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66

Compendium of Forest Hydrology and Geomorphology in British Columbia
Volume 1 of 2

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BRITISH COLUMBIA
The Best Place on Earth

Ministry of Forests and Range
Forest Science Program

FORREX
Supporting Sustainable Natural Resource Management Decisions

Compendium of Forest Hydrology and Geomorphology in British Columbia

Volume 1 of 2

Robin G. Pike, Todd E. Redding,
R.D. (Dan) Moore, Rita D. Winkler,
and Kevin D. Bladon (editors)

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Over the last two decades, hydrologists and geomorphologists have often discussed the need to document the history, scientific discoveries, and field expertise gained in watershed management in British Columbia. Several years ago, a group of watershed scientists from academia, government, and the private sector gathered at the University of British Columbia to discuss the idea of a provincially relevant summary of hydrology, geomorphology, and watershed management. Their main objectives were to bridge the sometimes disparate views in watershed science with an integrated understanding of forest hydrology and geomorphology and to create a "go-to" reference for this information. Through this meeting, the *Compendium of Forest Hydrology and Geomorphology* was born.

As a ***synthesis*** document, the *Compendium* consolidates our current scientific knowledge and operational experience into 19 chapters organized around six themes: the regional context, watershed hydrology, watershed geomorphology, water quality, stream and riparian ecology, and watershed management decision support. These chapters summarize the basic scientific information necessary to manage water resources in forested environments, explaining watershed processes and the effects of disturbances across different regions of the province. Some chapters incorporate case studies highlighting pertinent examples that move discussions from the abstract and theoretical to the applied and practical. Each chapter also presents a comprehensive list of references, many of which are electronically linked for reader convenience. In short, the *Compendium* is about British Columbia and is primarily intended for a British Columbian audience, giving it a uniquely regional focus compared to other hydrology texts.

To ensure that the *Compendium* presented reliable, relevant, and scientifically sound information, chapters underwent extensive peer review employing the standard double-blind protocol common to most scholarly journals. Each chapter was reviewed by three to five peers, several steering committee members, an English editor, and executive reviewers from the B.C. Ministry of Forests and Range and FORREX. With 67 authors and 84 peer reviewers, the *Compendium* embodies the spirit of partnerships—strengthening connections among colleagues, agencies, and disciplines. Although the *Compendium* focusses on British Columbia, its genesis and development involved hydrologists, geomorphologists, and related professionals from across Canada, the United States, and around the world. The "Authors" section (page v) lists all authors and their affiliations, and the "Peer Reviewers" section (page x) provides a record of all peer reviewers.

At over 800 pages, the *Compendium* showcases the rich history of forest hydrology, geomorphology, and aquatic ecology research and practice in British Columbia and sets the foundation for the future by showing us how much more we have yet to learn. We hope it will become a valuable resource for students, water resource professionals, and anyone else interested in water in British Columbia.

Robin G. Pike, Todd E. Redding, R. Dan Moore, Rita D. Winkler, and Kevin D. Bladon
June 2010.

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This *Compendium* could not have been produced without the significant efforts of numerous individuals and organizations. We extend our sincerest appreciation and thanks to all who contributed to this project.

FORREX staff provided leadership throughout the project's life. Their commitments to extension and project management ensured that the *Compendium* made that important leap from an idea to a reality. Christine Hollstedt (CEO), Julie Schooling (publications), and Shelley Church (publications) were instrumental in all aspects of the *Compendium*'s development. The Steering Committee also played a crucial role in shaping the *Compendium* by providing the technical direction necessary to keep the project focussed on its objectives. Past steering committee members included Dave Toews, Eugene Hetherington, Doug Golding, and Rob Scherer.

Undoubtedly, the largest contribution was from the *Compendium*'s 67 volunteer authors (see Authors). Words cannot express our deep appreciation of them and their organizations in the development of this project. In addition, 84 provincial, national, and international technical specialists from government, industry, and academia voluntarily peer reviewed all *Compendium* chapters to ensure that high standards were maintained (see "Peer Reviewers").

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We wish to formally thank our colleagues, employers, and funding agencies for their help, support, and patience throughout the *Compendium*'s evolution from concept to completion. We also thank the early hydrologists and geomorphologists who first formulated the idea of a comprehensive document to share their knowledge.

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CONTENTS

APPENDICES

TABLES

FIGURES

CHAPTER 1

Forest Hydrology in British Columbia: Context and History

DAVID A.A. TOEWS AND
EUGENE D. HETHERINGTON

INTRODUCTION

In British Columbia, water issues are frequently at the forefront of public concern over forest management. Of highest concern are activities that might affect drinking water quality and supply, flooding, landslides, and fish habitat. Over the past 40 years, our knowledge of the effects of resource management on forest hydrology and geomorphology has steadily increased in British Columbia. Forest hydrology research examines the ways in which forests and associated wildland vegetation affect the water cycle, including the effects on erosion, water quality, and microclimate (Hewlett and Nutter 1969). The study and practice of geomorphology is related to the processes that affect the development of landforms, and in a forest management context is concerned with landslides, erosion, and sediment transport. The two disciplines are therefore inextricably linked and the dynamic interactions between hydrologic and geomorphic processes in forested landscapes affect watersheds, water resources, aquatic habitat, infrastructure, and public safety.

Forest hydrology and geomorphology research began in response to the need to understand how to protect watershed, water resource, and public values while harvesting timber. The purpose of this compendium is to provide a comprehensive summary of what we have learned about forest hydrology, geomorphology, and stream ecosystems in British Columbia. To provide context for the chapters that follow, this chapter reviews the evolution of forest hydrology research and practice in British Columbia focussing on basic processes, research, and practice.

HISTORY OF WATERSHED MANAGEMENT IN BRITISH COLUMBIA

Because the Crown owns most of the forested land in British Columbia, the ultimate responsibility for land management lies with the provincial government. The forest industry is charged with planning and implementing timber harvesting according to policies set by government. Historically, the British Columbia Forest Service has been the lead agency in forest stewardship. Agencies[1] representing watershed and fish habitat protection were the British Columbia Ministry of Environment's Fish and Wildlife Branch

1 The names and responsibilities of the agencies involved with water and natural resources have continually changed over the years and currently include: the British Columbia Ministry of Environment, Ministry of Forests and Range, Ministry of Agriculture and Lands, Ministry of Health Services, Ministry of Healthy Living and Sport, and Ministry of Energy, Mines and Petroleum Resources; and Fisheries and Oceans Canada. First Nations organizations throughout British Columbia also play an active role in resource management.

(responsible for non-anadromous fish), the federal Department of Fisheries and Oceans (anadromous fish), and the provincial Water Rights Branch (community watersheds). Originally, fish habitat protection was the major focus of interactions among these agencies. Legislative authority provided by the federal *Fisheries Act* prohibited deposition of deleterious substances into streams and mandated the protection of fish habitat. Drinking water supply was (and continues to be) another important area of concern, particularly in watersheds designated as available for timber harvest. Conflicts between these values are most prominent in areas where inhabited private land and Crown forest land intermingle, particularly in the Okanagan, the Kootenays, and on Vancouver Island.

Conflicts between agencies regarding timber harvesting and silvicultural practices, along with a common interest in protecting watershed values, have provided the main stimuli for the development of watershed management guidelines and the initiation of forest hydrology and geomorphology research in British Columbia.

The 1960s

The 1960s—a time of growing environmental awareness—was a period of rapid expansion for the forest industry in British Columbia. At the same time, the need for forest hydrology research emerged in response to a general lack of knowledge about how to protect watershed values in areas of commercial timber harvesting activity. As operations and roads extended into watersheds, poor practices—for example, inappropriate skidding on steep slopes, operating machinery in stream channels, and constructing substandard roads and culverts—led to consequences such as excessive soil disturbance, muddy water, and landslides.

At a time when the forest industry focussed on harvesting and processing timber at the least cost, forest hydrology and geomorphology research in British Columbia was non-existent and therefore research from elsewhere was used to answer local questions. For example, the proceedings of an international hydrology symposium held in Pennsylvania were used extensively (Sopper and Lull [editors] 1967). Some of the articles explained basic forest hydrology concepts, including the effects of harvesting on water yield and the effects of roads on sedimentation in streams. Results from the Alsea study in coastal Oregon, which documented the effects of harvesting on water quality, streamflow, and aspects of fish habitat, were also beginning to emerge (Brown and Krygier 1971; Harr et al. 1975).

To the best of our knowledge, the first attempt to establish a watershed study in British Columbia was the Genesee Creek project. This project was initiated in 1968 by the Department of Fisheries and Oceans on a salmon-rich stream flowing into Owikeno Lake near Rivers Inlet; however, it was discontinued in 1972 owing to the expense of working in a remote location and the impracticality of harvesting the upland watershed. The Genesee Creek project demonstrated the challenges involved in conducting watershed studies, including the need for a rigorous approach and for government and industry cooperation.

Dr. Walt Jeffrey of the Faculty of Forestry at the University of British Columbia is credited with initiating forest hydrology research in British Columbia. Before Jeffrey's arrival at the university in 1966, hydrology was only a minor component in one of the faculty's courses. Jeffrey's passion for forest hydrology extended beyond teaching undergraduates to acquiring funding and recruiting a strong cadre of capable graduate students. Forest hydrology teaching and research continues at the university; the richest legacy of the 1960s is the training of graduate students who have carried on the work in forest hydrology in British Columbia and across Canada.

Dr. Jeffrey publicly advocated for stream protection. He described land use hydrology as a regional science and maintained that local studies were needed to appropriately apply lessons learned elsewhere (Jeffrey 1970). He set up research studies, encouraged others to do likewise, and supported land managers in applying results in their jurisdictions. In addition, he recruited an experienced forest hydrology researcher from the U.S. Forest Service, Dr. Bert Goodell, who continued the research and teaching following Jeffrey's untimely death in 1969. Faculty of Forestry researchers started the Jamieson/Elbow Creeks paired-watershed experiment (1968–1993) in the Seymour River catchment (Golding 1988, 1995), but it did not yield substantial results because of calibration and jurisdiction issues.

The 1970s

In the 1970s, governments and the forest industry responded to growing environmental awareness by hiring more staff and initiating new research projects. For example, the first forest hydrologists were hired by the Department of Fisheries and Oceans in 1971, the Canadian Forest Service in 1972, and the British Columbia Forest Service in 1975. In 1978, B.C. Forest Products Limited became the first forest company to hire a forest hydrologist. Key findings in forest hydrology were summarized in the proceedings of the 1970 symposium on forest land uses and stream environment (Krygier and Hall [editors] 1971) in Oregon. Although this document remained a significant reference for some time, pressure mounted within the research and forest management communities to produce local information that would be applicable to British Columbia.

A workshop on watershed management in Parksville in 1973 presented several important policy changes aimed at protecting water supplies. The most notable policy was the introduction of 11 stream protection directives that became known as "P" clauses. These were included in cutting permits as required. Key among the P clauses was the prohibition of the operation of machinery within stream channels and the deposition of polluting materials into streams. The clauses also required licensees to fell timber away from streams and construct bridges or culverts on every crossing. Although evidence suggests that operational staff worked towards improving watershed management, many unanswered questions remained.

The first major watershed study in British Columbia to produce published results was initiated at Carnation Creek in 1970. The site was chosen after a thorough search of watersheds throughout coastal British Columbia—even at this time, it was difficult to find an untouched drainage suitable for research. The federal Department of Fisheries and Oceans and the Canadian Forest Service took the lead in initiating the study, in co-operation with MacMillan Bloedel Limited and the British Columbia Forest Service. The Carnation Creek project, which went through pre-harvest calibration (1970–75; Figure 1.1), harvesting (1976–81; Figure 1.2), and post-harvest monitoring (1982–present; Figure 1.3) stages, is the longest-term and most highly published watershed study in British Columbia. The roles of the partners have changed over the years, with provincial agencies gradually assuming responsibility for the project's funding and direction.

During the 1970s, an increase in public concern for environmental protection prompted expansion in government regulation of forestry practices. Some of the new guidelines for harvesting in coastal areas (see Cameron 1972), such as the placement of a maximum limit of 200 acres (81 ha) on the size of clearcuts, were controversial and generated considerable debate. Some people argued that the net result would be increased environmental damage owing

FIGURE 1.1 *Carnation Creek tributary C weir, December 1973. (Photo: B.C. Ministry of Forests and Range)*

FIGURE 1.2 *Logging on the Carnation Creek floodplain, 1976. (Photo: Natural Resources Canada, Canadian Forest Service)*

to the extra roads required. The implementation of the guidelines led land managers to seek answers through research and thus they became more interested in studies such as those at Carnation Creek.

A second important 1970s study was the Slim-Tumuch project east of Prince George, which was initiated in 1972 by the British Columbia Fish and Wildlife Branch, with participation from the federal fisheries agency (Brownlee et al. 1988). This was a short-term study with no pre-harvest calibration. Harvested watersheds were compared with unharvested watersheds. The objective was to evaluate the effects of timber harvesting on water quality, fish habitat, and fish populations. From a water-quality standpoint, one of the main findings was that most of the harvesting-related sediment entering streams came from a single source of lacustrine material, a result that might have been avoided if the road had been located elsewhere.

The published studies from elsewhere (e.g., Krygier and Hall [editors] 1972) and the unpublished presentations of the Carnation Creek and Slim-Tumuch projects conducted in the early 1970s had considerable influence on forest practices. At the time, the forest industry was rapidly expanding, and as new road systems were built in drainages without previous anthropogenic development, questions were raised about the potential impacts. Watershed management was based on a relatively simple referral system that allowed other agencies (e.g., the Depart-

FIGURE 1.3 *Carnation Creek weather station N (Photo: D. Spittlehouse)*

ment of Fisheries and Oceans, British Columbia Fish and Wildlife Branch, and the provincial Water Rights Branch) to comment on harvesting proposals before implementation. Because little was known about local resource values, the habitat protection biologists and other resource professionals practising at this time were sometimes at a loss when asked to make site-specific comments on harvesting proposals.

Also in the 1970s, efforts were under way to develop appropriate resource planning tools. For example, the folio system, which was developed in the Prince George Forest Region but was applied widely to other areas of the province, represented one attempt to identify critical resource issues more efficiently. The system required each agency to identify its important resource interests on transparent maps that could be placed over forest development plan maps. In practice, however, this system was often ineffective because of a lack of inventory data. Another project undertaken at the same time by the British Columbia Forest Service on Vancouver Island developed site-specific, streamside management prescriptions (Moore 1978).

The 1970s also saw provincial agencies working on new multiple-resource inventories. The provincial Resource Analysis Branch of the Environment and Land Use Committee Secretariat was established to develop inventory methods for environmental resources other than timber. This program allowed considerable strides to be taken toward understanding the distribution and nature of various resources in British Columbia, including stream networks and surficial geology. Of note, surficial geologists in the Resource Analysis Branch and the forest industry developed methods for rating and mapping slope stability, and they trained many professionals in these methods. The resulting maps are now one of the main tools for those involved in watershed protection.

Innovations in slope stability mapping were also being made by the forest industry. In response to agency pressure and *Fisheries Act* charges related to landslides on steep land on Haida Gwaii (formerly the Queen Charlotte Islands), MacMillan Bloedel Limited pioneered a four-class slope stability mapping and classification system. These maps focussed on identifying and appropriately managing unstable terrain. The efforts of the Resource Analysis Branch and industry eventually culminated in a widely adopted terrain mapping system (Howes and Kenk 1988). Another initiative resulted in the addition of environmentally sensitive areas to British Columbia Forest Service inventory maps.

In 1979, a historic conflict arose on Haida Gwaii concerning the approval for harvesting in the Riley Creek watershed, an area particularly prone to landslides (Donnelly and Martin 1980). Initially, the Department of Fisheries and Oceans did not object to the original referral and the British Columbia Forest Service issued a cutting permit. The Department of Fisheries and Oceans reconsidered its stance following a significant storm in the fall of 1978 that resulted in landslides. Fisheries personnel notified the forest licensee that unless harvesting was suspended, the forest company would be charged under the *Fisheries Act*. When the company continued to log, a company official was charged with obstruction of justice and fallers were arrested. A major impasse resulted and senior politicians, including the Premier, the federal Minister of Fisheries, and the Prime Minister, became involved. Finally, all parties concurred that harvesting could take place in the area provided that a safe method of doing so could be found. A steep portion of the block directly above a tributary stream was deleted from the cutting permit and the road system was deactivated immediately after the completion of harvesting.

This incident at Riley Creek proved to be pivotal in shaping forest hydrology, geomorphology, and fish/forestry research in British Columbia. It led to the formation of the Fish-Forestry Interaction Program in 1981, which culminated in a final symposium 15 years later and led to the publication of the influential *Land Management Handbook 41* (Hogan et al. [editors] 1998). Under the Fish-Forestry Interaction Program, many researchers participated in projects on Haida Gwaii and this research contributed to the *British Columbia Coastal Fisheries/Forestry Guidelines* (B.C. Ministry of Forests et al. 1988). Among the many studies that were part of this project, the overall emphasis was on slope (Figure 1.4) and channel stability and the effects of harvesting on fish habitat in a geomorphically active environment (Figure 1.5).

As more information about forest hydrology became available in the 1970s, awareness of the effects of large wood in streams and the importance of the structure of stream channels also increased. Scientists learned that the traditional forest hydrology topics of quantity, regime, and quality of water could not be viewed in isolation. Rather, these topics should be considered in conjunction with the stream channel. It became apparent that timber harvesting adjacent to stream channels could affect long-term stability of channels, even when no instream ac-

FIGURE 1.4 *Example of the effects of pre-1980s harvesting and road-building practices in unstable/erodible terrain, coastal British Columbia. (Photo: B.C. Ministry of Forests and Range)*

FIGURE 1.5 *Debris from landslide in Carnation Creek, 1984. (Photo: B.C. Ministry of Forests and Range)*

tivities occur. In other words, it was not enough to simply keep harvesting equipment out of streams.

Another intense topic of discussion was the effect of harvesting on rain-on-snow events on the Coast. It was known that infrequent fall and winter storms triggered most landslides, but it was not known empirically what role harvesting might play. A study on Haida Gwaii outlining the theoretical effects of harvesting on peak flow (Toews and Wilford 1978) led to considerable debate, and the recommendation that harvesting be limited to one-third of a watershed within a 25-year period. This debate resulted in the initiation of a related study on Haida Gwaii,[2] but during the 3 years of field monitoring no rain-on-snow events occurred. Following a series of technical problems, the project was discontinued. Beaudry and Golding (1987) undertook a more successful study on the topic in the Vancouver area.

With the objective of coming together to discuss key issues in forest hydrology, an informal group called the "Friends of Forest Hydrology" was established in 1976. Although it has no constitution, executive, budget, or proceedings, the group has continued to meet, usually annually, for the past 30 years. These gatherings resulted in many lively, informative discussions and field observations that have helped shape the practice of hydrology in British Columbia (Figure 1.6).

The 1980s

The 1980s was a period of considerable research output. Projects initiated in the 1970s were producing results, and science-based watershed management policies were evolving. The *Guidelines for Watershed Management of Crown Lands Used as Community Water Supplies* (B.C. Ministry of Environment 1980) was released early in the decade. Developed by a multi-agency committee, the document identified 279 watersheds (excluding those in Vancouver and Victoria) that supplied water to 21% of the province's population. A primary effect of these guidelines has been to increase the planning and water-protection efforts in community watersheds. The guidelines have been of particular importance in the Southern Interior, where community watersheds make up a significant portion of the provincial forest. The Department of Fisheries and Oceans released a parallel document that dealt with fish habitat called *A Handbook for Fish Habitat Protection on Forest Lands in British Columbia* (Toews and Brownlee 1981). This

2 The results of these studies were not published.

FIGURE 1.6 *Demonstration of salt dilution gauging to measure streamflow in Fishtrap Creek, Friends of Forest Hydrology workshop, September 2005. (Photo: K. Turner)*

handbook summarized the relevant literature, documented planning procedures, and provided guidelines in a format designed for use by fisheries officers and habitat protection specialists.

The earlier investment in Carnation Creek started yielding significant results, which were presented at major workshops in 1981 (Hartman [editor] 1982) and 1987 (Chamberlin [editor] 1988). At Carnation Creek, complex relationships between biological (Figure 1.7) and hydrological parameters such as stream temperature, stream water nutrients, and channel stability (Figure 1.8) came to be better understood. Land managers called for the translation of these results into guidelines. A workshop in 1983 led to the development and implementation of the *British Columbia Coastal Fisheries/Forestry Guidelines;* these were subsequently refined in a second edition (B.C. Ministry of Forests et al. 1988). Both industry and government agreed to the guidelines, which were implemented with an intensive training program.

By the mid-1980s, the British Columbia Forest Service had hired research hydrologists in five of the province's six forest regions and had given them the dual role of undertaking research and providing operational advice. The British Columbia Ministry of Environment and forest companies were also hiring specialized staff with interests in forest hydrology. New programs were set up to broaden the geographic scope of forest hydrology research in British Columbia. In the Nelson and Kamloops forest regions, the focus was on snow hydrology and the effects of harvesting on streamflow in snow-dominated watersheds. The Kamloops Forest Region and Weyerhaeuser Company Limited initiated the Upper Penticton Creek Experiment in 1982 (Winkler et al. 2008b). The Prince George Forest Region specifically focussed on harvesting effects in northern British Columbia, and the Prince Rupert Forest Region looked at channel and slope stability issues, particularly as these related to fish and forestry. Common concerns in British Columbia included mass wasting, surface erosion and sediment from roads, peak and low flows, limits to logging and rate of cut, and fish–forestry interactions (including the role of riparian areas and large organic debris in streams). Hetherington (1987) published a

FIGURE 1.7 *Measuring fork length of Coho salmon fry from Carnation Creek. (Photo: B.C. Ministry of Forests and Range)*

FIGURE 1.8 *Demonstration of low-level aerial photography using helium-filled balloons to quantify changes in stream channels at Carnation Creek, 1989. (Photos: D. Toews)*

useful synthesis of the Canadian literature concerning many of these issues.

For most—but not all—of these technical issues, it has been possible to translate research findings into practice. The "limits to logging" issue has been an exception. Is it necessary to regulate the extent of logging in a watershed? If so, how should this be done? An industry/agency committee met for several years to address this issue on the Coast. A workbook for managers of coastal watersheds (Wilford 1987) illustrated some of the difficulties of combining the various factors in play within a watershed into a single analysis. The trials of this method promoted further discussion regarding the appropriate nature of a cumulative effects model.

Meanwhile, initial outcomes of the Fish-Forestry Interaction Program on Haida Gwaii were presented at a workshop in 1986. Findings helped expand the forest hydrology knowledge base and were used to revise the *British Columbia Coastal Fisheries/Forestry Guidelines* (B.C. Ministry of Forests et al. 1988; Hogan et al. [editors] 1998).

The 1990s

In the 1990s, research initiatives in forest hydrology continued. The Department of Fisheries and Oceans established another fish/forestry/hydrology watershed study in a sub-boreal section of the Stuart-Takla region north of Prince George (MacDonald [editor] 1994). Various attempts were also made over the years to develop or calibrate computer-simulation models that could be used to assess the effects of forest management scenarios on streamflow. Examples include the Hydrological Simulation Program–Fortran model at Carnation Creek, the University of British Columbia's Watershed Model at Upper Penticton Creek, the Water Resource Evaluation of Non-Point Silvicultural Sources model in the Okanagan, and other models such as the Distributed Hydrology Soil Vegetation Model at several locations across British Columbia. The challenge is that a watershed model must be able to link water balance changes on specific areas within the basin to downstream changes in streamflow. Although hydrologic models continue to be used and developed for research purposes, operational applications have been limited (Beckers et al. 2009).

Unfortunately, the *British Columbia Coastal Fisheries/Forestry Guidelines* (B.C. Ministry of Forests et al. 1988) were not always effective in protecting streams from damage (Moore and Bull 2004) (Figures 1.9, 1.10). A series of environmental audits reported that many small streams had been significantly affected by timber-harvesting activities, and this spurred efforts by both government and industry to tighten procedures and undertake directed research on gully assessments and debris removal (Tripp 1998).

In the early 1990s, a crucial event in the development of harvesting guidelines occurred in Clayoquot Sound on the west coast of Vancouver Island. An alliance of environmental groups, First Nations, and private individuals protested the harvesting activities in a specific area on the outer coast. The government responded by selecting a group of independent scientists to prepare a report to guide the protection of key resources in this area. In 1995, the Scientific Panel for Sustainable Forestry Practices in Clayoquot Sound released its recommendations (Scientific Panel for Sustainable Forest Practices in Clayoquot Sound 1995). In this report, the Panel recommended that harvesting should continue but on a very limited scale and under stringent guidelines.

Two very important forest policy initiatives were implemented in the 1990s. First, the provincial government called for the preparation of the *Forest Practices Code of British Columbia Act* and accompanying guidebooks, which came into effect in 1995. Second, in 1994, the government initiated Forest Renewal BC (Province of British Columbia 1994), an organization that launched several substantial programs to rehabilitate watersheds throughout the province. The previous decades of research paved the way for the preparation of the Forest Practices Code guidebooks and for the development of procedures for the Watershed Renewal Program of Forest Renewal BC.

Forest Renewal BC established a significant legacy with two endowed chairs in forest hydrology at the University of British Columbia and two at the University's Okanagan campus. In addition to the formal hydrology training at universities, the various road and stream rehabilitation field trials in all parts of British Columbia have resulted in an enhanced level of field expertise.

Before the implementation of the Forest Practices Code, environmental guidelines from the provincial government had been more discretionary, with considerable latitude permitted in adapting to a given situation. With the Forest Practices Code, however, the government set out to specify as precisely as possible, through legislation, regulations, and guidebooks, how forestry activities were to be carried out and how studies were to be undertaken. For example, the regulations in the Code required

FIGURE 1.9 *Investigating the Donna Creek landslide, 1992. (Photo: D. Hogan)*

FIGURE 1.10 *Logjam in the Similkameen River. (Photo: D. Hogan)*

that a watershed analysis be conducted on important streams, community watersheds, and other watersheds as directed by a Forest Service District Manager. Procedures previously conducted on an ad hoc basis were now to be undertaken exactly as specified in the guidebooks. The Coastal and Interior watershed assessment procedure guidebooks (B.C. Ministry of Forests and B.C. Ministry of Environment 1995a and 1995b) were prepared and analyses conducted on hundreds of watersheds throughout British Columbia (Carver and Teti 1998).

To professionals involved in producing the guidebooks, it was evident that many gaps and shortcomings existed in provincial information related to the new field of applied forest hydrology and geomorphology research. At best, the guidebooks offered a systematic way to evaluate watersheds based on a combination of the British Columbia experience and of similar efforts in the U.S. Pacific Northwest.

A fundamental problem was the necessity to base the guidelines on indicators, such as road length, percentage of clearcut area, percentage of streamside area harvested, and amount of harvesting on unstable terrain. As proxies rather than as direct measures of effects, these indicators are useful as screening tools to ascertain whether further investigation is warranted; however, these indicators are not sufficiently definitive to allow predictions of what will happen in every situation.

With the need to verify the information presented in the guidebooks, a number of research projects were put into place in the late 1990s. In Penticton, a workshop was held in 2000 to present these research findings and to facilitate dialogue among the professionals who had conducted assessments in the Interior as well as among the people applying the results (Toews and Chatwin [editors] 2001).

In 1999, the government introduced a new analysis procedure that gave hydrologists more latitude to use professional judgement in assessing potential effects of harvesting on watersheds (B.C. Ministry of Forests 2001). For example, although road length was previously used as an indicator for sediment in streams, the new procedures recommended direct observation and categorization of sediment sources (Figure 1.11). These new procedures were more accurate for identifying the risk to water quality. Also in 1999, the provincial government passed the Private

Land Forest Practices Regulation (Province of British Columbia 2000), which sets standards related to water quality and streamside management on privately owned forest land.

Today

In 2001, the provincial government decided to replace the Forest Practices Code with the new results-based approach of the *Forest and Range Practices Act*. The new legislation specifies only broad overall results for the forest industry and allows forest licensees considerable latitude in selecting methods to achieve those results (Figure 1.12). Government is now less involved in planning and approving forest development, but places more emphasis on enforcement.

From a watershed management perspective, the main challenge is how to evaluate watershed management "results." The research question is: Which measurements are reliable for determining watershed condition on an operational scale? Basic to implementing this new strategy is the need for a good understanding of watershed science within the context of our geographically diverse province. To address these questions, the Forest and Range Evaluation Program (FREP) has been developing monitoring protocols and indicators to provide the tools necessary to evaluate the effects of forest management activities on a range of resource values, including water and riparian/fish.

Most recently, changes in British Columbia's climate, and the consequences of those changes, have been a focus for researchers across the province. The infestation by the mountain pine beetle (MPB) during the late 1990s and into the 2000s introduced a new emphasis on understanding the potential hydrological effects of both MPB-induced pine mortality over large areas, and subsequent large-scale salvage harvesting (Figure 1.13). Several new research projects were initiated during this period, and new funding allowed the reinvigoration of pre-existing and long-term hydrology experiments. Field research has focussed on the effects of MPB canopy defoliation on snow accumulation and melt. In addition, a number of modelling projects were carried out to examine the potential effects on peak flows from MPB-affected watersheds. Most notably, a report completed by the Forest Practices Board (2007), which employed model simulations of the

FIGURE 1.11 *Assessing sediment source from channel erosion at Rembler Creek, October 2003. (Photo: P. Tschaplinski)*

FIGURE 1.12 *Example of current riparian management practices. (Photo: B.C. Ministry of Forests and Range)*

FIGURE 1.13 *Bonaparte Plateau north of Kamloops showing evidence of mountain pine beetle infestation on the landscape. (Photo: R.D. Winkler)*

Baker Creek watershed near Quesnel, raised awareness about the potential changes to flood frequencies caused by the MPB infestation and subsequent salvage harvesting. Winkler et al. (2008a) and Redding et al. (2008) have summarized the results of research on the effects of the MPB on hydrology. Similarly, many researchers are currently active in the climate change research field, analyzing historical data, developing modelling approaches, and extending information on how watershed processes in British Columbia will likely change in the future (see Chapter 19, "Climate Change Effects on Watershed Processes in British Columbia").

SUMMARY

Since the mid-1960s, a great deal has been accomplished—through research, practice, education, and policy development—in forest hydrology and geomorphology and in understanding the effects of forest resource management in British Columbia. Nevertheless, much remains to be learned. This compendium summarizes the extent of our knowledge to 2010 and also identifies the knowledge gaps that currently exist in our understanding. In so doing, the authors hope to stimulate further gains in the understanding of forest hydrology, geomorphology, and stream ecosystems and the practical application of this knowledge to facilitate effective watershed management in British Columbia.

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CHAPTER 2

Physiography of British Columbia

MICHAEL CHURCH AND JUNE M. RYDER

INTRODUCTION

British Columbia lies astride the Western Cordillera of North America, a broad and complex system of mountains and plateaus that has developed over nearly a billion years on the leading edge of the North American continental plate. In the northeast, the province extends beyond the mountains into the Alberta Plateau. Consequently, the contemporary landscape exhibits a wide variety of landforms and materials, ranging from high mountain peaks, through broad, rolling plateaus, to alluvial valleys. This chapter presents an overview of the contemporary landscape of British Columbia, highlighting rock and soil materials and their properties, and explains how the landscape has developed. Rather than elaborately describe the physiography, we instead consider the major elements that have created the topography and the principal attributes of the physical landscape—landforms, surface materials, and soils. Figure 2.1 shows the major physiographic divisions of the province, reflecting the present topography. The divisions are based on topographic boundaries.

The present topography is the consequence of the interaction of three principal factors: tectonic history, rock types, and climate. Tectonic history is the sequence of earth movements—mobilized by forces within the Earth—that has produced the large-scale arrangement and elevation of major topographic features. Hence, the major physiographic units of the landscape (Figure 2.1) are largely tectonically determined. Tectonics also accounts for the distribution of the rock types exposed at the surface, and these have influenced the character of the soil materials that mantle the surface (surficial materials) today. These, in turn, have a major influence on the terrestrial components of the hydrologic cycle.

Climate controls the hydrologic cycle and determines the processes of rock weathering, erosion, and sediment transfer at the surface of the Earth. These geomorphic processes create the landforms and soils within the landscape.

The temporal scale for tectonic processes is in the order of 1 million years and the spatial scale falls between 100 and 10 000 km. Geomorphic processes, insofar as they operate cumulatively to produce significant landforms, have temporal scales from 10 to 100 000 years, and work on spatial scales of between 0.10 and 1000 km. The upper end of the range of geomorphic processes intersects tectonic scales, and the lower end corresponds with the time and space scales of everyday human interest. At these scales, the processes act principally on the materials that mantle the Earth's surface, so the relation between these processes and surficial materials is of central interest. Important tectonic (e.g., earthquakes, volcanic eruptions) and geomorphic (e.g., landslides, major floods) events can occur almost instantaneously and have significant impacts on the landscape, on society, and on land management.

Although the British Columbia landscape has developed over a very long time, this chapter focusses

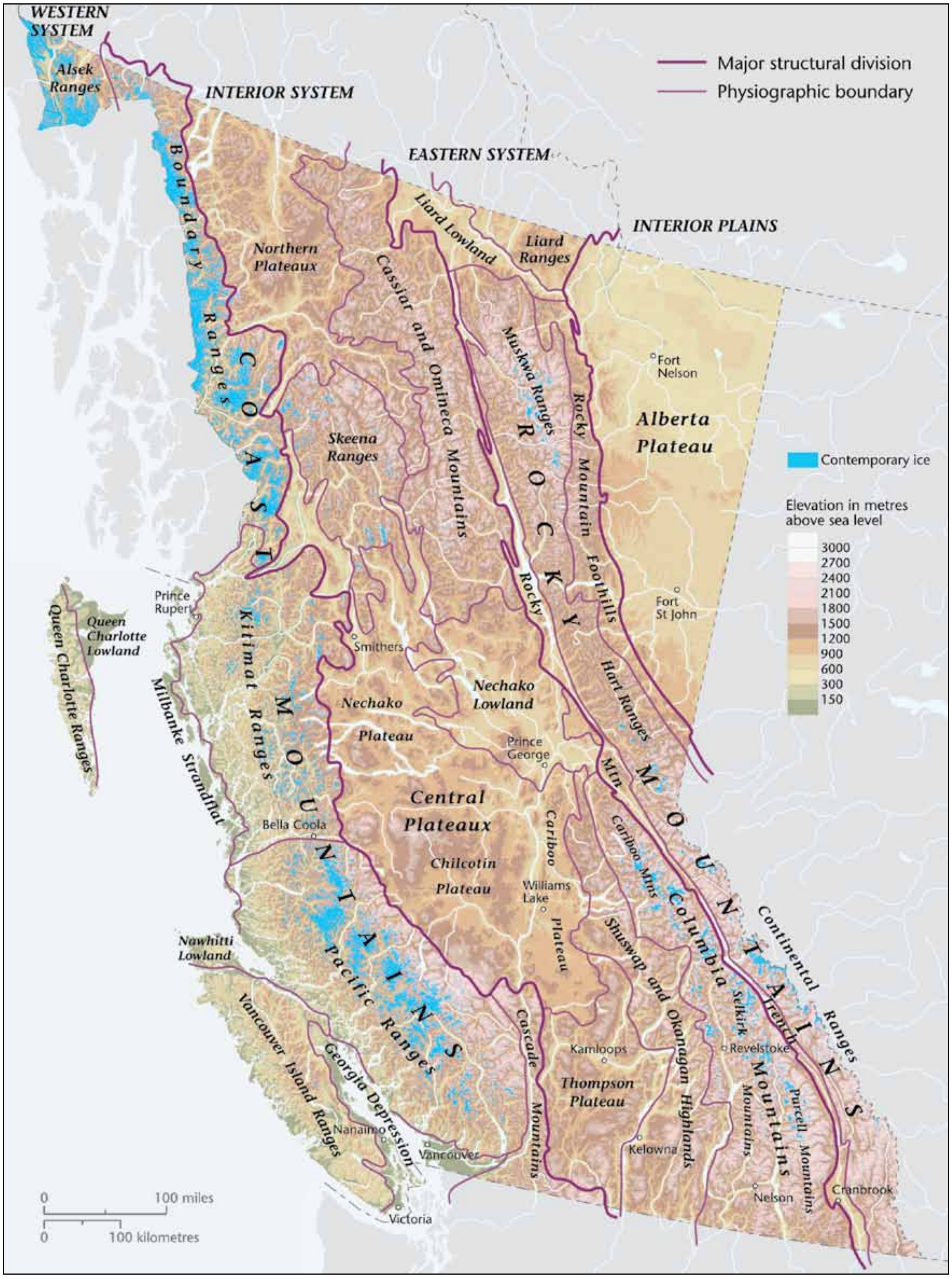

FIGURE 2.1 *Physiographic regions of British Columbia (simplified after Mathews 1986: Mathews' lowest-order units have been omitted; relief data are from digital files of the United States Geological Survey). These physiographic divisions are also superimposed on maps of geology (Figure 2.4), glacial features (Figure 2.11), glacial deposits (Figure 2.15), and soils (Figure 2.22) so that the reader can determine the principal attributes of the physical landscape in each topographic division. (Minister of Public Works and Government Services Canada)*

on the most recent periods of Earth history within which tectonic and geomorphic processes have created the contemporary landscape. Within this time frame, the dominant elements have been mountain building and episodic glaciation. The contemporary landscape is primarily the legacy of these two processes.

TECTONIC SETTING AND GEOLOGY

Tectonic History

Following extensive mountain building in late Mesozoic and earliest Tertiary times (approximately 70 to 35 million years ago), the bulk of the Tertiary Period was an interval of erosion of the British Columbia landscape. Topography west of the Rocky Mountains was reduced to gentle slopes and broad valleys with up to 500 m of local relief, forming an ancient eroded landscape, relicts of which remain as high surfaces throughout the Cordillera. The contemporary landscape began to develop during the most recent phase of uplift and river incision commencing about 10 million years ago (Mathews 1991).

The tectonic regime is controlled by the motions of the Pacific, Juan de Fuca/Explorer, and North American plates (Figure 2.2), all three of which come together at a "triple junction" off the north coast of Vancouver Island at the minor Winona Block. North of the junction, the Pacific Plate moves northwesterly relative to the continental North American plate along the Queen Charlotte Fault at about 55 mm/yr. To the south, the Juan de Fuca and Explorer plates are moving easterly at about 40 mm/yr; the subduction of these plates under the continental margin takes up part of this motion (Riddihough and Hyndman 1991). These plate movements have promoted tectonic, seismic, and volcanic activity on the continental margin. The main orientations of the major valleys in the Coast Mountains, including the fjords, define the principal conjugate failure planes for the stresses established by onshore movement of the Juan de Fuca/Explorer plates, demonstrating their influence over the topography of coastal British Columbia.

The Pacific Plate has moved toward the North America Plate for a long time, causing a sequence of "exotic terranes" (units of Earth's crust that originated elsewhere) to be carried to the margin of North America. These terranes today make up much of the basement rock west of the Omineca and Columbia mountains, an early continental margin terrane that developed on the ancient rim of the North American craton. The largest more recent body of locally formed rock is the sequence of Mesozoic intrusions that makes up the Coast Mountains. These tectonic features today define the principal physiographic divisions of the province (Figure 2.1).

The main locus of tectonic activity today lies offshore of British Columbia, along the Queen Charlotte Fault and the Cascadia subduction zone, but occurs onshore in the U.S. Pacific Northwest and in southern Alaska. The most impressive physiographic evidence of this activity is the line of continental stratovolcanoes that extends from southern British Columbia along the Cascade Mountains to northern California, and in the Wrangell Arc, extending from the St. Elias Mountains into the Aleutian Islands. In between, deeper-seated volcanism, yielding basaltic volcanic centres, occurs in the Anahim and Stikine volcanic belts of British Columbia.

Within the last 10 million years, the Coast Mountains have risen by up to 4 km (Mathews 1991), with average rates up to 0.5 mm/yr. In the latter half of this interval, the main activity has shifted from the northern Coast Mountains toward the south. More than half of the uplift is caused by isostatic compensation for erosion, which has resulted in the incision of most major rivers well below the general level of the landscape. Uplift is also attributable to primary tectonic effects and to thermal expansion of the crust and upper mantle associated with subduction. In the interior plateaus of British Columbia, extensive basalt flows were extruded in the Tertiary Period. An older series of lavas widespread in the Thompson Plateau—the Kamloops volcanics—dates from early Tertiary time. A younger series dating from the late Tertiary—the Plateau basalts—buried the earlier Tertiary erosion surface in the more northerly and westerly plateaus (Figure 2.3).

Today, the base of the lavas defines the ancient land surface over extensive areas. Both the erosion surface and overlying basalts are now warped up toward the Coast Mountains, indicating lesser uplift east of the mountains. In fact, the contemporary elevation of the plateaus is largely the isostatic

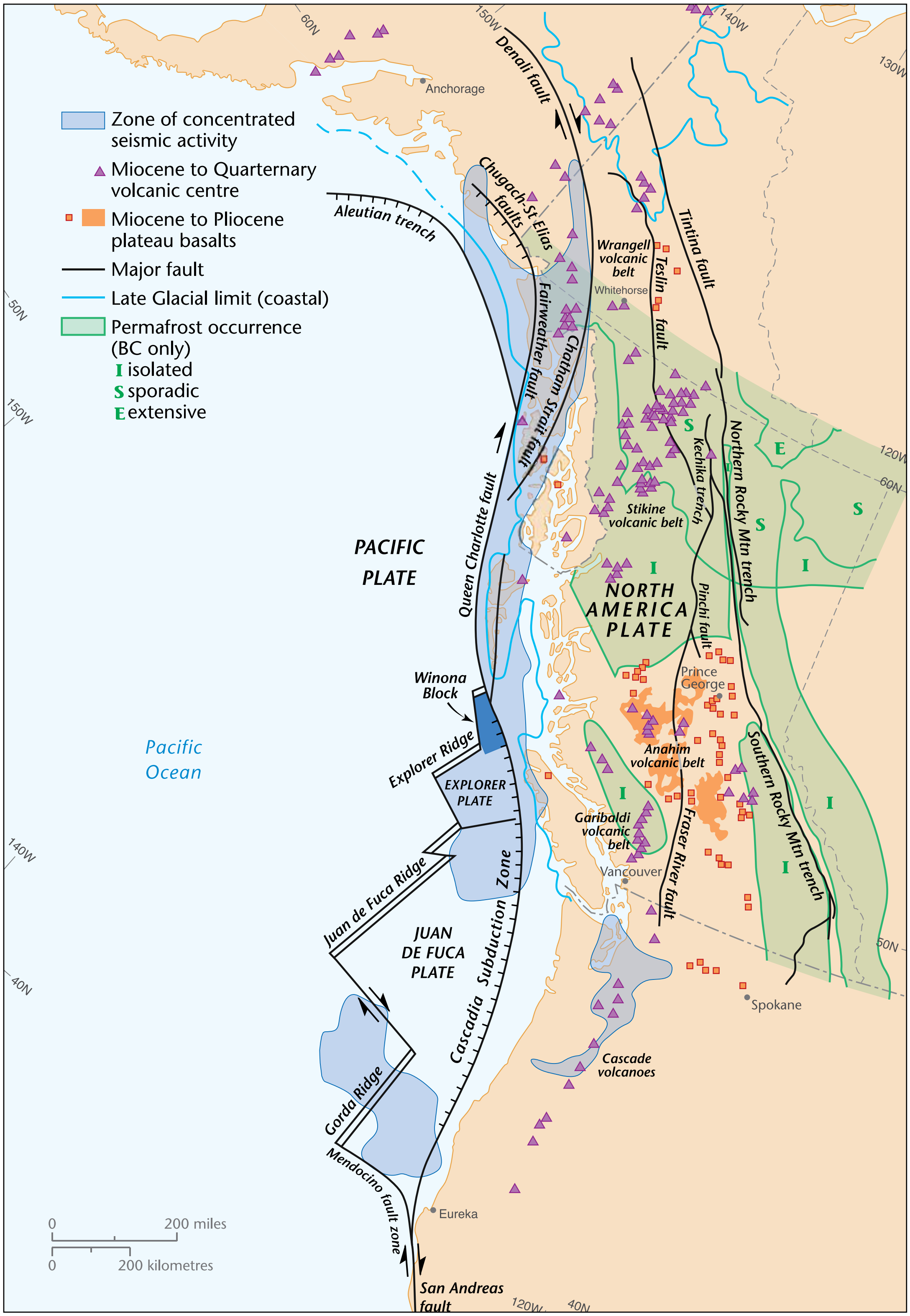

FIGURE 2.2 *Major geophysical features having physiographic expression in British Columbia (compiled from Gabrielse and Yorath 1991: plate tectonic features; Natural Resources Canada 1995: permafrost; Clague 1989: other features). (Minister of Public Works and Government Services Canada)*

FIGURE 2.3 *Tahltan Highland: view south along the east side of the Spectrum Range (composed of flat-lying Tertiary volcanic rocks), with the Little Iskut Valley to the left. A major landslide has occurred off the rock scarp in the middle ground (indicated by lighter vegetation). (Photo: J.M. Ryder)*

response to continued erosion rather than the result of tectonics. Contemporary uplift averages about 2 mm/yr along the outer coast of British Columbia and 3–5 mm/yr in the Saint Elias–Alaskan coast region. Part of this deformation is elastic strain in response to subduction that fault rupture will eventually relieve.

More details on the tectonic history of the Canadian Cordillera and its influence on the contemporary topography can be found in Gabrielse et al. (1991). An important feature of the tectonic history is that the development of the forests and biota of British Columbia occurred within the same 10 million years as the contemporary mountains and were undoubtedly conditioned by it.

Geology

At a more local scale, rock structure and lithologies determine major landforms, such as the orientation of mountain ranges and major valleys. Structural features (e.g., faults and folds, joints, and bedding) determine the details of the landscape, such as the asymmetry of individual ridges and the location of major gullies. The mineral composition of the rocks determines their susceptibility to weathering and the character of the weathering products.

Because of the complex tectonic history, even a highly generalized geological map of British Columbia presents a complex pattern (Figure 2.4). Although geological mapping usually emphasizes the age of the rocks and their stratigraphic sequence, this information does not indicate rock properties clearly, such as strength and weathering susceptibility, which are of considerable relevance to land managers. Unfortunately, these performance indicators rarely are systematically mapped and are, in any case, difficult to show on small-scale and hence generalized maps. We have nonetheless attempted to give some indication of likely rock performance in Figure 2.4 and in the following notes. The most general classification divides rock types into igneous, sedimentary, and metamorphic types, with igneous rocks subdivided into intrusive and extrusive types. To some extent, these classes correspond to rock character and rock performance.

