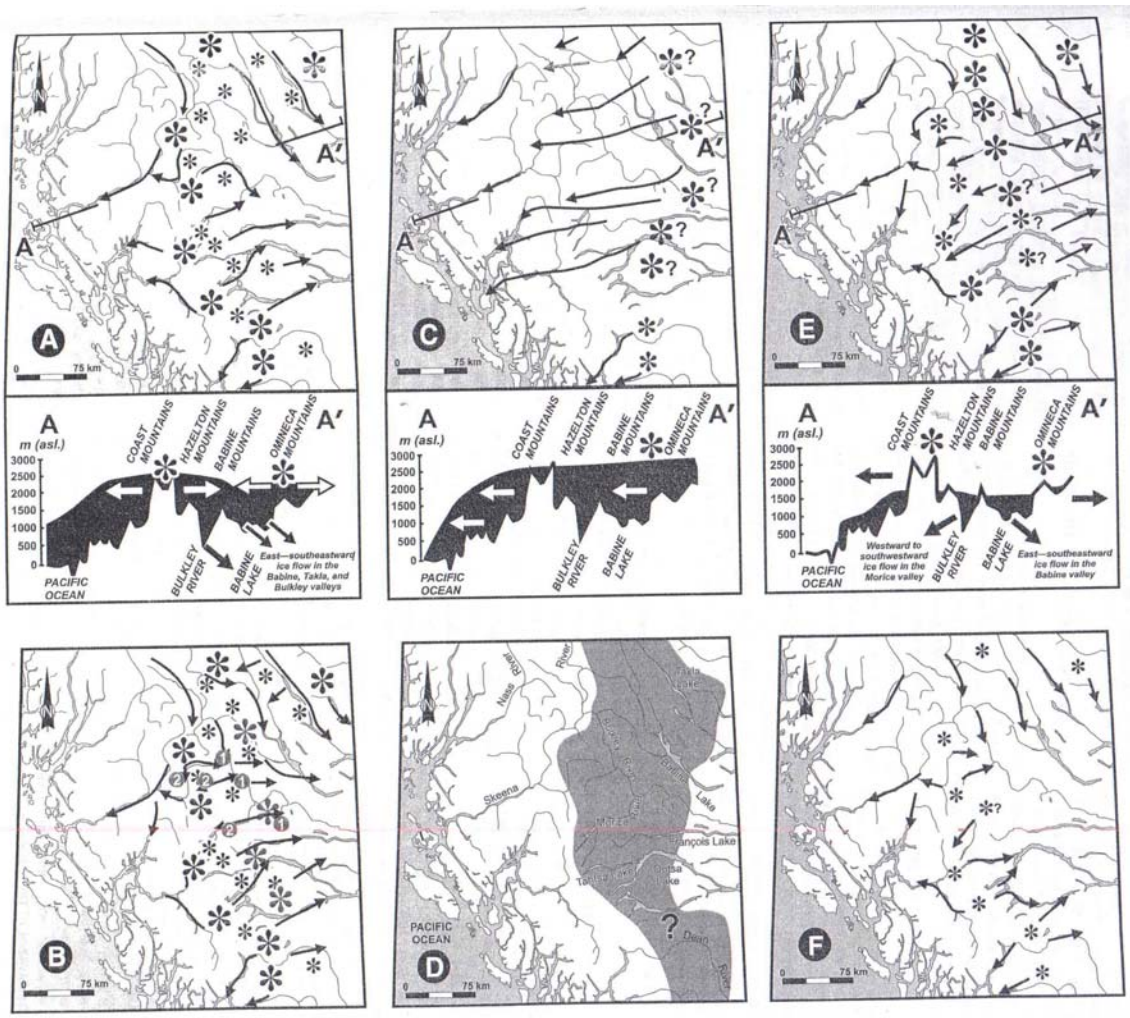

Figure 6. Models for ice expansion, maximum and late glacial phases of Fraser Glaciation. Arrows indicate ice flow directions while asterisks denote approximate locations of major and minor ice accumulation centers throughout Fraser glaciation (Stumpf 2000).