Igneous rocks

Intrusive igneous rocks (plutonic rocks) are formed in the Earth's crust by cooling of intruded magma.

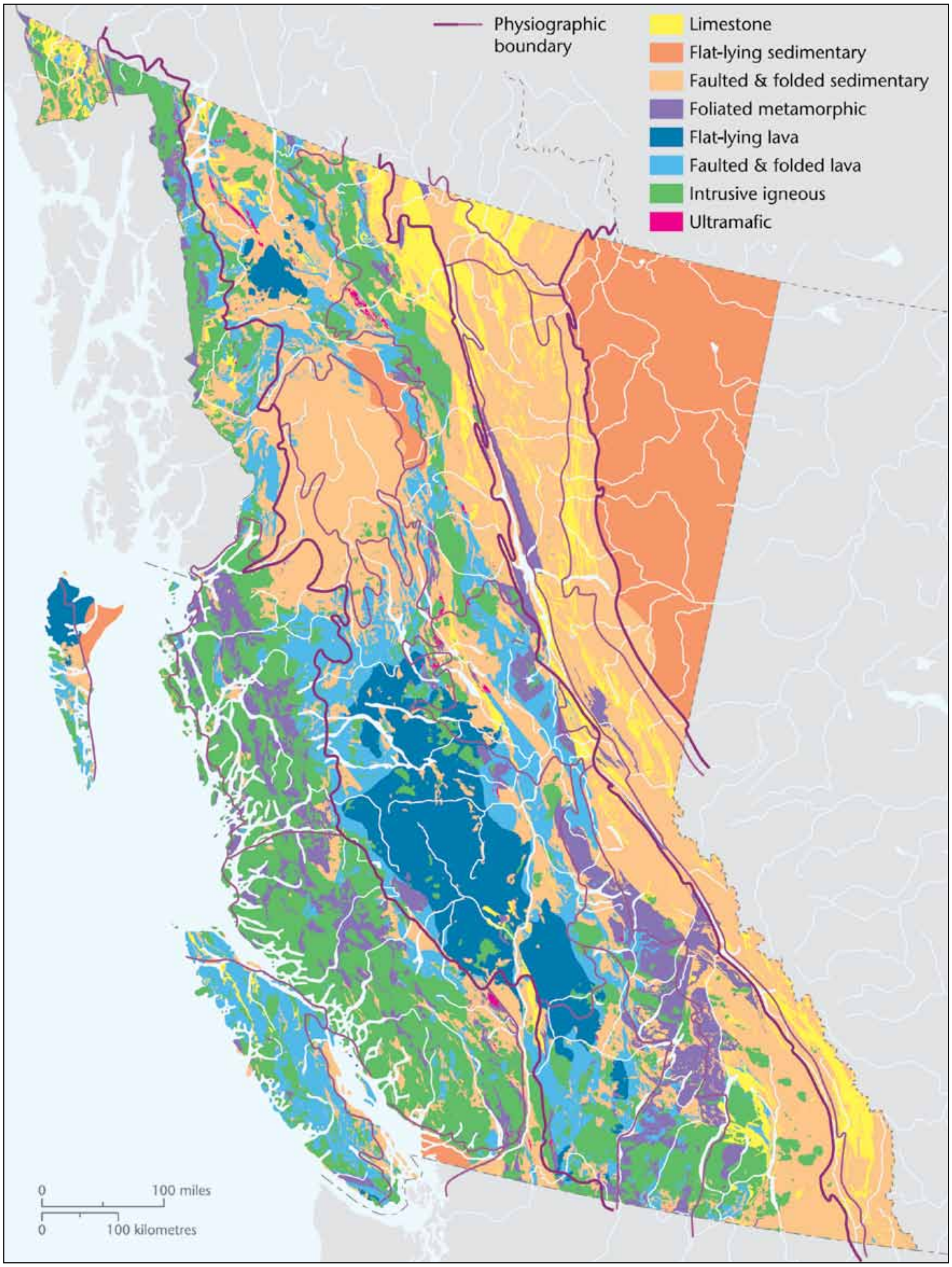

FIGURE 2.4 *Generalized geology of British Columbia: lithostratigraphical units (modified from Massey et al., 2005). (B.C. Ministry of Energy, Mines and Petroleum Resources)*

Typically, these rocks are coarsely crystalline and range from simple stocks occupying tens to hundreds of square metres to the huge plutonic complex of the Coast Mountains. Rock composition in British Columbia varies mostly from syenite through various granitic rocks to diorite (Woodsworth et al. 1991). Plutonic rocks are relatively resistant to weathering owing to a composition of predominantly durable minerals (quartz, feldspars, hornblende) in the form of large, interlocking crystals, and joint planes that are widely spaced. Hence, plutonic rocks support steep slopes and rugged topography (Figure 2.5).

In the landscape, lines of weakness (major joints and faults) are followed by depressions ranging in size from minor gullies to major valleys. These rocks are subject to mechanical breakdown, which typically involves the loosening and release of large, joint-bounded blocks giving rise to coarse, blocky colluvial slopes and large boulders in glacial drift.

Plutonic rocks are also subject to granular disintegration caused by the combined effects of mechanical and chemical weathering, giving rise to a sandy, gritty crystalline residue (termed "grus"), sandy till, and sandy, gravelly alluvium.

Extrusive igneous rocks (volcanics) are of two types: (1) magmas that solidified in the surface environment (i.e., lava flows), and (2) solidified accumulations of volcanic ash and cinders (pyroclastic rocks). Basaltic magma is very fluid so that flows spread readily to form large, but gently sloping, shield volcanoes or extensive lava sheets. The dissected remnants of the former are visible, for example, in and around the Rainbow Range of the Chilcotin Plateau (Figure 2.6).

The Tertiary Plateau basalts resulted from fissure eruptions of basaltic lavas that spread for great distances to form extensive lava sheets; subsequent uplift and dissection transformed these into plateaus bounded by stepped escarpments. Andesite lavas are less extensive in British Columbia than basalts. This magma is more viscous and commonly associated with pyroclastic deposits that, together with andesite lava, tend to form classic volcanic cones. All lavas are typically fine textured and joints are closely spaced because of crystallization and contraction

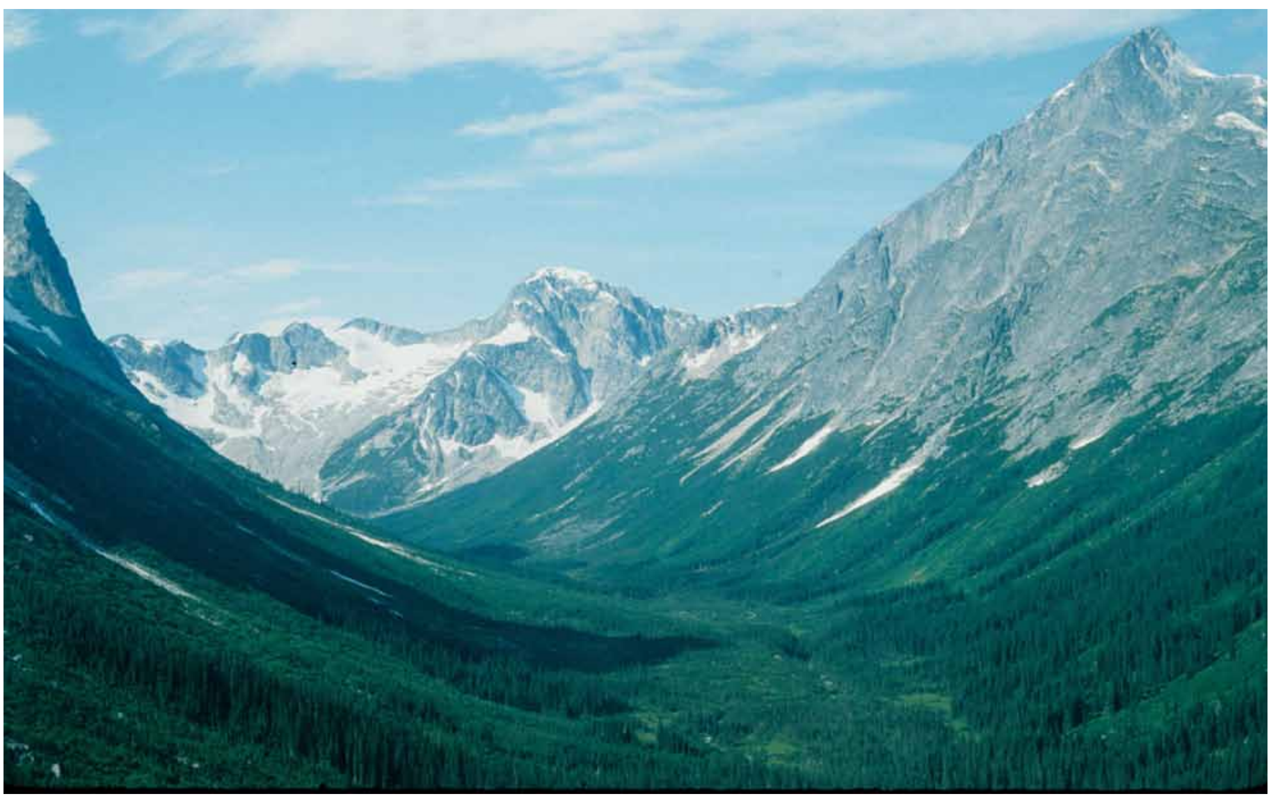

FIGURE 2.5 *Boundary Ranges of the Coast Mountains. View up the south branch of Vekops Creek to the Sawback Range, a spectacular U-valley. (Photo: J.M. Ryder)*

FIGURE 2.6 *Plateau edge along Taseko River (Chilcotin Plateau), showing the Tertiary Plateau basalts forming a lava flow escarpment. Block slump failures and extensive talus are present. (Photo: J.M. Ryder)*

during rapid cooling. Basalts typically display columnar jointing; jointing in andesite is often chaotic. Pyroclastic rocks formed by accumulations of ash, cinders, and shattered volcanic rocks may be welded and strong or scarcely consolidated and weak. Quaternary volcanic piles in the Garibaldi region of southwestern British Columbia include weakly indurated rocks that present major landslide hazards.

Mechanical weathering, especially by freeze–thaw processes, is important in volcanic rocks because of the joints and other openings that give ingress to water. The minerals in basalt—plagioclase, olivine, and pyroxene—are also susceptible to chemical weathering, so that this rock yields fine-textured weathering products as well as blocky or rubbly debris. In general, unweathered volcanic rocks can maintain steep slopes and substantial relief. Many volcanic sequences, however, include layers of severely weathered material. These may include soils that developed on one lava flow before it was buried during the next eruption or layers of volcanic ash (tuff) and other pyroclastic material that weather relatively rapidly to a clayey residue. Some of these clayey weathered layers are extremely weak and deformable, and give rise to slope instability. Major landslides related to these conditions are widespread on steep slopes around the edges of interior plateaus that are capped by volcanic rocks (e.g., the Thompson, Fraser, and Stikine). Relatively young volcanic rocks, such as the Pleistocene Garibaldi volcanics, are also subject to instability ranging from abundant rock fall through debris flows to rock avalanches.

Sedimentary rocks

Sedimentary rocks are formed by deposition of clastic fragments (particles of disintegrated pre-existing rocks) or by chemical precipitation mostly in the oceans. Initially deposited as horizontal strata, many sedimentary rocks have subsequently been compressed, folded, and fractured as a result of tectonic activity. This results in the development of joints and faults that, together with bedding planes, function as planes of weakness.

Erosion of horizontally bedded rocks gives rise to plateaus, escarpments, and mesas (Figure 2.7), the more-resistant beds forming the scarps. Such topography is well represented in the Alberta Plateau physiographic region of northeastern British Columbia where thick units of resistant sandstone alternate in sequence with weak, easily eroded shales.

Dipping sedimentary rocks form cuestas (asymmetric ridges) and impart a characteristic grain to topography, as in the Rocky Mountain foothills and the southern Gulf Islands. Strongly folded and faulted rocks form complex and rugged topography, although a distinct parallel alignment of ridges and ranges is usually still evident, as in the Main Ranges of the Rocky Mountains.

The induration (hardness and internal cohesion) of clastic sedimentary rocks depends on their history of loading (compaction) and degree of cementation of the grains. Subsequently, their weathering characteristics depend on the degree of induration, the chemical composition of the cement, and the spacing of joints and bedding planes that admit water into the rock mass.

Weakly indurated and thinly bedded rocks, such as most mudstones and shales, disintegrate readily, are easily eroded, and tend to form relatively low-

FIGURE 2.7 *The Alberta Plateau, view near Pink Mountain settlement (Beatton River headwaters). The flat-lying Cretaceous sediments erode to form mesas. The cap rocks are sandstones and conglomerates. (Photo: J.M. Ryder)*

lying areas. Well-indurated (cemented) rocks with widely spaced joints and bedding planes, such as many sandstones and conglomerates, resist erosion and tend to underlie topographic highs. Physical weathering releases rock fragments of a size determined by the spacing of planes of weakness in the parent rock. Thinly bedded shales thus produce small, platy clasts that readily disintegrate when transported. Massive sandstones and conglomerates with widely spaced planes of weakness produce blocks that may survive glacial transport. Chemical dissolution of cementing agents is an effective weathering mechanism that essentially reconstitutes the original sediments. Hence, residual soils and glacial sediments tend to be silty or clay-rich when derived from shales, or sandy when derived from sandstones and conglomerates.

Another sedimentary rock, limestone, is a chemical precipitate consisting of calcium carbonate. Although relatively resistant to erosion in our cool-temperate environment, it is subject to solution by acidic waters (the average pH of precipitation over British Columbia is about 5.6, and organic acids may make runoff waters more acidic). The result may include the formation of cliffed gorges on the surface and underground passages and caverns where groundwater moves along joints and bedding planes. Collapsed cavern roofs and swallow holes create steep-sided, closed depressions. Such features, collectively known as "karst terrain," occur in British Columbia on central and northern Vancouver Island and along the Main Ranges of the Rocky Mountains. Karst hydrology and geomorphology are discussed in detail in Chapter 11 ("Karst Geomorphology, Hydrology, and Management").

Metamorphic rocks

Metamorphic rocks are formed when pre-existing rock bodies are heated and compressed as a result of deep burial or the intrusion of igneous rocks, causing modification of their characteristics. The degree of metamorphism has a strong effect on the mineralogical composition and internal structure of the resulting rocks, and hence the rock's resistance to erosion. Many metamorphic rocks are foliated (minerals are segregated into bands in the rock [e.g., gneiss and schist]), while others develop a distinctive cleavage (preferred orientation of breakage planes unrelated to bedding). Subsequently, differential weathering may produce linear topography (e.g., the Omineca Mountains). Most gneisses are relatively resistant to erosion; schists may be resistant or weather relatively readily depending on their mineralogy. Some metamorphosed sedimentary rocks (e.g., marble, from limestone; quartzite, from sandstone) may be extremely resistant and so form the backbone of significant mountain ridges. All pre-Miocene rocks in the Cordillera have been metamorphosed; high-grade metamorphism, reflecting deep burial and heating, occurred in the Coast and Omineca belts, and lower-grade metamorphism, indicating burial, occurred elsewhere.

The geology of the province is analyzed in detail in Gabrielse and Yorath (editors, 1991).

GLACIATION

Glacial History

While tectonics has established the architecture of the province, the detail has been wrought by Pleistocene glaciation. Glaciers formed in southern Alaska more than 9 million years ago (Denton and Armstrong 1969)—that is, early in the late Tertiary mountain-building episode. In British Columbia, evidence exists for Cordilleran glaciation in the late Tertiary period (3 million years ago) in the form of tills interbedded with lava flows in the northern and central Interior (see Clague [compiler] 1989). Widespread ice cover may have developed only within the last million years (e.g., till underlies 1.2 million years old lavas on the Cariboo Plateau near Dog Creek, well outside the mountains). Glaciation has strongly sculpted the topography, particularly in the mountains (i.e., the Rockies, Cassiar-Columbia, Coast, and Insular mountains) (Figures 2.5 and 2.8).

The Pacific coast is highly dissected and indented by fjords extending up to 150 km inland. Local relief near the summit of the Coast Mountains is as much as 2500 m, with another 750 m of submarine relief. The fjords and terrestrial valleys represent the structurally controlled preglacial drainage lines, but today they exhibit remarkably steep, glacially eroded slopes (Figure 2.9). It has recently been demonstrated that, during the Pleistocene Epoch in the southern Coast Mountains, glacial erosion of the major valleys has been more than six times as effective as subaerial erosion (Shuster et al. 2005).

The characteristic pattern of Cordilleran glacia-

FIGURE 2.8 *Seven Sisters Mountain, Bulkley Range of the Hazelton Ranges on the south side of the Skeena Valley. Composed of Mesozoic sediments and exhibiting strong glacial erosional topography. (Photo: J.M. Ryder)*

tion (Clague [compiler] 1989) has been for ice first to develop in cirques high in the mountains, with an ice distribution not unlike that of today (Figures 2.10, 2.11). The onset of a glaciation is characteristically rather slow, so that alpine glaciation with its characteristic cirque and valley glaciers and high mountain icefields, might persist for thousands of years. Eventually, ice expanded beyond the moun-

FIGURE 2.9 *A deep, glacial trough cut through the Coast Mountains, Pacific Ranges, near Bute Inlet: valley of Orford River, view southwest. (Photo: J.M. Ryder)*

FIGURE 2.10 *Bishop Glacier, central Pacific Ranges of the Coast Mountains. Bishop Glacier drains the north side of the Lillooet Icefield. The rocks are heavily glaciated Mesozoic granitic intrusions. (Photo: J.M. Ryder)*

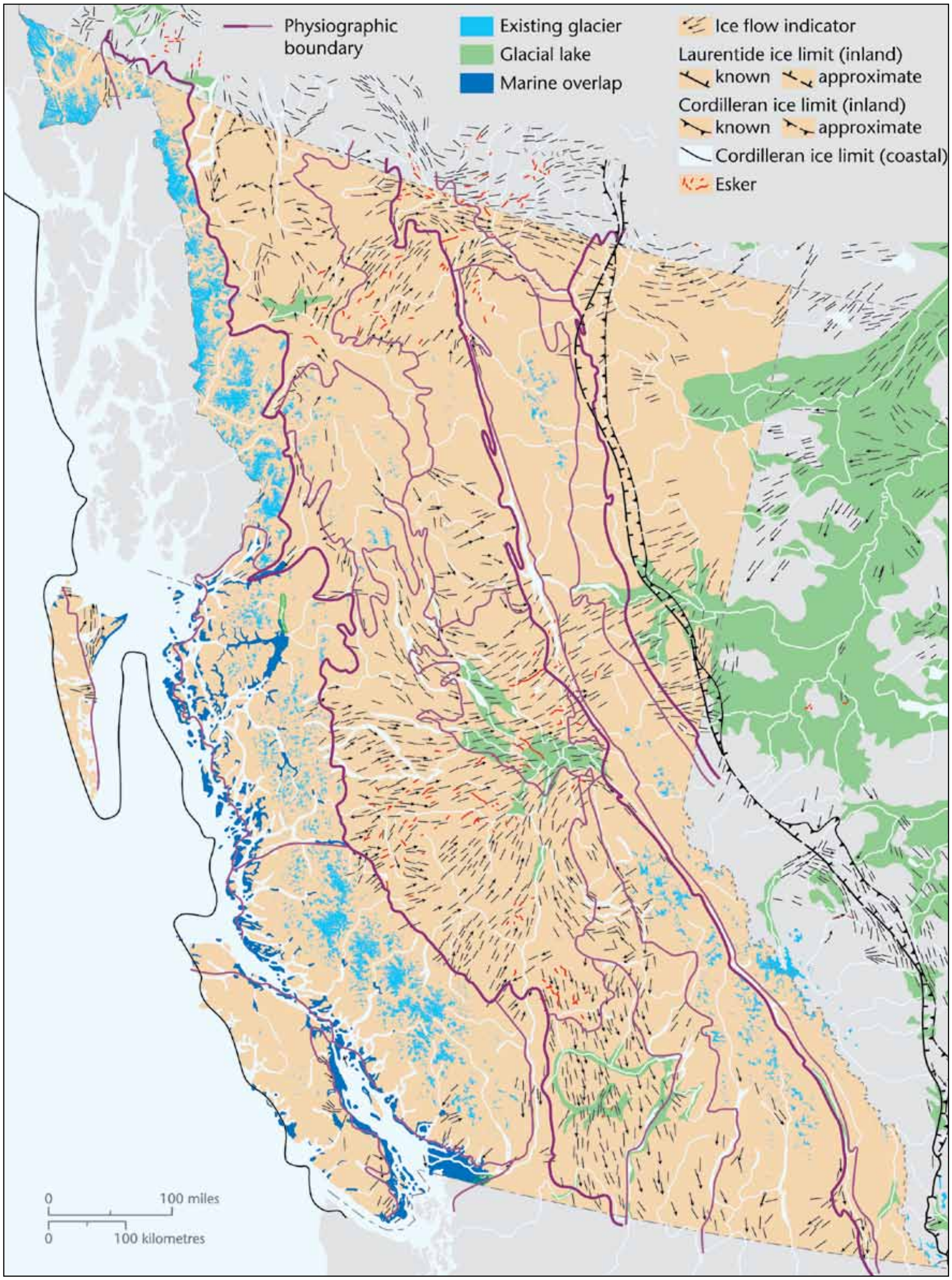

FIGURE 2.11 *Principal glacial features in British Columbia (modified from Geological Survey of Canada Map 1253A 1968; area of marine overlap from Clague [compiler] 1989, Figure 1.5). (Minister of Public Works and Government Services Canada)*

tains and coalesced to form an ice cover over the interior plateaus and plains, and extended offshore onto the continental shelf.

Global climate history and dated evidence for glacial events in the Cordillera more than 1 million years ago suggest that as many as 20 or more significant glaciations may have occurred in the mountains. Direct evidence of older glaciations is extremely rare, however, because each succeeding glacial advance destroys the evidence of its predecessor, and the more recent glaciations have been the most extensive. Until evidence from deep-sea marine sediments became available, the conventional wisdom was that only four major glacial episodes had occurred during the Pleistocene Epoch. Fortunately, abundant evidence of the nature of the last glaciation (Figure 2.11) allows us to explain the character of the contemporary landscape, which is our principal concern.

The latest Pleistocene glacial episode, known as Fraser Glaciation in British Columbia, began in the mountains more than 30 000 years ago. Because the ice expanded very slowly at first, it was only after about 25 000 years BP (before present)[1] that glaciers emerged from the mountains. After that, ice expanded rapidly, coalescing over the plateaus and lowlands, to form the Cordilleran Ice Sheet. Fraser Glaciation ice reached its maximum extent about 15 000 years BP. The ensuing glacier melt was rapid, so that parts of the Fraser Lowland were ice-free before 13 000 years BP. By 10 000 years BP, there was no more ice in the mountains than there is today. Minor fluctuations are superimposed on this overall history, so that seemingly local advance and retreat phases have been recorded in many places within the Cordillera.

Because of the difficulty in establishing absolute correlations between deposits in different places, field investigators have long observed a convention to give various chronostratigraphic units local names. A correlation chart of these names for several regions of British Columbia is shown in Figure 2.12. Reports of local glacial history and Pleistocene deposits will use the local names.

The cumulative evidence of Pleistocene glacial erosion is widespread in British Columbia. In the highest mountains, where peaks projected above the general level of the ice sheet, alpine glacial scenery is preserved in the form of horns, arêtes, and cirques (Figure 2.13). Where mountainous topography and high plateaus were overridden, summits and ridges are rounded (Figure 2.14), although major erosional landforms, chiefly cirques and glacial troughs, are well defined .

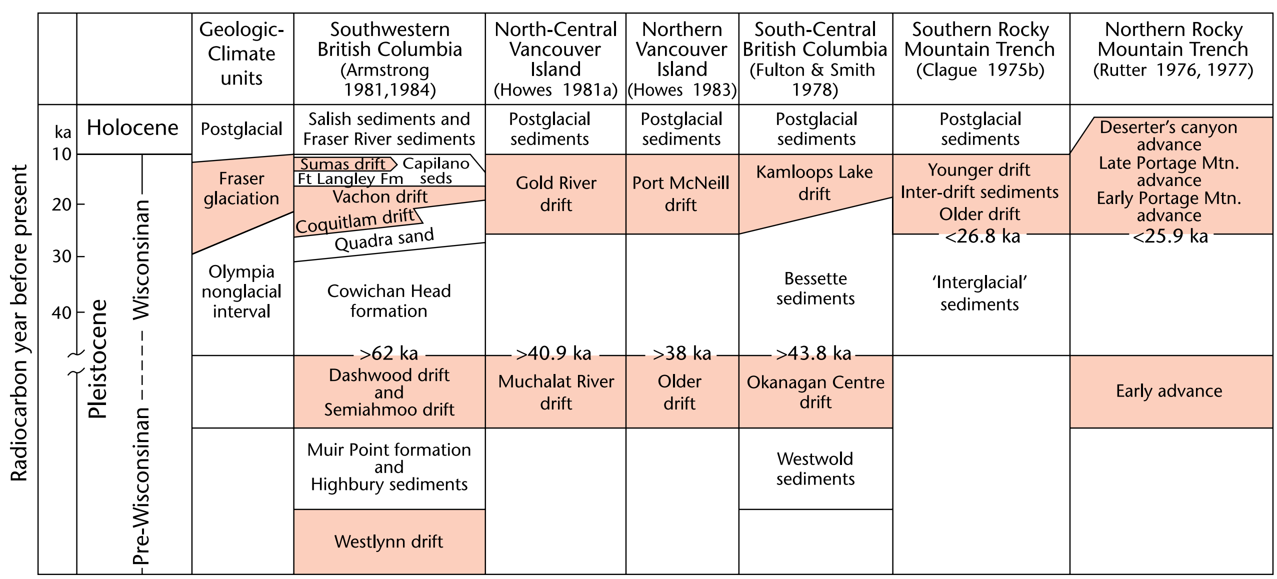

FIGURE 2.12 *Correlation chart of Fraser Glaciation history in British Columbia. (Adapted from Clague [compiler] 1989, Figure 1.17)*

1 Dates in this paragraph refer to ^{14}C years; see note under section on "Holocene History."

FIGURE 2.13 *View north from Idaho Peak in the Selkirk Mountains above New Denver. The long slopes are evident below alpine glacial topography, and there is also a steep inner valley. The mountains here are Triassic volcanic and sedimentary rocks (Idaho Mountain is located on Middle Jurassic granitic intrusives). (Photo: J.M. Ryder)*

FIGURE 2.14 *Glacially overridden ridge crests in the southern Cascade Mountains. These mountains were inundated by Fraser Glaciation ice, creating the smooth topography. Many of the valleys, however, remain V-shaped because ice flow crossed them rather than being directed along them, so erosion of the valley sides was limited. (Photo: J.M. Ryder)*

Valley erosion was greatest where the valleys were aligned parallel to the ice flow, such as along the valleys of the major interior lakes (e.g., Chilko, Quesnel) at the edge of the mountains. In contrast, valleys oriented across the direction of former ice flow commonly were little modified from their preglacial form. Some mountain valleys have well-developed troughs along their upper reaches that were occupied by valley glaciers, but become V-shaped farther downstream where ice sheet flow was transverse to the valley's trend (Figure 2.14).

Styles of deglaciation differed dramatically between the major mountains and the interior plateaus and plains—consequently, so do the resulting glacial depositional landforms. Within the mountains, in general, ice remained active (i.e., continued to flow) to the end of the glaciation, so that proglacial outwash features were constructed in the valleys. Recessional moraines are rare because ice retreat was rapid and more or less continuous. On the plateaus and plains, ice stagnated during the later stages of glaciation as the ice supply from the mountains declined and the emerging uplands—exposed as the level of the ice surface declined—cut off the ice supply.

As melting lowered the surface of stagnant ice, glacial debris melted onto the ice surface, eventually forming widespread ablation moraines. Abundant eskers, kames, and kame terraces were formed where sand and gravel were deposited by meltwater under and against the ice, and kettles formed where buried ice melted. Major valleys were commonly blocked by receding or stagnant ice, impounding major but temporary glacial lakes (see, for example, Fulton 1975).

A singular consequence of Pleistocene glaciation was that large volumes of water were transferred from the ocean basins to the land surface (in the form of ice), so that sea level fell significantly (by 140 m at the climax of Fraser Glaciation). At the same time, the weight of the ice depressed the land surface isostatically (by more than 250 m in the Cordilleran region). Hence, the relative levels of sea and land were changed. At the close of glaciation, sea level rose in accordance with ice melt, whereas the physical properties of the Earth's mantle govern isostatic recovery. The latter occurred relatively slowly so that, as the ice melted, the land was still depressed. Coastal areas up to about 200 m above sea level were inundated, and marine sediments accumulated on what is now dry land. Isostatic rebound continued for several millennia (the time varying with location) as relative sea level gradually came to its present level.

British Columbia glacial history is summarized in Clague (compiler, 1989). Fulton et al. (2004) summarized the Fraser Glaciation and provided an extensive list of surficial geology maps, including NTS map-sheet references.

Glacial Deposits [2]

Our chief interest in the glacial history lies in understanding the nature and performance of the surface materials that have been left behind (Figure 2.15).

The direct deposits of glacier ice are termed till. In general, two kinds of till are recognized: basal till and ablation till. Basal till was deposited as the base of the glacier melted from frictional and geothermal heat. In some places, the basal debris-rich layer was sheared off the main ice mass, and then melted in place; in others, the overpassing ice might deform a subglacial soil. Ablation till is melted-out glacial debris that accumulated on the ice surface as it was lowered by solar heating. Basal till is non-sorted and massive (not bedded), although in large exposures it may display faint stratification caused by shearing at the ice base. The texture of till depends on the nature of the source material. For example, till in and near the Coast Mountains tends to be sandy because the granitic bedrock weathers to a sandy residue. Till derived from the basalt of the interior plateaus contains a high proportion of silt and fine sand, while till derived from shale, for example in northeastern British Columbia, contains much clay; however, where tills are derived from "recycled" sediments such as earlier glacial deposits, their texture and mineralogy may not reflect the local rocks. In the Thompson River valley, for example, tills consist of rounded pebbles derived from fluvial deposits in a sandy-silt matrix derived from glaciolacustrine sediments.

Basal tills are dense—denser and stronger than any other surficial material, typically overconsolidated (i.e., compacted by forces greater than the existing overburden weight), relatively cohesive, and of low permeability. These tills form good foundation

2 In this section, many glacial landforms are named without further definition. Terms may be found in the glossary (Appendix 1) accompanying the Terrain Classification System for British Columbia (Howes and Kenk [editors] 1997). These definitions, written from the perspective of identification for terrain analysis, are not all complete or authoritative. Further references include Benn and Evans (1998), Martini et al. (2001), Evans (editor, 2003), and Evans and Benn (editors, 2004).

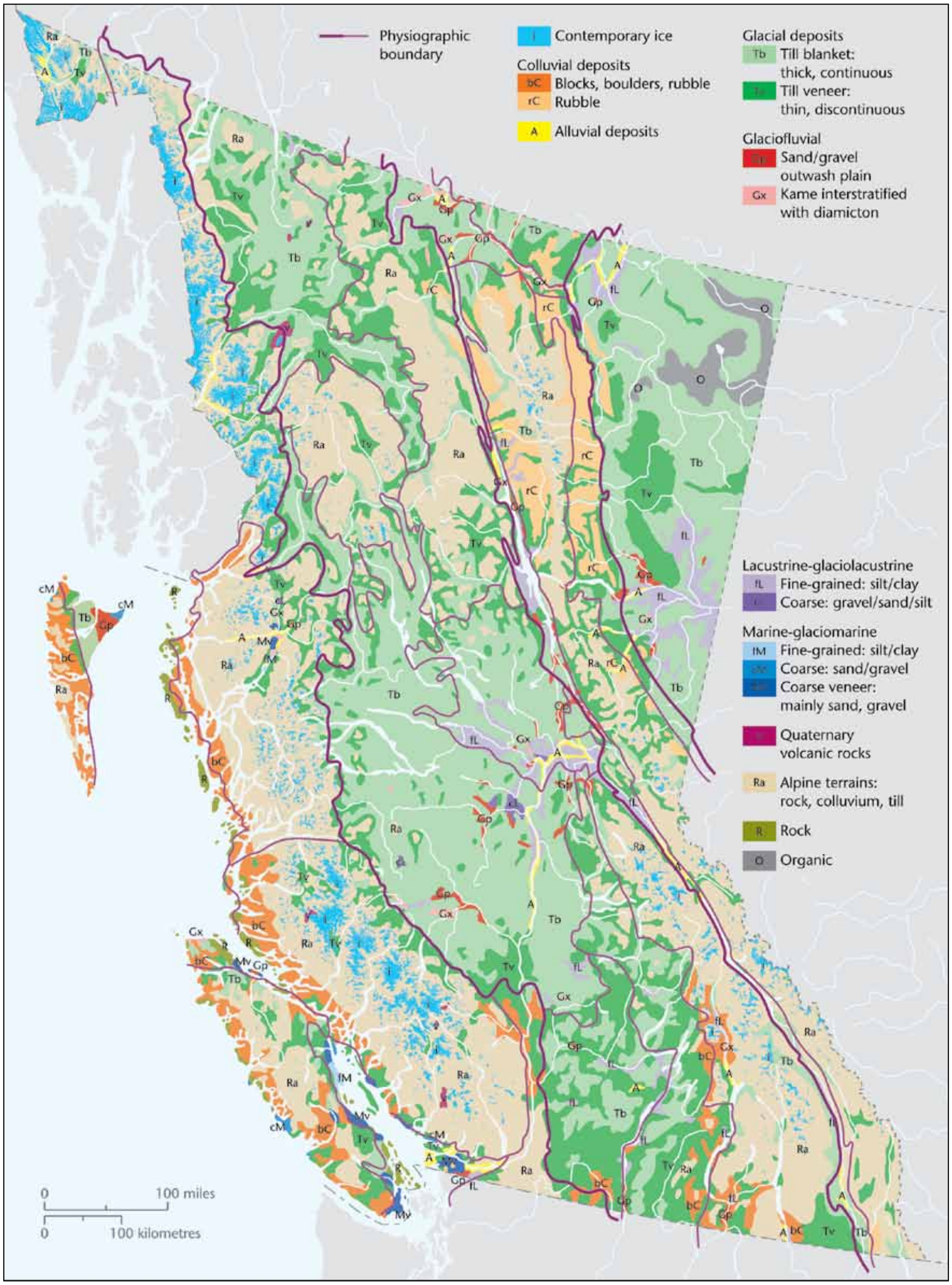

FIGURE 2.15 *Generalized glacial deposits in British Columbia (from R.J. Fulton 1995). (Minister of Public Works and Government Services Canada)*

material, but drain slowly. In general, ablation till tends to include fewer fines (silt, clay) than basal till and more angular clasts, both characteristics resulting from the absence of abrasion and grinding that modified debris at the ice base. The material is massive, except where debris flows off melting ice (called "flow till") may have resulted in localized bedding. This material tends to have high bearing strength, but lower cohesion and higher porosity and permeability than basal till. Ablation till may be similar to the material in kames (see below), and sometimes the two are hard to distinguish. At glacier margins, tills may accumulate to form moraines, which are almost always complex deposits incorporating tills of varying character, and often meltwater deposits as well.

Kames and eskers consist of meltwater-transported sand and gravel that was deposited in contact with glacier ice (ice-contact glaciofluvial deposits). Eskers are sinuous ridges that were formed in glacial tunnels. Kames include mounds and hummocks, irregular terraces, and delta-terraces, and typically exhibit conspicuous slump and fault structures owing to partial collapse of the original deposit when the supporting ice melted. The materials are bedded and more or less well sorted within beds—often appearing no different than ordinary fluvial deposits—but the bedding is highly variable. Lenses or layers of silt and, less commonly, clay may be present, having accumulated in backchannels or kettle ponds. Eskers often reveal limited or little bedding, since they may have been quickly deposited under strongly decelerating currents.

Kame deposits are widespread on the interior plateaus, where they are often associated with meltwater channels and drainage divides. These deposits form a significant proportion of the surficial deposits in many interior valleys where the last ice stagnated, but are less common within the mountains. Kames tend to have good bearing strength, are permeable and generally well drained, although the highly variable bedding with included silt or clay layers may create perched water tables locally.

Proglacial outwash deposits (glaciofluvial deposits), laid down by meltwater streams draining away from the glaciers, are similar to postglacial fluvial deposits, and are found along many valleys as plains, terraces, raised deltas, and large fans (Figure 2.16).

FIGURE 2.16 *View toward the Muskwa Ranges (northern Rocky Mountains) near Liard Hot Springs. In the foreground is a glaciofluvial delta issuing from the tributary valley and, below that, river terraces. (Photo: J.M. Ryder)*

Where meltwater transported chunks of floating ice, which were then buried, the outwash is pitted by kettles. These may be dry, or the depressions may contain ponds or marshes, forming small but ecologically significant wetlands. Outwash texture ranges from sand to boulders, sorting from none to well sorted. Stratification is mainly tabular, with foreset beds locally representing the preserved slipfaces of former bars. Where deposition was rapid and continuous, bedding may be absent. Deltas consist primarily of foreset beds. The landforms originated as braidplains, which may have remained active well after the end of glaciation, but most have been dissected and transformed into terraces in Holocene time as changing hydrologic and sediment supply regimes have caused the rivers to degrade. Outwash deposits (Figure 2.17) have good bearing strength and relatively high porosity and permeability. These deposits tend to be well drained unless the local water table is high. Some thick outwash bodies are important aquifers, such as the Abbotsford aquifer in the Lower Mainland.

FIGURE 2.17 *Quaternary sediments in Coquitlam Valley near Vancouver: interbedded outwash gravels and sands. Similar sediments underlie much of the Surrey upland of the Fraser Lowland. (Photo: J.M. Ryder)*

Glaciolacustrine (glacial lake) sediments were deposited in ice- or moraine-dammed lakes during deglaciation (see Figure 2.11 for major late-glacial lakes). Silt and fine sand are the common materials, but coarser materials are found close to inflow points, including gravel fan-deltas at the mouths of mountain streams. Clays occur locally, particularly in parts of the Nechako Lowland. Bedding ranges from massive to laminated, and varves (annual rhythmic beds) are common in some areas. Where the thickness of the sediments is great, the upper surface (i.e., the old lake floor) is level. Elsewhere, fine sediments are draped over and reflect the pre-existing topography of the lake floor.

After the ice-dammed lakes drained, rivers and streams cut down rapidly into the glaciolacustrine sediments, leaving the remaining parts of the former lake floor standing as terraces (e.g., in parts of the Fraser, Thompson, and Columbia valleys) (Figure 2.18). These tend to be gently undulating, not as flat as fluvial terraces, and commonly slope gently upward toward the valley sides.

Glaciolacustrine sediments are moderately cohesive, especially where clay content is high, and will stand in vertical scarps controlled by vertical desiccation joints where they are relatively dry. They also display some degree of plasticity (depending on clay content and mineralogy) when wet. These materials are highly erodible, so extensive gully systems developed rapidly after streams dissected the lake floors. Fine silts and clays have relatively low permeability, which can give rise to perched water tables where piping (subterranean erosion) may initiate gullying. Perched water tables may also induce high pore pressures, reducing slope stability and initiating slump and flow failures at scarps and gully heads. The bearing strength of glaciolacustrine deposits is less than that of till and gravelly materials, but may be satisfactory for light structures. Drainage control is critical for the security of any activities on glacial lake terraces, especially near the crests of scarps. Irrigation has induced massive failures of glaciolacustrine terraces in interior British Columbia.

Glaciofluvial and glaciomarine sediments were deposited in coastal areas when the sea level was relatively high at the end of Fraser Glaciation. Glaciomarine sediments range from stony, silty clay (the stones dropped from melting, floating ice), which is

FIGURE 2.18 *View upstream in Thompson Valley from Elephant Hill (Ashcroft). Colluvial cones overlie glaciolacustrine silts in the valley. The silts are prone to piping and mass failure when wet; significant landslides followed the inception of irrigated agriculture on the terraces. (Photo: J.M. Ryder)*

typically massive but may show faint stratification, to laminated silt and clay. In either case, mollusc shells are diagnostic, and distinguish stony material from similar basal till. As sea level fell, beach deposits—thin layers of sand or gravel—were developed by wave action on these glaciomarine deposits. Fine-grained glaciomarine deposits have very slow drainage. Some clayey deposits retain very high pore water content in an openwork structure of clay minerals originally bonded by salt. These may be super-sensitive ("quick clays") and susceptible to liquefaction when strongly disturbed, as by seismic shaking, slope failure, or vibration of persistent heavy traffic.

Maps of surficial materials have been completed at scales between 1:20 000 and 1:250 000 for many areas in British Columbia. Many of the maps are not formally published. Mapping follows the *Terrain Classification System for British Columbia* (Howes and Kenk [editors] 1997).

HOLOCENE EPOCH

Holocene History

The Holocene Epoch was designated to demarcate postglacial time and conventionally represents the last 10 000 years. The definitive end of the last glaciation remains controversial because the date varied in different parts of the world. The figure is probably a reasonable generalization for British Columbia, in terms of radiocarbon years, but probably should be approximately 11 500 calendar years.[3]

During and following deglaciation of the British Columbia landscape, widespread glacial deposits remained on hillslopes and along valleys in positions that were unstable in the new subaerial environment.

3 The chronology of the latest part of the Quaternary Period is often specified in "radiocarbon years" BP, since ^{14}C assays on fossil organic material were the first, and have for over 50 years been the principal, means for estimating the absolute age of events within the last 50 000 years (encompassing the Fraser Glaciation). Radiocarbon years and calendar years diverge, however, because of variations over time in the ^{14}C content of the atmosphere. The divergence is large, ^{14}C age being about 1250 years younger at the time of the final deglaciation, and approaching ~2500 years about 15 000 years ago. The discrepancy, while not critical for ordering events, becomes important for those attempting to establish mean rates of postglacial landscape change. It is necessary to obtain the calendar age of the limiting reference surfaces to develop rate information. Correction tables are available. (See Walker 2005 for further details.)

Substantial mass wasting and fluvial reworking of these deposits created a major "pulse" of sediment movement in the early postglacial landscape. This has been referred to as "paraglacial sedimentation" (see Ballantyne 2002). It worked its way through the smaller and steeper upland basins first, where supplies of movable sediment were quickly exhausted. As the output from these basins moved downstream, larger drainage basins experienced a prolonged paraglacial sediment pulse. When the establishment of forests and grassland stabilized postglacial slopes, paraglacial sedimentation largely ended. The most dramatic part of the process was completed before 7000 years BP, but its signal still affects the largest drainage basins, with the principal source of sediments today being the erosion of Pleistocene valley-fill deposits. Rock slopes debuttressed by glacial erosion also experienced an increased rate of failure in early Holocene times, so that many major rock slides occurred. The net result of this phase was the early Holocene development of talus slopes (Figure 2.19), colluvial cones, alluvial fans, and early Holocene floodplains, many of which are now dissected.

The Holocene climate has by no means been constant. Warm, dry conditions characterized the earlier half of the period, but a continuing modulation of climate with cooler, wetter conditions recurred approximately every 2500 years. In many parts of the province, landscape conditions are sufficiently stable that these fluctuations have left no major evidence. However, in more extreme environments at higher elevations, effects are preserved especially in records of renewed alpine glaciation. A resurgence of glaciers occurred at the end of the Pleistocene about 11 500 ^{14}C years BP (13 000+ calendar years) and is known locally as the "Sumas stade" (globally, it represents the end-Pleistocene "Younger Dryas" cold interval). Following that, known cold phases are approximately centred on 8000, 5500, and 3000 years BP; and within the last millennium. Each of these cold phases has been more severe than the last, so that the "Little Ice Age" of the last millennium, culminating in the 18th century, represents the most

FIGURE 2.19 *The Rocky Mountains in the Continental Ranges near Elko. Cliff-forming Cambrian limestones stand above the glaciated valley. The cliffs are hemmed by talus slopes that are also subject to snow avalanches below the gullies cut into the upper slopes. (Photo: J.M. Ryder)*

severe conditions of the Holocene Epoch. Within the last three centuries, alpine glaciers attained their greatest Holocene extent but, since the early 19th century, increasingly rapid retreat has been the norm. This implies that a transient pulse of paraglacial sedimentation has occurred in our mountain landscapes within the past three centuries. Some of the geomorphic processes that we observe today in mountain valleys—including mass wasting and fluvial processes—may represent continuing relaxation from this recent "Neoglaciation."

Neoglacial effects are not the sole complicating mechanism in the pattern of Holocene erosion and sedimentation. Fire and forest infestations, by affecting the vegetation cover of the land surface, may have substantial transient effects on sediment mobilization. The space and time scales of these disturbances are much more local than those of regional climatic trends, but may be hundreds of kilometres and many decades in the boreal forests. The history of such disturbances is regionally variable, as well, so that Holocene histories of landscape stability may be complex and will differ in their details in different parts of the province. The historical record is best preserved in sites of persistent sedimentation—lakes and bogs—and has not been studied in detail.

The Contemporary Landscape

The result of the tectonic, geological, and glacial history described above is the landscape in which we live. Because the province has a wide range of elevations, and because local climate is elevation dependent, hydrologic and geomorphic processes exhibit vertical zonation in the contemporary landscape (Figure 2.20a). Furthermore, because water and sediment move downhill, process regimes at higher elevations influence landscape processes and landscapes at lower elevations. In general, the steepest gradients are high in the landscape and near the origin of drainage, so that geomorphic "process domains" may be mapped in a space defined by gradient and increasing area (Figure 2.20b). Hillslope processes of mass wasting dominate the steep topography at the highest elevations. Stream channels originate in humid mountains at drainage areas of about 0.01 km^2 (1 ha) and debris flow processes dominate these steep channels. Fluvial processes dominate the movement of sediment in the landscape at areal scales greater than 1 km^2 and on gradients of less than 10% (Figure 2.20b, plot 1). In glaciated landscapes, however, process distribution is modified by the glacial topographic inheritance of upland glaciated valleys, which constitute a low-gradient sediment trap above the main valleys. Hence, decoupled high altitude and low altitude process systems may both exist (Figure 2.20b, plot 3). The headmost cycle of erosion, mass wasting, and sediment deposition may occur within cirques and hanging valleys.

Sediment yield patterns reflect this complex topographic system (Figure 2.21). Relatively high sediment yields driven by mass wasting are experienced in headwater areas, followed by a sharp decline in sediment yield with a low near 1 km^2, indicating sediment trapping and aggradation in upland valleys—mainly high-elevation glacial troughs.

After this point, a renewed increase occurs in sediment mobilization on the main valley walls below, where debris flows continue to be a prominent phenomenon in drainage basins of up to 10 km^2. In larger drainage basins, fluvial processes become dominant. Locally, this transition is indicated by renewed deposition on valley floor colluvial fans and reduced sediment yield. Regionally, however, renewed uptake of sediments by direct fluvial erosion of Pleistocene valley fills creates increasing specific sediment yield and consequent landscape degradation in drainage basins of between about 10 and 30 000 km^2.

In British Columbia, the highest elevations remain the domain of rock, snow, and ice. Contemporary alpine glaciers and glacial processes continue to erode mountains above a glaciation limit that varies from about 1800 m above sea level (asl) on the southwestern outer coast to near 3000 m asl in the Rockies. The glaciation limit, which represents the lowest elevation of the mountain at which a glacier can persist, lies slightly below the regional average equilibrium line for the annual balance of snow accumulation and ablation. Below this limit lies a zone of rock and alpine tundra that is dominated by freeze–thaw and nival processes.

In areas with modest winter snow cover—the consequence either of modest winter precipitation or of snow removal by wind—lies an alpine periglacial zone (Figure 2.2) in which the ground may remain at subzero temperatures year-round. The severe climate promotes active freeze–thaw weathering and the production of rock fragments. Since the zone is mountainous and steep, it tends also to be a region of active mass wasting by rock fall and rock slides.

This zone is also characterized by the presence of neoglacial deposits that may be eroded during

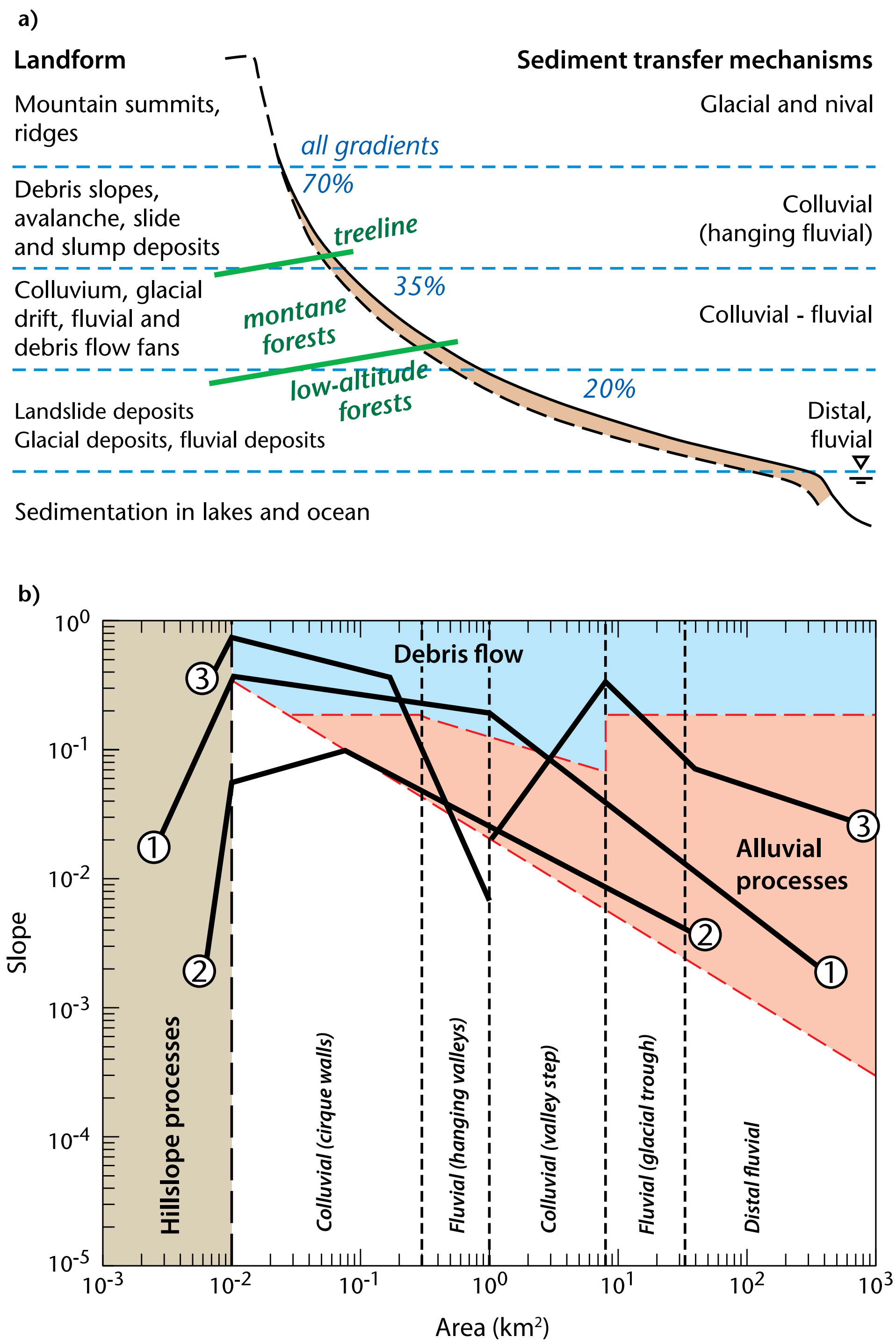

FIGURE 2.20 *(a) Vertical zonation of hydrologic and geomorphic processes in the British Columbia landscape. (b) Geomorphic process domains and the distribution of topography in slope-area space. The basic concept is from Montgomery and Foufoula-Georgiou (1993), modified for glaciated terrain by Brardinoni and Hassan (2006), and further generalized for this presentation. Plot 1 represents Montgomery and Foufoula-Georgiou's original concept for equilibrium (unglaciated) montane landscapes; plot 2 is a conjectural trajectory for regions of modest relief; plot 3 represents Brardinoni's findings for the British Columbia Coast Mountains. Notice the sharp gradient decline centred on 1 km², indicating hanging glacial troughs.*

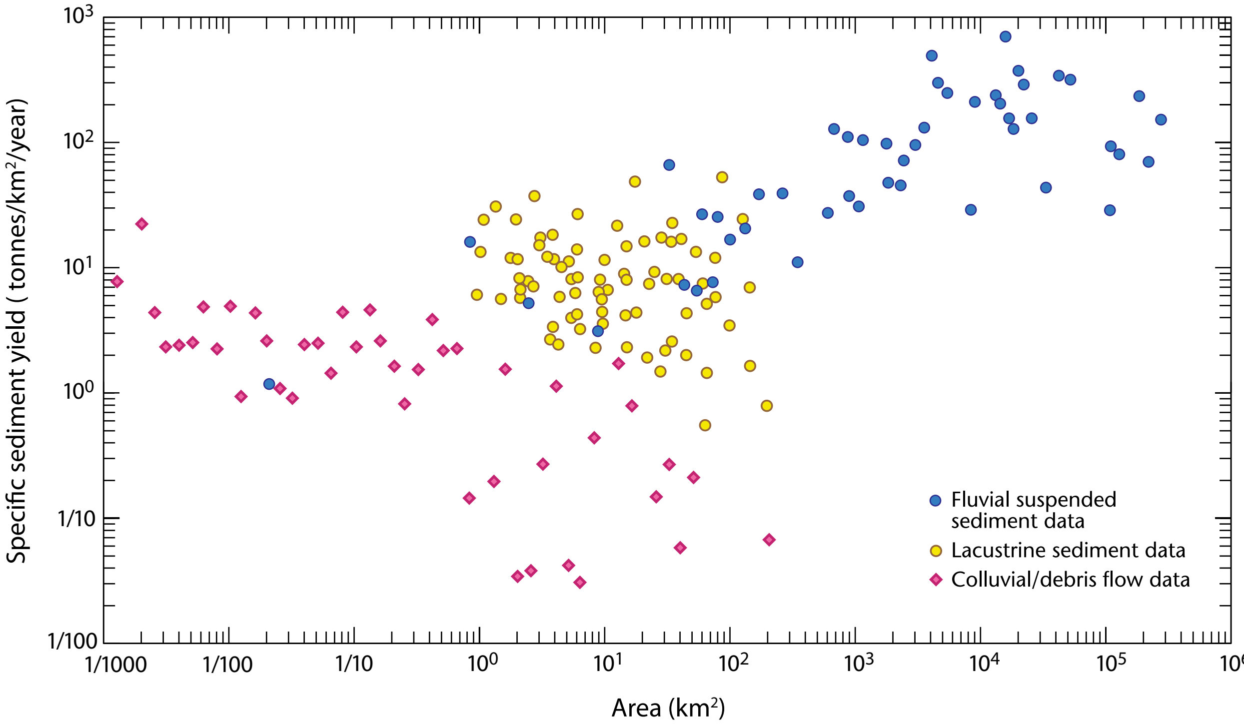

FIGURE 2.21 *Regional sediment yield pattern (after Church et al. 1989 and Brardinoni et al. 2009). The pattern of sediment yield is recognizably correlated with the slope-area distribution of landscape area, but the data exhibit a disjunction near 1 km²: the mass wasting data represent a sample from a restricted region, whereas the fluvial sediment transport trend is based on province-wide data.*

summer by debris flows and running water. The possibility for major rock avalanches and for sudden drainage of moraine- or ice-dammed lakes lends a catastrophic possibility to geomorphic processes in this zone.

In the mountains, glacial valley floors below the summits define the lower limit of the nival–colluvial zone and are the site of a high-altitude fluvial domain, often with wetlands. These systems are often "hanging" above the main valleys and are decoupled from sediment transfers. Below this hanging system, sediments are mobilized and fed into the main valley system by colluvial processes, including rock fall, debris slides, and soil creep, and by debris flows descending gullies on the steep valley walls. The occurrence of debris-contributing streams varies with climate; they may be separated by as little as 50 m along mountain sides on the exposed coast, and may decline to effective non-occurrence in the driest portions of the interior plateaus. Along the major valleys, clear fluvial dominance of drainage is established at areas of about 10 km². In the transition zone (1–10 km²), significant active landforms include colluvial fans and cones formed at the base of the valley walls by mountainside debris flows, and alluvial fans emanating from larger, fluvially dominated tributary valleys where slope declines and confinement is lost. Debris flow appears to be the dominant process by which sediments are delivered from hillsides to the main valley floors in the mountains today, although other colluvial processes—such as rock fall—may be locally important.

Along the valley floors, glacial deposits continue to be modified in the contemporary landscape by fluvial erosion and by slumps and slides that result from the consequent removal of lateral support. Many of the rivers have become incised within their Pleistocene valley fills during Holocene time, so scarps in glacial sediments are common along many valleys. In fine-grained (e.g., glaciolacustrine) sediments, altered drainage conditions—including irrigation—can produce piping, gullying, and slump–earth-flow type failures. In comparison, land surface erosion by running water is relatively rare since soil infiltration capacities generally exceed rates of precipitation. Exceptions occur on exposed mineral soils such as landslide scars, severely burned areas (where soils have become hydrophobic), and soils strongly compacted by machinery.

In the interior plateaus, an important variation occurs in the distribution of process domains (Figure 2.20b, plot 2). Here, the geomorphic process system often begins in an upper-level landscape of gentle gradient developed on glacial soils and featuring abundant small lakes and wetlands. It then spills down through gorges in the bounding scarps into the major valleys and well-organized fluvial system. This "gentle-over-steep" terrain has no headmost colluvial system, but colluvial activity may be important on the side slopes of the major valleys (Figure 2.20a).

Soil Development

The most locally variable feature of the landscape is the soil, the final and often most important of the landscape features we have been considering. Soils form the matrix for plant germination and growth and the base foundation for most human-made structures. Soils influence water retention and drainage and disturbed soils present significant problems of erosion.

Soils (in the pedological sense) develop at the Earth's surface as the consequence of mineral weathering and the admixture of organic material from vegetation in a particular climate. Climate is important both regionally and locally because characteristic thermal and moisture conditions in the soil significantly influence the biotic and chemical processes that are important in soil formation. The word equation:

soil = ***f*** [parent material, topography, climate, biota, time]

is often used to summarize the factors controlling soil formation. The first two of these factors have been directly considered in this chapter. Climate in British Columbia is strongly influenced by topography and by distance inland (see Chapter 3, "Weather and Climate"), and biota is strongly influenced by climate.

Therefore, we may expect some general correlation between the geological and topographical features we have been considering and regional soils. In turn, soils and the biogeoclimatic zonation of the province have a discernable correlation, since biota is regionally influenced by the same topographic and climatic factors that determine soil types.

Soils, however, exhibit a great variety of morphology in response to subtle variations in site conditions, particularly moisture and drainage. A detailed map of soils would be considerably more complex than those of geology or surficial materials. To deal with this circumstance, soil classifications are generally hierarchical in nature. The Canadian system of soil classification employs five levels, the most general of which are order and group (or great group) (Soil Classification Working Group 1998). The nine soil orders reflect regional factors of soil formation, chiefly climate and vegetation and hence organic matter accumulation, in determining soil character. Soil groups make these distinctions more specific and introduce factors based on drainage. Soil groups for British Columbia are mapped in Figure 2.22 and brief descriptions follow. The nine major soil orders are Podzols, Brunisols, Chernozems, Luvisols, Gleysols, Organic soils, Cryosols, Solonetz soils, and Regosols.

Podzolic soils are formed in cool, humid to perhumid climates beneath coniferous forests. Hence, they are widespread along the coast of British Columbia and, except at the lowest elevations, throughout the mountain regions. Formed as a result of intense leaching, these soils are characterized by relatively high acidity and strong profile development. Podzols also tend to develop on volcanic ash that is distributed over parts of southern British Columbia.

Ferro-Humic Podzols occur on the perhumid western slopes of the Coast and Insular Mountains and feature accumulation of organic matter, iron, and aluminum in a Bhf horizon. Humo-Ferric Podzols occur in slightly less moist mountainous areas along the lee side of the Coast and Insular Mountains and in the Interior mountains. They display a red horizon (Bf) where iron and aluminum have accumulated, and less organic accumulation.

Brunisols developed as a result of relatively minor modification of parent material, and the soil horizons are less strongly developed than in soils of other soil orders. This is attributed to long, cold winters in the north and at high altitudes, lack of moisture in more southerly and interior areas, and, in some cases, because parent materials are young (such as recent alluvium). Consequently, Brunisols are often thought of as immature soils. They are characterized by a thick, pale reddish-brown (Bm) horizon and are found in slightly drier landscapes than Podzols. Two groups can be mapped on a regional scale. Dystric Brunisols (pH < 5.5) are found in relatively humid areas on somewhat acid parent materials, and are similar to, but less well developed than, Humo-Ferric Podzols. Eutric Brunisols (pH > 5.5) occur in

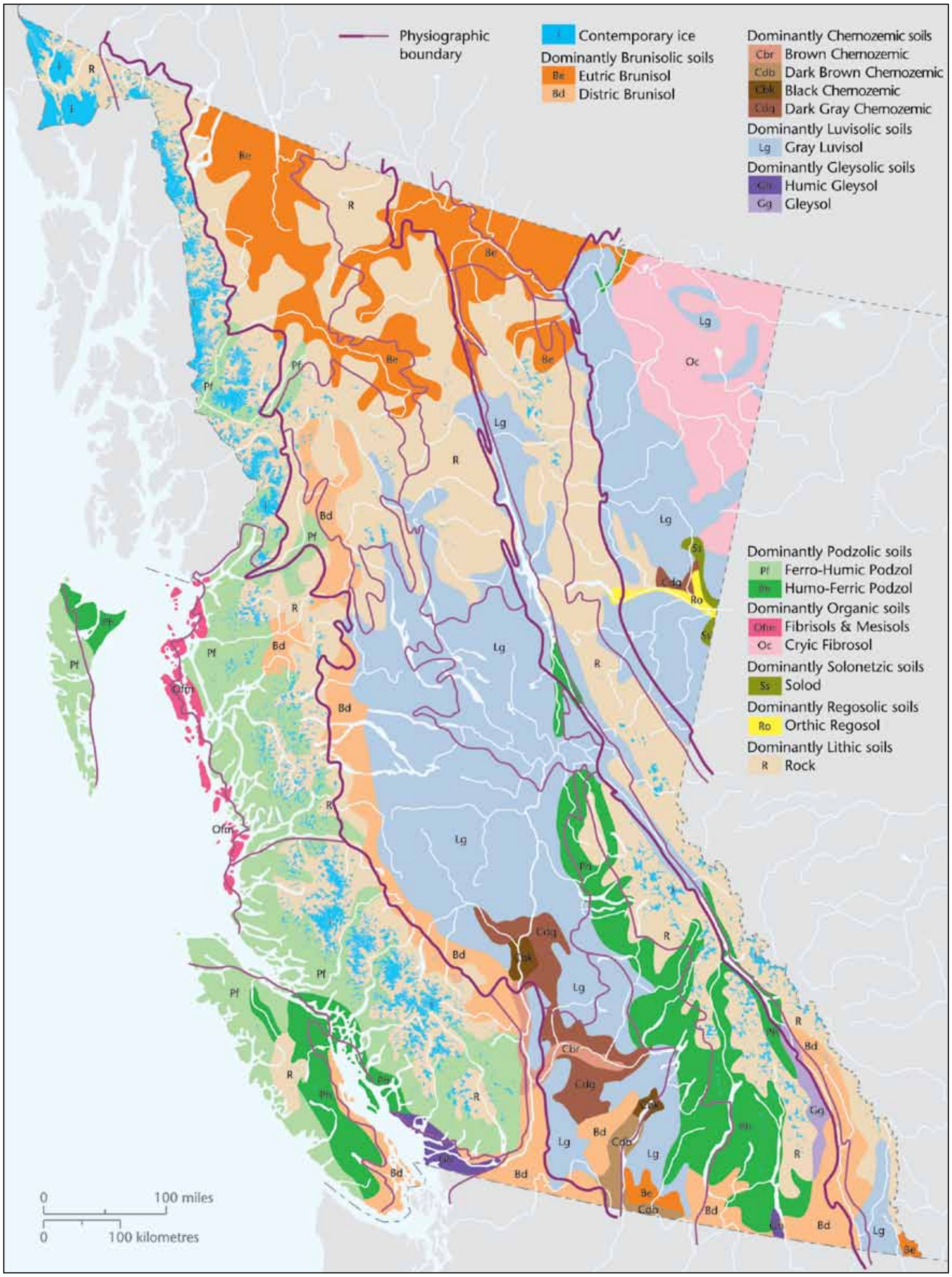

FIGURE 2.22 *Generalized map of Soil Groups for British Columbia (modified after Soil Research Institute 1972). (Minister of Public Works and Government Services Canada)*

slightly drier areas under forest or shrub vegetation. Sombric Brunisols, which are patchily distributed in subalpine and alpine areas, are distinguished by a thick dark surface horizon (Ah) enriched by organic matter.

Chernozems are grassland soils of semiarid to subhumid regions. These soils are widespread in the valleys and on the plateaus of the southern Interior and are limited in the Peace River region. Their principal feature is a thick black surface horizon (Ah) resulting from accumulation of organic matter in the topsoil. Soil groups are Brown, Dark Brown, and Black Chernozemic soils, in order of increasing organic matter content and increasing soil moisture, and Dark Gray Chernozems, which are forest–grassland transition soils.

Luvisols are associated with forest vegetation (including mixed forest and boreal forest), parent materials that are base-saturated and fine-grained, and regions of moderate precipitation. These soils are characterized by leaching, which results in the accumulation of clay to form a Bt horizon. Gray Luvisols are distinguished by the presence of an Ae (white, eluviated) horizon and are typically more acidic than other Luvisols. This soil type is widespread on the interior plateaus of British Columbia south of 56°N and in the Rocky Mountain foothills of the Peace River region, where it commonly develops on clayey tills and glaciolacustrine sediments. Luvisols are found in more humid or colder regions than the Chernozems, so that profile development by leaching is more active.

Gleysols are characteristic of poorly drained areas. These soils are saturated for extended periods and exhibit reducing conditions; hence, mineral and organic matter transformations in the soil are curtailed. They are common on fine-grained parent materials, such as clayey till and glaciolacustrine sediments. Gleysols occur in association with other soils wherever seasonally wet areas occur and are grouped according to organic matter accumulation. Humic Gleysols have a thick humic horizon owing to the accumulation of untransformed vegetative matter. Orthic Gleysols lack this heavy accumulation. Luvic Gleysols have significant clay accumulation in the B horizon.

Organic soils develop in the wettest parts of the landscape, are saturated for most of the year, and are composed mainly of organic matter. These soils are distinctive because formation is related to in-situ accumulation of organic matter rather than modification of pre-existing mineral material. Organic soils are found in wetlands across the province, ranging from small kettle marshes to extensive peat bogs. Normally saturated organic soils are Fibrisols, in which the organic matter is poorly decomposed; these typify peat bogs in the northeast and on the Coast. Mesisols in the central Interior exhibit moderately decomposed organic matter, and locally occurring Humisols are highly decomposed. The other major class of organic soils is Folisols, which consist of organic matter draped over thin mineral soil or rock on mountainsides in very wet areas such as the windward slopes of the Coast and Insular Mountains.

Cryosols are associated with permafrost (Figure 2.2), and are most widespread in the drier parts of the high northern mountains (where insulation from snow cover is least), and in the far northeast. Turbic and Static Cryosols are mineral soils of the mountains, with horizons disrupted or not disrupted by frost heaving, respectively.

Organic Cryosols are peat soils in the northeast with sphagnum peat that insulates and preserves the underlying frozen ground. These soils fall within the southern limit of the discontinuous permafrost zone (Figure 2.2).

Solonetz soils have significant sodium and (or) magnesium salt accumulations and are characterized by the presence of Bn (sodium accumulation) or Bnt horizons. Small areas of these soils—mostly in closed depressions—are scattered throughout the semiarid southern Interior, the salt having been transported into the depressions by drainage. In the Peace River region, Solonetz soils are moderately common on sedimentary rocks that have high salt content. These are mostly of the Solod order, which is characterized by highly advanced leaching and salt concentration in the B horizon.

Regosols show very little pedological development, and are distinguished by the absence of any significant (>5 cm) B horizon, although there is usually some organic material in the surface horizons. These soils are found wherever local conditions have precluded soil development; that is, where parent materials are very young (sand dunes, river bars) or where weathering and soil formation occur very slowly, as on resistant rock. Thus, Regosol occurrence is relatively extensive in the high mountains and on steep slopes. Bare rock, also widespread in the mountains, is classified as "Lithic soil." Lithic soils are often intimately interspersed with Podzols, Folisols, and Regosols.

Human activities have become a significant factor in the geomorphic environment of British Columbia. Today, humans are the principal landscape-modifying agent in the province at all but tectonic scales. This is a recent development, both here and across the world. Humans have had a significant impact on the landscape for a long time, simply because of cumulative effects beginning with the first agrarian societies. Industrial societies, however, can also concentrate sufficient energy locally in the landscape to effect major changes in relatively short periods of time.

Communication routes create artificial barriers in the landscape and often reorganize local drainage. Forest clearance for agriculture or for forest exploitation modifies hydrology and land surface conditions, often creating conditions for accelerated soil erosion, and urban land conversion produces the most dramatic changes of all.

In British Columbia, the most widespread impact on the land—though by no means the most severe, locally—arises from forestry activity. Effects on slope stability, soil erosion, and downstream sedimentation are a long-standing concern. Table 2.1 gives order of magnitude summaries of soil erosion rates measured in British Columbia and in the adjacent Pacific Northwest states for undisturbed terrain and for terrain subject to forest harvesting.

Much of the soil that is mobilized by both natural and human agencies does not travel beyond the base of the slope on which it originated. This picture is consistent with the distribution of geomorphic process domains discussed previously. Headward colluvial processes result in the accumulation of sediments on lower slopes and on alluvial fans. Hence degradation occurs on upper slopes and aggradation at lower levels. Denudation, measured locally, declines from hillslopes to headmost drainage basins (Figure 2.21). Farther downstream, sediments are mobilized along the rivers, which continue to re-entrain Pleistocene valley fill materials. In the British Columbia landscape, sediment yields increase steadily with drainage basin area, up to about 30 000 km^2 (representing a linear distance of about 175 km). Significant sediment accumulation is occurring only on downstream floodplains and at deltas of large rivers. This situation is relatively exceptional in the world: most rivers worldwide appear to be aggrading because sediment mobilized from severely disturbed land surfaces is delivered to the rivers and deposited in them. The most pervasive source of severe disturbance is agricultural activity: in comparison, the mainly forested wildland landscape of British Columbia remains relatively little disturbed.

TABLE 2.1 *Annual sediment mobilization and yield from hillside slopes*

	Mobilization rate		Yield rate to stream channels	
Process	**Forested slopes**	**Cleared slopes**	**Forested slopes**	**Cleared slopes**
Normal regime				
Soil creep (including animal effects)	1 m^3/km[a]	2×	1 m^3/km[a]	2×
Deep-seated creep	10 m^3/km[a]	1×	10 m^3/km[a]	1×
Tree throw	1 m^3/km^2	–	–	–
Surface erosion: forest floor	< 10 m^3/km^2	–	< 1 m^3/km^2	–
Surface erosion: landslide scars, gully walls	> 10^3 m^3/km^2 (slide area only)	1×	> 10^3 m^3/km^2 (slide area only)	1×
Surface erosion: active road surface	–	10^4 m^3/km^2 (road area only)		10^4 m^3/km^2 (road area only)
Episodic events				
Debris slides	10^2 m^3/km^2	2–10×	≤ 10 m^3/km^2	≤10×
Rock failures: falls, slides	No consistent data: not specifically associated with land use.			

a These results are reported as m^3/km channel bank. All other results reported as m^3/km^2 drainage area. Generalized results after Church and Ryder (2001, Table 1).

SUMMARY

This chapter has introduced the physiography and associated physical features of the province. The approach has been to present the major physiographic divisions of the province (Figure 2.1), a consequence of the historical development of the landscape, and to focus on the events that have influenced the recent development of the British Columbia landscape. Attention has also been paid to landforms, materials, and soils, the features that make up the detail and the fabric of the landscape. In this way, the reader may understand the factors that have created the physiography and landscape of today in British Columbia.

The physiography has strongly influenced some important features of the natural history, and even of the cultural history, of the region that are important to consider when practicing resource and land management. This topic will be discussed in detail in many of the following chapters.

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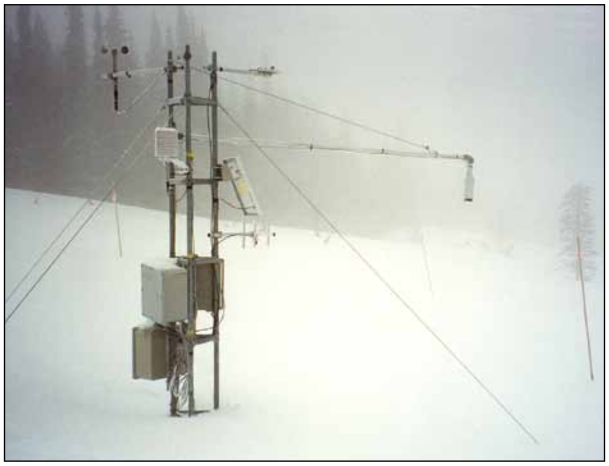

Weather and Climate

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PAUL H. WHITFIELD, AND KERSTIN STAHL

INTRODUCTION

This chapter introduces British Columbia's weather and climate. "Weather" refers to the specific condition of the atmosphere at a particular place and time. It is measured in terms of variables including wind speed and direction, air temperature, humidity, atmospheric pressure, cloudiness, and precipitation. Weather can change from hour to hour, day to day, and season to season. "Climate" is a statistical characterization of the weather, averaged over many years; it is usually represented in terms of means, variability, and extremes of the various weather elements. The following sections focus on (1) the large-scale context of British Columbia's weather and its expression at the scale of synoptic atmospheric circulation; (2) a description of regional climatic variations resulting from the interaction of weather systems with the topography of British Columbia; (3) an overview of large-scale ocean–atmosphere interactions, including El Niño, and their influence on the province's weather and climate; (4) an overview of the climatology and meteorology of extreme hydrogeomorphic events; and (5) a discussion of historic climatic variability and projected future climate change.

ATMOSPHERIC CIRCULATION PATTERNS INFLUENCING BRITISH COLUMBIA

Alternating sequences of west-to-east flowing low-pressure (cyclonic) and high-pressure (anticyclonic) systems dominate the weather in British Columbia. The low-pressure systems form off the Aleutian Islands in the north Pacific and then migrate easterly to approach the province. These systems are generally associated with fronts, which are the boundaries between warm and cold air masses. Low-pressure systems occur more frequently in winter. During summer, the North Pacific high-pressure system shifts northward into the central-north Pacific and tends to dominate, often producing extended spells of fine weather, particularly in southern British Columbia.

Although the weather situation each day is unique, it is possible to group synoptic-scale conditions into relatively distinctive types based on patterns of sea-level pressure or other atmospheric variables. Stahl et al. (2006b) used a computer-assisted pattern recognition approach to classify daily sea-level pressure patterns into 13 circulation types (Figure 3.1). Airflow can be interpreted from the patterns of the pressure contours (isobars) using the following rules: (1) air tends to flow parallel to the isobars, with anticlockwise flow around low-pressure centres; and (2) wind speed is related to the isobar spacing, with tighter spacing indicating stronger winds. The monthly variation in synoptic

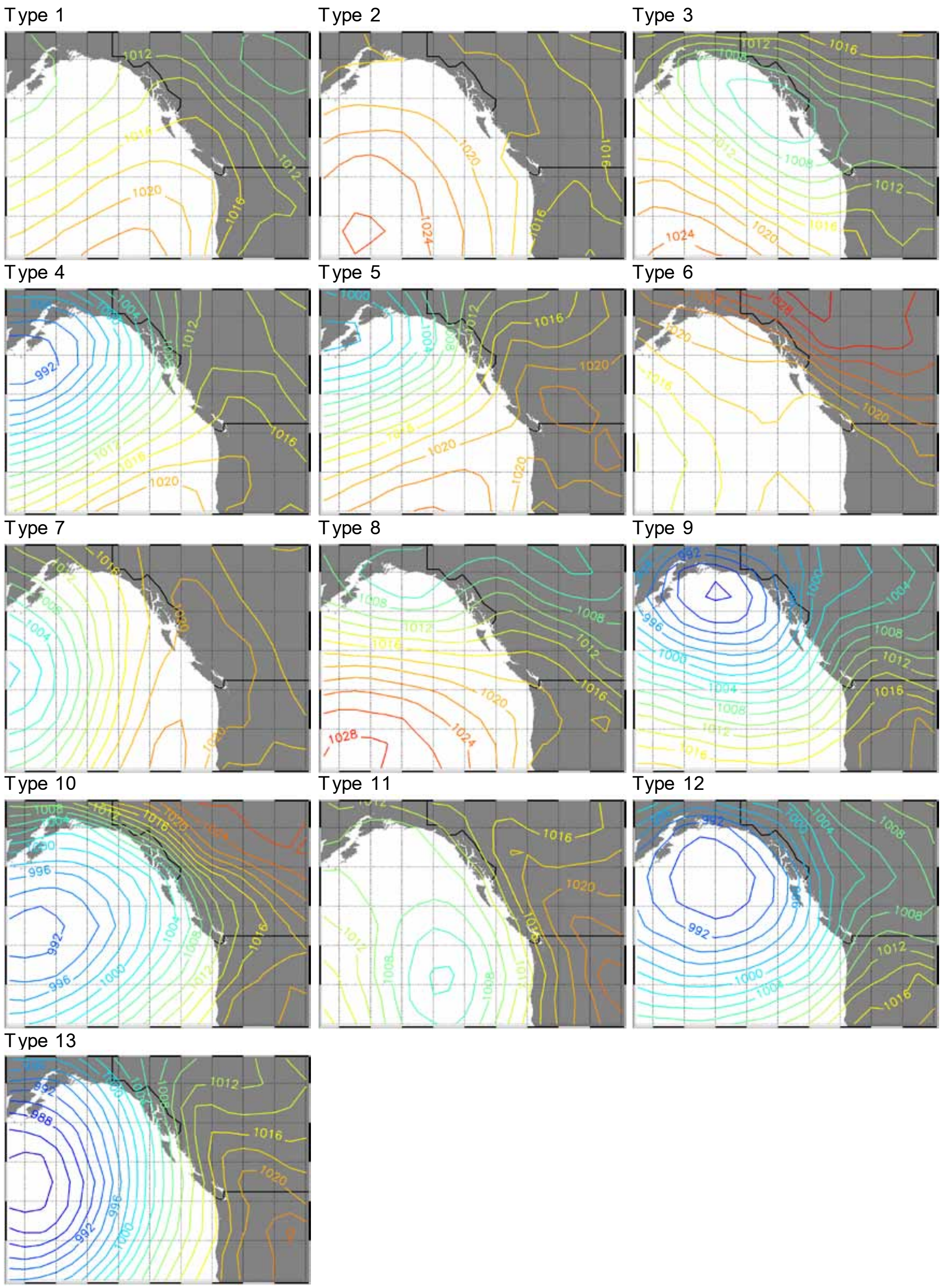

FIGURE 3.1 *Composites of sea-level pressure patterns associated with the 13 synoptic types. (Adapted from Stahl et al. 2006b)*

type frequency is shown in Figure 3.2. Types 1 and 2 are the most frequent circulation types. They occur year-round, but dominate British Columbia's climate during the summer months. Together with Type 8, these types represent systems dominated by the presence of the Pacific High just off the coast. The mainly meridional (south–north) flow over British Columbia (Types 1 and 2) is typical for dry and fine summer weather, whereas the southern position of the High in Type 8 causes more zonal (west–east) flow with unstable weather.

During autumn, winter, and spring, British Columbia's synoptic climatology is more variable, and most of the circulation types occur with a similar frequency between November and February. Circulation Types 3–7 are distributed most evenly throughout the year, but all have a minimum in summer. These types represent a mix of patterns and influences on the provincial climate. The Aleutian Low is visible in all of these patterns, except for Type 6. All involve some influence of continental high-pressure systems over British Columbia. Perhaps with the exception of Type 6, which has a high persistence, these patterns are not very stable and tend to be transitional within the typical mid-latitude circulation characterized by a series of frontal systems moving from west to east. Types 9 through 13 occur mainly in winter, and are all characterized by an Aleutian Low of varying location and depth. Southwesterly flow over British Columbia is common during these situations.

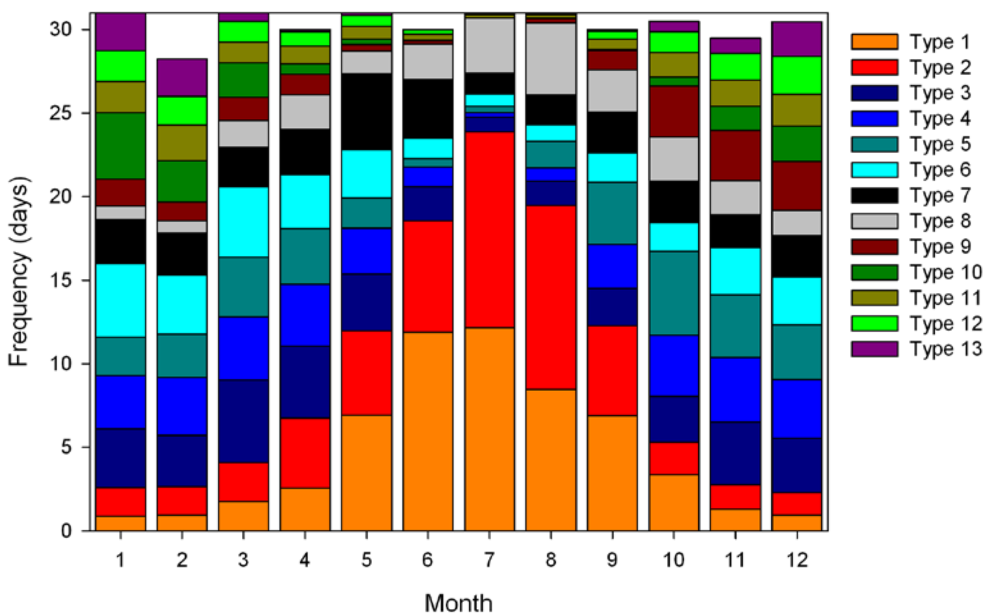

FIGURE 3.2 *Mean frequencies of synoptic types by month. Types 1, 2, and 8 dominate in summer, Types 3–7 occur year-round, and Types 9–13 dominate in winter. (Adapted from Stahl et al. 2006b)*

REGIONAL CLIMATES OF BRITISH COLUMBIA

The interaction of large-scale weather systems with the topography of British Columbia's land surface produces distinctive climatic patterns that vary with elevation, distance from the coast, exposure to the prevailing winds, and season. This section describes the spatial and temporal variability of climate within British Columbia.

Climatic Zones

British Columbia can be divided into five physiographic regions that have distinct macroclimatic regimes (Figure 3.3) (Valentine et al. 1978; Phillips 1980; Chilton 1981). The numerous parallel mountain ranges and extensive plateaus, plains, and basins

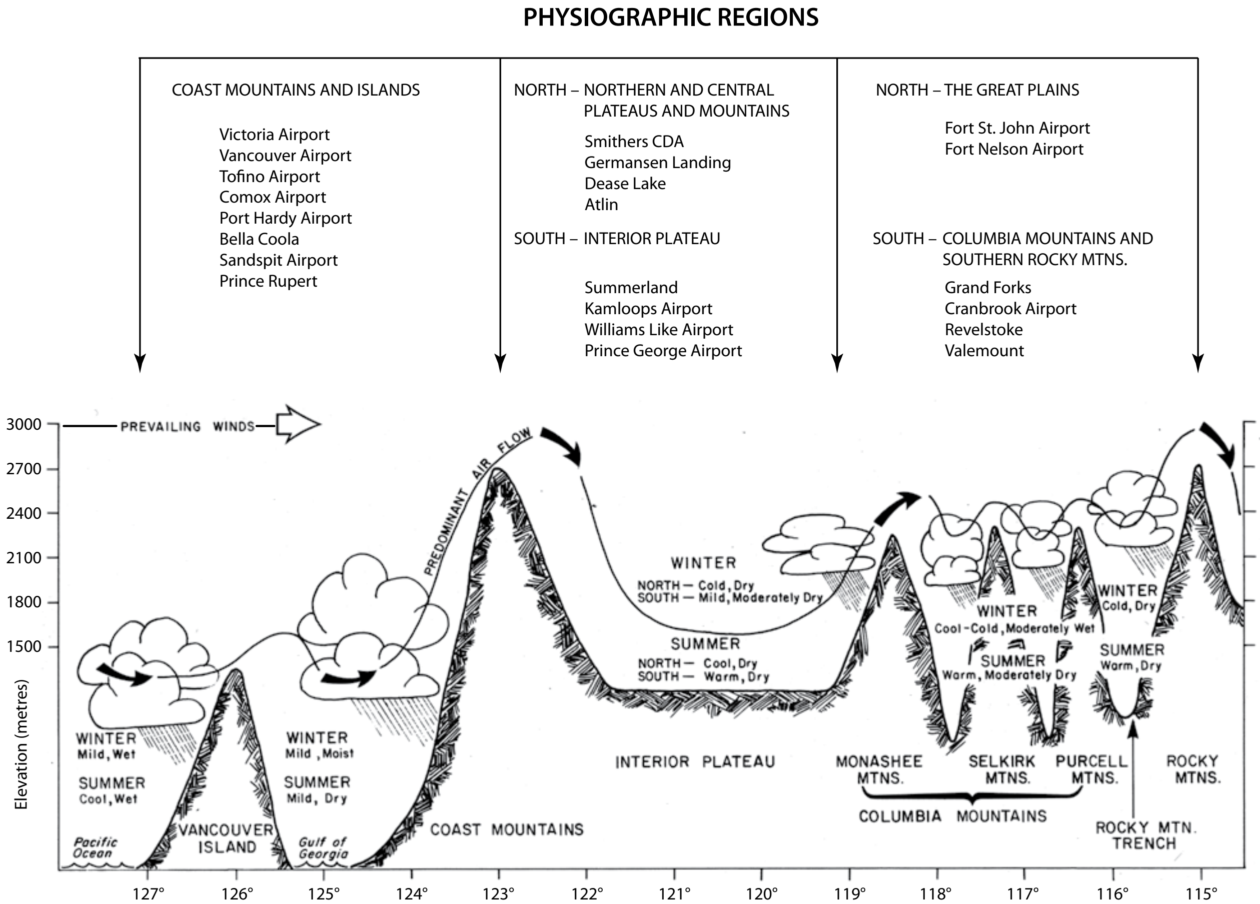

FIGURE 3.3 *Latitudinal cross-section through southern British Columbia illustrating physiographic diversity and resulting climatic regimes. Listed in the upper part of the figure are the five major physiographic regions and associated weather stations within these climates. (Adapted from Chilton 1981)*

produce a range of climatic regimes that is reflected in 14 vegetation zones named from the dominant tree species of the zone (Figure 3.4, Tables 3.1 and 3.2). The zones range from cool, moist coastal forests to warm, dry interior forests, and from Garry oak parkland to black spruce muskeg. Mediterranean-type, semi-arid, subarctic, and alpine climates also occur, supporting extensive areas of open forest, grassland, scrub, and tundra.

The mountain ranges, which lie roughly perpendicular to the dominant large-scale airflow, largely determine the overall distribution of precipitation and the balance between Pacific and continental air masses in the various regions of British Columbia. The wettest climates of British Columbia (and Canada) occur on the outer coast, especially near the mountains on the windward slopes of Vancouver Island, the Queen Charlotte Islands, and the mainland Coast Mountains (Figures 3.5 and 3.6). Here, moist air carried by prevailing westerly winds drops large amounts of rain or snow as it is forced up the mountain slopes. The air descends over the eastern slopes and is warmed by compression, causing the clouds to thin out. The driest climates of British Columbia, located in the valley bottoms of the south-central Interior, are in the rain shadow of the Coast Mountains (Figure 3.5). Travelling farther eastward, the air releases additional moisture as it ascends successive ranges, including the Columbia and Rocky Mountains in the south, and the Skeena, Omineca, Cassiar, and Rocky Mountains in the north. The mountains restrict the westward flow of cold, continental arctic air masses from east of the Rocky Mountains, thus moderating the winter climate in the southern Interior. In spring, the Interior has little precipitation, but early summer is often relatively wet. By midsummer, however, interior storms and precipitation decline again. In middle and late summer, the Pacific

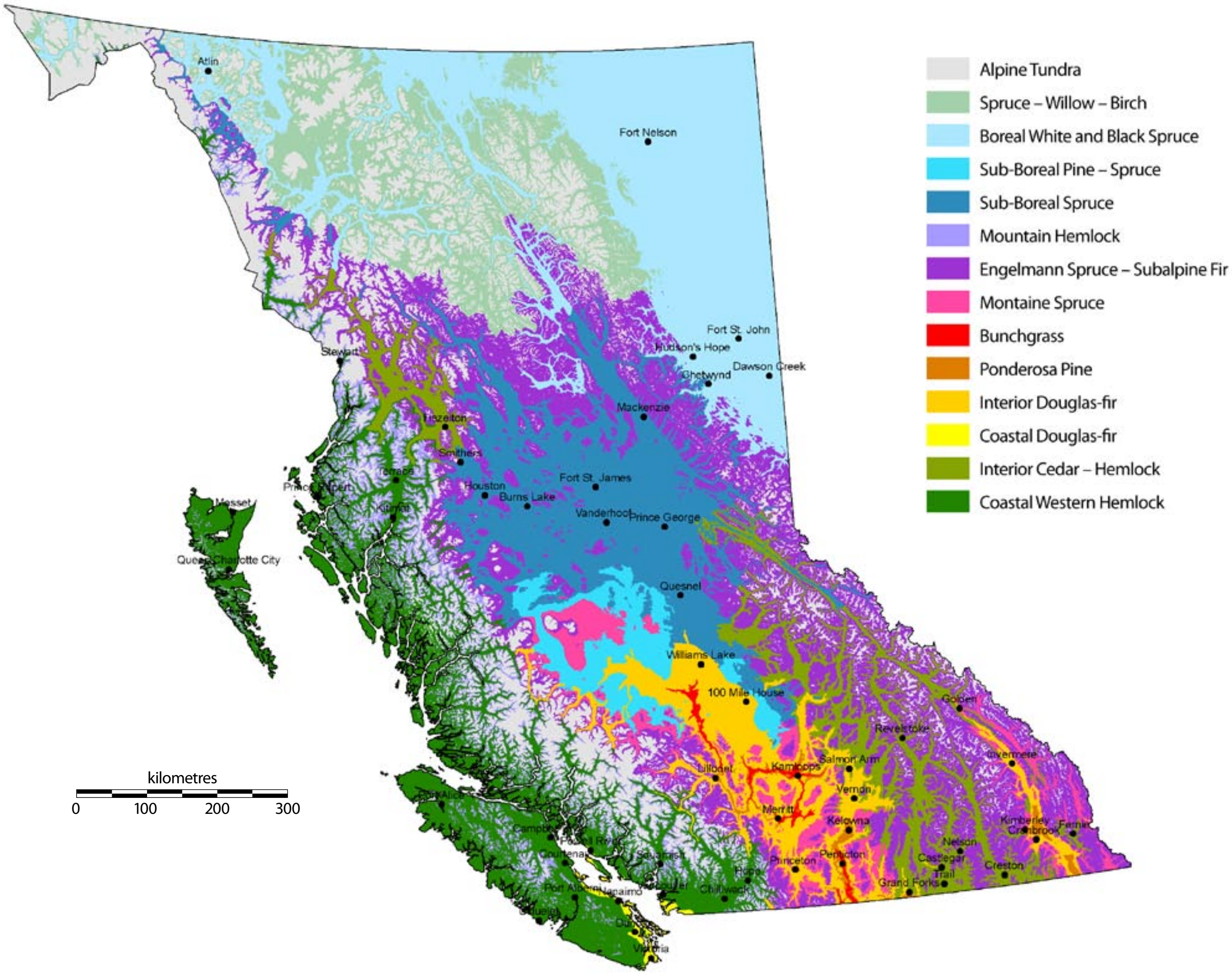

FIGURE 3.4 *Biogeoclimatic zones of British Columbia (B.C. Ministry of Forests and Range 2008).*

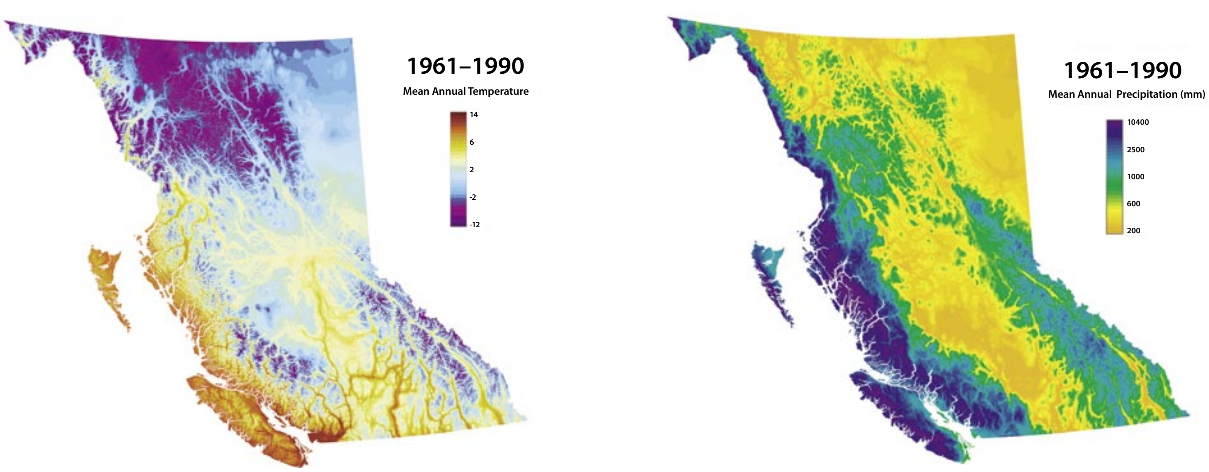

FIGURE 3.5 *Mean annual temperature and mean annual precipitation for the 1961–1990 period. Data interpolated using ClimateBC (Spittlehouse 2006, 2008; Wang et al. 2006).*

TABLE 3.1 *1961–1990 climate normals for zones of the biogeoclimatic ecosystem classification system. Climate variable and zone abbreviations are explained below the table. Data are based on information at ftp://ftp.for.gov.bc.ca/HRE/external/!publish/Climate/.*

Zone[a]	MAP[b] (mm)	MSP (mm)	PAS (mm)	MAT (°C)	MCMT (°C)	MWMT (°C)	xTmin (°C)	FFP (days)	DD<0	DD>5	SH:M
BAFA	1090	447	598	–2.6	–13.4	9.1	–44.6	15	2071	340	22
BG	342	161	100	6.1	–6.3	17.5	–35.8	118	575	1717	114
BWBS	514	308	178	–0.3	–16.0	14.3	–46.5	77	2090	1023	48
CDF	1091	201	61	9.6	3.0	16.9	–15.4	204	31	1965	88
CMA	3198	816	1795	–0.3	–9.7	9.6	–40.0	43	1364	440	14
CWH	2893	651	427	6.7	–0.4	14.5	–22.1	151	191	1339	27
ESSF	1096	404	566	0.3	–10.6	11.5	–41.8	51	1413	650	30
ICH	920	342	379	3.3	–8.4	14.7	–38.9	88	922	1152	45
IDF	493	210	178	4.0	–7.7	15.1	–38.6	84	813	1238	74
IMA	1539	473	959	–2.0	–11.3	8.4	–43.1	20	1791	301	19
MH	3119	730	1198	2.8	–5.7	12.0	–33.2	76	690	781	19
MS	648	261	292	1.9	–9.3	12.8	–40.8	62	1101	848	51
PP	382	165	112	6.3	–5.9	17.9	–35.2	120	517	1762	112
SBPS	473	228	191	1.7	–10.3	12.6	–43.2	35	1176	843	57
SBS	657	280	274	2.2	–10.3	13.6	–41.9	75	1169	988	19
SWB	691	352	322	–1.8	–13.9	10.9	–44.6	37	2038	525	12

a BAFA = Boreal Alti Fescue Alpine; BG = Bunchgrass; BWBS = Boreal Black and White Spruce; CDF = Coastal Douglas-fir; CMA = Coastal Mountain-heather Alpine; CWH = Coastal Western Hemlock; ESSF = Engelmann Spruce–Subalpine Fir; ICH = Interior Cedar–Hemlock; IDF = Interior Douglas-fir; IMA = Interior Mountain-heather Alpine; MH = Mountain Hemlock; MS = Montane Spruce; PP = Ponderosa Pine; SBPS = Sub-boreal Pine–Spruce; SBS = Sub-boreal Spruce; SWB = Spruce–Willow–Birch.

b MAP = mean annual precipitation; MSP = mean summer precipitation (May to September); PAS = precipitation as snow (water equivalent); MAT = mean annual temperature; MCMT = mean coldest month temperature (January); MWMT = mean warmest month temperature (July); xTmin = extreme minimum temperature; FFP = frost-free period; DD<0 = degree-days less than 0°C; DD>5 = degree-days greater than 5°C; SH:M = summer heat:moisture index.