Figure 6 from Stumpf (2000) visually illustrates ice accumulation centers and the movement of ice during the ice expansion phase, ice maximum phase, and late glacial phase. During ice expansion phase (A), ice center locations (black asterisks) were located over high mountains. Continued glacial growth caused ice accumulation centers to shift east of the Hazelton and Coast Mountains (B). During maximum glaciation, ice centers shifted towards the interior (C). In (D), shifts in ice flow direction occurred throughout the maximum phase. Rapid thinning of ice during late glacial phase resulted in ice accumulation centers shifting westward (E and F). Early in late glacial phase, ice sheets were still thick enough to to flow over low mountains and ridges, but forced to flow around high mountains (E). Later in late glacial phase, glacial flow again became locally controlled by topography (F).

Ice Expansion Phase

This phase incorporated the time, which led to the growth of the Wisconsinan Cordilleran ice sheet and lasted well into the last major glaciation period called the Fraser glaciation. Within this large block of time, there were a few subphases as glaciers developed and moved. Alpine glaciation dominated early on within the phase, while more intense Alpine glaciation and eventually mountain ice sheet phases dominated later (Stumpf 2000). Initially, the alpine glaciers confined to valleys expanded outward from major high mountain accumulation centers into interior and coastal areas. Ice expansion led to glaciers flowing westward to the Pacific Ocean from the Coast and Skeena Mountains, and southeastward along the Babine, Takla and Bulkley valleys within the Skeena and Omineca Mountains (Stumpf 2000). As this ice development progressed, most areas were influenced and controlled by major topographic features. Throughout the interior, valley glaciers integrated into a system of plateau ice sheets while continuing to flow along major valleys. In some regions, glaciers became thick enough to flow over low ridges.

Ice flow data from Stumpf (2000) indicate that late in the ice expansion phase, ice was thick enough to flow over low mountains and along passes at 1700 meters above sea level. Ice flow data from Stumpf (2000), based on erosional features found in southeast trending valleys, indicate that the ice expansion phase was the most erosive event during the Fraser Glaciation time period. These data also imply that later glacial flows were unable to extensively modify existing landforms or destroy much of the early ice expansion flow record. West of the Bulkley River, indicators of ice expansion were not present suggesting that a later flow with westerly erosional patterns destroyed some of the ice expansion flow record (Stumpf 2000).

Maximum Ice Flow Phase

Additional glacial development and growth within the region forced individual glaciers to coalesce into one primary ice sheet known as the Cordilleran Ice Sheet. During this time, the glacial maximum of the Fraser Glaciation period occurred. Throughout this phase, major accumulation centers of ice shifted inland, further altering and changing flow directions across the region (Stumpf 2000). Prior to the absolute maximum of glaciation, ice flow maintained its southwestern movement throughout the Skeena, Hazelton, and Coast Mountains; however, once the maximum peak occurred, ice flows split into two primary directions.

Throughout glacial maximum, the Cordilleran Ice Sheet incorporated a minimum surface elevation of over 2500 meters above sea level, overlying topography in the interior while moving upslope over higher mountains closer to the ocean to the west (Stumpf 2000). The ice flows were derived from several major centers of ice accumulation that formed an ice divide located east of the Babine Lake valley (Stumpf 2000). This divide potentially extended both southward and northward from this region forcing ice to flow in two opposite directions. To the west, ice flows headed through fjords to the Pacific Ocean. East of the divide, ice flows moved eastward into the interior. Theoretically, this divide must not have been constant lengthwise over time. Some regions of ice within the divide likely moved and shifted, while other portions remained relatively stationary. While varying parts of this divide shifted throughout the maximum ice flow phase, the divide generally separated ice flow into two dominant flow patterns to the east and to the west on both sides of the divide.