TABLE 3.2 *One standard deviation on mean values of 1961–1990 climate normals for zones of the biogeoclimatic ecosytem classification system presented in Table 3.1. Climate variable and zone abbreviations are listed in Table 3.1.*

Zone	MAP (mm)	MSP (mm)	PAS (mm)	MAT (°C)	MCMT (°C)	MWMT (°C)	xTmin (°C)	FFP (days)	DD<0	DD>5	SH:M
BAFA	524	150	345	1.2	1.6	1.2	1.5	19	330	114	7
BG	30	19	12	0.5	0.5	0.6	1.1	9	66	134	16
BWBS	61	37	24	0.7	2.1	0.7	1.4	8	259	107	7
CDF	166	47	13	0.3	0.5	0.4	1.3	19	14	87	18
CMA	1252	397	737	2.2	3.5	1.9	5.0	32	554	218	7
CWH	785	186	205	0.9	1.4	0.9	3.2	23	88	181	8
ESSF	251	76	156	0.8	0.9	0.8	1.2	15	176	116	6
ICH	197	59	111	1.0	1.0	1.0	1.7	14	162	175	9
IDF	82	27	39	0.7	0.7	0.9	1.3	13	101	160	11
IMA	351	113	251	1.2	1.2	1.4	1.6	20	278	133	6
MH	888	234	446	1.3	2.0	1.3	3.8	28	254	192	7
MS	128	46	74	0.6	0.7	0.7	1.0	11	103	110	10
PP	43	17	19	0.7	0.6	0.8	1.3	10	78	161	15
SBPS	55	32	31	0.6	0.7	0.7	1.0	18	90	104	11
SBS	107	38	58	0.6	0.8	0.6	1.0	11	108	104	3
SWB	134	77	89	0.7	1.5	0.8	1.2	18	191	97	3

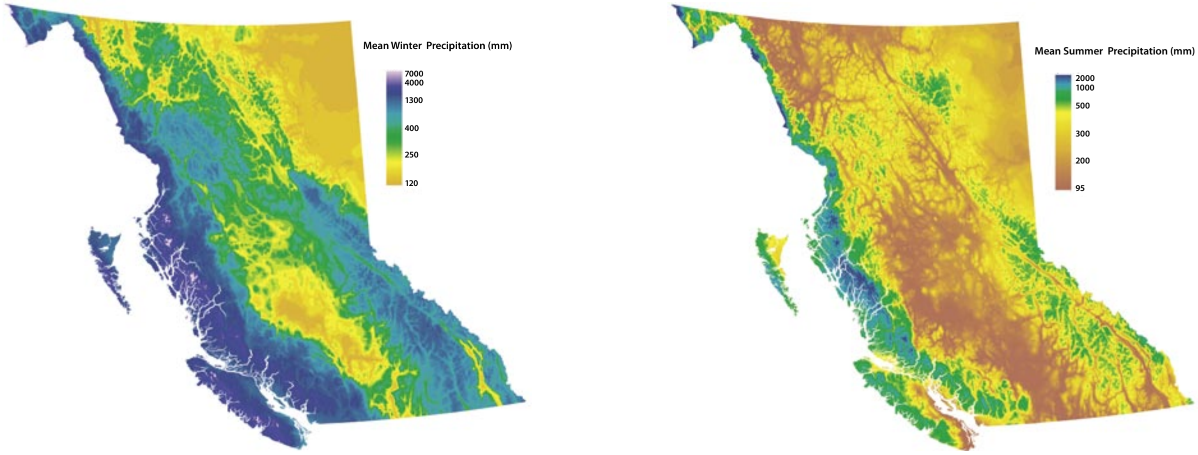

FIGURE 3.6 *Mean October to April and mean May to September precipitation for the 1961–1990 period. Data interpolated using ClimateBC (Spittlehouse 2006, 2008; Wang et al. 2006).*

High often dominates western North America, giving warm, clear weather to much of British Columbia, except where it generates convective cells and precipitation in the afternoon and early evening.

Air temperatures vary with distance from the coast and with latitude (Figures 3.5 and 3.7). Higher mean annual temperatures are found on the coast and inland in valley bottoms. Mean annual temperatures generally decrease to the north and (or) with increasing elevation. Summer temperatures are often higher in the Interior than on the Coast, with the reverse pattern holding for winter temperatures. The annual temperature range is greatest in the northern Interior of the province. Temperature variations from day to day tend to be greater in winter, and are largely controlled by the origin of the dominant air mass (e.g., arctic continental vs. maritime). Diurnal temperature ranges are generally greater in summer than winter, reflecting the seasonal variation in solar radiation.

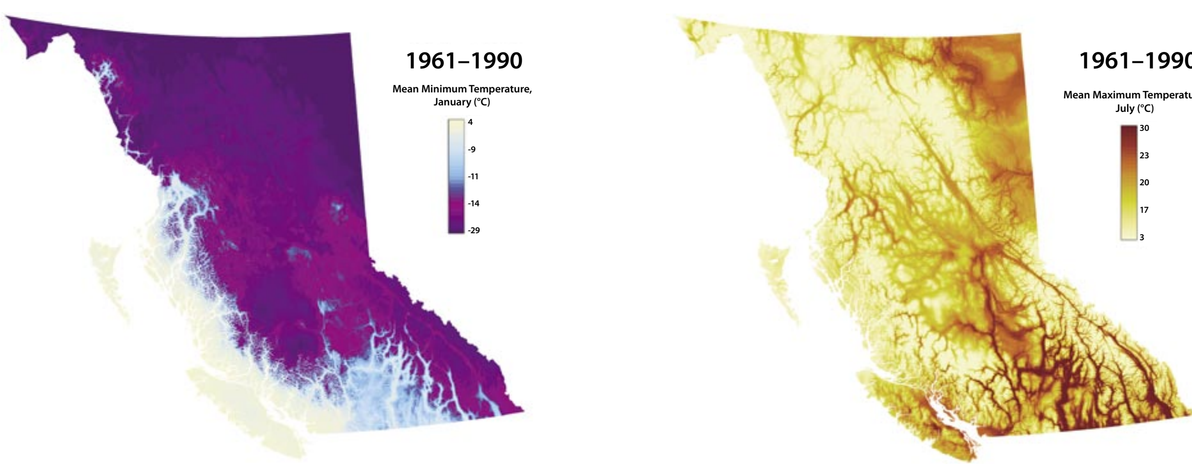

FIGURE 3.7 *Mean January minimum and mean July maximum air temperature for the 1961–1990 period. Data interpolated using ClimateBC (Spittlehouse 2006, 2008; Wang et al. 2006).*

Air temperature, precipitation, and other climatic elements vary seasonally because of the combined effects of varying solar radiation and weather-type frequency. For example, coastal stations are dominated by a winter-wet, summer-dry pattern because of the increased frequency of high-pressure systems and associated reduction in frontal and cyclonic weather (e.g., Figure 3.8; Nanaimo and Prince Rupert). In the lee of the Coast Mountains and throughout much of the Interior, precipitation exhibits weaker or little seasonality (e.g., Figure 3.8; Penticton, Upper Penticton Creek, Dease Lake, and Prince George).

Of particular importance to hydrogeomorphic processes is the climatic moisture regime, which reflects the balance between the precipitation falling on an area and the evaporative demand for water. The evaporative demand is a function of solar radiation, air temperature, humidity, and wind speed, and is represented by a monthly reference evaporation (E_{ref}). The climatic moisture deficit (CMD) equals the monthly E_{ref} minus the monthly precipitation; CMD is zero if precipitation exceeds E_{ref}, in which case there is a moisture surplus. During periods of high CMD, streamflow will typically drop to low levels, sustained by the discharge from surface storage (lakes, ponds, and wetlands) and groundwater. The exception to this situation occurs in catchments with more than about 5% glacier cover, where increased glacier melt during hot, dry weather can increase streamflow (Stahl and Moore 2006).

To illustrate variations in British Columbia's climatic moisture regimes, moisture deficits and surpluses were determined for eleven Meteorological Service of Canada weather stations with data from 1971 to 2000, one high-elevation research station (Winkler et al. 2004), two long-term weather station records for coastal British Columbia (Spittlehouse 2003, 2004), and three locations with interpolated values (1961–1990 normals). Calculations for other locations show that the shorter-period data yield values that are within 5% of those for the longer period. The calculated CMD and E_{ref} were then averaged by month over the calculation period. Note that the use of monthly averages, although common, obscures some details of the seasonal variations in climatic conditions. Site factors, such as the vegetation cover, topographic shading, soil characteristics, and location on a slope, may exacerbate or ameliorate climatic moisture conditions. The influence of site factors is discussed in Chapter 17 ("Watershed Measurement Methods and Data Limitations").

The reference evaporation rate (E_{ref}) was calculated with the Penman-Monteith equation for a 0.12 m tall grass surface (Allen et al. 1998) on a monthly time step from 1970 to 2005 (see footnote in Table 3.3 for details). The Thornthwaite equation (Thornthwaite 1948; Mather and Ambroziak 1986) was used to determine E_{ref} for stations where only temperature and precipitation data were available. A comparison of the Thornthwaite with the Penman-Monteith equation showed that it tended to overestimate the monthly evaporation and CMD, consistent with the analysis of Mather and Ambroziak (1986). In both methods, the monthly E_{ref} is zero if the monthly average temperature is less than or equal to zero. This approach, while appropriate for estimating soil moisture deficits, ignores sublimation and evaporation of intercepted snow on trees or from open areas. These processes are difficult to estimate, however. Values of E_{ref} are zero for large parts of the winter, but can be 0.5 mm/d for forests under favourable conditions. Events such as chinook (foehn) winds can produce high rates of loss from forests and open areas. In the southern Interior of British Columbia, winter interception loss of snow by forests is 20–60 mm depending on precipitation regime. Sublimation and evaporative losses from openings will be much lower than canopy interception losses in the forest. These processes are discussed in more detail in Chapter 7 ("The Effects of Forest Disturbance on Hydrologic Processes and Watershed Response").

The climatic moisture regimes are summarized in Figure 3.8 and Tables 3.3 and 3.4. As expected, warmer and drier environments have a higher evaporative demand and greater CMD. Cool, wet coastal locations, such as Prince Rupert, show no deficit in most months and only in some years. The cloudy environment reduces solar radiation and the ocean moderates summer temperature such that, although evaporation can take place over most of the year, there is a lower E_{ref} than in the cold areas, such as Dease Lake and Fort St. John. Dry interior sites, such as Penticton, Kamloops, and Cranbrook, have high E_{ref} that, combined with a low summer precipitation, results in high CMD. Penticton and Upper Penticton Creeks (Figure 3.8) show the influence of elevation. The lower air temperature and slightly reduced

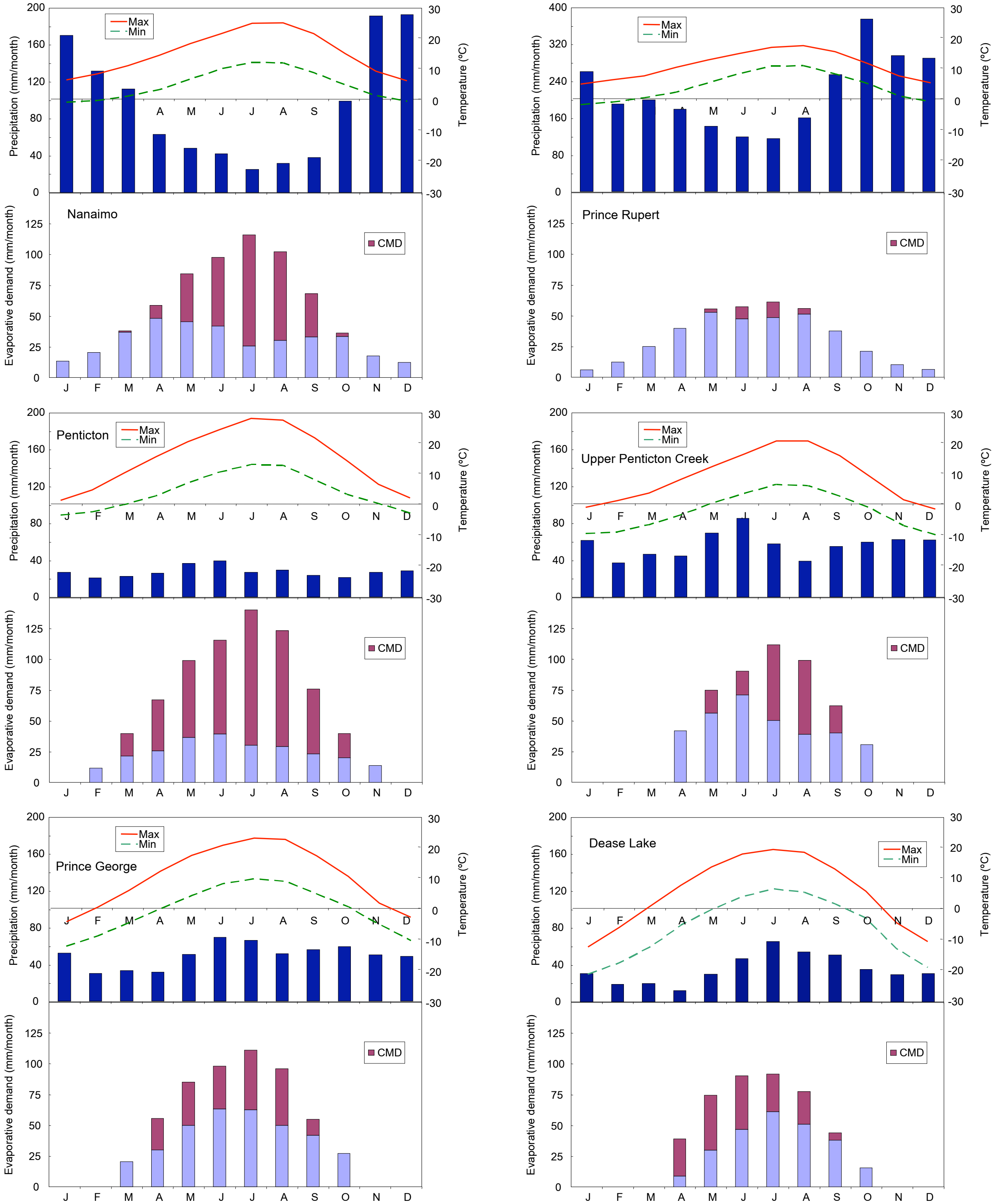

FIGURE 3.8 *Climate diagrams for Nanaimo, Prince Rupert, Penticton, Upper Penticton Creek, Prince George, and Dease Lake. Data are averages for 1970–2005 except Upper Penticton Creek, which is for 1992–2005. The upper panel of each station shows the mean monthly maximum (red solid line) and minimum (green dashed line) air temperature and the monthly precipitation (dark blue bars). The precipitation scale for Prince Rupert is double that of the other locations. The lower panel shows the climatic moisture deficit (purple bar) and the reference evaporation (light blue plus purple bar). See Tables 3.3 and 3.4 for more details on each location.*

solar radiation result in a lower E_{ref} that, combined with high precipitation, reduces the CMD at Upper Penticton Creek. The higher-elevation site also has a shorter season in which evaporation can occur.

Variation in E_{ref} between years is usually about 5%; however, variations in precipitation are much greater and result in the monthly CMD varying by up to 100% (Table 3.3). Some of the early and late months in the year have a small CMD even though the average precipitation is greater than the average E_{ref}. This is because the variability in annual precipitation results in a deficit in some years and not in

TABLE 3.3 *Reference evaporation (E_{ref}) and climatic moisture deficit (CMD) for selected locations and the biogeoclimatic ecosystem classification (BEC) zone in which they occur. SD is standard deviation of the average. Data are calculated using monthly data from 1971 to 2000, or monthly data from 1992 to 2004. For an explanation of BEC zone abbreviations, see Table 3.1.*

Location[a]	BEC zone	Latitude	Longitude	Elevation (m)	E_{ref}[b] (mm)	SD (mm)	CMD (mm)	SD (mm)
Cranbrook	IDF	49°37’	115°47’	939	680	20	422	68
Dease Lake	BWBS	58°25’	130°00’	816	433	25	182	54
Fort St John A	BWBS	56°14’	120°44’	695	487	36	199	70
Kamloops A	BG	50°42’	120°27’	345	735	30	531	68
Nanaimo A	CDF	49°03’	123°52’	30	664	28	307	62
Penticton A	BG	49°28’	119°36’	344	732	23	479	66
Upper Penticton Creek	ESSF	49°39’	119°24’	1620	514	29	191	74
Prince George A	SBS	53°53’	122°41’	676	551	27	209	72
Prince Rupert A	CWH	54°18’	130°26’	34	388	18	9	16
Smithers A	SBS	54°49’	127°11’	523	503	31	233	65
Tofino	CWH	49°05’	125°46’	24	522	19	62	43
Victoria A	CDF	49°39’	123°26’	20	633	24	336	61

a Locations with an “A” following the name indicate measurements made at airports

b E_{ref} was calculated using the Penman-Monteith equation for a 0.12 m tall grass surface with a solar radiation reflectivity of 0.23, and a stomatal resistance of 70 s/m (Allen et al. 1998). The weather data used are monthly sunshine hours or solar radiation, mean maximum and minimum air temperature, and total precipitation. Mean wind speed at 2 m is assumed constant at 2 m/s (this assumption has little effect on the calculations). Procedures for converting sunshine to solar radiation, and solar radiation and air temperature data to net radiation, can be found in Allen et al. (1998) and Spittlehouse (2003, 2004). Using solar radiation instead of sunshine hours gave virtually the same values for E_{ref}. Normals were used for missing monthly sunshine hours. E_{ref} calculated using normals agrees with to within 5% of the average E_{ref} determined annually and then averaged over the period of the normals.

TABLE 3.4 *Averages for 1970–2005 of annual values for weather stations listed in Table 3.3. Precipitation (PPT) as snow is for 1961–1990 for all stations except Upper Penticton Creek, which is 1992–2005.*

Location[a]	Solar radiation (MJ/m^2)	Maximum T_{air} (°C)	Minimum T_{air} (°C)	Average T_{air} (°C)	Annual PPT (mm)	May-Sept PPT (mm)	PPT as snow (mm)
Cranbrook A	4712	11.7	–0.1	5.8	382	199	148
Dease Lake A	3474	5.0	–6.5	–0.7	426	248	227
Fort St John A	3948	6.9	–2.8	2.1	463	295	198
Kamloops A	4490	14.5	3.4	9.0	279	144	86
Nanaimo A	4255	14.8	4.9	9.9	1147	186	93
Penticton A	4386	14.7	3.7	9.2	327	155	73
Upper Penticton Creek	4438	7.4	–3.6	1.9	700	305	320
Prince George A	3942	9.4	–1.4	4.0	597	293	234
Prince Rupert A	3005	10.5	3.8	7.2	2576	789	143
Smithers A	3598	9.2	–1.2	4.0	510	228	216
Tofino A	3994	12.9	5.4	9.2	3230	600	53
Victoria A	4581	14.2	5.4	9.8	876	141	47

a Locations with an “A” following the name indicate measurements made at airports other years. Calculating E_{ref} and CMD using climate normals misses this variability and also tends to underestimate the mid-summer values. The bulk of the CMD occurs during summer. The annual variation in E_{ref} and CMD for May to September is shown in Figure 3.9 for Victoria Airport. In this warm, sunny, and dry environment, E_{ref} and CMD are high. Annual variability is mainly driven by the variation in precipitation. Two of the wettest and driest summers in the 35-year record have occurred in the last decade.

Figure 3.10 shows 105 years of climatic moisture deficits for the Campbell River Airport on the east coast of Vancouver Island (Spittlehouse 2004). The CMD varied from less than 100 to over 350 mm. In the upper panel of Figure 3.10, higher than average CMD years are indicated by a positive value and lower than average by a negative value. A period with extremely high CMD occurred between the 1930s and 1960. Occasionally, there are two consecutive very dry years (e.g., 1951 and 1952; 2002 and 2003). How this influences the vegetation will depend on the soil type and soil depth (Spittlehouse 2003). The between-year variation in CMD for Campbell River is similar to that for Victoria (Figure 3.9) because both are coastal sites on the lee side of Vancouver Island, about 250 km apart. The main difference is the higher CMD at Victoria, but differences in dryness between years also occur.

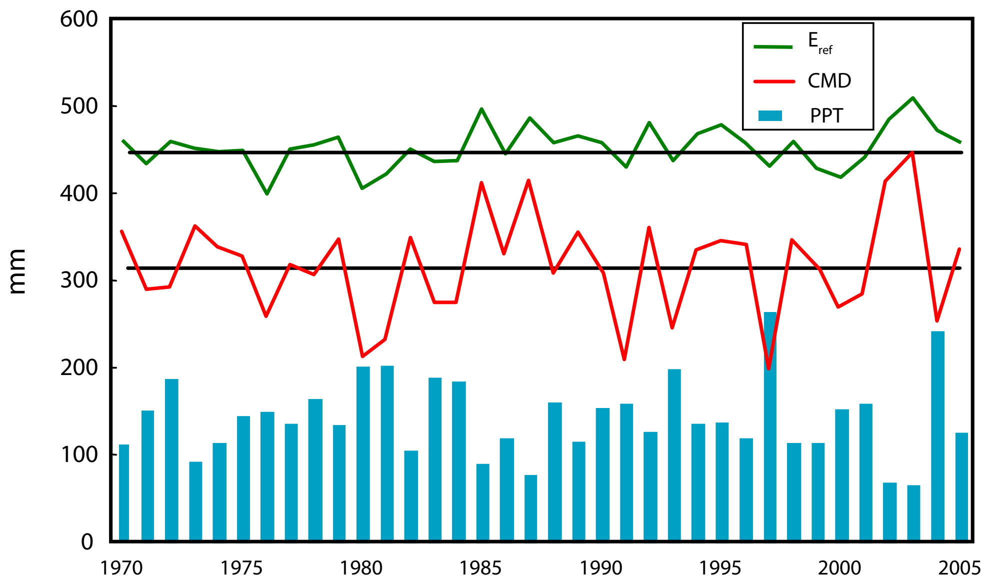

FIGURE 3.9 *May to September reference evaporation (green solid line), climatic moisture deficit (red dashed line), and total precipitation (blue bar) for Victoria Airport 1970–2005. The solid black lines show the mean for the period for E_{ref} and CMD. More information on Victoria can be found in Tables 3.3 and 3.4. E_{ref} was calculated using the Penman-Monteith equation.*

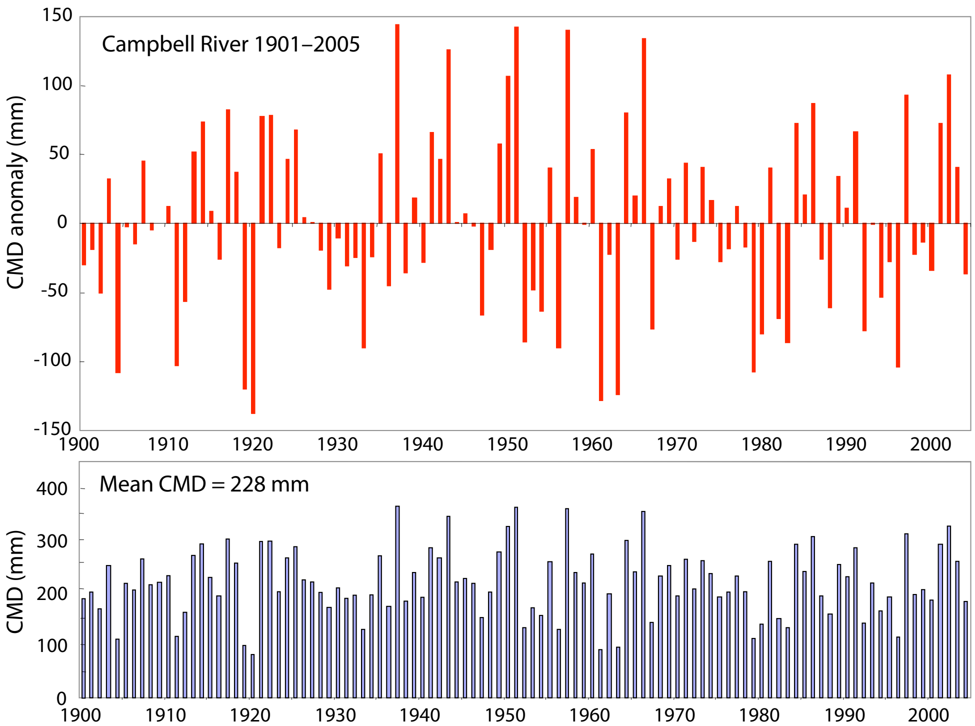

FIGURE 3.10 *105 years of climate moisture deficits for the Campbell River airport. The lower panel shows the May to October climatic moisture deficit (CMD, mm) for Campbell River Airport for 1901–2005. The upper panel shows the between-year variation in CMD (anomaly in mm) obtained by subtracting the period mean CMD (228 mm) from the annual values. Negative values indicate that the CMD was lower (wetter season) than the average and positive ones higher (drier season) than average.* E_{ref} *was calculated using the Penman-Monteith equation.*

INFLUENCE OF ELEVATION

Temperature tends to decrease with increasing elevation while precipitation increases; however, the vertical gradients vary in both space and time (e.g., Stahl et al. 2006c). Particularly in coastal regions, the variation of air temperature with elevation can influence whether precipitation falls as rain or snow. The relative influence of snow accumulation and melt on streamflow in coastal regions tends to increase with increasing basin elevation. For more details, see Chapter 4 ("Regional Hydrology").

Evaporative demand tends to decrease with elevation, which, combined with the positive relation between precipitation and elevation, results in a decrease in CMD with increasing elevation. This elevational dependence is illustrated with data for the east coast of Vancouver Island. Because of the lack of high-elevation weather stations in this area, temperature and precipitation data were determined from the high spatial resolution interpolation of climate data for British Columbia (Spittlehouse 2006; Wang et al. 2006). Mean monthly temperature and precipitation were determined at four separate elevations on the east side of central Vancouver Island: 100, 600, 1100, and 1800 m. The 100-m values are equivalent to those for Campbell River Airport (106 m). Evaporative demand and CMD were calculated for each of the four elevations and summed to give annual totals (Table 3.5). Evaporative demand and CMD decrease with elevation as a function of a decrease in temperature and increase in precipitation. The high precipitation in coastal British Columbia from October through April restricts the CMD to summer. A CMD occurs from May to September at 100 m of elevation, June to August at 600 and 1100 m, and July and August at 1800 m.

TABLE 3.5 *Evaporative demand (E_{ref}), climatic moisture deficit (CMD), mean annual temperature (MAT), and precipitation (MAP) at four elevations on the east coast of Vancouver Island near Campbell River. Evaporative demand for 1800 m is restricted to June to September because snow would be covering the vegetation outside of this period. Evaporation is assumed to occur in forests when the air temperature is above 0 C° even though there is snow on the ground below the canopy. Evaporative demand was calculated using temperature data and the Thornthwaite method (Thornthwaite 1948; Mather and Ambroziak 1986).*

Elevation (m)	100	600	1100	1800
E_{ref} (mm)	605	555	495	340
CMD (mm)	200	135	95	55
MAT (°C)	8.4	6.9	5.1	2.0
MAP (mm)	1450	1860	2180	2590

INFLUENCE OF SYNOPTIC-SCALE CIRCULATION

The synoptic-scale circulation types are characterized by distinctive surface climate anomalies, which can vary seasonally. Figure 3.11 illustrates these anomalies for selected synoptic types. A positive anomaly indicates that conditions associated with a given synoptic type are wetter (for precipitation) or warmer (for air temperature) than the overall average for a given location. In Type 1, a ridge of high pressure results in relatively dry conditions over southern British Columbia, while northern British Columbia tends to be warmer than average because of the advection of maritime air via the southwesterly Pacific airflow. Type 6 is typical of an arctic outbreak, in which a high-pressure system extends south from the Yukon, accompanied by the movement of cold, dry arctic air over most of the province. The cold, dense air is confined by the topographic barriers of the Coast Mountains and Rocky Mountains, with cold air drainage through deep coastal valleys resulting in cold, windy conditions at coastal sites. Temperature anomalies associated with Type 6 are strongly negative throughout the province, and precipitation also tends to be lower than average. Type 12 is dominated by a low-pressure cell in the Gulf of Alaska, which brings a strong southwesterly flow of warm moist air over southern British Columbia, resulting in warm, wet conditions. The source of the air mass is the tropical Pacific Ocean, and extreme versions of this pattern are called a "pineapple express" or "tropical punch." Type 13 has a low-pressure cell located southeast of the Gulf of Alaska, in conjunction with a ridge of high pressure extending into British Columbia from the south. This pattern results in strong southerly airflow along the Coast, with generally warmer than average conditions throughout most of British Columbia. The southerly airflow brings wet conditions to coastal areas, but the high-pressure ridge suppresses precipitation in the Interior.

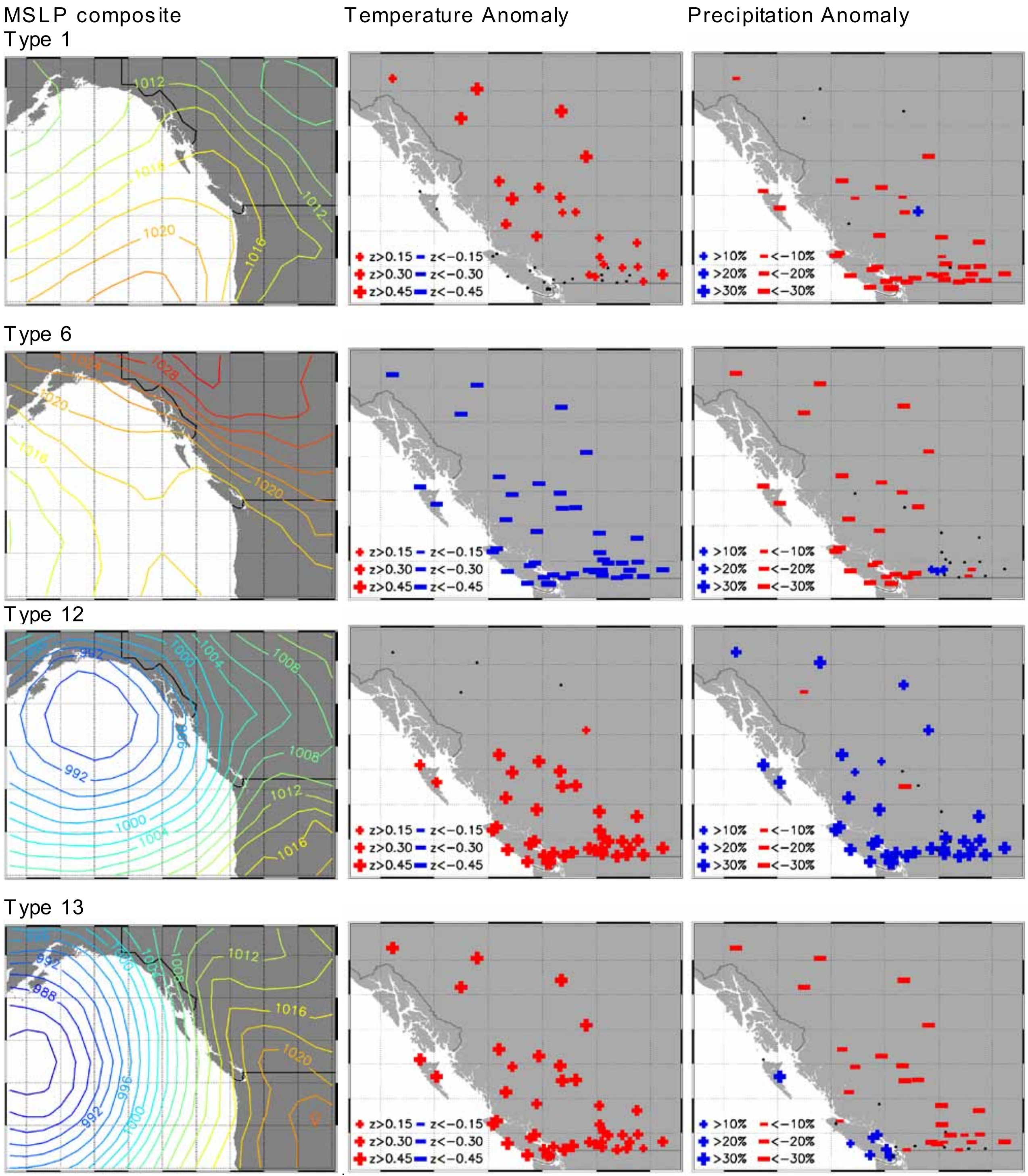

FIGURE 3.11 *Winter (December to February) surface climate anomalies associated with selected synoptic types. (Adapted from Stahl et al. 2006b)*

Sea surface temperatures and atmospheric circulation patterns over the Pacific Ocean strongly influence the weather and climate of British Columbia. These large-scale phenomena are often called "teleconnections," in reference to the correlation of climatic phenomena over large distances. For example, sea surface temperatures in the equatorial Pacific Ocean influence weather and climate throughout large parts of North America, including much of western Canada. The dominant modes of atmosphere–ocean variability relevant to British Columbia include El Niño–Southern Oscillation (ENSO), Pacific Decadal Oscillation (PDO), Pacific North American (PNA) Pattern, and Arctic Oscillation (AO).

El Niño–Southern Oscillation

El Niño events (also known as ENSO warm phase) are characterized by a warming of the equatorial ocean surface off the coast of South America. La Niña conditions (ENSO cool phase) involve upwelling of cold water in the same region (Figure 3.12). The ENSO is a coupled phenomenon in which the sea surface temperatures in the tropical Pacific set up the atmospheric circulation and surface winds, which in turn determine the sea surface temperatures, so that positive feedbacks are strong. However, changes below the ocean surface, and the slow response of the ocean off the equator, provide a delayed signal that can reverse the changes and cause the oscillation from El Niño to La Niña (Trenberth 1997). El Niño winters favour a split in the jet stream in the mid-Pacific to produce both low-latitude and high-latitude storm tracks that avoid southern British Columbia (Shabbar et al. 1997). In British Columbia, winters following the onset of an El Niño event are generally warmer and drier than normal, and La Niña winters are generally cooler and wetter (Shabbar and Khandekar 1996; Shabbar et al. 1997; Stahl et al. 2006b) (Figure 3.13).

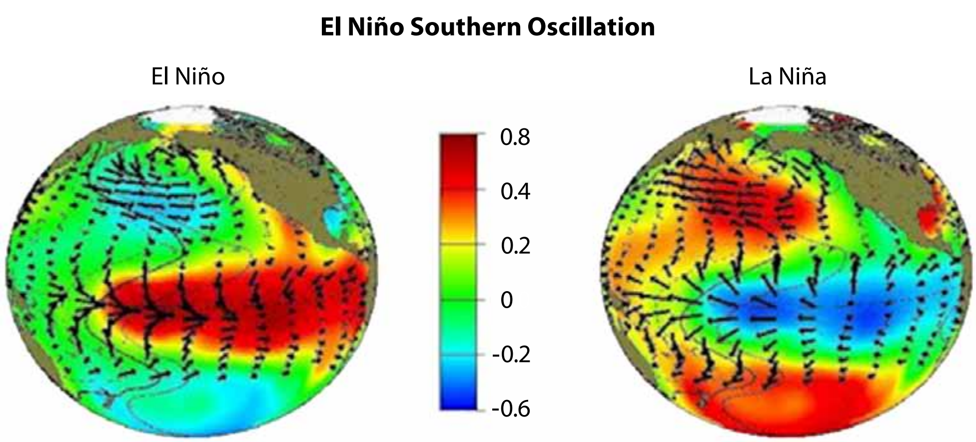

FIGURE 3.12 *Sea surface temperature and surface wind stress anomalies (differences from long-term average) during El Niño and La Niña conditions. Colours indicate magnitude of temperature anomalies; length and direction of arrows indicate wind stress magnitude and direction. (www.wrh.noaa.gov/fgz/science/pdo.php)*

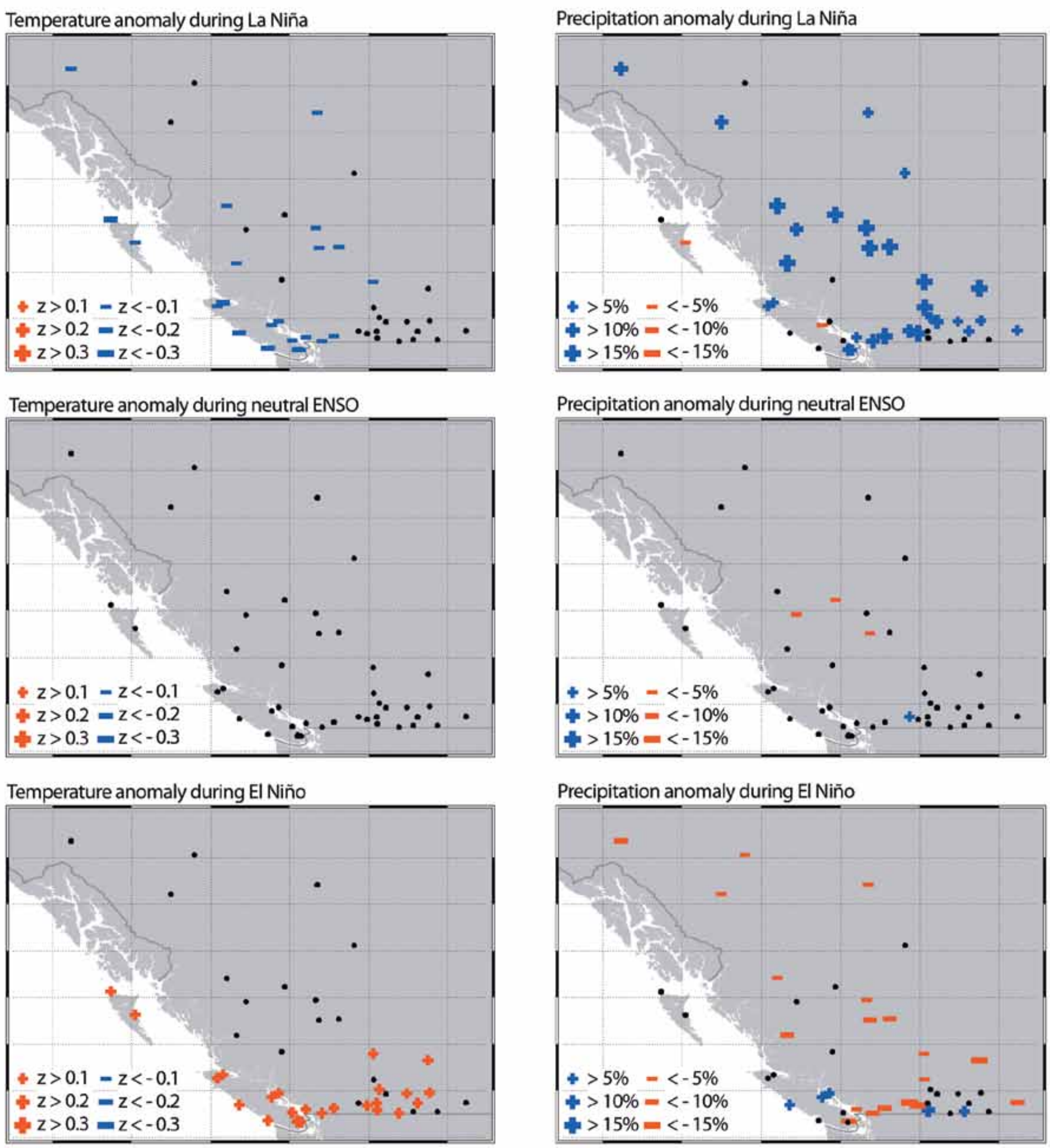

FIGURE 3.13 *Winter (December to February) temperature and precipitation anomalies associated with ENSO cool (La Niña), neutral, and warm (El Niño) phases. Temperature anomalies are z-scores (i.e., standardized by dividing by the standard deviation of the anomalies); precipitation anomalies are percentage of the long-term average (adapted from Stahl et al. 2006b).*

Pacific Decadal Oscillation

The Pacific Decadal Oscillation (PDO) involves shifts between two dominant patterns of sea surface temperatures in the North Pacific Ocean (Mantua et al. 1997). The warm (positive) phase of the PDO is characterized by below-normal sea surface temperatures in the central and western north Pacific and unusually high ones along the west coast of North America (Figure 3.14). The PDO cold (negative) phase produces the reverse distribution. Positive PDO phases are associated with positive winter temperature anomalies throughout western Canada and with negative precipitation anomalies in the mountains and interior, which also reduce the snowpack (Figure 3.15). Temperature impacts over Canada tend to be

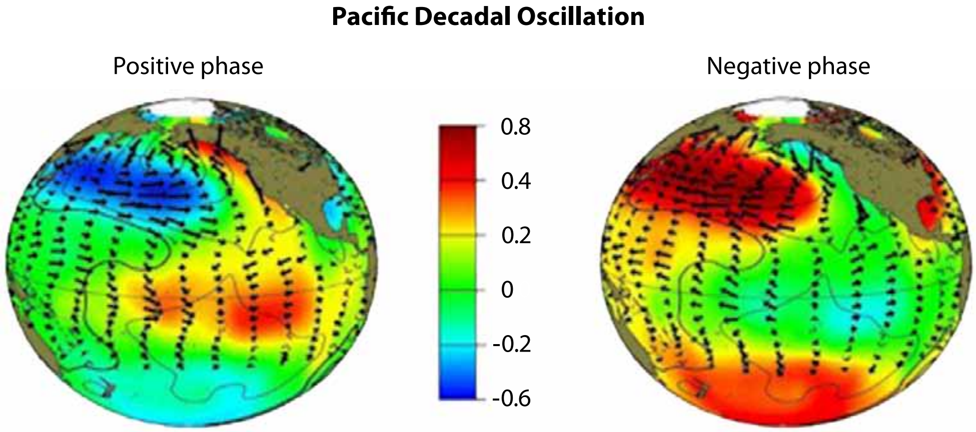

FIGURE 3.14 *Sea surface temperature and surface wind stress anomalies (differences from long-term average) during positive and negative phases of the PDO. Colours indicate magnitude of temperature anomalies; length and direction of arrows indicate wind speed and direction. (www.wrh.noaa.gov/fgz/science/pdo.php)*

enhanced when ENSO and PDO are in phase (Bonsal et al. 2001a). El Niño and positive PDO winters are characterized by a deepening of the Aleutian Low and an enhanced high-pressure ridge over the Rocky Mountains. This situation represents a positive Pacific North American pattern and indicates a northward displacement of the polar jet stream, which inhibits the outflow of cold arctic air, and thus accounts for the positive temperature anomalies (Bonsal et al. 2001a). During negative PDO phases, the jet stream is displaced southward, which results in more frequent arctic outflow events and hence lower temperatures.

The PDO is often described as a long-lived El Niño–like pattern of Pacific climate variability, but two main characteristics distinguish the PDO from the ENSO: (1) 20th-century PDO "events" persisted for 20–30 years, while typical ENSO events persist for 6–18 months; and (2) the climatic fingerprints of the PDO are most visible in the north Pacific / North American sector, while secondary signatures exist in the tropics. The opposite is true for ENSO (Mantua et al. 1997). Only two full PDO cycles appear to have occurred in the past century: "cool" PDO regimes prevailed from 1890 to 1924 and again from 1947 to 1976, while "warm" PDO regimes dominated from 1925 to 1946 and from 1977 through (at least) the mid-1990s (Hare and Mantua 2000). In British Columbia, PDO cool and warm phases are associated with temperature and precipitation anomalies that are similar to the corresponding ENSO phases (Figures 3.13 and 3.15) (Fleming 2006; Fleming et al. 2007). The PDO is not fully understood, and it is not possible to predict when shifts will occur. Despite this limitation, the state of the PDO may be used to improve season-to-season and year-to-year climate forecasts for North America because of its strong tendency for multi-season and multi-year persistence. Recognition of the PDO is important because it shows that "normal" climate conditions can vary over time periods comparable to the length of a human's lifetime.

Pacific North American Pattern

The Pacific North American (PNA) pattern is a natural, internal mode of atmospheric circulation variability over the North Pacific and North America. The strong or enhanced (positive) phase is characterized by a strong Aleutian Low, with southerly airflow along the west coast of North America and a ridge of high pressure over the Rocky Mountains. The weak (negative) phase is dominated by a weaker Aleutian Low and westerly, zonal flow. The strong phase tends to coincide with warmer winters and reduced snowpack accumulation throughout British Columbia, especially in the southern Coast Mountains (Moore 1996; Moore and McKendry 1996). Although both enhanced and weak PNA patterns can occur under any sea surface temperature conditions, positive values of the PNA pattern tend to be associated with the warm (positive) phase of PDO and El Niño events (Hsieh and Tang 2001).

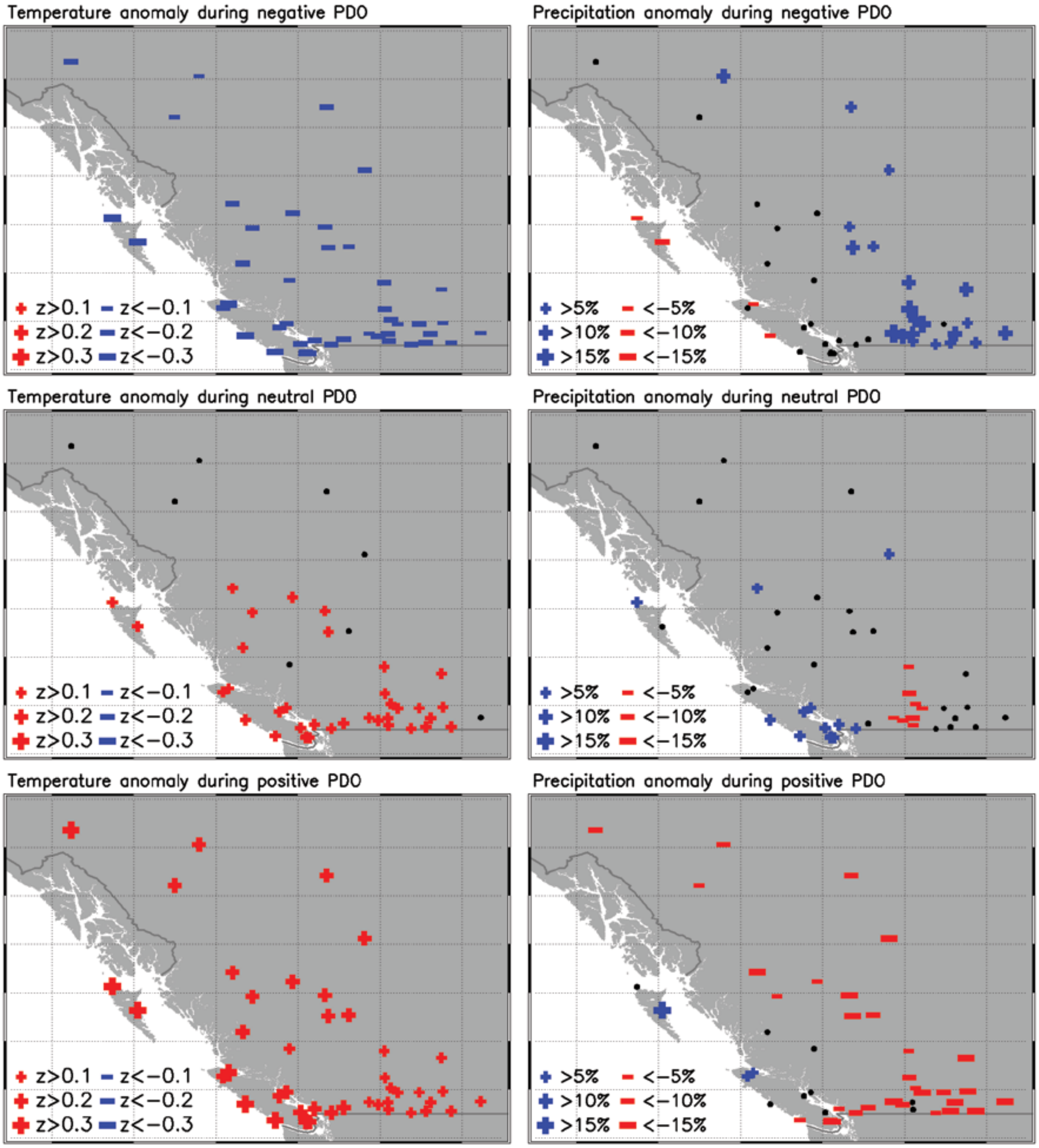

FIGURE 3.15 *Winter (December to February) temperature and precipitation anomalies associated with PDO cool, neutral, and warm phases (after Stahl et al. 2006b). Temperature anomalies are z-scores (i.e., standardized by dividing by the standard deviation of the anomalies); precipitation anomalies are percentage of the long-term average.*

Variations in climatic patterns associated with the ENSO, PDO, and PNA tend to be strongest in winter; however, some effects, such as increased air temperature associated with ENSO/PDO warm phase, can extend into spring and early summer (Fleming 2006; Fleming et al. 2007).

Arctic Oscillation

The Arctic Oscillation (AO) is associated with fluctuations in the strength of the winter stratospheric polar jet. A positive AO index value generally indicates negative and positive sea-level pressure anomalies in the Arctic and mid-latitudes, respectively, and relatively strong 55°N (surface) westerlies; a negative index value indicates the opposite pressure anomalies and weaker westerly flow (Thompson and Wallace 1998). Some evidence indicates that the negative AO phase is linked to the occurrence of extreme cold-weather systems (arctic outbreaks) that can cause mountain pine beetle mortality. For example, the arctic outbreak events in autumn 1984 and autumn 1985 appear to have ended the Chilcotin beetle outbreak that began in the late 1970s (Stahl et al. 2006a).

The synoptic climatology presented earlier provides an intuitive framework for understanding the influence of large-scale teleconnections on the surface climate of British Columbia. For example, synoptic types associated with warmer winter weather, such as types 10, 12, and 13, tend to be more frequent in warm-phase ENSO and PDO conditions (Stahl et al. 2006b). In addition, for some synoptic types, weather conditions at some stations exhibit systematic within-type variability among phases. For example, in warm-phase PDO years, not only are synoptic types associated with arctic outbreaks less frequent, but the minimum air temperatures tend not to be so extreme even when those types do occur. The combination of these effects suggests that air temperatures low enough to cause cold-mortality of the mountain pine beetle are less likely to occur in warm-phase PDO and warm-phase ENSO years (Stahl et al. 2006a).

EXTREME EVENTS

The hydrogeomorphology of catchments is affected not only by the regional climate, which represents the collective effect of multiple weather systems, but also by singular events, such as rainfall or snowmelt events of sufficient intensity and duration to generate floods, slope failures, or widespread surface erosion. Along the coast, hydrogeomorphically significant events often occur under synoptic situations similar to Type 12 (Figures 3.1 and 3.16). The southwesterly airflow advects large quantities of warm, humid air from the tropical Pacific region, creating intense rainfall along the entire coast, which can also penetrate inland to some extent. Such events have been called a "pineapple express" or "tropical punch." For example, a "pineapple express" in October 2003 generated significant flooding and landslide activity all along the Coast Mountains (Jakob et al. 2006).[1]

Extreme winds often accompany extreme precipitation events. Wet soils result in reduced soil-root adhesion and soil shear strength and, combined with extended periods of very strong winds, can cause extensive damage through blowdown of trees, which can then trigger landslides. Snow and ice on the crown also increase the risk of stem breakage and windthrow (Stathers et al. 1994). Extreme winds are associated with the passage of frontal storms originating in the Pacific Ocean, and occur most frequently in the winter. Strong winds can also accompany summer thunderstorms, which are more likely to occur in the Interior than on the Coast. Local terrain can direct airflow and increase wind speed; areas where valleys converge can experience extremely strong winds. Cutblock boundaries such as the edge of riparian reserves can be highly susceptible to blowdown.

The role of antecedent conditions complicates the definition of hydroclimatic thresholds for landslide initiation or flooding. For example, rainfall events are more likely to generate landslides and floods when soils are already moist (Toews 1991; Jakob and Weatherly 2003). In addition, "rain-on-snow" events occur when heavy rain in conjunction with warm air occurs in the presence of accumulated snow on the ground. Melting of the snowpack can augment rainfall and generate widespread landslides and (or) flooding. Egginton (2005) found that a range of conditions generated major landslides in northern British Columbia, including large cyclonic events and convective rainfall during summer. Mass-wasting events are often generated by intense, localized cells of rainfall that are not recorded by a climate station (e.g., Church and Miles 1987). Egginton (2005) found satellite images useful to help identify locations of heavy rainfall in the interpretation of landslide triggers in northern British Columbia. Landslides and their causal factors are discussed in detail in Chap-

1 McCollor, D. 2003. Summary of weather for the heavy rain event of 15 October through 23 October, 2003. BC Hydro, 27 Oct. 2003. Unpubl. internal memo.

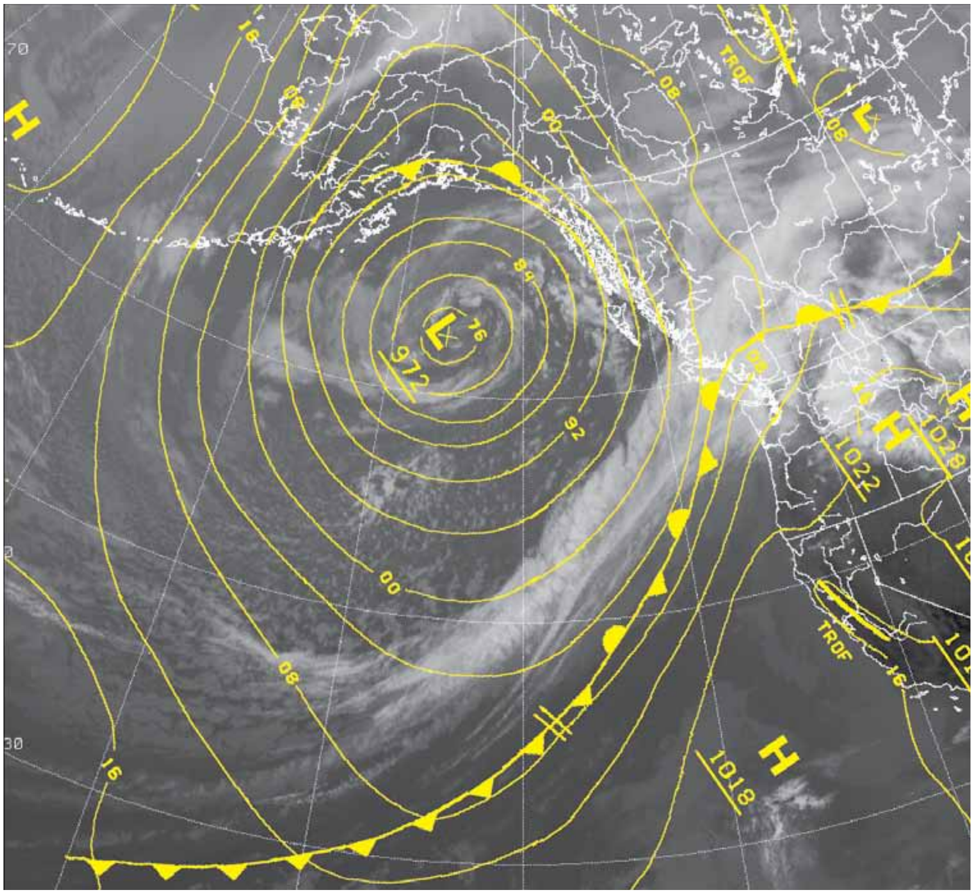

FIGURE 3.16 *Depictions of the October 2003 "pineapple express" event that caused widespread flooding and mass-wasting events throughout coastal British Columbia. Weather map showing sea-level pressure isobars and fronts has been overlaid on a GOES satellite image showing pattern of water vapour. (Figure by Eric Leinberger)*

ter 8, "Hillslope Processes," and Chapter 9, "Forest Management Effects on Hillslope Processes."

Droughts can be considered another form of extreme event. Extended periods of well below normal precipitation, either as winter snow or summer rain, can result in low streamflow with implications for water supply and fish habitat. Low summer flows and high summer air temperatures can produce high stream temperatures (Moore et al. 2005). Extremely high summer temperatures along with low precipitation also increase the risk of forest fires. This, in turn, has implications for water quality and streamflow through the increased risk of surface erosion and landslides because of the loss of ground cover, hydrophobicity of soils, and the effects on snow accumulation and melt.

Finally, extreme minimum temperatures, or the lack of them, can have indirect hydrologic implications. The prime example in British Columbia is the current mountain pine beetle infestation, which is partially caused by a lack of cold conditions late in the fall that would kill the insect. Stahl et al. (2006a) showed that mild winters and a decreased frequency of cold mortality for the pine beetles are associated with El Niño and PDO warm-phase conditions, but a decrease in extreme winter cold weather is also consistent with projected effects of climate warming. The reduction in transpiration and precipitation interception associated with tree mortality can increase soil moisture and water to streams (Hélie et al. 2005).

Climatic Variability and Change: Nature and Detection

Climatic variability and change occur on a range of time scales. In addition to apparently random variability from year to year, systematic variations relate to phenomena such as the ENSO and PDO because of their coherent influences on air temperature and precipitation patterns (Figures 3.13 and 3.15). For example, there was a notable shift from dominantly negative PDO/PNA from about 1947 to 1976, to dominantly positive PDO/PNA from 1977 into the late 1990s (Figure 3.17). This apparent step change shows up in many hydroclimatic records in southern British Columbia, including surface climate and snow accumulation (Moore and McKendry 1996; Fleming et al. 2007), streamflow (Moore 1991, 1996; Fleming et al. 2007), and glacier mass balance (Moore and Demuth 2001).

Of increasing concern are climatic changes over longer time scales and the attendant consequences for water resources. Scientific consensus has been building over the last two decades that global climate is currently undergoing an accelerated warming, largely caused by anthropogenic greenhouse gas emissions (Intergovernmental Panel on Climate Change 2007). Concern is growing about the effects of global change on regional and local climates and the consequences of these effects for terrestrial ecosystems, hydrology, and freshwater ecosystems. Climate change and the effects on watershed processes are discussed in detail in Chapter 19 ("Climate Change Effects on Watershed Processes in British Columbia").

The last major ice age ended roughly 10 000 years ago, and since then the global climate has experienced both warm and cool periods (Rosenberg et al. 2004). The last global cool period, the "Little Ice Age," ended in the mid-19th century. Although instrumental climate records (e.g., temperature and precipitation measurements) in British Columbia mainly cover the most recent century or so, the Little Ice Age is recorded in proxy climate records, such as tree rings and glacier terminal moraines,

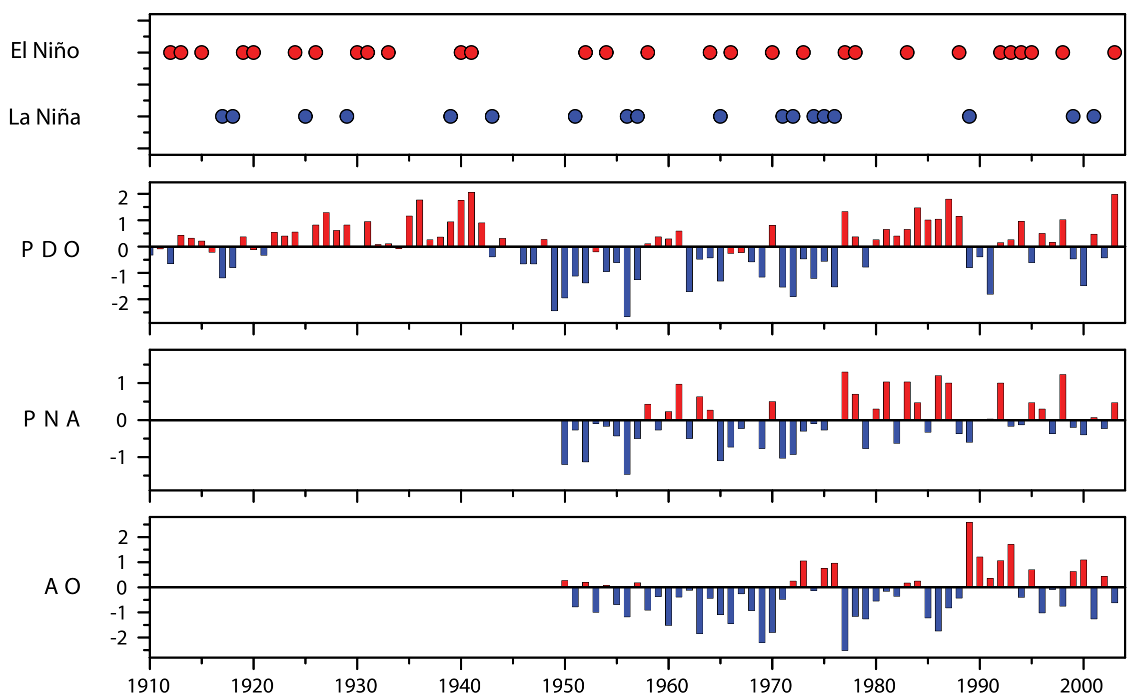

FIGURE 3.17 *Time series of large-scale climatic indices. (www.wrcc.dri.edu/enso/ensodef.html)*

which clearly show that significant glacier retreat has occurred (B.C. Ministry of Water, Land and Air Protection 2002; also see Chapter 2). Substantial effort has identified recent trends in the instrumental climate record. Work has also focussed on the use of general circulation models (also increasingly known as "global climate models") to simulate future climate under different assumptions regarding greenhouse gas emissions and land use (Intergovernmental Panel on Climate Change 2007).

Resolving climatic changes within the instrumental climate record requires data at a temporal resolution that is appropriate to the spatial domain of interest (Zwiers and Zhang 2003). At a global scale, it may be appropriate to examine time series of annual mean temperatures. Annual means, however, are not as relevant when the interest is in regional impacts, where changes at seasonal and sub-seasonal scales can be important for ecological and hydrological processes. For example, even monthly data may be too coarse to detect changes in air temperature that can influence the onset and timing of snowmelt (Whitfield 2001).

Trend detection is strongly affected by the length of the record, the magnitude of the trend relative to inherent between-year variability, and the window of data relative to decadal shifts such as the PDO. For example, a station with data spanning the period 1920–1965 could have an apparent trend dominated by the shift from negative to positive PDO about 1947. If the record for the same station spanned the period 1960–1995, the apparent trend computed from the data might reflect the shift from negative to positive PDO in 1976–1977. For more details on climate variability and trends see Chapter 19 ("Climate Change Effects on Watershed Processes in British Columbia").

Past Climatic Variability and Change in British Columbia

A range of studies has looked for climatic trends in British Columbia, often in the context of specific locations or regions, or for specific climatic signatures. For example, Egginton (2005) examined climatic trends in northern British Columbia, particularly in relation to mass wasting, and Stahl et al. (2006a) focussed on the occurrence of winter temperatures low enough to cause mortality of mountain pine beetle. The general consensus from these studies is that British Columbia has been warming over the period of record, consistent with the current trend to increasing global temperatures. Vincent and Mekis (2006) found general trends across Canada, and particularly in southern British Columbia, to less extreme cold temperatures and more extreme warm temperatures. Regional variability, however, is significant in the magnitudes of trends, and the seasonal expression of these trends, which limits our ability to generalize at the provincial scale. For example, Bonsal et al. (2001b), Whitfield and Cannon (2000a, 2000b) and Whitfield et al. (2002a) found trends across southern British Columbia to significantly warmer springs, falls, and winters, but not summers. The pattern in northern British Columbia is different, with cooler falls and warmer winters. Variability also exists among studies. For example, Egginton (2005) found warming trends during fall in northern British Columbia. These differences among studies arise from the use of different periods of records, different methods of analysis (e.g., simple linear regression with time, Mann-Kendall non-parametric trend test), and differences in delineating regions.

Zhang et al. (2000) found that precipitation in western Canada increased by 5–35% from 1950 to 1998. Whitfield et al. (2002b) showed that in southern British Columbia the winter wet periods were now more wet; early fall has become somewhat drier and late fall wetter. In northern British Columbia, in contrast, falls and early summers have become wetter.

Figures 3.18–3.20 illustrate trends in daily minimum and maximum air temperature and precipitation for the period 1900–2003, based on gridded estimates of climate data. The warming trend is stronger for daily minimum than daily maximum air temperatures, and is strongest in winter. Precipitation generally increased over the province in most seasons except autumn, for which the south coastal region tended toward decreasing precipitation, though the trends are not statistically significant. Figures 3.21 and 3.22 provide results from a trend analysis focussed on the period 1950–2003 that used station data rather than gridded values. These analyses suggest that the warming trend has been particularly expressed through a reduction in the frequency of cold nights, and to a lesser extent an increased frequency of warm days. For more details on climate variability and trends see Chapter 19 ("Climate Change Effects on Watershed Processes in British Columbia").

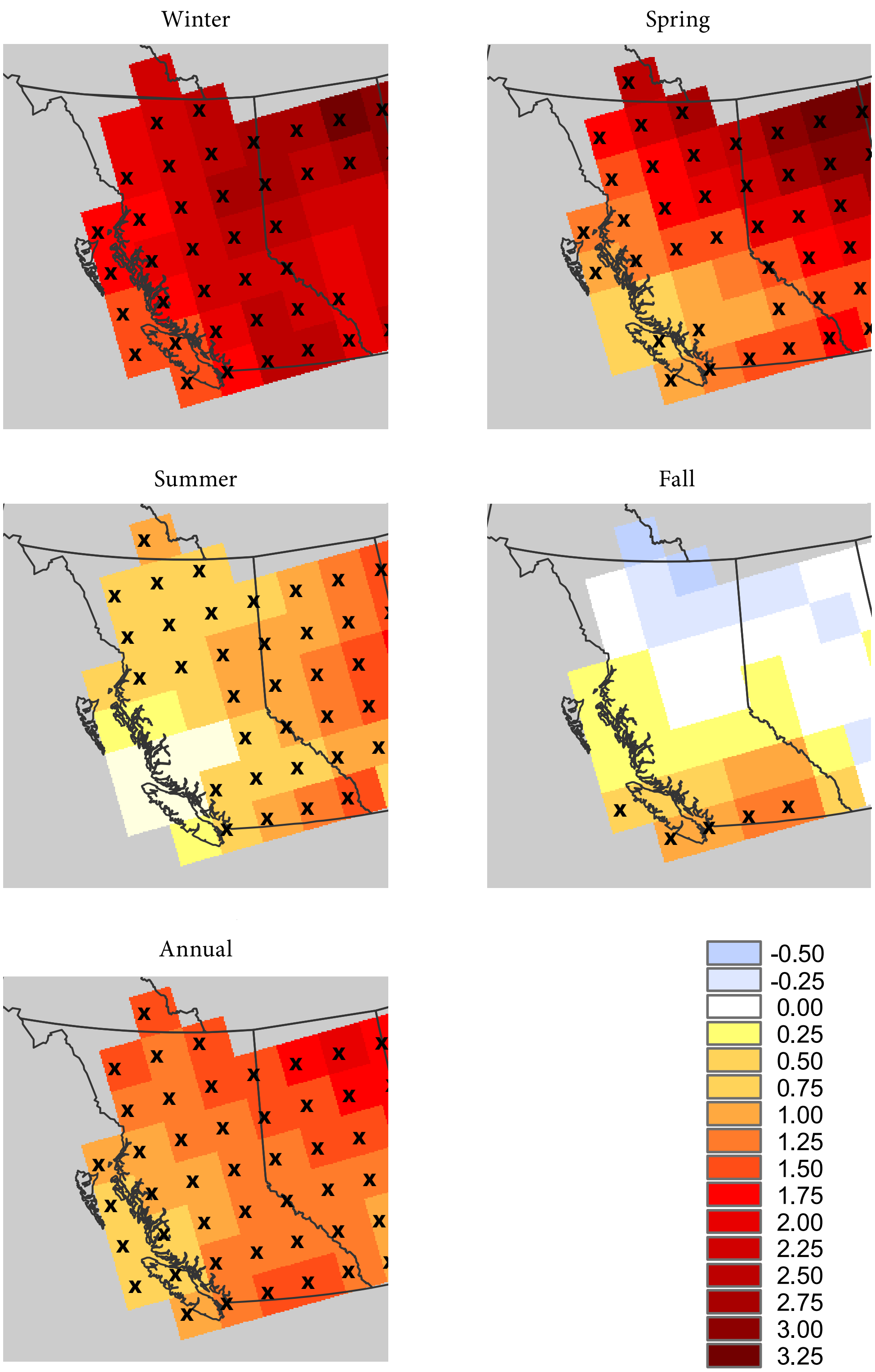

FIGURE 3.18 *Trends in daily minimum air temperature for the period 1900–2003 for western Canada. Units are degrees Celsius over the 104-year period. Grid cells with crosses indicate trends that are significant at a 5% significance level. Grey cells indicate areas with insufficient data to estimate gridded temperatures. (Adapted and updated from Zhang et al. 2000)*

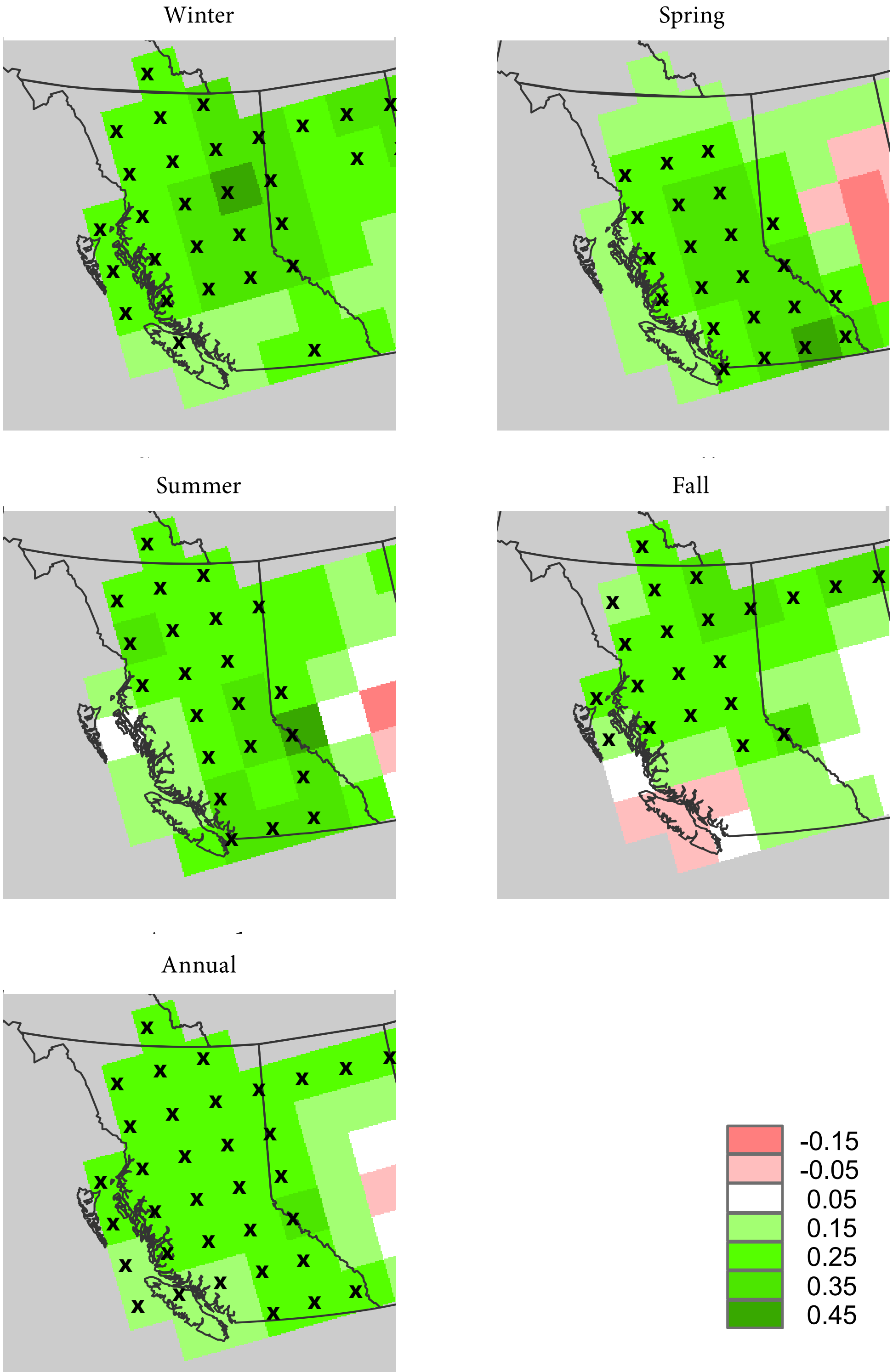

FIGURE 3.19 *Trends in precipitation totals for the period 1900–2003 for western Canada. Units are percentage change per year over the 104-year period. Grid cells with crosses indicate trends that are significant at a 5% significance level. Grey cells indicate areas with insufficient data to estimate gridded temperatures. (Adapted and updated from Zhang et al. 2000)*

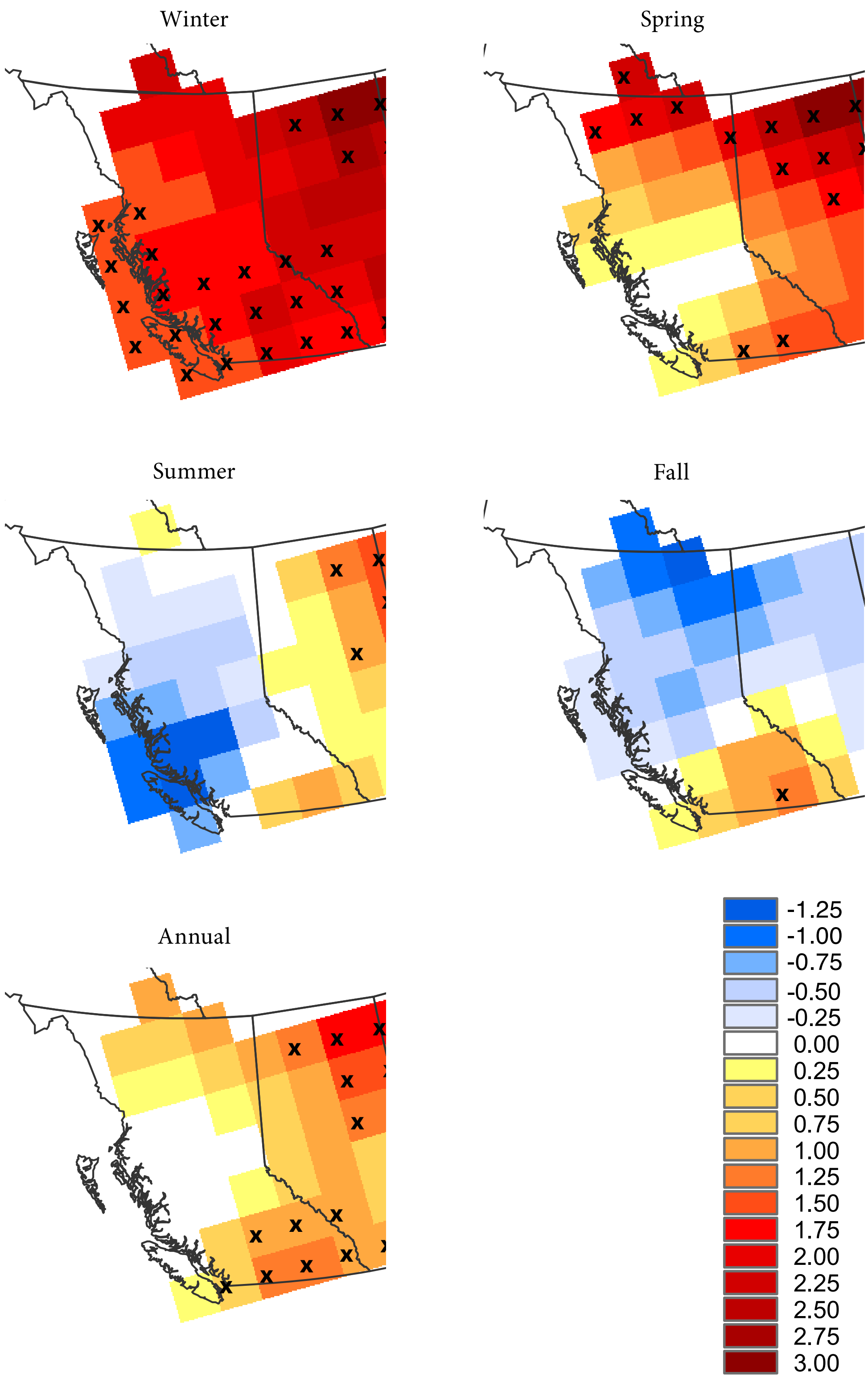

FIGURE 3.20 *Trends in daily maximum air temperature for the period 1900–2003 for western Canada. Units are degrees Celsius for the 104-year period. Grid cells with crosses indicate trends that are significant at a 5% significance level. Grey cells indicate areas with insufficient data to estimate gridded temperatures. (Adapted and updated from Zhang et al. 2000)*

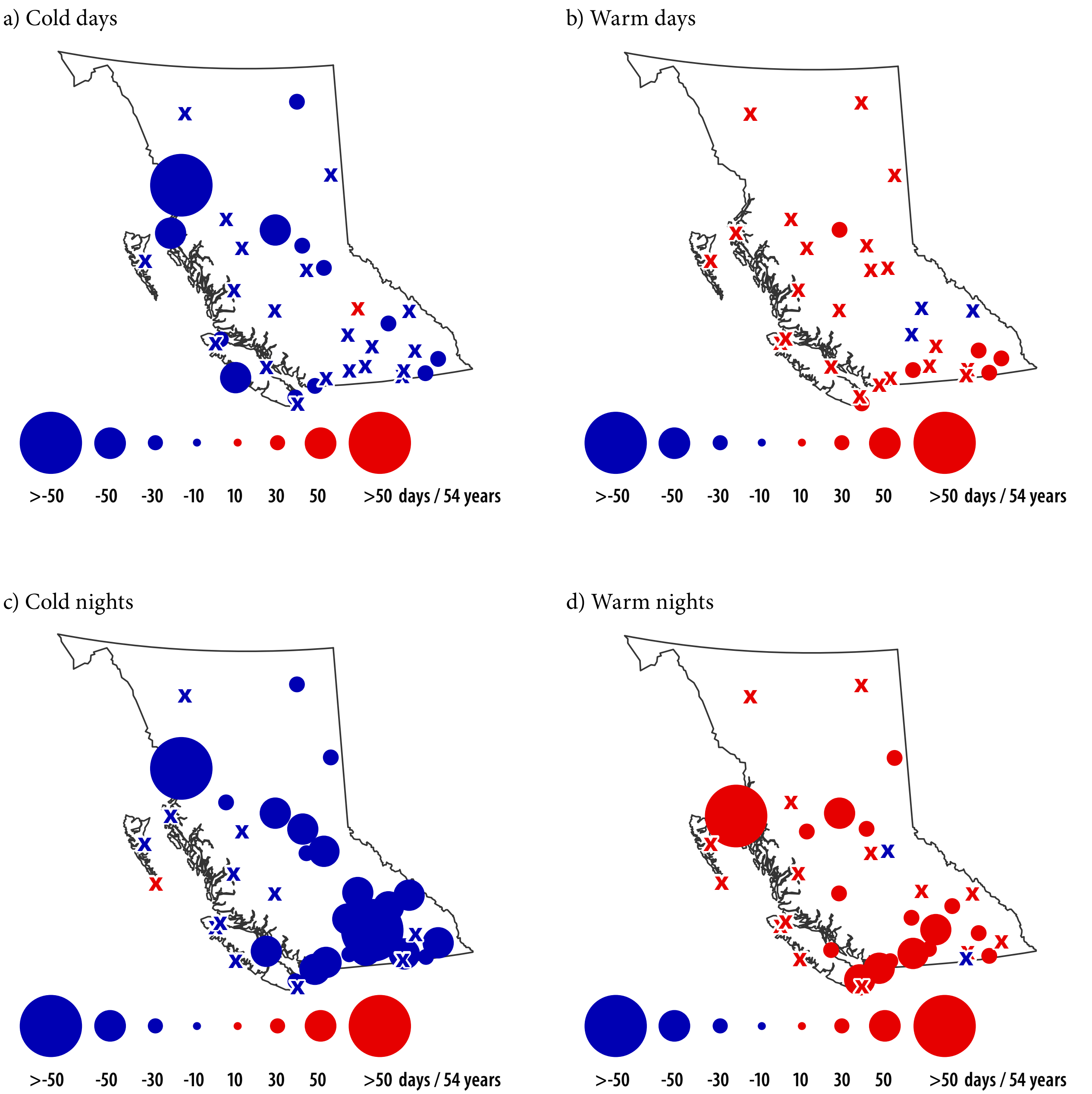

FIGURE 3.21 *Trends in four temperature metrics at climate stations within British Columbia for the period 1950–2003. Metrics are the frequencies per year of cold/warm days/nights, indicated by temperatures below/above defined thresholds. Dots indicate trends that are significant at a 5% significance level. Dot area is scaled to indicate the magnitude of the trend. (Adapted and updated from Vincent and Mekis 2006)*

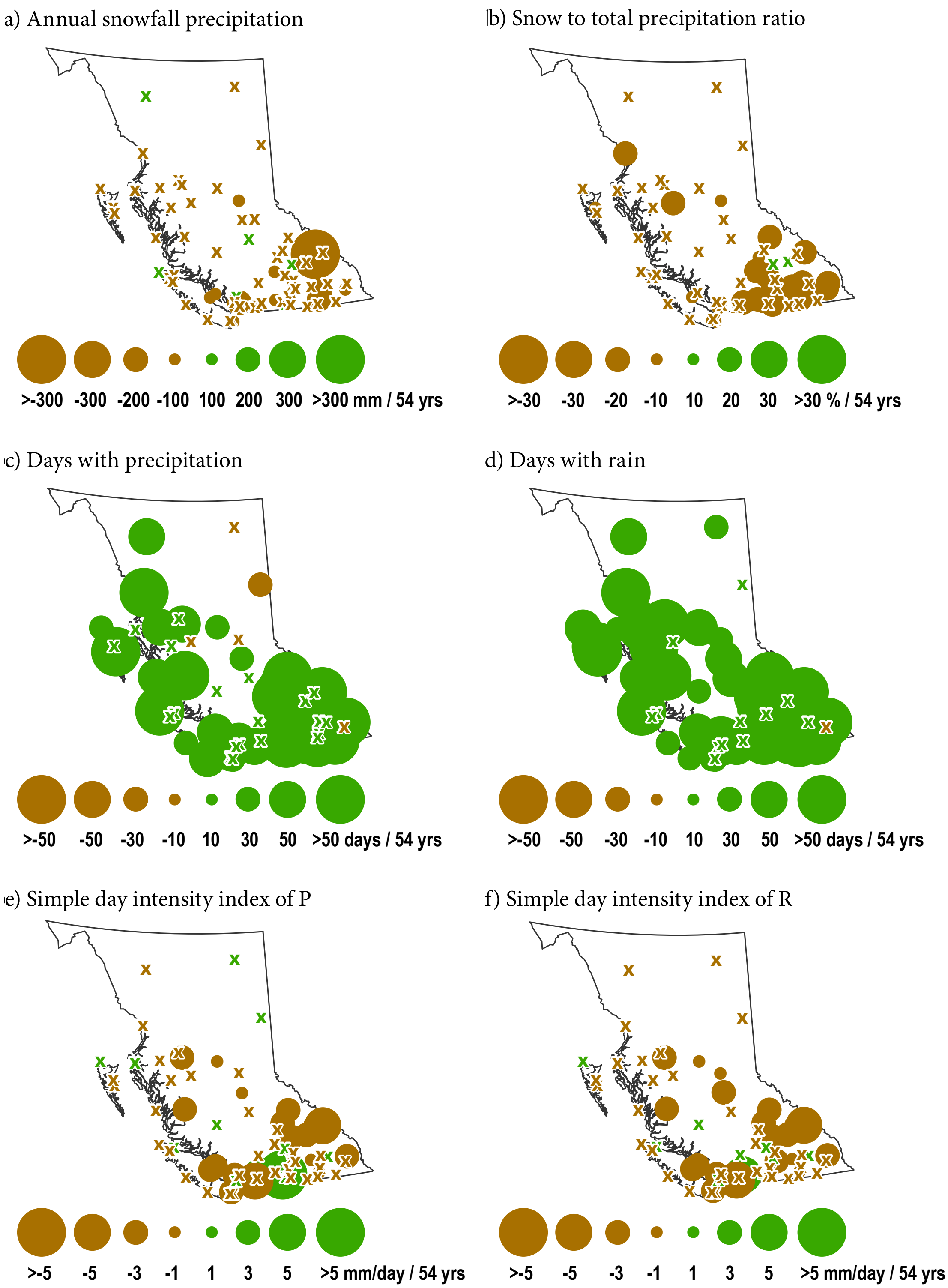

FIGURE 3.22 *Trends in six precipitation metrics at climate stations within British Columbia for the period 1950–2003. Brown and green dots indicate trends significant at the 5% level. Dot area is scaled to indicate the magnitude of the trend. Crosses denote non-significant trends. (Adapted and updated from Vincent and Mekis 2006)*

The Intergovernmental Panel on Climate Change uses multiple scenarios to generate possible future climatic conditions. Each scenario is based on a "story line" about developments in technology, economic growth, and international cooperation (Intergovernmental Panel on Climate Change 2007). For each scenario, emissions of greenhouse gases are estimated and used as input to global climate models (GCMs). This approach explicitly acknowledges the inherent uncertainties in predicting future socio-economic conditions. The GCMs that generate projections of future climatic conditions are also a source of uncertainty because of differences in how certain atmospheric and oceanic processes are modelled. Consequently, a single future climate projection is not considered more likely than another, but the models allow bounds to be placed on the range of possible outcomes. For more details on climate variability and trends see Chapter 19 ("Climate Change Effects on Watershed Processes in British Columbia").

Figure 3.23 shows a range of future climate projections for three emission scenarios and a number of GCMs. The A2 scenario assumes that emissions will continue to increase without any significant efforts to reduce them. The B1 scenario assumes that the rate of emission increases will slow down soon and begin to decrease by the middle of the century. Scenario A1B is intermediate between the A2 and B1 scenarios (Intergovernmental Panel on Climate Change 2007). The gold line shows that even if all emissions ceased immediately, an additional 0.5°C warming would occur because of the effects of a lag in heat distribution in the oceans. Changes in precipitation will accompany changes in temperature.

Model projections indicate that British Columbia will have greater warming and changes in precipitation than the global average (Intergovernmental Panel on Climate Change 2007). A range of future climates is possible, depending on the future emissions and the GCM (Table 3.6). The main feature illustrated in Table 3.6 is that all models and emissions scenarios produce an increase in temperature that grows with time. Not shown in Table 3.6 is that warming tends to be greater for minimum temperatures than for maximum temperatures. Projected changes in precipitation are quite variable. Generally, southern and central British Columbia are expected to get drier in the summer, and northern British Columbia is more likely to be wetter. Winters will, in general, become wetter across British Columbia, with a greater increase in the north. Figures 3.24–3.28 illustrate what such changes in climate would mean for British Columbia using the A2 emission scenario (minimal control of emissions) as simulated by the Canadian Global Climate Model version 2. Table 3.7 presents historic climatic conditions for selected locations in British Columbia, along with projections based on the A2 scenario. Other possible future climates are equally likely.

The projections in Tables 3.6 and 3.7, and Figures 3.23–3.28, are based on multi-year averages. However, as noted previously, extreme events are often more important than long-term climatic means. In the simulations of future climate, changes in warm extremes follow changes in the mean summertime temperature. Cold extremes warm faster, particularly in areas that experience a retreat of snow with warming. There is also an increased intensity of precipitation regimes and a reduction in return periods of current extreme events (Kharin et al. 2007). Note that, while only projections for temperature and precipitation are discussed here, GCMs also generate projections for humidity, wind speed, and solar radiation, which users can access if needed.