Late Glacial Phase

Starting around 13,000 years ago, climate began to influence glaciation within the Cordilleran Ice sheet. With an increase in temperature, rapid thinning of the Cordilleran Ice Sheet occurred. Major accumulation ice centers shifted west towards the Skeena, Omineca, Hazelton and Coast Mountains, back to similar zones prior to Fraser Glaciation maximum (Stumpf 2000).

As time passed, glacial thinning persisted. As the elevation of the top of the Cordilleran Ice Sheet decreased, glacial movements became controlled again by topography. West of the Cordilleran Ice Divide, glacial movement persisted westward through large valleys, but was locally controlled by mountain passes and low-lying ridges. In particular, eastward slopes where ice had flowed upslope during the glacial maximum, flows either ceased or returned to downslope directions.

An explanation for the thinning of the Cordilleran Ice Sheet during the late glacial phase can be found in the equilibrium line theory. As discussed above, the location of the equilibrium line within a glacier influences accumulation and ablation rates. When the equilibrium line is high on a glacier, the ablation zone dominates increasing the rate of melting. When the equilibrium line is low on a glacier, the accumulation zone is dominant increasing the rate of growth and size of the glacier. Throughout the late glacial phase, the equilibrium line shifted upwards throughout the ice sheet and in some places rose above the elevation of the surface. Ablation was dominant and widespread throughout this phase causing the Cordilleran Ice Sheet to thin. Figure 7 visually illustrates the large distribution of sediments deposited during pre-fraser glaciation, during fraser glaciation, during the retreat of fraser glaciation and during postglacial time.

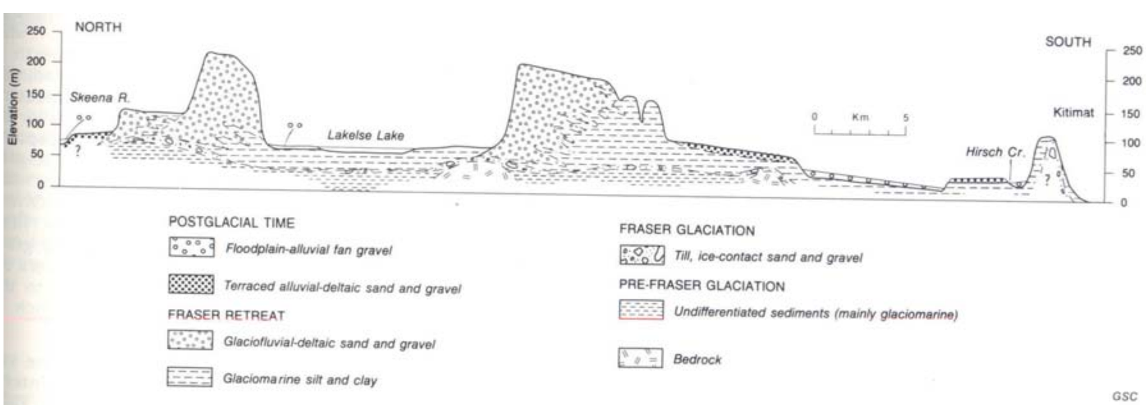

Figure 7. Schematic diagram illustrating stratigraphic relationships of Quaternary sediments in the Kitsumkalum-kitimat trough between the Skeena River and Kitimat. Coupling of deltaic and glaciomarine sediments is visually represented (Clague, 1984).

Outlets of Glacial Flow

Throughout the Fraser Glaciation, ice flows entered the Pacific Ocean and the Canadian interior through dominant outlets. Along the Pacific Coast of Canada, five major outlets for ice flows existed throughout Fraser Glaciation: the Nass River Valley, the Skeena River Valley-Dixon Entrance, the southern part of the Kitsumkalum Kitimat Trough, the Gardner Canal, and the Dean and Burke Channels (Stumpf 2000).