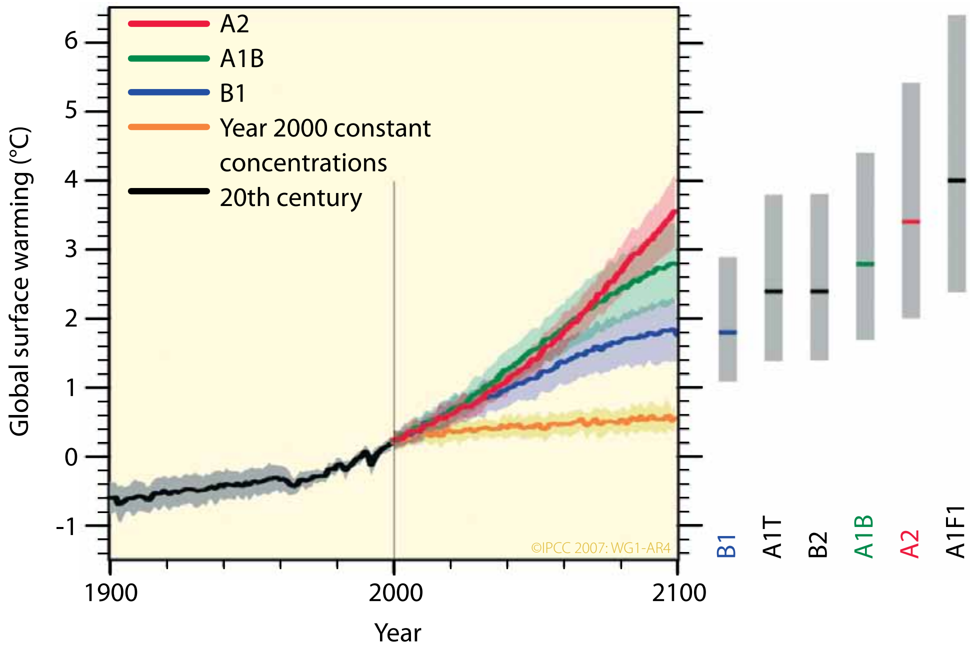

FIGURE 3.23 *A range of future climate projections for three emission scenarios and a number of global climate models. Solid lines are multi-model global averages of surface warming (relative to 1980–1999) for the scenarios A2, A1B, and B1, shown as continuations of the 20th-century simulations. Shading denotes the ±1 standard deviation range of individual model annual averages. The orange line is for an experiment in which concentrations were held constant at year 2000 values. The grey bars at right indicate the best estimate (solid line within each bar) and the likely range assessed for the six Special Report on Emission Scenarios marker emissions scenarios. The assessment of the best estimate and likely ranges in the grey bars includes the coupled atmosphere ocean general circulation models in the left part of the figure, as well as results from a hierarchy of independent models and observational constraints. (Intergovernmental Panel on Climate Change 2007)*

TABLE 3.6 *Climate change scenarios (seven models and eight emission scenarios). Data are the changes from 1961 to 1990 climate as a change in mean temperature or a percentage change in total precipitation (PPT). Values are based on data from the Canadian Climate Change Scenarios Network (www.cccsn.ca) and Pacific Climate Impacts Consortium (www.pacificclimate.org).*

	2020		2050		2080	
	Temperature (°C)	PPT (%)	Temperature (°C)	PPT (%)	Temperature (°C)	PPT (%)
Southern British Columbia						
Winter	0 to 2	–5 to +15	1.5 to 3.5	0 to +20	2 to 7	0 to 25
Summer	0.5 to 2	–30 to +5	1.5 to 4	–35 to 0	2.5 to 7.5	–50 to 0
Central British Columbia						
Winter	0 to 2	–5 to +15	1.5 to 4	0 to +30	2.5 to 6	+5 to +40
Summer	0.5 to 1.5	–10 to +5	1.8 to 3.5	–20 to 0	2.5 to 6.5	–20 to +5
Northern British Columbia						
Winter	0 to 2.5	0 to 20	1.5 to 5.5	0 to +25	2.5 to 9	0 to +45
Summer	0.5 to 1.5	–10 to +10	1.5 to 3.5	–10 to +15	2 to 6	–15 to +25

TABLE 3.7 *Climate change scenarios for five locations in British Columbia based on the Canadian global climate model simulations for the A2 emission scenario (CGCM2 a2x; obtained using the ClimateBC software [Wang et al. 2006; Spittlehouse 2006, 2008]).*

	MAT[a] (°C)	MWMT (°C)	MCMT (°C)	MAP (mm)	MSP (mm)	FFP (days)	SH:M
Castlegar							
1961–1990	8.3	19.9	–3.2	732	284	153	70
2020s	9.5	21.1	–1.8	750	279	169	76
2050s	10.6	22.2	–0.4	744	263	181	84
2080s	12.2	23.6	0.6	767	255	210	93
Fort Nelson							
1961–1990	–1.1	16.7	–22.0	449	303	106	55
2020s	0.1	18.2	–20.3	466	312	118	58
2050s	1.5	19.3	–17.9	477	313	130	62
2080s	3.3	20.8	–14.5	499	329	145	63
Kelowna							
1961–1990	7.4	18.8	–4.5	366	171	125	110
2020s	8.6	19.9	–3.4	374	170	139	117
2050s	9.7	21.0	–2.1	369	161	152	130
2080s	11.2	22.4	–1.0	375	158	176	142
Prince George							
1961–1990	3.7	15.3	–9.9	615	287	93	53
2020s	4.9	16.5	–8.2	615	280	110	59
2050s	6.1	17.5	–6.4	630	285	127	61
2080s	7.7	18.8	–4.9	635	275	149	68
Port Hardy							
1961–1990	8.1	13.9	3.0	1871	410	183	33
2020s	9.1	14.8	3.9	1885	394	208	38
2050s	10.1	15.9	4.9	1935	387	259	41
2080s	11.4	17.2	6.1	2035	373	324	46

a MAT = mean annual temperature; MWMT = mean warmest month temperature (July); MCMT = mean coldest month temperature (January); MAP = mean annual precipitation; MSP = mean summer precipitation (May to September); FFP = frost-free period; SH:M = summer heat:moisture index.

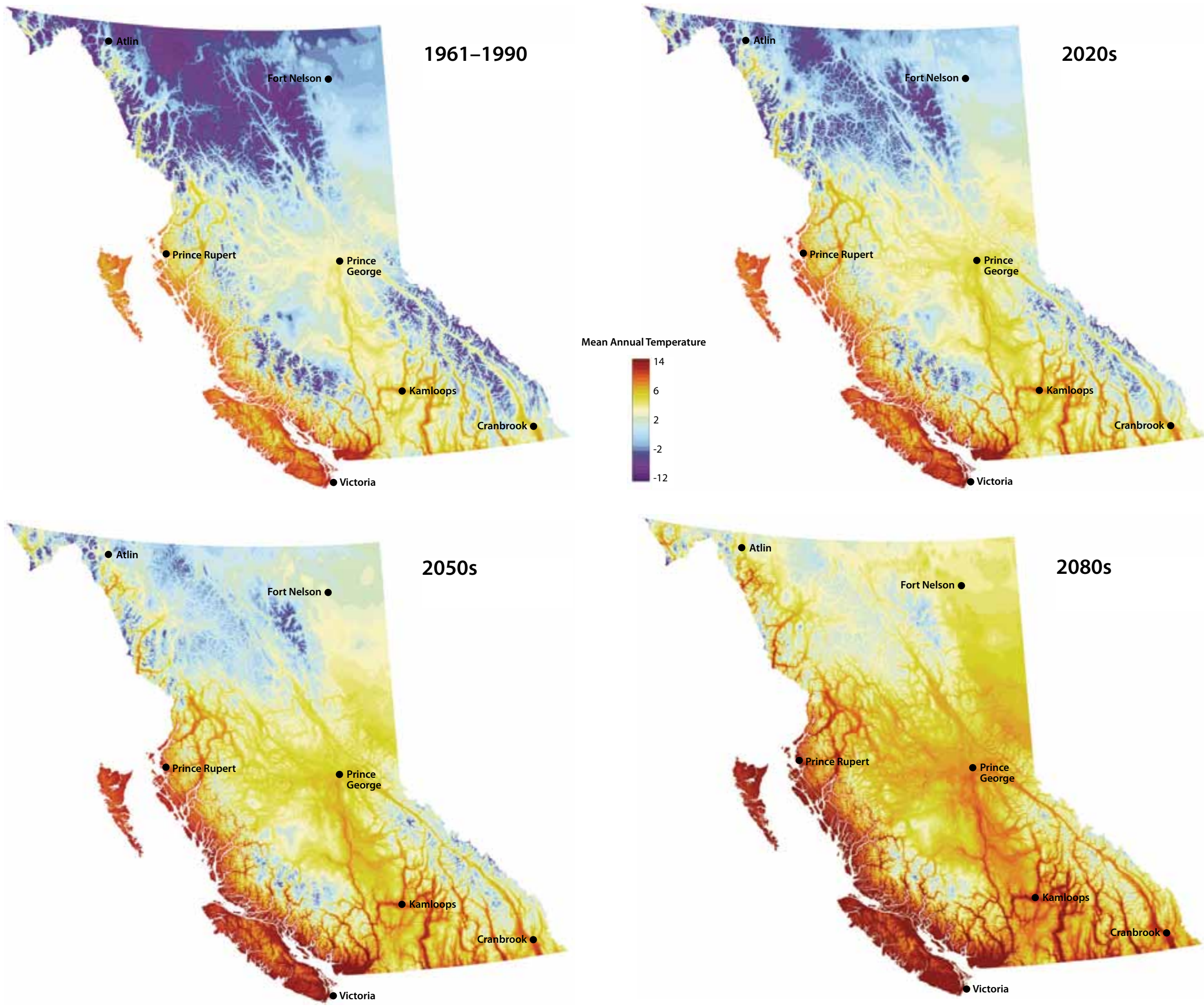

FIGURE 3.24 *Mean annual temperature for British Columbia for current climate (1961–1990 average) and that predicted for British Columbia in 2020s, 2050s, and 2080s for the A2 scenario from Canadian Global Climate Model version 2. Downscaling was done with ClimateBC (Spittlehouse 2006, 2008; Wang et al. 2006).*

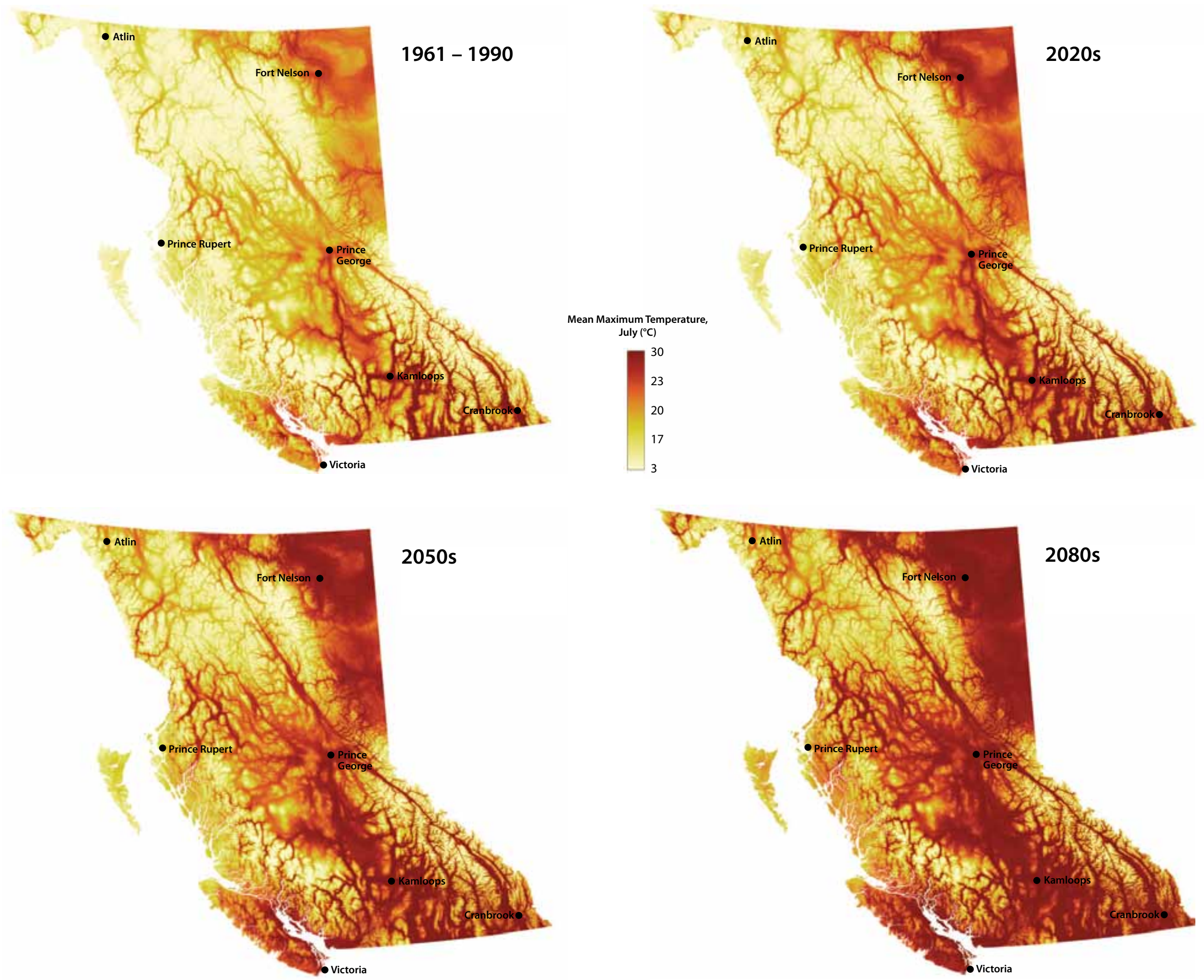

FIGURE 3.25 *Mean maximum July temperature for British Columbia for current climate (1961–1990 average) and that predicted for British Columbia in 2020s, 2050s, and 2080s for the A2 scenario from Canadian Global Climate Model version 2. Downscaling was done with ClimateBC (Spittlehouse 2006, 2008; Wang et al. 2006).*

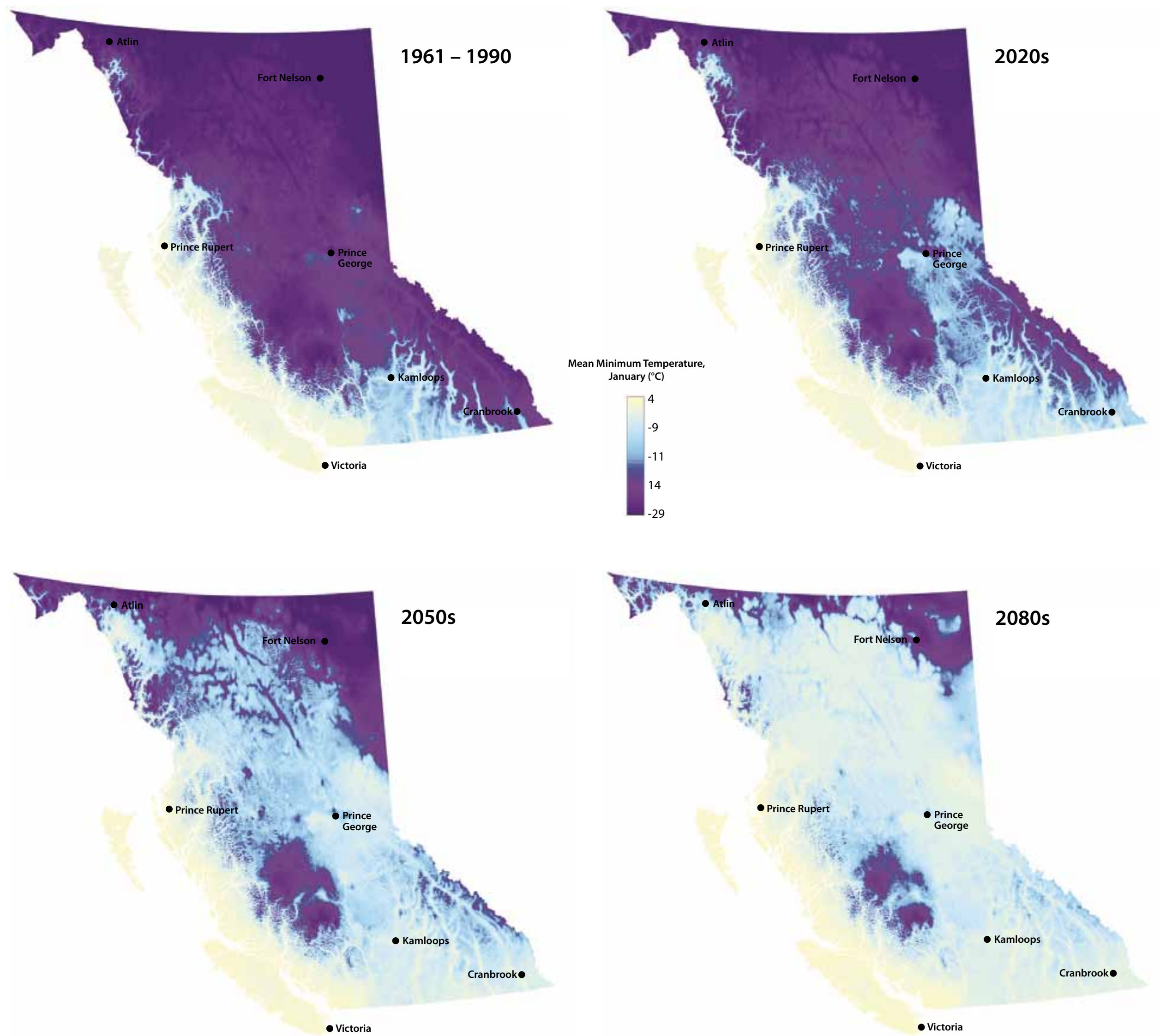

FIGURE 3.26 *Mean minimum January temperature for British Columbia for current climate (1961–1990 average) and that predicted for British Columbia in 2020s, 2050s, and 2080s for the A2 scenario from Canadian Global Climate Model version 2. Downscaling was done with ClimateBC (Spittlehouse 2006, 2008; Wang et al. 2006).*

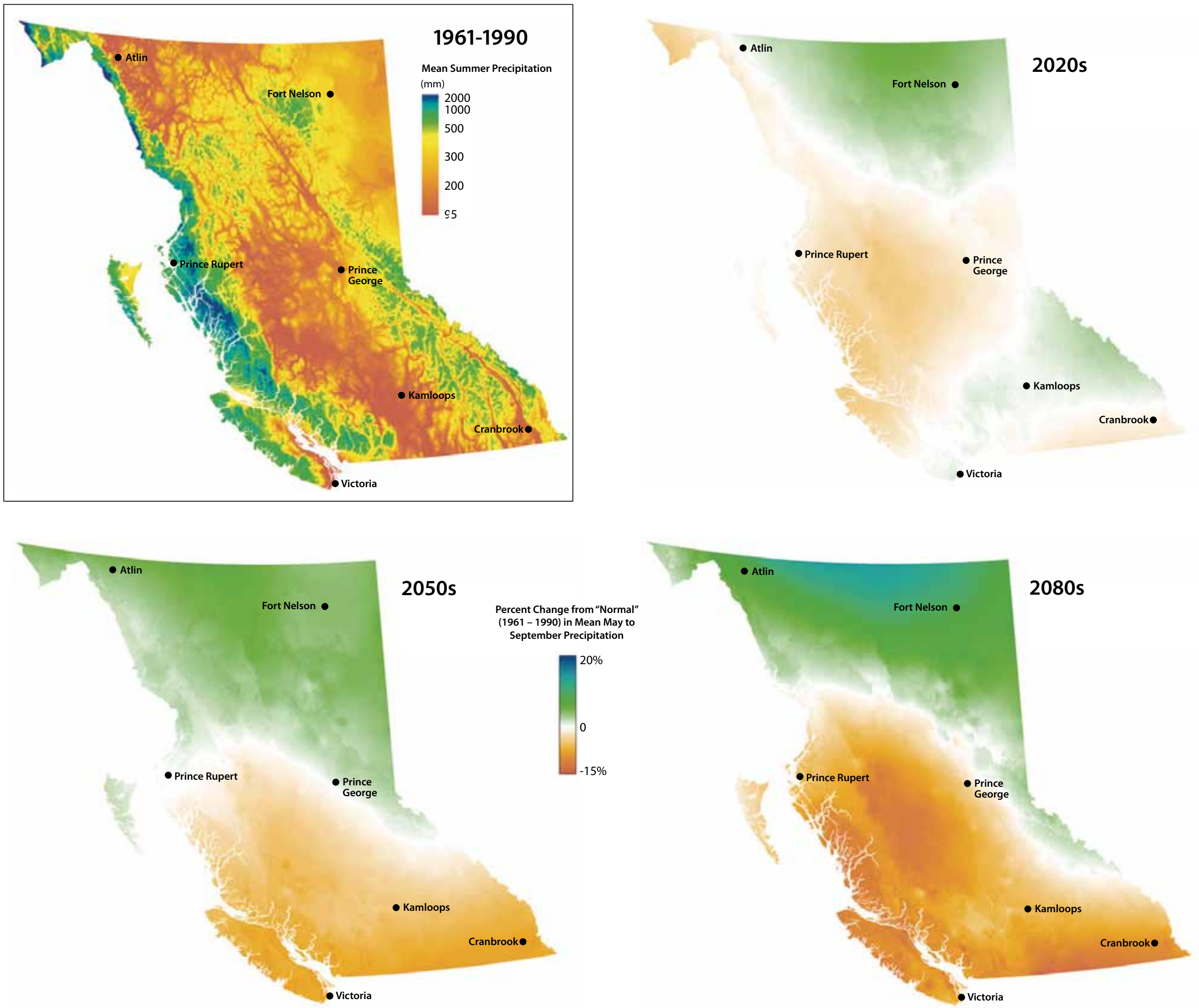

FIGURE 3.27 *Mean May to September precipitation for British Columbia for current climate (1961–1990 average) and the percentage change predicted for British Columbia in 2020s, 2050s, and 2080s for the A2 scenario from Canadian Global Climate Model version 2. Downscaling was done with ClimateBC (Spittlehouse 2006, 2008; Wang et al. 2006).*

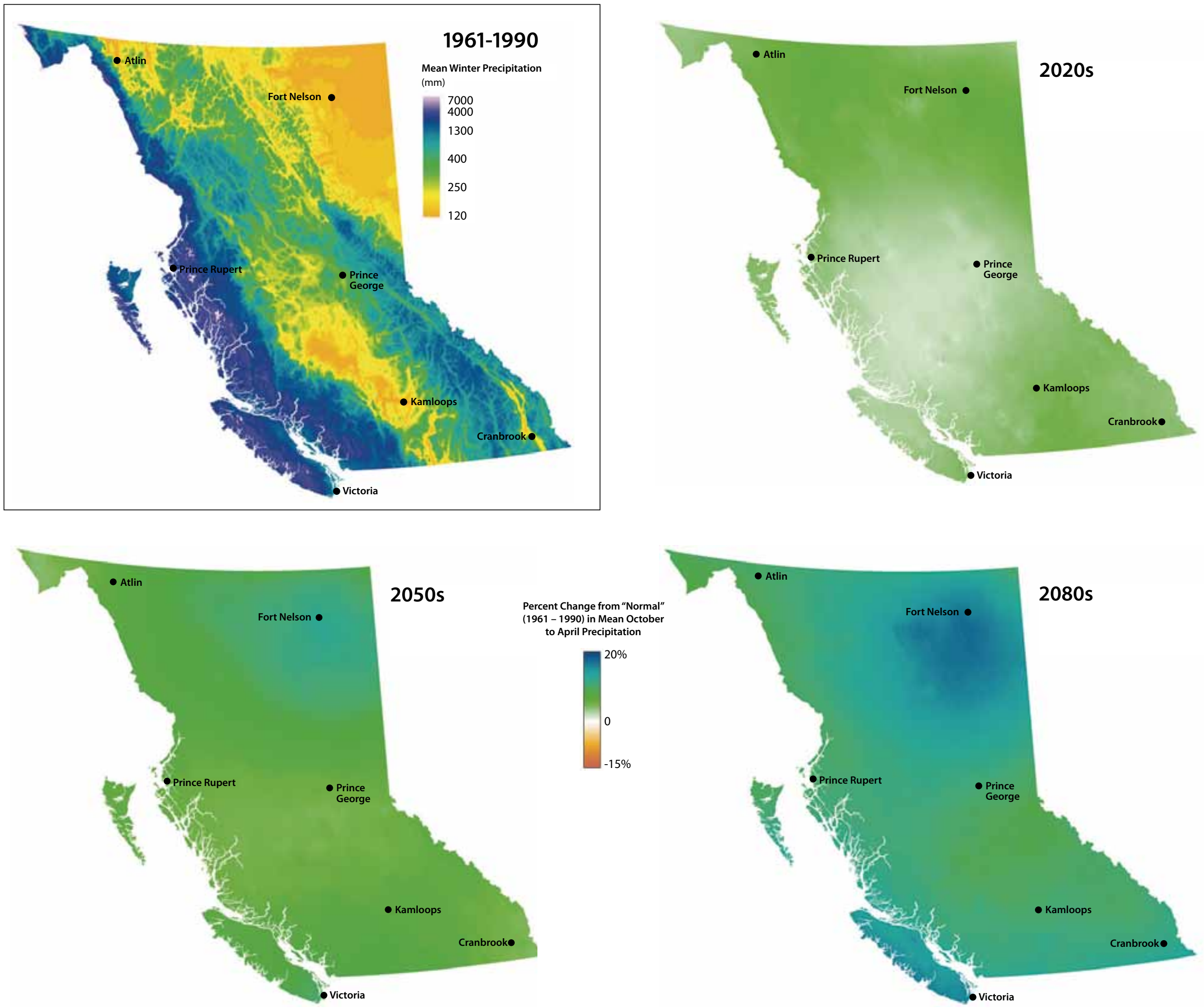

FIGURE 3.28 *Mean October to April precipitation for British Columbia for current climate (1961–1990 average) and the percentage change predicted for British Columbia in 2020s, 2050s, and 2080s for the A2 scenario from Canadian Global Climate Model version 2. Downscaling was done with ClimateBC (Spittlehouse 2006, 2008; Wang et al. 2006).*

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CHAPTER 4

Regional Hydrology

BRETT EATON AND R.D. (DAN) MOORE

INTRODUCTION

Stream water is a valuable resource. It is withdrawn for irrigation in many parts of British Columbia, and is commonly the dominant source for drinking water. Aquatic ecosystems also depend on streamflow, since the timing, magnitude, and temperature of the flows determine the health of the aquatic ecosystem. The current decline of the populations of some indicator and economically important species, such as coho salmon, highlights the importance of streamflow conditions. The water carried by the streams may also pose a significant hazard. Peak flows can destroy stream crossings and other instream engineered structures, erode land adjacent to the stream, and flood the surrounding areas.

The flow carried by a stream varies over a range of time scales. Over short time scales (e.g., less than a day to a few days), it is controlled by the passage of weather events and their influence on rainfall, snow and ice melt, and evapotranspiration. Over longer time scales (years to decades), larger-scale climatic processes such as El Niño play an important role. However, streamflow tends to exhibit typical seasonal patterns, or regimes, which can be broadly classified in relation to the dominant sources of streamflow: rainfall, snowmelt, and glacier melt. Understanding the variability of regimes both within and among regions is an important first step in assessing the state and potential sensitivity of a watershed to land use practices. For example, forest harvesting has different influences in rain- and snow-dominated watersheds. For this reason, the *Forest Practices Code of British Columbia Act* imposed different watershed assessment procedures for coastal and interior regions of British Columbia.

This chapter explores the typical spatial variations in the seasonal flow regimes, providing generalized explanations for the observed patterns. It also explores the magnitude, timing, and processes producing peak flows across British Columbia. Practitioners should be able to use this chapter to characterize the seasonal flow regimes and peak flow regimes for ungauged drainage basins across British Columbia. Furthermore, the analysis that we present can act as a template for subsequent analysis of basins that are gauged, and also provide a context for interpreting the results of such an analysis.

Rain-dominated regimes are found primarily in coastal lowland areas and at lower elevations of the windward side of the Coast Mountains. In these regimes, the temporal variability of streamflow closely follows that of rainfall, though smoothed somewhat by the effects of water storage and movement in soils, groundwater, lakes, and wetlands. As a result, these regimes typically exhibit the highest monthly discharge values in November and December, since this is when the most intense frontal weather systems move over the British Columbia coast (Figure 4.1, Carnation Creek). The lowest monthly flows occur in July and August, when a blocking high-pressure system typically directs precipitation-generating weather systems away from at least the southern half of the province.

Snowmelt-dominated (or nival) regimes occur in the interior plateau and mountain regions, and in higher-elevation zones of the Coast Mountains. In these zones, winter precipitation dominantly falls as snow and remains in storage until spring melt. As a result, these regimes exhibit low flows through winter, and high flows in May, June, and July (Figure 4.1, Fishtrap and Redfish Creeks). Low flows may also occur during the late summer and fall as a result of low precipitation inputs and the exhaustion of the snowpack water supply. The regime for Coquihalla River is mostly nival, but the drainage basin has sufficient low-elevation area and proximity to maritime influences that it has a stronger influence of autumn rains than Fishtrap and Redfish Creeks.

Many basins, especially in coastal and near-coastal regions of British Columbia, exhibit characteristics of both rain- and melt-dominated streamflow regimes. These are referred to as mixed-regime or hybrid-regime basins. For example, Capilano River has rain-dominated high flows from October to January that are of a similar magnitude to the melt-dominated flows in April to June (Figure 4.1). For mixed-regime basins, the relative importance of the rainfall influence decreases inland from the coast or northwards up the coast; in both cases, the mean temperature tends to decrease, promoting the occurrence of snow rather than rain during the winter. The relative importance of the rain regime also decreases with increasing mean basin elevation for the same reason. Even on the coast, winter temperatures are low enough that a significant proportion of the precipitation will fall as snow at elevations greater than about 1000 m above sea level (asl).

A distinct population of mixed-regime basins also occurs in the northeastern part of the province, where snowmelt produces a peak in early May owing to the relatively low relief in the area. In the northeast, precipitation peaks in the summer months as a result of convective precipitation (see Chapter 3, "Weather and Climate"). This produces an increase in the seasonal runoff in June and July, which has the appearance of a snowmelt-dominated regime, but is the product of snowmelt combined with rain-related runoff.

Drainage basins with more than two to five percent of the area covered by glaciers have regimes similar to nival regimes, except that the period of high flows extends from about May to August or September, and low-flow conditions occur only when precipitation is accumulating in the snowpack, usually from December to March (Figure 4.1, Lillooet River). The extended melt freshet is partly associated with the higher elevations typical of glacierized drainage basins (i.e., basins that currently have glaciers covering more than two to five percent of the watershed) that hold snow later into the summer, but is primarily associated with the presence of glaciers which, during the melt season, act as inexhaustible reservoirs of water.

Although streamflow regimes are dominantly controlled by seasonal patterns of temperature and precipitation over a drainage basin, a range of watershed characteristics also influence the temporal distribution of streamflow. For example, drainage basins on the interior plateaus tend to have low relief, and have a snowmelt period that is synchronized over most of the basin, resulting in a shorter spring freshet period. Mountainous drainage basins, in contrast, typically have an extended freshet caused by the sequential melting of snow from lower to higher elevation bands. This contrast is illustrated in Figure 4.1 by the regimes for Fishtrap Creek (plateau) and Coquihalla River and Redfish Creek (mountain drainages). Presence of lakes, ponds, and wetlands can attenuate flows, decreasing high flows as water goes into storage and augmenting low flows through storage depletion. The role of geology, through its influence on groundwater storage and discharge, can be particularly important. For example, a drainage

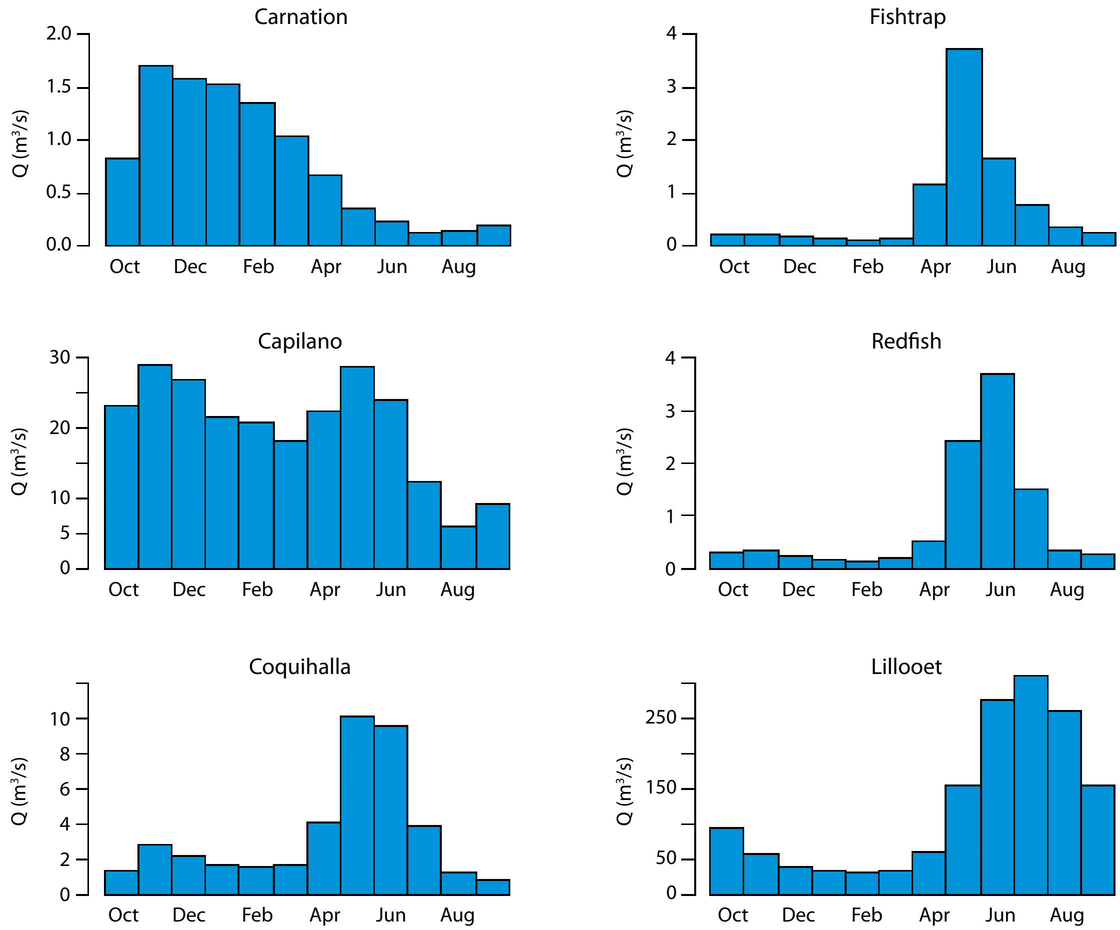

FIGURE 4.1 *Examples of long-term average streamflow regimes drawn from southern British Columbia. Values are monthly average discharge (Q) in m³/s.*

basin underlain by highly permeable bedrock will tend to have more water recharging the groundwater during rain and melt events, thus decreasing storm runoff but moderating flow during dry weather. Basins underlain by relatively impermeable bedrock such as granite will tend to have less groundwater influence on the stream hydrograph; however, even for relatively impermeable bedrock, flow through fracture systems can be important, particularly for smaller, headwater streams (Whyte 2004).

Glaciation has left a significant legacy in the hydrogeomorphic form and function of British Columbia watersheds (Brardinoni and Hassan 2006), especially the U-shaped valleys with oversteepened side slopes, as well as the surficial materials left behind as glaciers retreated. For example, many glaciated areas have shallow soils underlain by compacted till, which is relatively impermeable and promotes runoff generation as shallow saturated subsurface flow (Hutchinson and Moore 2000).

Streamflow regimes within a region may also depend on drainage basin scale, particularly in areas with high relief. Smaller basins, particularly those less than a few square kilometres in area, will tend to have higher mean elevations than larger drainage basins, leading to a stronger nival influence in regions with mixed streamflow regimes, and a delayed seasonal melt freshet in regions dominated by nival regimes.

The hydrographs in Figure 4.1 represent averages over a number of years; however, streamflow regimes can vary significantly in both the intra-annual pattern and overall magnitude of streamflow. Rain-dominated streamflow regimes closely reflect the input of precipitation to a drainage basin with relatively little smoothing or lagging associated with long-term storage within the basin. As a result, year-to-year variation in the seasonal distribution of the mean monthly discharge is significant, reflecting the sequence and intensity of the weather systems that affect a drainage basin in a given year (Figure 4.2). The monthly discharge pattern for a given year is therefore unlikely to closely resemble the average pattern for a number of years.

Snowmelt-dominated systems integrate precipitation inputs over the winter and spring within the snowpack, and then release the stored water during the spring-summer melt period. Consequently, the monthly discharge pattern for a given year is generally similar in shape, if not magnitude, to the long-term mean monthly discharge pattern (Figures 4.3 and 4.4).

For mixed-regime rivers, the relative strengths of the rain and snowmelt influences on the annual hydrograph vary from year to year, primarily depending on air temperatures during winter storms. For storms that are associated with relatively high air temperatures—for example, during El Niño years or the Pacific Decadal Oscillation (PDO) warm phase—

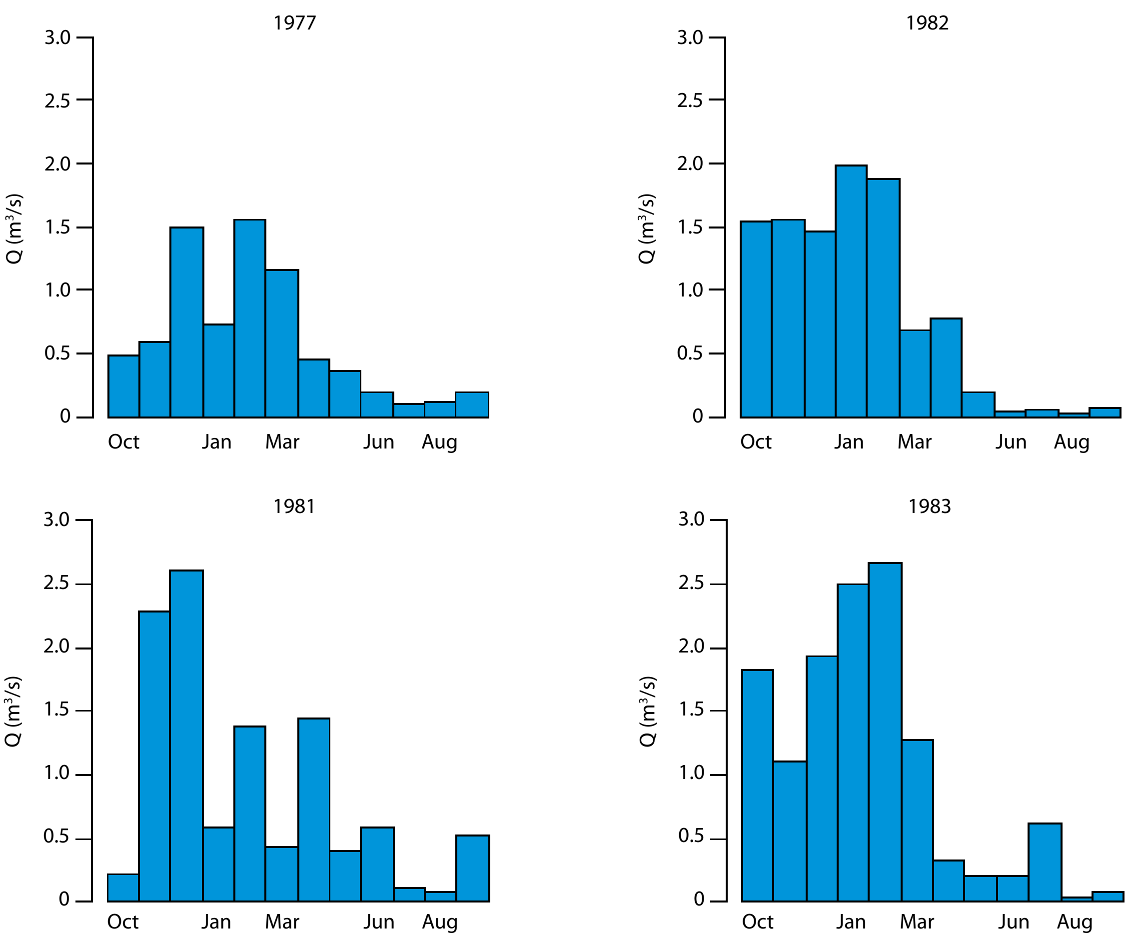

FIGURE 4.2 *Mean monthly discharge (Q) for Carnation Creek in 1977, and 1981–1983. Over southern British Columbia, 1977 and 1981 had generally warm, dry winters; 1982 had a cool, wet winter; and 1983 had a warm, wet winter.*

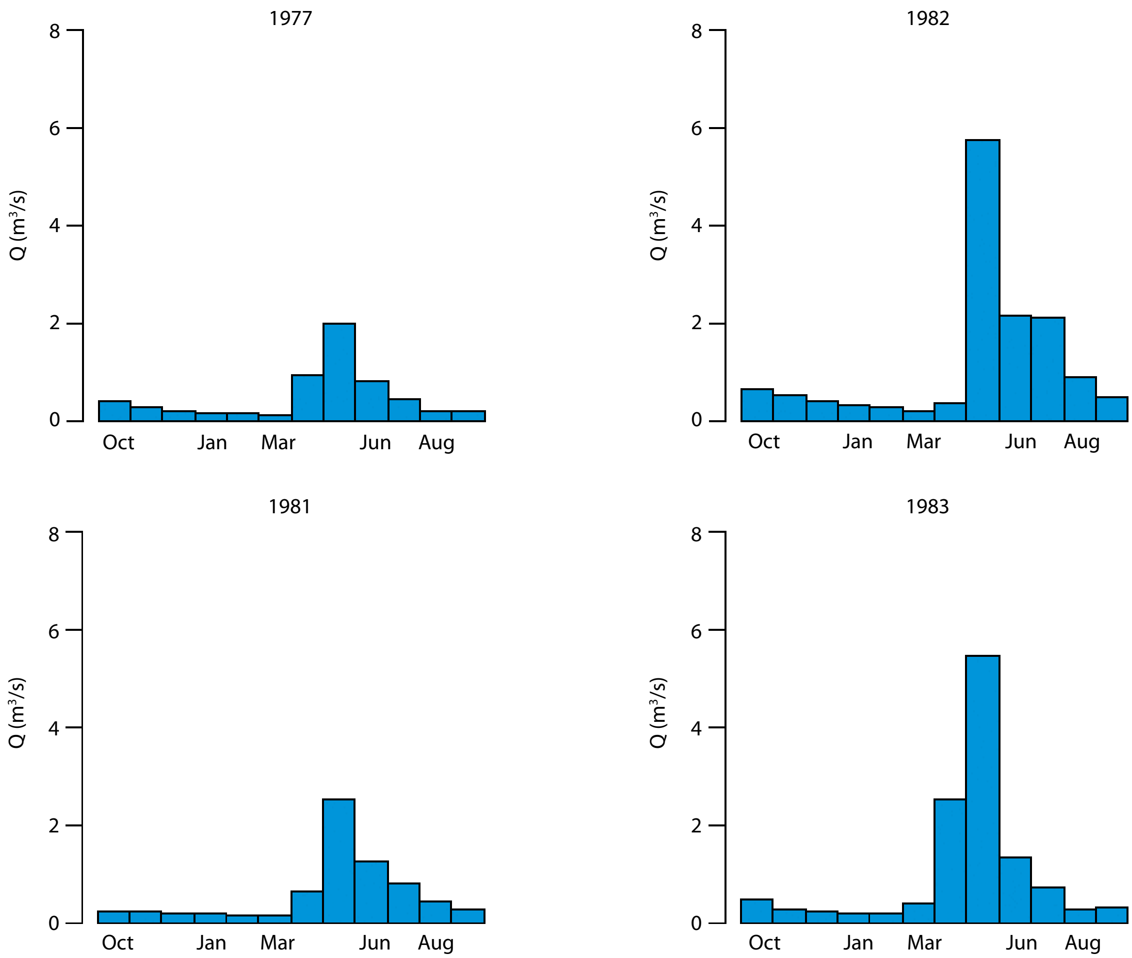

FIGURE 4.3 *Mean monthly discharge (Q) for Fishtrap Creek in 1977, and 1981–1983.*

precipitation falls mainly as rain. Conversely, in years dominated by cold storm systems (e.g., during La Niña years or PDO cold phase), more precipitation falls as snow and contributes to higher flows during the spring melt period. For example, Capilano River spring flows were relatively high during 1982, following a relatively cool winter that produced a deep snowpack (Figure 4.5).

In 1983, relatively warm winter conditions caused high rainfall and mid-winter snowmelt, producing high mid-winter streamflow. Coquihalla River also exhibits temperature-related responses. The relatively cool winter of 1982 produced an almost purely nival hydrograph, whereas the relatively warm winters of 1977 and 1981 produced stronger rainfall contributions (Figure 4.6). Like snowmelt-dominated streams, glacier-fed streams have relatively stable flow regimes (Figure 4.7). These regimes may be even more stable than purely nival regimes because glaciers effectively regulate the between-year variability in meltwater runoff by maintaining a source of meltwater after the seasonal snowpack has disappeared (Moore 1992; Moore and Demuth 2001).