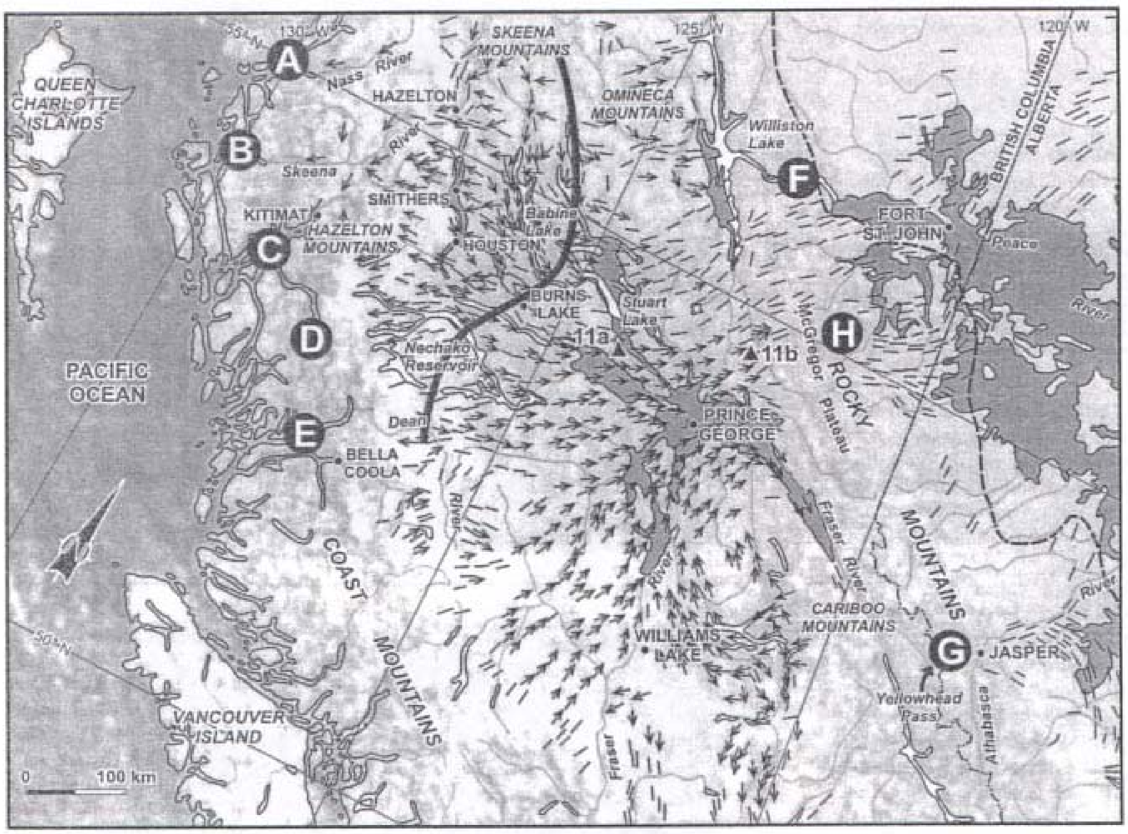

Figure 8. Shown are major outlets of Fraser Glaciation; the Nass River Valley (A), the Skeena River Valley-Dixon Entrance (B), the southern part of the Kitsumkalum Kitimat Trough (C), the Gardner Canal (D), and the Dean and Burke Channels (E) (Stumpf 2000).

Throughout eastern British Columbia and western Alberta, glaciers generally flowed eastward through mountain valleys and passes. The primary outlets were Williston Lake-Peach River, Fraser River-Yellowhead Pass-Athabasca River, and over low ridges and mountains in the Rocky Mountains (Stumpf, 2000).

DISCUSSION

Continental and mountain glaciation are some of the most significant land sculpting processes in high latitude regions. As a result of glaciation, massive amounts of glacier-derived sediments are eroded away and deposited in new areas affecting the topography and water drainage patterns. When glaciation is combined with active tectonic uplift, the magnitude of glacial erosion is intensified (Spotila in press). Climate flux has controlled the extent and retreat of glaciation thereby affecting the degree of erosion while sculpting valleys in British Columbia. The history of the Wisconsinan Fraser glaciation physically altered the British Columbia province while leaving behind physical records of their existence.