In a year with low snow accumulation, snow will tend to melt off a glacier earlier in the season, exposing the less reflective glacier ice, which melts at a higher rate than snow exposed to the same meteorological conditions. Thus, the warm and dry conditions that produce relatively low summer streamflow from glacier-free portions of a basin increase meltwater runoff from glaciers and, in turn, increase streamflow (Stahl and Moore 2006). Conversely, the streamflow regimes for glacierized basins may vary over longer time frames as the glaciers in the basin advance or retreat in response to climatic variations and change. Even if climate remains relatively stable

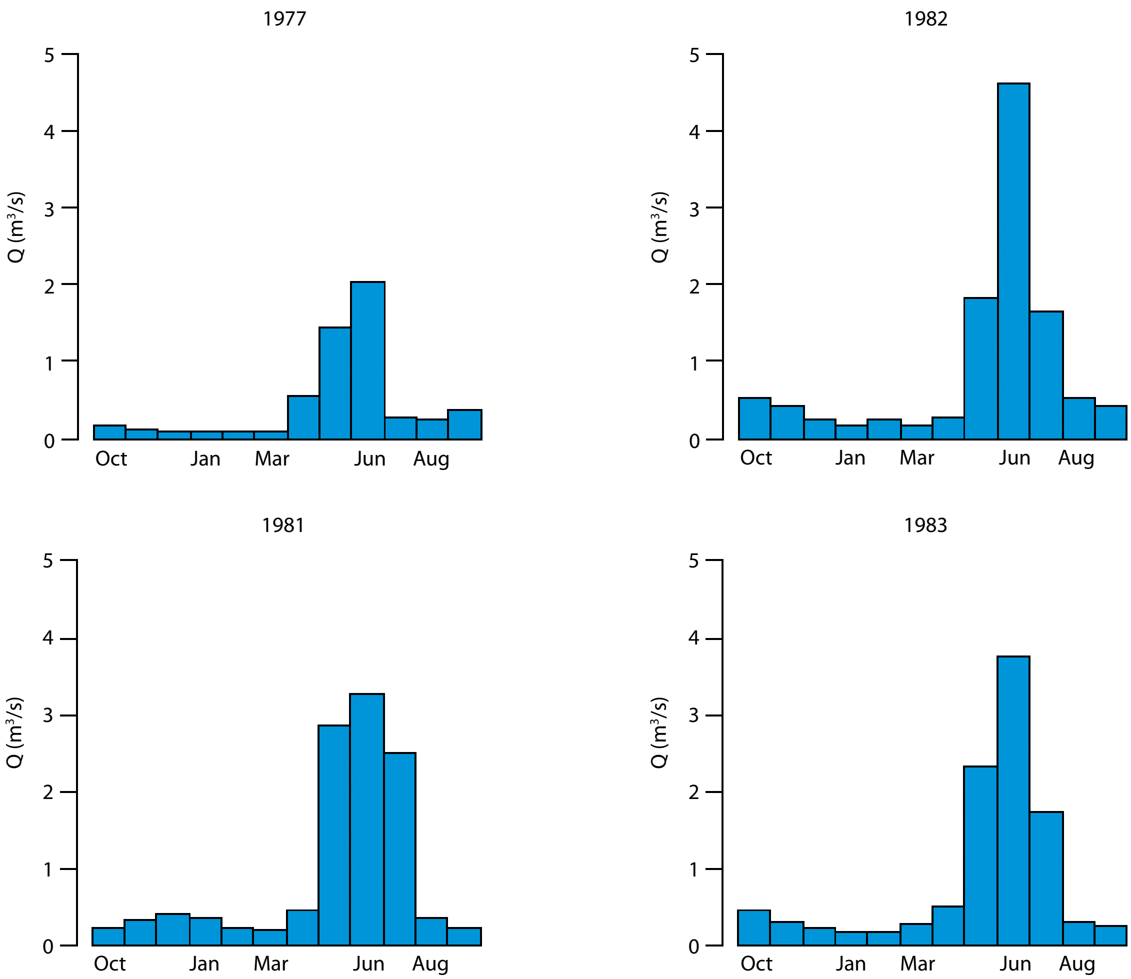

FIGURE 4.4 *Mean monthly discharge (Q) for Redfish Creek in 1977, and 1981–1983.*

within the period of interest, the streamflow regime may shift systematically due to the lag times in the response of the glaciers to climatic fluctuations.

Streamflow can also vary on decadal time scales in response to climatic fluctuations such as the PDO. For example, Moore (1996) showed that summer streamflow in the Capilano River displayed apparent step-shifts in the mean that were consistent with the timing of the PDO shifts. The lower winter temperatures associated with PDO cool phases result in more snow than rain. This augments snowmelt inputs and produces higher summer streamflow. Fleming et al. (2007) showed that streams with hybrid regimes were particularly sensitive to the effects of large-scale climatic variations such as the PDO and El Niño–Southern Oscillation (ENSO). In contrast,

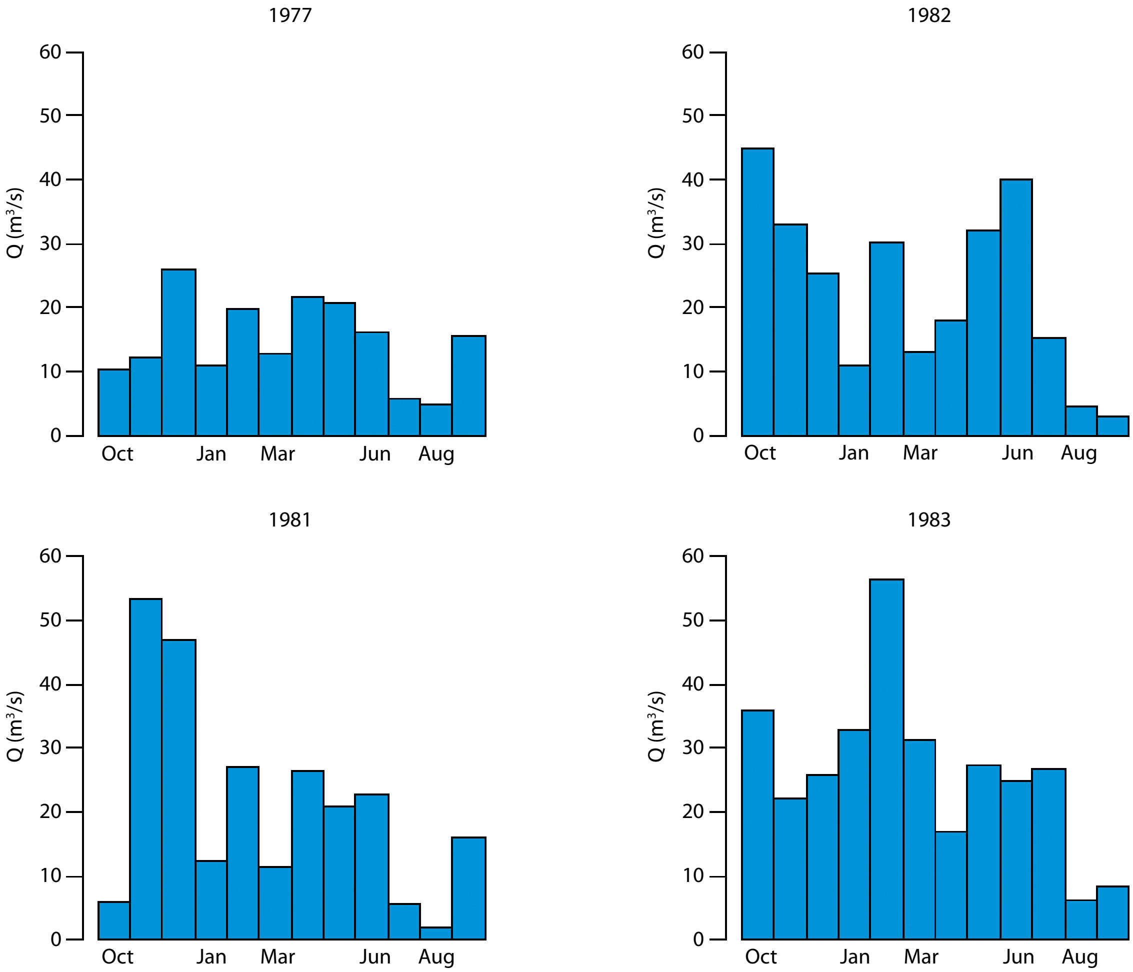

FIGURE 4.5 *Mean monthly discharge (Q) for Capilano River in 1977, and 1981–1983.*

glacier-fed streams, such as Lillooet River, appeared less sensitive to these influences.

Unfortunately, few rivers in British Columbia have records that capture all four phases of the PDO over the last century, limiting our ability to investigate geographic variability in response. Moore (1991), however, examined flow changes within the Fraser River drainage basin associated with the 1976–1977 shift, and found that the post-shift flows had decreased by about 20% on average, consistent with a coincident decrease in snow accumulation over southern British Columbia (Moore and McKendry 1996). Therefore, the PDO influences hydroclimatic patterns over at least the southern half of the province.

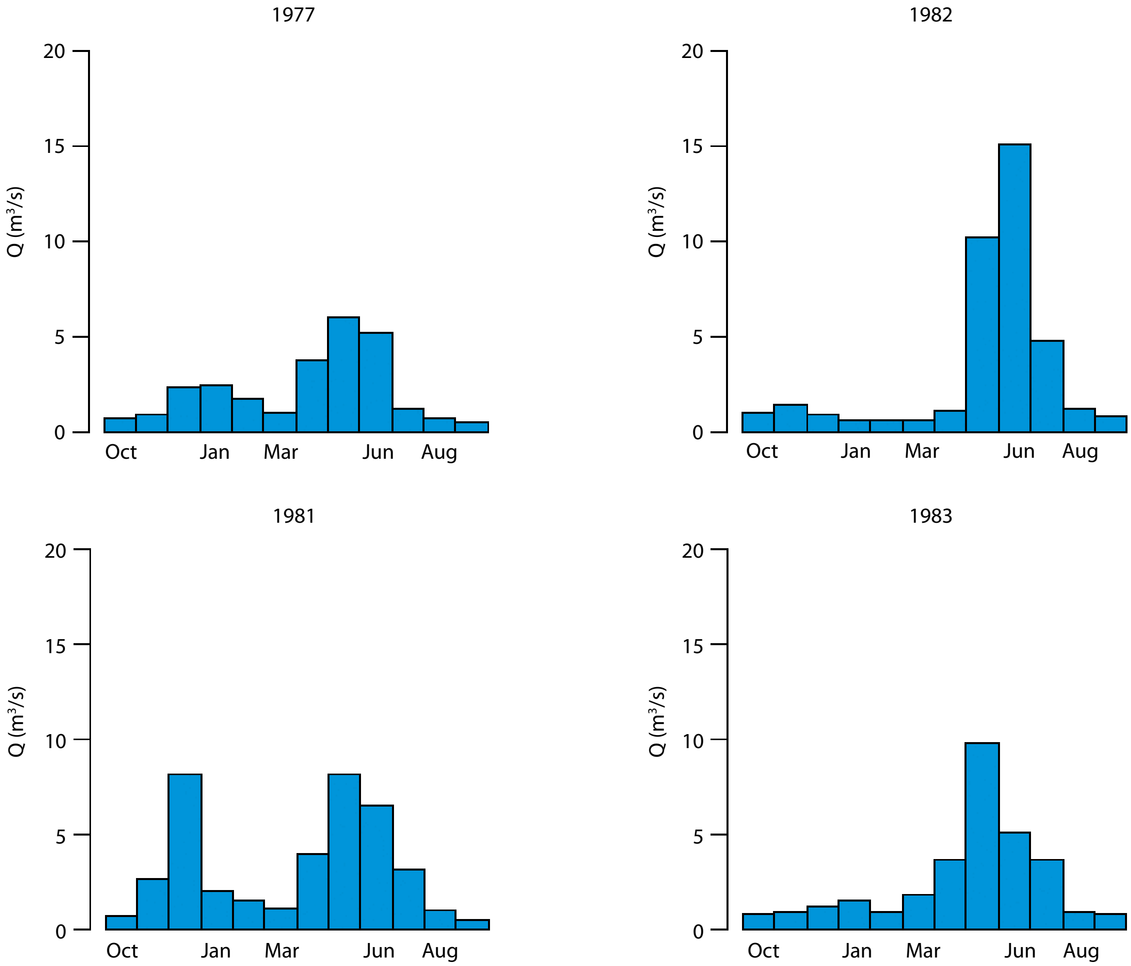

FIGURE 4.6 *Mean monthly discharge (Q) for Coquihalla River in 1977, and 1981–1983.*

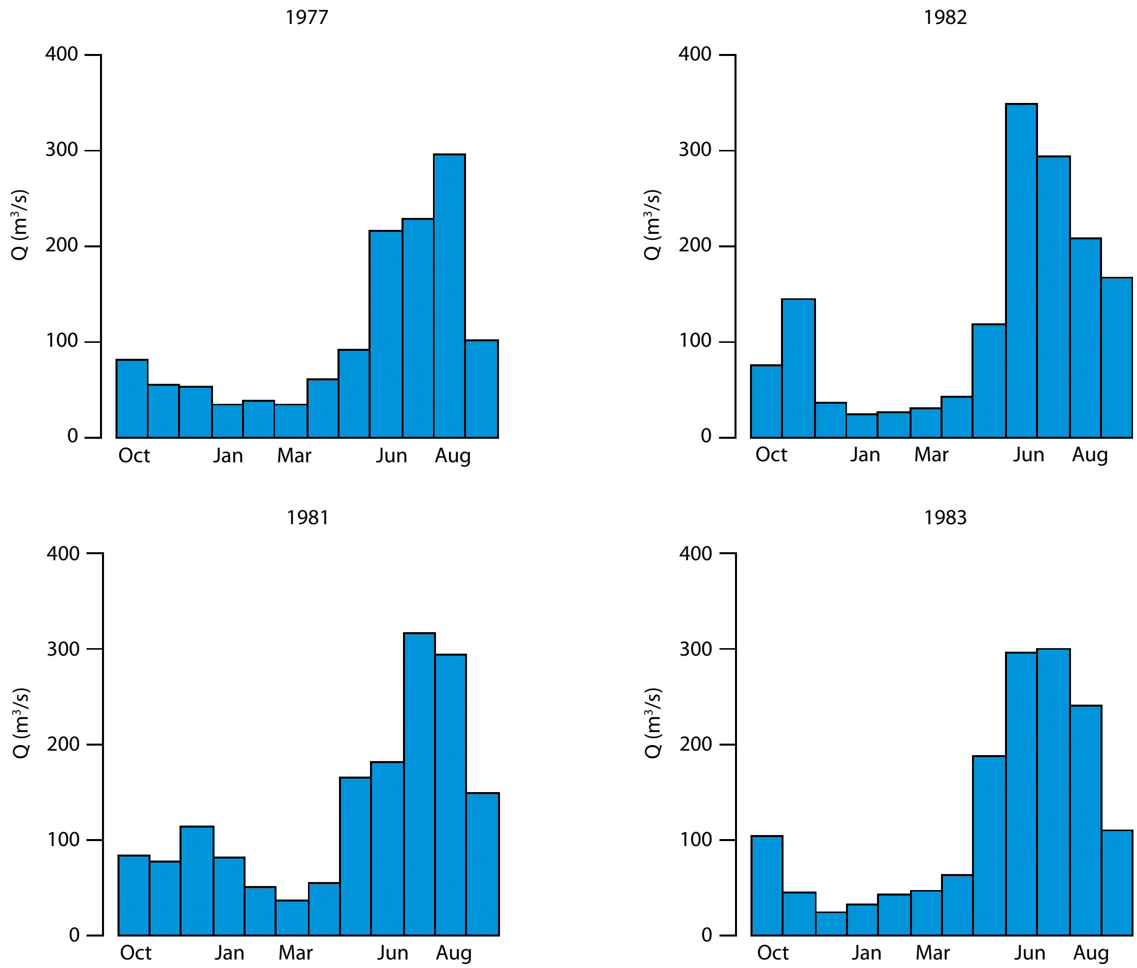

FIGURE 4.7 *Mean monthly discharge (Q) for Lillooet River in 1977, and 1981–1983.*

Seasonal streamflow regimes vary systematically across the province in conjunction with climate and physiography. Although coastal British Columbia exhibits various seasonal flow regimes as a result of the mixed streamflow-generating mechanisms, most of the Interior is characterized by snowmelt regimes that have similar streamflow patterns. The key differences between different regions of the Interior are: (1) the volume of streamflow per unit area, and (2) the between-month and between-year variability of the flow regime. In this section, the patterns are illustrated using mean monthly flow data for selected stations having similar drainage basin areas, organized into three west-to-east transects across the province (Figure 4.8; Table 4.1).

FIGURE 4.8 *Hydrometric stations in British Columbia used to illustrate the general spatial variations in monthly and peak flow characteristics. (Original map data provided by The Atlas of Canada http://atlas.gc.ca/ © 2007)*

TABLE 4.1 *Water Survey of Canada gauging sites related to each transect in Figure 4.8*

Station number	Station name	Latitude	Longitude	Drainage area above gauge (km²)	Mean basin elevation (m)	Period of record examined
Southern British Columbia						
08HB048	Carnation Creek at the mouth	48°54'56"N	124°59'52"W	10.1	453	1972–2003
08HB014	Sarita River near Bamfield	48°53'34"N	124°57'54"W	162.0	535	1948–2003
08HB025	Browns River near Courtenay	49°41'33"N	125°05'07"W	86.0	982	1960–2003
08GA026	Capilano River above Eastcap Creek	49°27'14"N	123°06'33"W	69.9	1042	1926–2003
08MG025	Pemberton Creek near Pemberton	50°19'02"N	122°48'05"W	31.9	1550	1987–2003
08MH056	Slesse Creek near Vedder crossing	49°04'16"N	121°41'58"W	162.0	1300	1957–2003
08NL036	Whipsaw Creek below Lamont Creek	49°22'09"N	120°34'11"W	185.0	1553	1964–1998
08NM173	Greata Creek near the mouth	49°47'40"N	119°51'04"W	40.7	1435	1970–2003
08NM171	Vaseux Creek above Solco Creek	49°14'58"N	119°19'16"W	117.0	1829	1970–2003
08NE117	Kuskanax Creek at 1040 m contour	50°20'36"N	117°31'05"W	113.0	1761	1973–1996
08NH016	Duck Creek near Wynndel	49°12'10"N	116°31'56"W	57.0	1624	1921–2003
08NK026	Hosmer Creek above diversions	49°35'03"N	114°57'14"W	6.4	1784	1981–2003
08NK022	Line Creek at the mouth	49°53'29"N	114°50'00"W	138.0	2110	1971–2003
08NP004	Cabin Creek near the mouth	49°05'39"N	114°33'11"W	93.2	1919	1977–2003
Central British Columbia						
08OA002	Yakoun River near Port Clements	53°36'50"N	132°12'35"W	477.0	356	1962–2003
08FF003	Little Wedeene River below Bowbyes Creek	54°08'11"N	128°41'24"W	182.0	881	1966–2003
08JA015	Laventie Creek near the mouth	53°39'09"N	127°32'13"W	86.5	1900	1976–2003
08JA016	MacIvor Creek near the mouth	53°48'02"N	126°21'36"W	53.4	n/a	1976–1995
08JA014	Van Tine Creek near the mouth	53°15'48"N	125°24'30"W	153.0	1394	1974–2003
07EE009	Chuchinka Creek near the mouth	54°31'45"N	122°36'00"W	311.0	1089	1975–2003
08KE016	Baker Creek at Quesnel	52°58'23"N	122°31'11"W	1570.0	1200	1963–2003
08KH019	Moffat Creek near Horsefly	52°18'50"N	121°24'20"W	539.0	1348	1964–2003
08KA009	McKale River near 940 m contour	53°26'41"N	120°13'10"W	252.0	1919	1971–2003
08NC004	Canoe River below Kimmel Creek	52°43'41"N	119°24'30"W	298.0	2050	1971–2003
Northern British Columbia						
08CG006	Forest Kerr Creek above 460 m contour	56°54'56"N	130°43'15"W	311.0		1972–1994
10CD005	Adsett Creek at km 386.0 Alaska Highway	58°06'22"N	122°42'56"W	109.0		1983–2003
07FB005	Quality Creek near the mouth	55°08'45"N	120°55'24"W	29.5		1978–2001

We have attempted to focus on basins with drainage areas less than a few hundred square kilometres, as those are typically of greatest interest in forest management. The best data come from a transect through the southern part of the province at a latitude of about 49°30'N (Table 4.1; Figures 4.8 and 4.9).

The transect through central British Columbia (latitude 53°N; Figures 4.8 and 4.10) includes a similar number of stations, but their drainage basins are larger, on average, and less well distributed across the province. The third and final transect (Figures 4.8 and 4.10) describes the northern part of the province; however, few data can be found and only three suitable hydrometric records were identified.

Southern British Columbia

A generalized pattern of seasonal flow regimes for the southern part of British Columbia is described with reference to the monthly hydrographs for selected hydrometric stations (see Figures 4.8 and 4.9). To facilitate comparison between stations, the mean monthly discharge for each station is expressed as a depth of water (in millimetres) averaged over the basin area. The maximum and minimum recorded monthly values at each station for the period of record are also shown in Figure 4.9 to indicate the between-year range of mean monthly flows. To quantify the month-to-month variation in the mean monthly flows, we defined a "threshold low-flow" value. The threshold value is arbitrarily defined as 25% of the mean annual streamflow, expressed in millimetres per month (i.e., the mean annual streamflow divided by 12 months). All months where the mean monthly flow is less than the threshold low-flow value are indicated by a shaded area on the hydrograph for the relevant station in Figure 4.9. Regions where low flows fall below the threshold are likely susceptible to water shortages for human consumers and for the aquatic ecosystems that these rivers support; however, a detailed assessment of instream flow requirements requires a much more detailed analysis than can be achieved in this survey of the regional hydrology of British Columbia.

Vancouver Island

The hydrologic regime of the outer coast of Vancouver Island is described by hydrographs from two adjacent stations, Sarita River and Carnation Creek, both of which have rain-dominated regimes typical of relatively low-elevation, coastal basins. Carnation Creek has a smaller drainage area (10.1 km^2) than Sarita River (162 km^2), and a lower mean basin elevation (453 vs 535 m). The difference in their annual streamflow (3850 mm/yr for Sarita River compared with 2600 mm/yr for Carnation Creek) is consistent with the difference in mean elevation. The mean monthly flows are highest in November and December, when the discharge is about 400–600 mm. Mean monthly flows are lowest in July, August, and September (approximately 40–50 mm per month), when the discharge is about 10% of the mean flows recorded in November or December. The mean flows in July, August, and September fall below the low-flow threshold defined above. Basins on the outer coast, with a significant proportion of the basin area at high elevation (above approximately 1000 m asl), have annual streamflows similar to Sarita River, but exhibit a mixed-streamflow regime with a similar seasonal pattern to that for Browns River on the east coast of Vancouver Island. Since most of the mountain peaks on Vancouver Island are less than about 1800 m asl, no large basins have entirely snowmelt-dominated streamflow regimes. Many of the drainage basins on eastern Vancouver Island (e.g., characterized by Browns River) produce a mixed regime. Browns River exhibits high mean monthly flows in November and December as a result of storm-related precipitation, as well as a snowmelt-related peak in May. Low flows occur on average during August and September. The eastern part of Vancouver Island is effectively in the rainshadow of the island's outer coast, producing lower annual and monthly flows. For example, the annual streamflow for Browns River is about 2000 mm/yr, and the largest monthly flows (in November, December, and May) are between 200 and 300 mm. The effect of the snowpack on the seasonal regime of Browns River is clear in the comparatively high flows during June and July. By August, however, the mean monthly flows in Browns River are similar to those for the outer coast stations, averaging about 30 mm.

The stations on Vancouver Island all exhibit a high between-year range in flows for all months, which is typical of rain-dominated regimes. The minimum recorded monthly flow for all three stations is less than 10 mm in July, August, and September, and is less than 100 mm in all other months except for November and December, illustrating that low-flow conditions can occur in any summer month, and that the monthly flows in the early fall, winter, and spring can be well below the mean values for those months. The maximum recorded monthly flows for these three stations also indicate that any

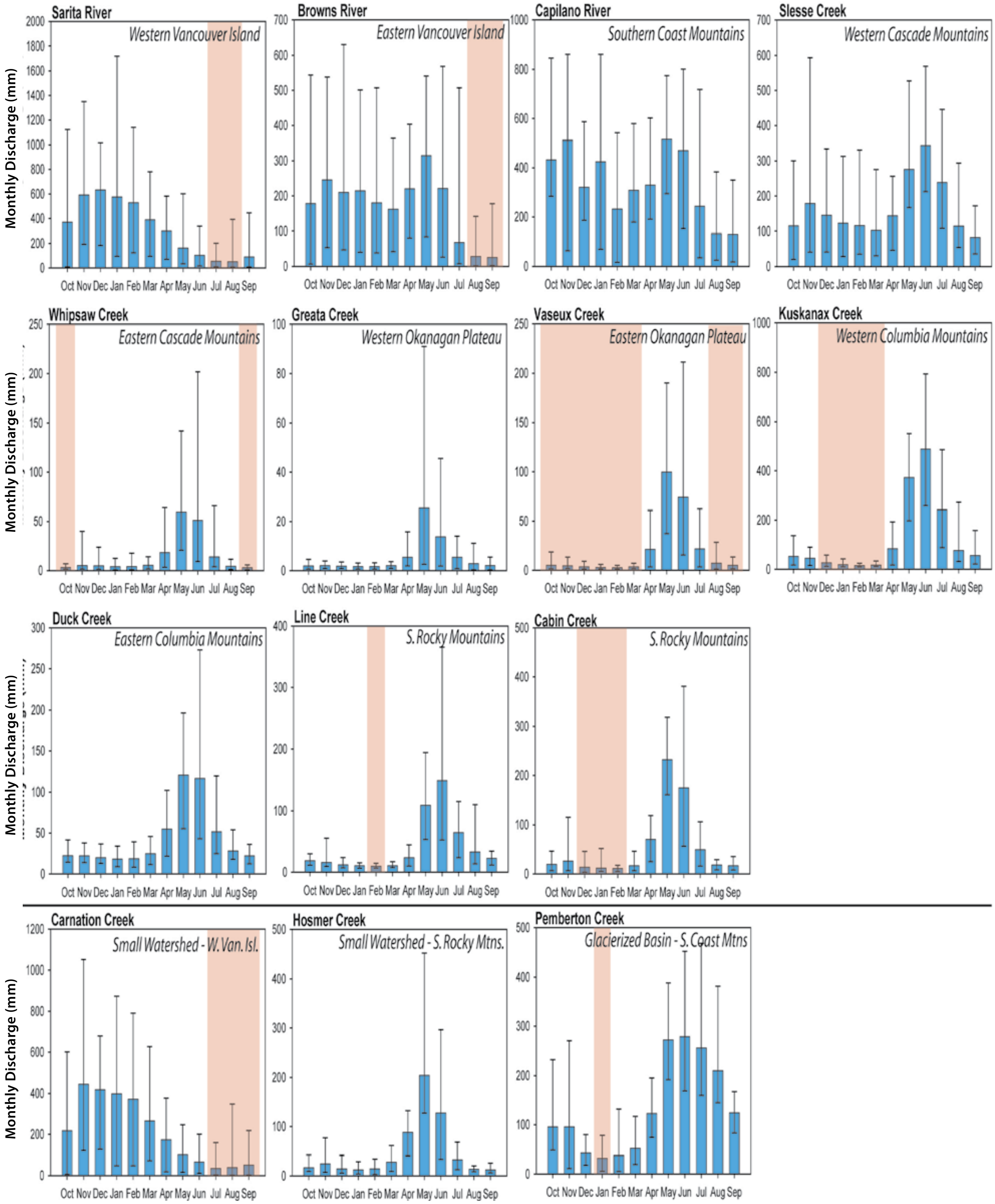

FIGURE 4.9 *Examples of average annual hydrographs for a transect of drainage basins across southern British Columbia. The height of the bars represents the multi-year mean streamflow for each month (millimetres per month); the error bar indicates the lowest and highest recorded monthly discharge in any year for the period of record.*

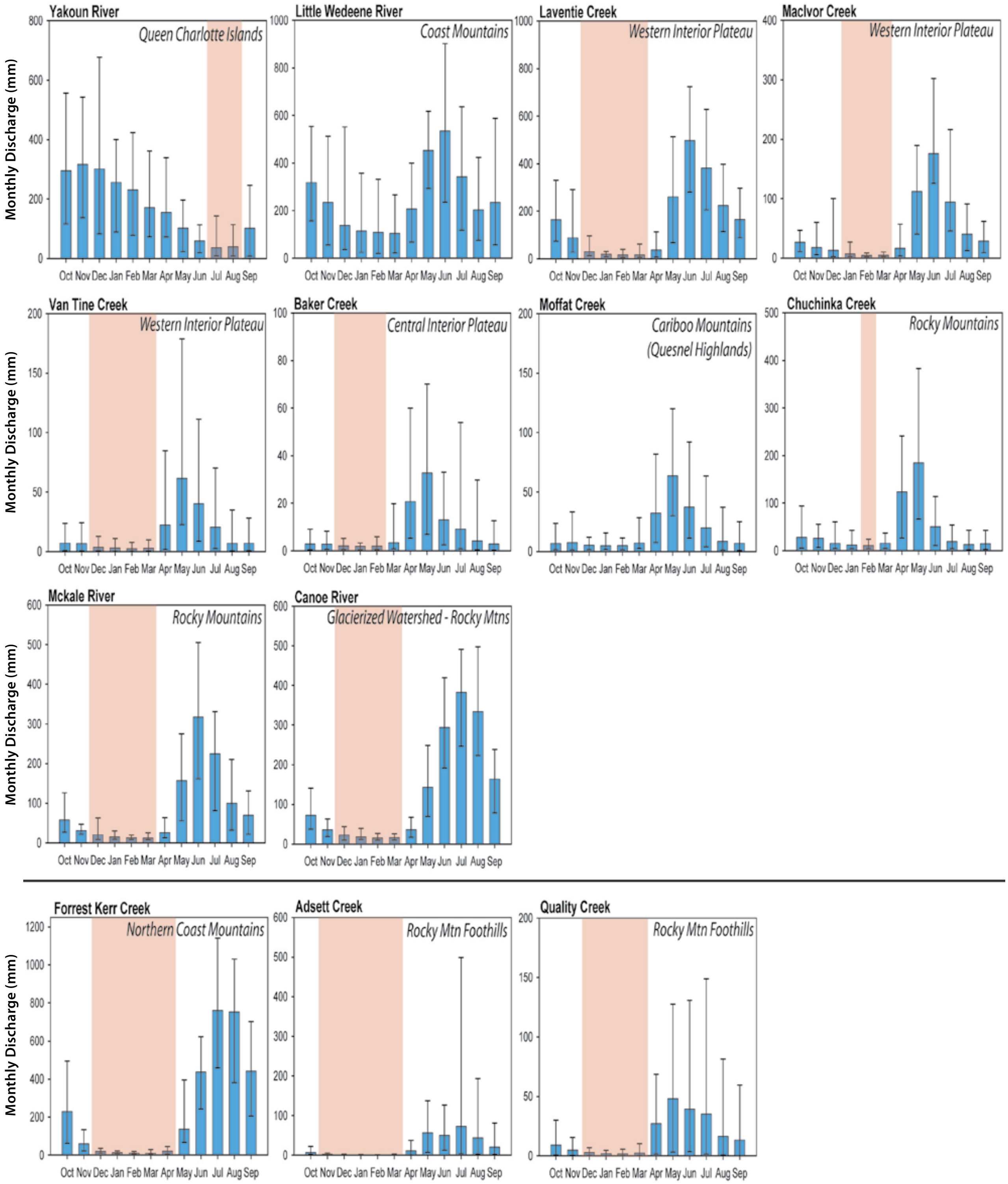

FIGURE 4.10 *Examples of average annual hydrographs for transects of drainage basins across central and northern British Columbia. The height of the bars represents the multi-year mean streamflow for each month (millimetres per month); the error bar indicates the lowest and highest recorded monthly discharge in any year for the period of record.*

month between October and March can be a "high-flow" month, having a flow greater than the highest mean monthly flow for the station. For the mixed-regime station (Browns River), any month except August and September can be a high-flow month.

Coast Mountains and Cascades

Across the Strait of Georgia, the height of the mountain peaks increases to 2000 m asl or more, resulting in an increase in orographic precipitation, a decrease in temperature, and an increase in the importance of the snowpack in the seasonal streamflow regime. Capilano River (above the Cleveland Dam) is a good example of the mixed-streamflow regime typical of basins in the Coast Mountains of southern British Columbia. The annual streamflow for this relatively high-elevation station (mean basin elevation is 1042 m) is similar to the outer coast of Vancouver Island (approximately 4000 mm/yr), as is the magnitude of the highest mean monthly flows (approximately 500 mm).

As expected for a mixed regime, the highest mean flows occur in November and May, in relation to heavy rainfall associated with frontal weather systems in the first instance and melting snow in the second. Low flows occur in August and September (having values of about 100 mm per month), but these flows do not typically fall below our definition of a threshold low flow. The between-year range is similar to that for the stations on Vancouver Island, and the highest recorded monthly flows for all months except August and September are greater than 500 mm. The record low monthly flows fall below 60 mm only in February, July, August, and September, indicating a reduced frequency of severe low flows, relative to Vancouver Island.

Slesse Creek occupies a similar environment, but is further inland from the coast and has a higher mean elevation (approximately 1300 m asl). Despite the higher basin elevation, the mean annual streamflow is about half that for Capilano River (approximately 2000 mm), similar to eastern Vancouver Island, reflecting the rapid decline in total precipitation with distance from the coast. The streamflow regime is dominated by snowmelt, producing a mean monthly peak of about 350 mm in June, although a smaller (approximately 180 mm), rain-related peak is evident in the mean monthly flow for November. The snowpack persists longer in the Slesse Creek drainage than in the Capilano watershed, and maintains relatively high flows (approximately 100 mm per month) even in August and September. The maximum recorded monthly flows exceed 350 mm only in November, May, June, and July, and the record low flows never fall below about 30 mm per month, indicating that the between-year range in seasonal streamflow is relatively low compared with stations that are more strongly affected by rain-related streamflow.

Some areas of the Coast Mountains are high enough and cold enough to be covered by glaciers. In these areas, snow and glacier melt dominate the seasonal streamflow pattern, though a weak rain-related streamflow component is evident in monthly discharge data. For example, Pemberton Creek (mean elevation of 1500 m asl) has a glacier covering roughly 25% of its drainage area. As a result, average high flows of between 200 and 275 mm per month occur in May, June, July, and August because of the seasonal melting of snowpack and glacier ice in the basin. Low flows typically occur from December to March, as precipitation contributes to snowpack storage and not streamflow. The record low flows indicate that only during the winter months, when most of precipitation is stored as snow and ice, do the flows ever drop below about 50 mm per month. The record high monthly flows for Pemberton Creek indicate that, although the highest mean monthly value (278 mm for June) is only ever exceeded during the glacier melt period, rainfall runoff generates smaller secondary monthly peaks in October and November.

The Cascade Mountains lie to the east of the Coast Mountains. Because precipitation declines rapidly with distance inland, as does temperature to a lesser degree, the Cascade Mountains are much drier than the Coast Mountains, particularly on their eastern aspects. For example, the annual streamflow for Whipsaw Creek is an order of magnitude less than any of the coastal drainage basins discussed above (approximately 180 mm/yr). The streamflow regime is strongly dominated by snowmelt, with more than half the total annual streamflow occurring in May and June. From August to March, the mean monthly flows are less than 6 mm, with record low flows below 2 mm for the same period. The mean monthly flows for September and October fall below the low-flow threshold. The record high flows exceed about 50 mm per month during April, May, June, and July alone, reflecting the variable timing of seasonal snowmelt and possibly the varying role of rainfall during the snowmelt freshet. The record monthly flow for November, about 40 mm, is the only evidence of a rain-related streamflow regime.

Interior Plateaus

The Interior Plateaus west of the Okanagan Valley are even drier than the eastern slopes of the Cascade Mountains. Greata Creek, draining off the plateau into Okanagan Valley near Summerland, has a mean annual streamflow of only 70 mm, despite having a mean basin elevation similar to Whipsaw Creek. High flows caused by snowmelt occur during May and June, falling to less than 3 mm per month between August and March. The record low monthly flows for every month are less than 3 mm, indicating that in some years, almost no snowmelt-generated streamflow occurs. Despite this, none of the mean monthly flows falls below the low-flow threshold because the annual streamflow is so low. Conversely, the record high monthly flow for May (91 mm) exceeds the mean total annual streamflow. In this relatively arid environment, the between-year range is large in the streamflow regimes.

The Interior Plateaus east of the Okanagan Valley are slightly higher, and tends to accumulate deeper snowpacks, producing slightly higher annual streamflow. Vaseaux Creek has a mean annual streamflow of about 250 mm, similar to Whipsaw Creek in the Cascades. The mean seasonal pattern and between-year range are similar to Greata Creek, with record high flows for May and June approaching the mean total annual streamflow volumes and a fairly narrow range of low flows between August and March; however, the mean monthly flows for all months between August and March fall below the threshold low flow, indicating persistent and severe water limitations for most of the year in this part of British Columbia.

Columbia Mountains and Rockies

East of the Interior Plateaus, the Columbia Mountains rise to more than 2500 m asl. These ranges induce another phase of orographic precipitation, resulting in higher annual streamflow values than on the plateau. This region of British Columbia is represented by Kuskanax Creek, which has an annual streamflow of about 1500 mm, of the same order as values at the coast and much greater than those for the Interior Plateaus. The monthly discharge for May and June is on average between about 400 and 500 mm, which is equivalent to the highest mean monthly for the coastal stations. Unlike the coast, however, flows of this magnitude only ever occur during the snowmelt season between May and July. Typically, flows decline more or less continuously after the peak in June, reaching threshold low-flow levels between December and March.

For Kuskanax Creek, the between-year range of monthly flows in May and June is relatively small, and is typically about ± 50%. Since the coastal basins with a dominant snowmelt component have similar ranges for May and June, this relatively conservative behaviour seems typical of humid, snowmelt-dominated streamflow regimes. The more arid parts of the province may have a greater year-to-year variation in snowpack, and consequently in the range of monthly flows, as at Greata Creek. The annual streamflow regimes for Duck Creek (520 mm), Line Creek (480 mm), and Cabin Creek (660 mm) illustrate the west-to-east trend through the Columbia and Rocky Mountains. Duck Creek flows west into Kootenay Lake near Creston; Cabin Creek flows east into the Flathead River near the British Columbia–Alberta border; and Line Creek flows west from the border into the Elk River north of Sparwood. The streamflow regimes for all three stations are similar in pattern and magnitude, and indicate a snowmelt-dominated streamflow regime with a fairly small year-to-year variation in the monthly flows. Although the highest mean monthly flow for these stations occurs variously in May or June, the highest monthly value on record for all stations invariably occurs in June. This implies that the largest monthly flows in snowmelt regimes are a result of the delayed melt of a substantial snowpack. In fact, the highest monthly flow on record occurs in June for all of the snowmelt-dominated stations except for Greata Creek, which has the lowest mean elevation of all the snowmelt-dominated stations on the southern transect.

The streamflow regime for Hosmer Creek (near Line Creek) is presented to illustrate the effect of scale on snowmelt-dominated seasonal streamflow regimes. The main difference between Hosmer Creek and Line Creek is the occurrence of both the highest record and highest mean monthly flow in May rather than in June, a result of the somewhat lower mean basin elevation (Table 4.1). In addition, Hosmer Creek's basin has a maximum elevation of about 2200 m, over 1000 m less than that for Line Creek. Thus, the timing of the peak monthly flows probably reflects the earlier onset of snowmelt at lower elevations.

Central British Columbia

The seasonal flow regimes for Central British Columbia are described with reference to the monthly hydrographs for the hydrometric stations shown in Figure 4.10. The variation from west to east is roughly similar to that for the southern transect, but reflects systematic differences in climate, linked to the higher latitude, and in the physiography.

Haida Gwaii

Streamflow regimes on Haida Gwaii (previously known as the Queen Charlotte Islands) are described by the monthly hydrograph for Yakoun River, which has an annual streamflow of 2060 mm. The pattern of peak and low flows and the between-year range for each month are similar to those for Sarita River and Carnation Creek on Vancouver Island to the south. The monthly flows (both mean and maximum) are slightly lower for Yakoun River, reflecting its lower mean basin elevation and its position some distance inland from the exposed west coast. The basins on the exposed west coast likely have monthly discharge values at least as high as that for Sarita River. A shift is also apparent in the beginning of the rainy season for Yakoun River, where the monthly flows for October are almost as high as those for November and December. Given the relatively low elevation of the Queen Charlotte Ranges, the seasonal streamflow regimes for nearly all of the drainage basins on Haida Gwaii are likely dominated by rain.

Coast Mountains

East of Haida Gwaii, the coast of the British Columbia mainland rises steeply, culminating in mountain peaks slightly lower than those in the southern Coast Mountains. Little Wedeene River, near Kitimat, has a relatively high annual streamflow of about 2990 mm, similar to Capilano River in the south. Little Wedeene River exhibits a mixed regime, with both rain- and snowmelt-related monthly peak flows. Despite having a lower mean elevation than the Capilano basin (881 vs 1042 m asl), Little Wedeene River exhibits a stronger snowmelt component, and the highest monthly flows (mean and record) occur in June. A shift in the timing of the rain-related peak monthly flows is also apparent, with the maximum occurring in October rather than November. These differences are attributable to climatic gradients associated with latitude, which result in a colder climate with earlier onsets of both wet weather and the shift from rainfall to snowfall in central British Columbia.

Interior Plateaus

Moving inland, the terrain grades into a high plateau with fairly low relief that slopes gently toward the Fraser River to the east. The mean basin elevations for Laventie, MacIvor, Van Tine, and Baker Creeks reflect this general physiography. The seasonal streamflow patterns for all of these stations are dominated by snowmelt. The total mean annual streamflow declines from west to east, with values of 1900, 540, 180, and 100 mm for Laventie, MacIvor, Van Tine, and Baker Creeks, respectively. The timing of the maximum monthly flows also changes. For the higher-elevation stations (Laventie and MacIvor Creeks), the maximum occurs in June, fed by relatively high-elevation snowpacks. For the lower-elevation stations (Van Tine and Baker Creeks), the peak occurs in May. All of these stations typically experience extended periods of low flows below the low-flow threshold during the winter months. In fact, this is generally true for all snowmelt-dominated regimes at this latitude. The between-year range of variability (based on the record high and low flows for each month) also varies consistently, and tends to become relatively larger (as a percentage of the mean) as the streamflow declines from west to east.

Cariboo Mountains and Rockies

East of the Fraser River, the terrain rises again, forcing an increase in the amount of precipitation. Moffat Creek, in the Quesnel Highlands, has a mean basin elevation similar to Van Tine Creek. It also has a very similar seasonal streamflow regime, and total annual streamflow (200 mm for Moffat Creek vs 180 mm for Van Tine Creek). McKale Creek in the Rocky Mountains is much higher and wetter, with an annual streamflow of 1040 mm. The Rocky Mountains at this latitude are not shielded by high mountains to the west, as they are in the south, and therefore the mean annual streamflow is about twice that for basins in the south. As expected for basins at relatively high elevation, the high flows for McKale Creek occur in June. Where glaciers exist (see Canoe River, annual streamflow of 1540 mm), the annual streamflow is higher—the result of an orographic precipitation effect—and the highest flows occur in July, with high flows occurring from May right through to the end of September.

There are local, physiographically related deviations from this west to east pattern. For example, Chuchinka Creek (54°30'N) is in a physiographic position similar to Moffat Creek to the south. Despite a lower mean basin elevation, however, it has nearly twice the mean annual streamflow (510 mm) of Moffat Creek. This difference is likely because the Coast Mountains decline in height from Bella Coola (52°30'N) to Prince Rupert (54°30'N), thereby having a reduced effect on weather systems tracking in from the North Pacific at the higher latitudes.

Northern British Columbia

The physiography of northern British Columbia differs from that to the south. The Coast Mountains increase in height north of Prince Rupert, which, combined with the latitudinal effect on the mean annual temperature, results in extensive alpine glacier cover. The seasonal streamflow regime for Forrest Kerr Creek (2890 mm) is probably typical of the drainage basins near the western border of northern British Columbia (Figure 4.10). The highest monthly discharge of these basins occurs in July and August, with virtually no flow during the cold winter months when most of the incoming precipitation is stored in the basin as snow and ice.

Unlike the southern parts of British Columbia, no well-defined interior plateau occurs east of the Coast Mountains, except for the relatively limited Stikine Plateau in the northwest corner of the province. Instead, one encounters first the Omineca and Cassiar Mountains, and then the northern Rocky Mountains. Glacier cover is limited in these mountain ranges, which typically exhibit a snowmelt-dominated regime. The available data are too sparse to describe the pattern of variation across these mountain ranges, but the mean annual streamflow on the eastern slopes of the Rocky Mountains falls to about 200–250 mm. The seasonal streamflow patterns for Adsett Creek (270 mm/yr) and Quality Creek[1] (200 mm/yr), which are both relatively low-relief drainage basins verging on the Alberta Plateau, indicate that the peak streamflow months occur between May and July, which is slightly later than for stations to the south. In fact, the highest monthly discharge on record for both stations occurs in July, whereas it occurs consistently in May or June for the snowmelt-dominated regimes in southern and central British Columbia. Snowmelt in these low-relief basins consistently peaks in May, and the high flows in June and July are attributable primarily to frequent rain produced by frontal systems moving south from the Arctic Ocean. The winter flows are also lower than to the south, with almost no streamflow occurring from November to March.

PEAK FLOWS

Stream channel dimensions are largely formed during the highest flows regularly carried by a stream channel. One common way of indexing these flows is by using the peak daily or peak instantaneous discharge occurring in a given year. The design of any stream crossings or instream structures, such as fish habitat enhancements, must consider the magnitude and frequency of these so-called channel-forming flows. The mechanisms that produce peak flows vary systematically across the province. Since land use activities are often managed based on assumptions about the way in which human activities (e.g., forest harvesting or road building) influence peak flows, it is important to understand how, when, and why these peak flows actually occur. The differences between the Coastal Watershed Assessment Procedures and the Interior Watershed Assessment Procedures described in the original *Forest Practices Code of British Columbia Act* were based primarily on the assumption that rain or rain-on-snow events produced floods at the coast, whereas snowmelt events produced floods in the Interior. Because more than one flood-generating mechanism can affect a given basin, it is probably inappropriate to classify watersheds as simply “coastal” or “interior” types, particularly in northern British Columbia where summer rain events seem to produce nearly all of the largest peak flows.

1 Although Quality Creek is relatively close to latitude 53°N, it is on the east side of the continental divide and is controlled by weather patterns typical of northern rather than central British Columbia. The scarcity of data in the northern part of the province requires that this station be included in what is an already sparse data set.