The Skeena River is located where most of the development and progression of the Wisconsinan Fraser glaciation took place. Glacial scour and glacial-derived sediments deposits show past directional flows of glaciation. This river drainage system is one of the largest drainage systems in the region, and even today is still geomorphologically altering the landscape.

The Fraser Glaciation, which most recently dominated British Columbia was composed of three main phases: 1) ice expansion phase, 2) ice maximum phase, and 3) late glacial phase. Through time, oscillating precipitation conditions and a change in climate altered the extent of the Fraser Glaciation. This glaciation drastically altered the surface geology of British Columbia, resulting in the topography and the river drainages we see today.

REFERENCES

American Geological Institute and the National Association of Geoscience Teachers. 2003. Laboratory Manual in Physical Geology (Sixth Edition). Pearson Education, Inc.

Bailey, Ronald H. 1982. Planet Earth: Glacier. Time-Life Books, Alexandria, Virginia.

Bates, Robert L., and Jackson, Julia A. 1984. Dictionary of Geological Terms-Third Edition-Prepared by the American Geological Institute, Random House, Inc.

Benn, Doug I. and Evans, Doug J.A. 1998. Glaciers and Glaciation. Arnold Publishing.

Benson, C.S. 1962. Stratigraphic Studies in the snow and firn of the Greenland Ice Sheet. U.S. Army CRREL Rept. 70: 1107.

Cannings, Sydney and Cannings Richard. 1999. Geology of British Columbia: A Journey Through Time. Greystone Books, Vancouver / Toronto.

Clague, John J. 1984. Quaternary Geology and Geomorphology, Smithers-Terrace-Prince Rupert Area, British Columbia. Geological Survey of Canada.

Garrison, Tom. 2002. Oceanography, An Invitation to Marine Science (Fourth Edition), Brooks/Cole Thomson Learning Publishing.

Haeberli, W. 1992. Glaciers and the Environment. UNEP/GEMS Environment Library No 9.

Hooke, R. LeB. 1998. Principles of Glacial Mechanics. Prentice Hall, Engelwood Cliffs, NJ.

Hutchison, W.W. 1982. Geology of the Prince Rupert-Skeena Map Area, British Columbia. Geological Survey of Canada.

Knight, P.G. 1992. Glaciers. *Progress in Physical Geography*. 1: 85-89.

Koppes, M.N. and Hallet, B. 2002. Influence of Rapid Glacial Retreat on the Rate of Erosion by Tidewater Glaciers: *Geology*. 30: 47-50.

Lawson, Daniel E. 1993. Glaciohydrologic and Glaciohydraulic Effects on Runoff and Sediment Yield in Glacierized Basins. American Society for Testing and Materials.

McCuaig SJ, and Roberts MC. 2002. Topographically-independent Ice Flow in Northwestern British Columbia: Implications for Cordilleran Ice Sheet Reconstruction. *Journal of Quarternary Science* 17: 341-348.

Menzies, John et al. 2002. Modern & Past Glacial Environments. Butterworth Heinemann Publishers.

Oerlemans, J. 2001. Glaciers and Climate Change. A. A. Balkema Publishers.

Owen, Lewis A. 1998. Mountain Glaciation. *Supplement 1 to Journal of Quaternary Science*. 13.

Petit, J.R. et al. 1999. Climatic and Atmospheric History of the Past 420,000 years from the Vostok ice Core, Antarctica. *Nature*. 399: 429-436.

Sharp, Robert P. 1988. Living Ice; Understanding Glaciers and Glaciation. Cambridge University Press, Cambridge.

Spotila, James A. et al. in press. Long-term Erosion of Active Mountain Belts: Example of the Chugach-St. Elias Range, Alaska.

Stumpf AJ, et al. 2000. Multiphase Flow of the Late Wisconsinan Cordilleran Ice Sheet In Western Canada. *Geological Society of America Bulletin* 112: 1850-1863.

Van Der Veen, C.J. 1999. Fundamentals of Glacier Dynamics. A.A Balkema publishers, Rotterdam / Brookfield.