Regional Variations

The factors controlling peak flow magnitude are the same as those controlling the seasonal streamflow regime—that is, precipitation inputs and storage of water within a drainage basin. In British Columbia, peak flow events can be produced by rainfall, snowmelt, and glacier melt, and by combinations of these mechanisms. For example, rain-on-snow events commonly occur at intermediate elevations in coastal montane areas. In these events, melting of snow accumulated during earlier, colder storms contributes to streamflow in addition to rainfall, typically increasing the resulting peak flow. The main difference between the mechanisms generating peak flows and the factors controlling the seasonal streamflow regime is the time scale over which the processes act, rather than the processes themselves. In addition, glacier outburst events and breaching of moraine-dammed lakes can cause peak flows (Church 1988); however, these events are not influenced by forest management and are not addressed further. For more detailed coverage of peak flow generating mechanisms and the effects of forest harvesting, see Chapter 6 ("Hydrologic Processes and Watershed Response") and Chapter 7 ("The Effects of Forest Disturbance on Hydrologic Processes and Watershed Response").

Comparisons of peak flows among basins of different drainage areas are difficult because the mean annual peak flow per unit area is not constant, even in hydrologically homogeneous zones. Smaller drainage basins typically exhibit higher peak flows per unit area than larger ones. To make a scale-independent comparison across the province, this scale effect needs to be removed by expressing the mean annual peak flow per unit area as a *k* factor, defined as:

$$k = \frac{Q_{peak}}{(A_d)^{\beta}}$$

where: Q_{peak} is the peak flow, A_d is the drainage area, and β is an exponent to account for the scale effect; in British Columbia, β is about 0.75. Figure 4.11 presents a map of *k* factors for the mean annual peak flow across British Columbia. See Eaton et al. (2002) for the details of the underlying analysis and a discussion of the limitations, as well as maps of *k* factors for the 5- and 20-year return period peak flows. For our purposes, it is sufficient to interpret the *k* factors in Figure 4.11 as measures of the mean annual peak flow per unit area for basins with the same drainage area, specifically 1 km^2. The maps for the 5- and 20-year floods are similar to those for the mean annual flood (k_{maf}), varying primarily in the absolute magnitude of the *k* value.

An analysis of the average ratio of k_{20} to k_{maf} by region shows that the largest differences occur where rain and rain-on-snow events produce floods and the smallest differences occur where snowmelt is the primary flood-generating mechanism (Eaton et al. 2002).

The text in the following paragraphs describing the *k* factor map are reproduced from Eaton et al. (2002) and only slightly modified. Figure 4.11 illustrates some fundamental aspects of British Columbia regional hydrology. The major pattern of peak flows follows the structural–topographic grain of the province that controls the distribution of precipitation (see precipitation map in Chapter 3, "Weather and Climate") and is also reflected in regional patterns of annual streamflow (e.g., see Waylen and Woo 1982a, for a map of streamflow patterns in the Fraser River basin, which encompasses most of southern British Columbia). The coast has substantially higher streamflow than anywhere else, resulting from frequent storms originating in the Pacific Ocean, which subsequently encounter the mountainous coast.

The gradients in the parameter *k* are steep along the entire coast, where values range from about 1 to more than 8. This steep hydrologic gradient is primarily due the relatively intense orographic uplift associated with the Coast Mountains.

The plateaus of the southern and central Interior have relatively low peak flows, with flows nearly two orders of magnitude less than in the wettest parts of the coast. The interior plateaus have characteristically low relief, and are strongly influenced by the rain-shadow produced by the Coast Mountains; however, the hydrologic gradients in the interior plateaus are even steeper than along the coast. The *k* values range from less than 0.1 in the centre of the plateaus to 1 along the eastern and western margins. Low peak flows also occur on the Alberta Plateau in the northeast part of the province.

The eastern Cordillera (Rocky Mountains and Columbia Mountains), where substantial orographic uplift occurs again, has *k* values between 1 and 2. The floods generated in this part of British Columbia are more strictly limited in between-year variability (ultimately, by the energy available for snowmelt) than are rainstorm-generated floods typical along the coast. The hydrologic gradients are much less strong here, reflecting primarily the influence of the relatively high (and wet) Columbia Mountains.

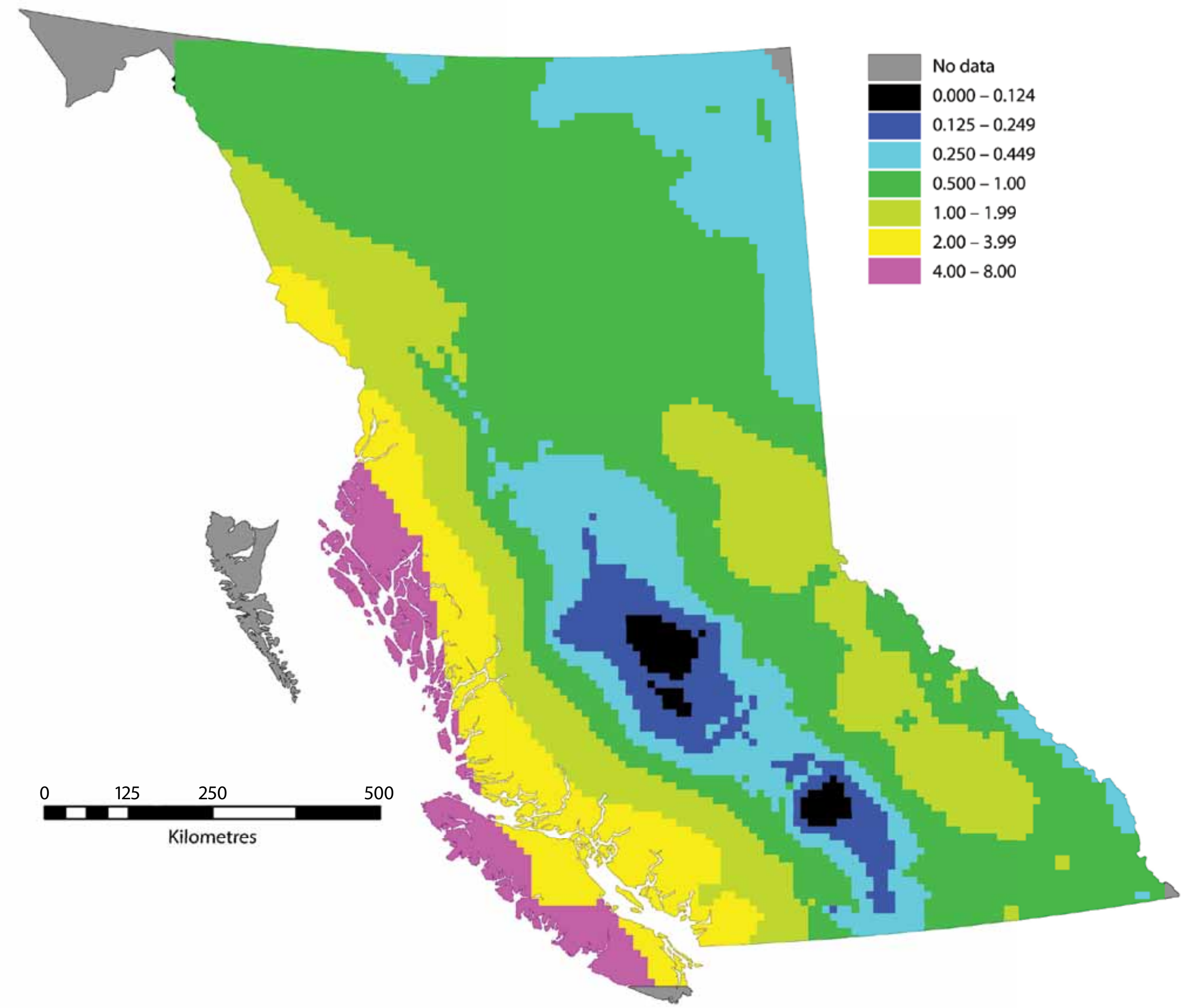

FIGURE 4.11 *Map of* k *factors, representing the pattern of mean annual peak flows over British Columbia.*

Peak Flow Timing and Mechanisms

Low-relief coastal basins

For basins where the seasonal streamflow regime is dominated by rainfall inputs (e.g., Carnation Creek and Sarita River in the south; Yakoun River in central British Columbia), the annual peak flows typically occur in the fall and winter (see Figure 4.12). On the central coast of British Columbia, annual peak flows may occur earlier than on the south coast, often in mid- to late September; in the south, these flows typically do not occur before mid-October. Annual peak flows in these rain-dominated drainage basins generally occur no later than the end of February, both in the southern and central coastal regions of British Columbia. The peak flow dates appear to be fairly uniformly distributed throughout this period, and the annual peak flow magnitude shows no systematic trend.

The highest annual peak flows in many coastal drainage basins most likely result from warm frontal systems combined with a wet snowpack; thus, melting of the snow and the inputs of large volumes of rain both contribute to the peak flow. These "rain-on-snow" events will be more common for basins at moderate to higher elevation, but will occasionally occur in the low-elevation coastal basins.

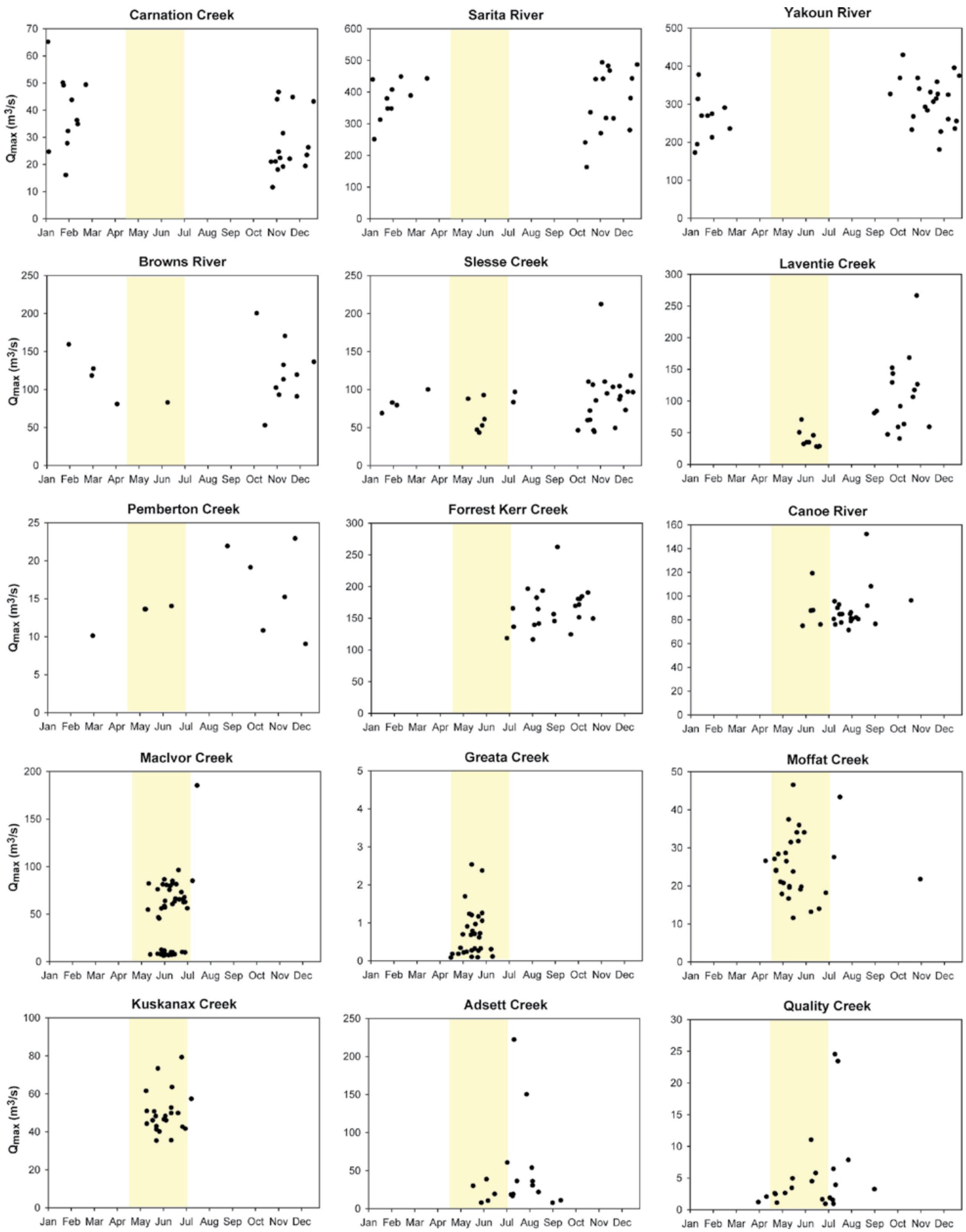

FIGURE 4.12 *Temporal distribution of annual peak flows for a selection of British Columbia drainage basins.*

High-relief coastal basins

The coastal stations with significant snowmelt contributions to their seasonal streamflow regimes (Browns and Slesse Creeks in the south, Laventie Creek in central British Columbia) may be subject to annual peak flows over many months in response to various peak flow generating mechanisms. Annual peak flows may occur in the fall in these environments as a result of rain events, occurring as early as September in central British Columbia and October in southern British Columbia. These cyclonic rain events occur less frequently in the winter, especially in central British Columbia, because much of the incoming precipitation is stored in high-elevation snowpacks and so does not immediately produce streamflow. Another distinct population of peak flows occurs in May, June, or July in response to seasonal snowmelt, when the water stored in the snowpack is released; however, the highest annual peak flows on record for Browns, Slesse, and Laventie Creeks invariably occurred in October or early November. These peak flows are likely the result of an early snowfall, followed by a warm frontal system producing rain-on-snow events that are much larger than those that could be generated by either rain or snowmelt events alone. Since various mechanisms can generate peak flows, the timing and magnitude of the peak flow in any given year are difficult to predict. Note that some coastal basins contain significant glacier cover, the effect of which is described separately below under the heading of glacierized basin.

Interior basins

The interior parts of British Columbia that are not influenced by glaciers exhibit a more predictable peak flow timing that usually depends on both elevation and latitude. The peak flows are frequently generated by snowmelt, although peak flows related to convective rainstorms do occur occasionally, and may indeed produce some of the largest flood events. For snowmelt events, the peak flow magnitude depends on the amount of water stored in the snowpack and the climatic conditions producing melt; for floods produced by convective storm cells, the degree of surface heating and availability of moisture determine the intensity and duration of the rainstorm and hence the magnitude of the flood.

In general, snowmelt-dominated peak flows in central and southern British Columbia occur between early May and late June (see MacIvor Creek, Figure 4.12), although both mean basin elevation and snowpack depth modulate the peak flow timing. Lower-elevation stations, such as Coldstream Creek (Figure 4.13), may experience annual peak flows as early as April and typically no later than late May (Greata and Moffat Creeks, Figure 4.12). The earlier peak flows tend to be smaller, reflecting a relatively small winter snowpack that is quickly exhausted with the onset of temperatures high enough to initiate seasonal snowmelt. The largest events in these basins occur in mid- to late May, and are generated from a snowpack that persists long enough to experience higher daily temperatures and solar radiation, leading to accelerated melt rates. Higher-elevation basins that develop deeper snowpacks typically experience peak flows between early May and early July (see Kuskanax Creek, Figure 4.12; Harper Creek, Figure 4.13). This shift in peak flow timing is the product of lower mean daily temperatures (and thus melt rates) at higher elevations and deeper snowpacks that persist into the warmer summer months. In many of these basins, convective storms in June, July, and August may either augment snowmelt-generated streamflow (Coldstream Creek, Figure 4.13), or produce entirely rainstorm-driven floods (Vaseaux and Van Tine Creeks, Figure 4.13).

In the northern part of the province, the timing of the peak flows is more variable, usually occurring between May and July, and sometimes as late as August, as illustrated by Adsett Creek (Figure 4.13). This shift is attributable to climatic differences, whereby frequent, large convective rainstorms in the northeastern part of British Columbia typically produce most of the significant flood events.

Glacierized basins

In drainage basins that have significant glacier cover, intense melting of glacier ice in the summer drives a distinct population of floods, which are superimposed on the rain, rain-on-snow, or snowmelt-generated peak flow regimes that are otherwise typical of the region. For instance, the annual peak flows at Pemberton Creek can occur in May or June in response to snowmelt, in August or September in response to glacier melt, or in October, November,

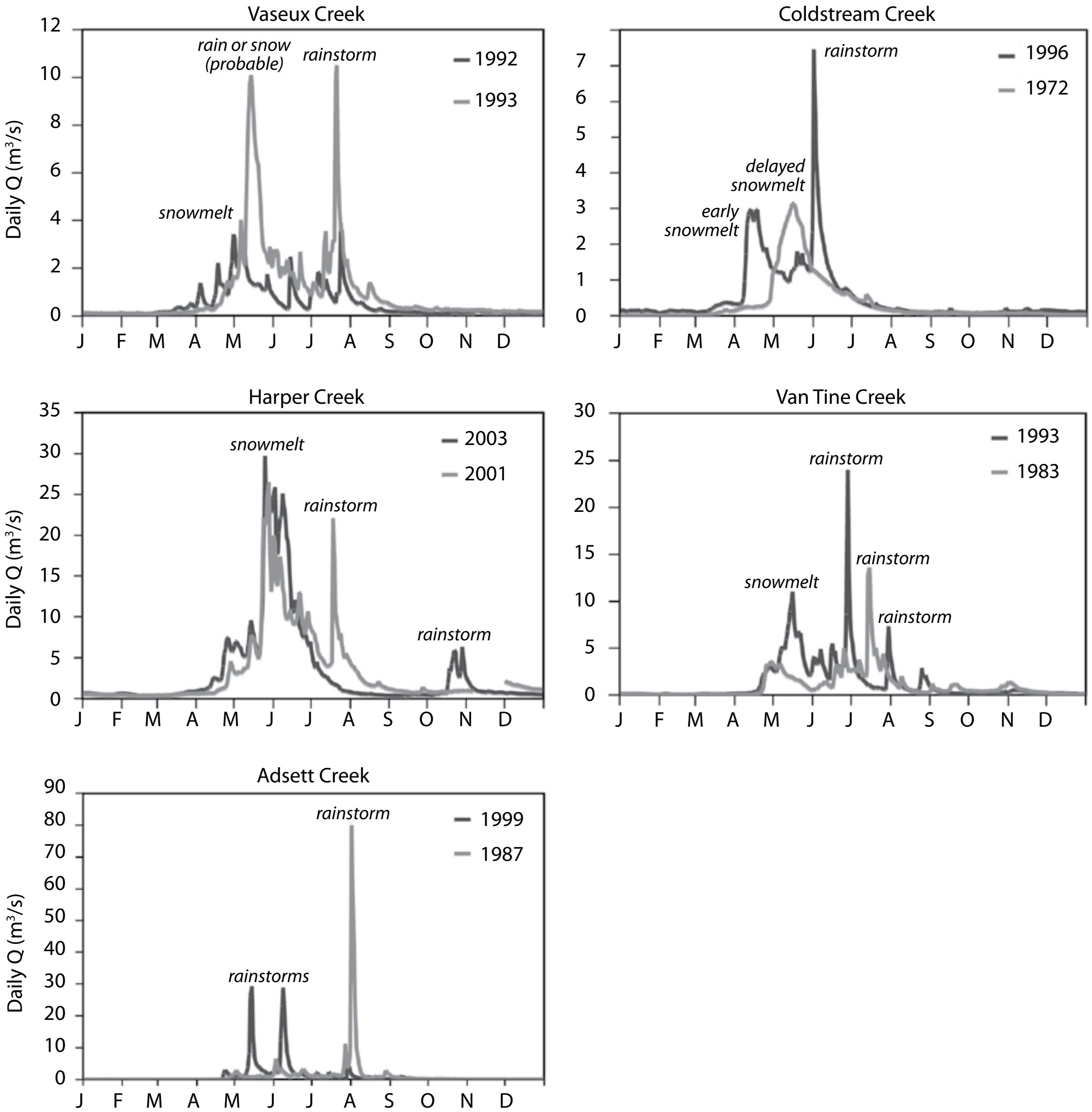

FIGURE 4.13 *Daily data for various hydrometric stations illustrating the range of possible flood-generating mechanisms throughout interior British Columbia.*

and December as a result of warm frontal systems producing rainfall with melting of either glacier ice or an early snowpack. In the northern part of the province, Forrest Kerr Creek also experiences annual peak flows between June and late October, likely as a result of snowmelt, glacier melt, or rain-on-snow peak flow generating mechanisms.

In the Rocky Mountains, the timing and generation of peak flows in glacier-covered basins is similarly variable: Canoe Creek may peak in May or June as a result of snowmelt, or peak in July, August, or September as a result of glacier melt. A rain or rain-on-snow event occurred in Canoe Creek during October.

Several studies have examined the potential effects of future climate change, usually employing scenarios based on output from general circulation models (also increasingly known as "global climate models"). As noted in Chapter 3 ("Weather and Climate") and Chapter 19 ("Climate Change Effects on Watershed Processes in British Columbia"), the models all generally predict warming, though they vary greatly in the amount of warming and its seasonal expression, and do not agree in relation to precipitation.

The effects on rain-dominated systems should broadly reflect the changes to rainfall patterns (Loukas et al. 2002). For nival regimes, especially in southern British Columbia, the warming trend should result in a shorter snow accumulation season with an earlier freshet, resulting in lower flows in late summer and early autumn (Loukas et al. 2002; Merritt et al. 2006). Indeed, some model scenarios suggest that Fraser River, currently a nival regime, may become more rainfall-dominated as a result of future warming (Morrison et al. 2002). For hybrid regimes, the nival component should become weaker or even non-existent, resulting in more winter rainfall and lower spring–summer flows, attributable to the loss of seasonal snowmelt contributions. Glacial regimes may be less affected by climatic change in the early stages of warming, as increased melt could enhance glacier runoff. In the longer term, however, glacier retreat would result in less ice area available for melt, and thus declining streamflow contributions. In fact, Stahl and Moore (2006) found that glacier-fed streams in British Columbia dominantly exhibited negative trends in August streamflow since the 1970s, a period marked by significant glacier recession in many regions.

SUMMARY

The seasonal streamflow regime and peak flow characteristics vary systematically across British Columbia as a result of systematic variations in precipitation and the way in which precipitation is stored and released once it reaches the ground surface. The seasonal streamflow regimes can be broadly classified into regimes dominated by inputs of precipitation as rain, regimes dominated by runoff generated by melting snow or glacier ice, and regimes that reflect both rain and melt contributions.

Rain-dominated regimes characteristically have higher flows during the fall and winter, when a persistent low-pressure system brings storms over British Columbia from the Pacific Ocean. Low flows occur during the summer as a result of the development of a persistent, blocking high-pressure system that directs storms to the north of British Columbia, and basins with rain-dominated regimes appear most susceptible to water shortages overall, especially during the growing season. Since the sequence of weather systems that occurs drives streamflow in these basins, the seasonal streamflow pattern varies substantially from year to year, and seasonal streamflows for a given year are no more predictable than the weather.

Snowmelt-dominated regimes characteristically exhibit high flows in May or June as a result of melting snowpacks, and low flows in the winter, when precipitation is accumulating on the ground as snow. Low flows can also occur in late summer or early autumn. Glacier-melt regimes are similar to snowmelt regimes, but melt-related high flows persist into the hottest summer months. The seasonal streamflow pattern for these basins does not vary much from year to year because the basins effectively integrate the weather-driven precipitation inputs over the fall and winter, and then release the stored water in the spring and summer, often overwhelming any weather-driven inputs of rain.

The spatial distribution of these regime types throughout British Columbia is determined primarily by the elevation of a drainage basin and the distance of the drainage basin from the coast; it is influenced to a lesser degree by the latitude of the drainage basin. Generally, the low-elevation coastal basins have a seasonal streamflow regime dominated by rainfall whereas the drainage basins in the Interior exhibit a snowmelt regime. High-elevation drainage basins near the coast often exhibit a mixed regime, showing components of both the rain and snowmelt regimes. The highest-elevation basins on the coast and in the Interior tend to contain glaciers, producing a glacier-melt regime. The local physiography also affects the characteristic streamflow regimes. For example, basins on the plateau tend to accumulate modest snowpacks that melt everywhere at nearly the same time, producing small annual streamflow values, but a short, steep snowmelt hydrograph. In contrast, mountainous drainage basins tend to intercept considerably more precipitation than the plateaus because of orographic uplift; however, snow melts first at the lower elevations and later in the season at higher elevations, producing high annual streamflow values with a longer, gradual snowmelt hydrograph.

Peak flows follow a similar geographic pattern, and can be generated by rain events, snowmelt, glacier melt, or some combination of rain and melt inputs of water. Typically, rain and rain-on-snow events are dominant on the coast, whereas melt or rain-on-snow events are dominant in the Interior. Rain events produce peak flows in nearly all drainage basins throughout British Columbia, and in some areas (particularly in northern British Columbia) may produce almost all of the largest flood events. Most drainage basins in British Columbia exhibit a mix of flood-generating mechanisms, which is significant for flood frequency analysis.

Conventional methods as described in standard textbooks (e.g., Linsley et al. 1982) assume that a single, common mechanism (e.g., seasonal snowmelt) generates all floods. It is more appropriate, although less common, to apply methods that acknowledge and account for heterogeneous flood-generating mechanisms (Waylen and Woo 1982b; Alila and Mtiraoui 2002).

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CHAPTER 5

Forest Practices

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INTRODUCTION

The forests of British Columbia have had, and continue to have, a powerful influence on the lives of the province's residents. Most British Columbians live surrounded by forests. Their houses are largely built of wood and their employment often depends on some aspect of the forest industry, whether they live on a farm, in a logging town, or within the metropolis of Vancouver. Young and old recreate in the forest, whether on ski runs or by lakeshores. Creative artists have drawn inspiration from the wild forest and the busy goings-on of a forest industry set within the stunning backdrop of ocean and mountains. Children learn that many provincial emblems are derived from the forest, including the dogwood flower, the western redcedar tree, and the Steller's Jay, a noisy, blue-black, forest-dwelling bird. For all British Columbians, their experience of the forest is inextricably linked with water.

From a human perspective, water supplies for communities of every size flow directly from forested mountain slopes. Watercraft fashioned from trees were used by Aboriginal communities, and then European explorers, to traverse the lakes and rivers of the Interior and the inlets of the Coast. Early commercial use of the forest relied on water for power and cheap transportation both by the ocean and along the inland waterways. Flumes, river driving, paddlewheelers, and ocean tugs all played a major role in British Columbia's forest history.

Past exploitation of the forest and the transition to forest management is also closely linked with water. Where there is too little water, or too much, forests do not grow. And within the range of adequate water, small changes have profound influences on the forest, its structure, and the constellations of organisms that live within its enveloping sphere of influence. Those same water supplies from forested mountain slopes support fish populations of immense size. As management of forests has evolved, care for the water resource, and its associated fisheries, has moved from a marginal issue to a touchstone for efforts to modify forest practices (Chapter 1, "Forest Hydrology in British Columbia: Context and History"). Public concerns about drinking water and fish habitat played a major role in this transformation. Contentious issues ranged from the widely publicized and intense battles over logging in the Greater Vancouver and Victoria watersheds (which ended in victory for the anti-logging side) to confrontations over logging and water quality in smaller communities throughout the Interior and, in particular, in the Slocan Valley where they persist. Very recently, the impact of the mountain pine beetle on forests and hydrological processes has reawakened concerns about the management of water throughout the province.

This chapter shows how some forest management practices in British Columbia may affect hydrological processes. Harvesting and regeneration systems are described followed by the transportation of wood from the forest to processing sites. The chapter ends with a brief overview of provincial forest policy.

Forest practices affect the forest's hydrological regime in three main ways (see Chapter 6, "Hydrologic Processes and Watershed Response," and Chapter 7, "The Effects of Forest Disturbance on Hydrologic Processes and Watershed Response").

- Forest cover removal: logging (patterns of logging), natural disturbances (fire, insects, diseases, windthrow, landslides / earth movements), fire prevention and suppression
- Regenerating the forest: natural regeneration, artificial regeneration through planting
- Transporting wood: movement of wood to mills and timber products to markets via water, rail, and road

Forest Cover Removal

Forest cover influences a number of important hydrologic processes, including interception of precipitation, snow accumulation and melt, and evapotranspiration (see Chapter 6, "Hydrologic Processes and Watershed Response"). Both natural and human-caused disturbances can remove forest cover, affecting hydrologic processes. Natural disturbances include wind, landslides, insects, diseases, and wildfire. Human activity can initiate or modify all of these natural disturbances, and cause direct disturbance through activities such as logging, road building, and settlement construction. Hydrologic processes can influence and be influenced by all of these disturbances to varying degrees. For example, excess soil moisture can intensify the effects of wind, leading to increased windthrow and (or) contribute to landslides; insect and disease activities are frequently associated with either an excess or shortage of precipitation and soil moisture; and fires often start during droughts and in certain situations may only be extinguished by prolonged precipitation.

Forest cover removal by logging is controlled through the amount of timber cut within a management unit and through the method of harvesting. The amount of timber volume harvested on Crown land in the Provincial Forest and Tree Farm Licences is determined by the Chief Forester, and controlled by limits placed on individual licensees under Section 8 of the *Forest Act* and through subsequent legislation, standards, and guidebooks that govern forest practices and planning.

The allowable annual cut (AAC), the rate at which timber is made available for harvesting, is determined for each provincial management unit, which includes 37 Timber Supply Areas (TSAs) and 34 Tree Farm Licences (TFLs). The actual cut calculation is based on analyses of timber supply and the availability of timber over time, and is influenced by many social, economic, and environmental considerations. Some of the factors considered include: inventory information on the actual amount of land available; reductions in the timber land base owing to the creation of new parks or the building of transmission lines; the many types of forest stands and their volume and growth rates; the depletion of timber by agents such as fire, insects, and diseases; the timber utilization level (or the amount of a tree that can be used); and the accessibility of timber using current technology. The AAC has risen from approximately 15 million m^3 in the mid-1960s to its current level of approximately 60 million m^3. The most recent AAC rose to a temporary level of approximately 75 million m^3 because of the mountain pine beetle infestation.

The actual amount of timber cut depends on market conditions. The individual licensees have some flexibility over the amount actually cut in any given year within their periodic cut control licence agreement (usually 5 years). This flexibility accounts for the discrepancy between published allowable cut and actual cut figures.

The timber supply of each management unit is reviewed at 5-year intervals, with more frequent reviews following large-scale disturbances such as major wildfire events or large-scale pest outbreaks (e.g., mountain pine beetle infestation). In each review, the data and assumptions used to determine the AAC are made available to the public for comment before a final decision is made.

Once the level of cut is determined for a given management unit, the way in which forest cover is removed is determined by the choice of silvicultural system (Figure 5.1). A silvicultural system is defined as a systematic program of silvicultural treatments over the life of a stand. Clearcutting is the best-known example. It involves removal of the entire forest stand in one operation with regeneration of the forest obtained artificially or from seeds germinating after the harvest. Until relatively recently, almost all harvesting in British Columbia followed the clearcutting system. In the early 1900s, all forest

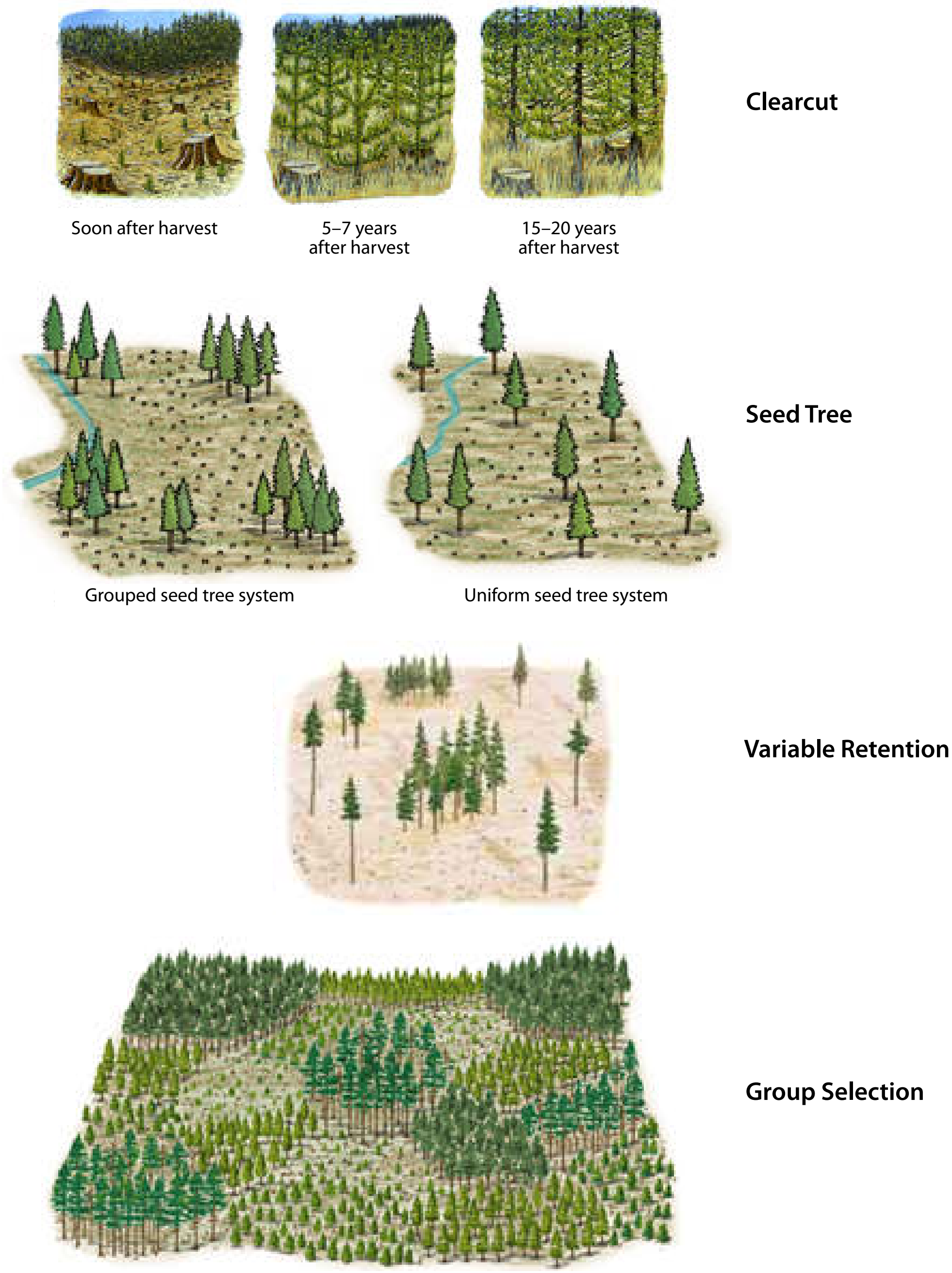

FIGURE 5.1 *Schematics of various silvicultural systems. (Adapted from B.C. Ministry of Forests and Range 1999)*

cover was removed from thousands of hectares of land over relatively short periods of time with no deliberate regeneration efforts. The effects of this "continuous" clearcutting on hydrologic processes was dramatic and, together with the poor aesthetics of hillsides laid bare from horizon to horizon, led to large-scale public protests and eventually limits on the amount of cover that could be removed within watersheds at any one time. Complex rules now govern the extent of clearcutting within a watershed to limit these effects. Much recent hydrological research has focussed on developing more effective and less costly restraints on logging operations.

Silvicultural systems other than clearcutting can be employed (Figure 5.1). Currently, the most common is clearcut with reserves, with the reserves covering 10% of the clearcut area. This is a minor modification of the clearcutting system that leaves some structural heterogeneity in varying amounts, types, and spatial patterns of living and dead trees to address a broad array of forest management goals. The retention of later seral conditions is assumed to sustain ecosystem functions and biological diversity at the stand level. Variable retention is designed to increase the amount of cover left on the logged area and the amount can range from 30 to 70% across the cutblock. Seed tree and shelterwood systems are used to increase natural regeneration on the cutblock. The retained cover can vary from 5 to 50%. Continuous cover or selection systems permanently retain a high level of forest cover. The amount of cover retained will vary between 50 and 80% depending on the intensity of the first and subsequent cuts.

Protection of the forest, and thus forest cover, from the destructive effects of fire, insects, and diseases has been a major preoccupation of forest managers and a focus for forest management since the early 1900s. Firefighting capabilities have developed to a high level of efficiency to protect timber values, communities, and lives. Ironically, this level of efficiency led to an accumulation of older stands and fine and coarse fuels, which is now thought to increase threats from both fire and insects; however, evidence for this in British Columbia is not strong and has been disputed. Whatever the set of causal factors, high fuel loads in the Okanagan and Thompson drainages contributed to the large wildfires of 2003. These fires removed well over 100 000 ha of forest cover in a few weeks and led to significant, but localized, hydrological effects (Redding 2008). Despite considerable wildfire–watershed research conducted over the past several decades, many unknowns still exist regarding the magnitude of watershed impacts and the factors governing watershed recovery (Ice and Stednick 2004; Neary et al. 2005). Similarly, forest ecosystem processes in the dry pine forests of the Interior have led to the build-up of bark beetle populations and the immense mountain pine beetle outbreak that covered 16.3 million ha in British Columbia, according to the 2009 survey. Significant hydrological effects are expected as a result of this unprecedented destruction of forest cover (Winkler et al. 2008). Future forest management in the province will be strongly influenced by these events.

Regenerating Forest Cover

Some form of natural regeneration occurs on almost all forest lands in British Columbia (Figure 5.2). Natural ingress of lodgepole pine in the Interior is abundant and western hemlock on the Coast regenerates rapidly, even in large openings. Over the past 40 years, however, various land cultivation and plant propagation techniques were developed and applied to speed up regeneration and improve the quality of regeneration for future harvesting. Some form of site preparation followed by planting is common in almost all forest types. Weeding is often employed to ensure that the desired species are free to grow rapidly. Thinning or spacing is also used as a mechanical method to control species competition and growth. The effect of most of these actions is to reduce the amount of time required to achieve a forest that has hydrological characteristics similar to those of the forest that was harvested.

Occasionally, practices can have negative effects. For example, mechanical site preparation can redistribute water flow and increase soil erosion. Disposal of logging debris by prescribed burning can negatively affect infiltration rates if the temperature is too high, causing hydrophobic soils to develop (Chapter 8, "Hillslope Processes"). Inappropriate tree species selection decisions in planting operations may lead to heavy mortality in young stands.

Transporting Wood

From the 1860s through the first half of the 1900s, the forest industry relied primarily on water, and to a lesser extent rail, for transport of timber products from the forest operation to the mills and markets. The transportation of wood includes three phases: primary, intermediate, and market transportation. Primary transportation (e.g., yarding or skidding) is the movemement of timber products (logs) a relatively short distance from the "stump" to a point where a secondary form of transport begins. Intermediate transportation occurs within a forestry operation where products are accumulated (stored) for more cost-efficient movement over longer distances (e.g., truck hauling). Market transportation is the movement of larger inventories of wood from the forest operations to the processing site (e.g., mill).

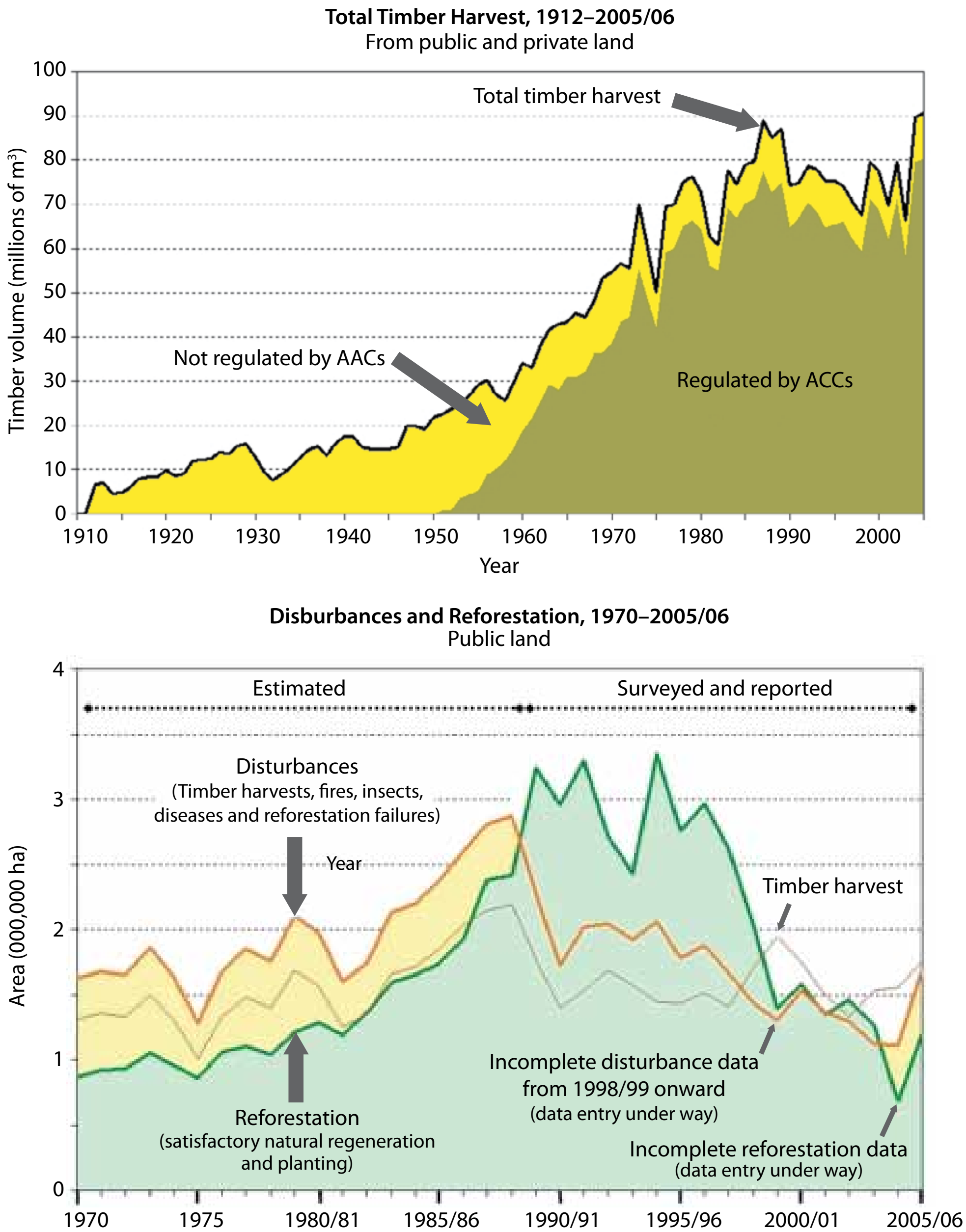

FIGURE 5.2 *Long-term timber harvesting (top) and regeneration (bottom) for British Columbia (B.C. Ministry of Forests and Range 2006).*

These modes of transportation were the only way that a large, heavy product could be moved efficiently. Market transportation thus dictated that the coastal forest industry developed in proximity to the ocean or navigable rivers and lakes.

Within forest operations, primary transportation of logs was not an immediate concern as hand logging at the water's edge was expanded upslope by the use of oxen or horses on the earliest form of road—the skid road. Skid roads were simply cleared earth trails, usually stabilized with log skids lying in the direction of travel or puncheon laid perpendicular (Figure 5.3). Animals were replaced with cable systems that were able to move logs more quickly over extreme terrain and without roads. Where economically suitable timber existed along major rail lines, a similar sequence of forest development could be found through other parts of British Columbia.

River transport of logs

Log driving (also known as "river driving") is the process of transporting cut timber by flotation along creeks, streams, and rivers and across lakes (Sedell et al. 1991; Figure 5.4). This was a common practice in the late 1800s and early 1900s throughout western North America. In many regions, roads and railways had not yet been built to facilitate the transportation of materials by land, and using water was less expensive than road building and maintenance (Cowell 1981; McCaig 1989). Log-driving activities were more intense in the Interior of British Columbia than on the Coast because logs tended to be of a smaller diameter. Thus, timber from the Interior was less likely to jam during log driving and was easier to manipulate. In addition, waterways in the Interior tended to be deeper, wider, and slower flowing with less erratic water levels than those on the Coast (Sedell et al. 1991; Drushka 1992). Tributaries of the Fraser River, including the Thompson, Adams, Shuswap, Nadina, and Stellako Rivers, as well as the Bull, Slocan, and Kootenay Rivers, are examples of waterways that were driven extensively in the first half of the 1900s. On the Coast, log drives on the Cowichan, Tsolum, Somass, Puntledge, Lillooet, and Coquitlam Rivers, for example, met with varying degrees of success (Drushka 1992; Mackie 2000). Logs were floated down the Fraser River through Hell's Gate to Hope until the 1960s and along a slow-moving stretch of the Fraser near Quesnel until at least the 1980s (Cowell 1981). Water transport continues today, but

FIGURE 5.3 *Skid road with log "skids" laid perpendicular to the direction of haul (circa 1890). (Photo: Royal BC Museum Archives, B-03827)*

FIGURE 5.4 *Log drives on Kootenay rivers: the Slocan River near Passmore (top) and Bull River near Cranbrook (bottom). (Photos: Royal BC Museum Archives, C-09790, F-09172)*

only in coastal marine waters and on major lakes in the Interior.

Log drives in the Interior were generally launched during spring breakup, as meltwater swelled the river, and continued until the spring freshet began to subside (Drushka 1998). Logs that had been cut during the winter and decked at the sides of creeks and rivers were transported downstream when water levels were considered sufficiently high to carry timber to its destination. Log drives would continue until the water levels in creeks and rivers had fallen too low to float the timber effectively (McCaig 1989; Sedell et al. 1991; Drushka 1998). In the early 1900s, for example, log drives of 37 000–43 000 m^3 of wood were commonly moved down the Shuswap River in the space of 3–4 weeks (McCaig 1989; Drushka 1998). In later years, log drives were restricted to the interval between the inward migration of spawning salmon and the outward migration of young salmon headed for the Pacific, in order to reduce the direct impact of floating logs on fish (Cowell 1981).

Damage caused by log driving

Even before a single piece of timber was floated downstream, significant damage to the ecology of waterways was caused by "river improvements" undertaken to prepare waterways in advance of the drive. "Fin booms" (long, floating chains of logs with one end anchored onshore) or log cribbing were used to block off any areas where slow-moving water might trap or redirect logs; this included side channels sloughs, swamps, sand and gravel bars, or low banks along wider parts of streams and rivers (Cowell 1981; Sedell et al. 1991; Drushka 1998). Obstructions within the main stream channels, such as boulders, pools above rapids, abrupt changes in gradient, leaning trees, and sunken logs, would also be cleared, frequently with dynamite, and later with bulldozers (Cowell 1981; Sedell et al. 1991; Drushka 1998). Occasionally, water was redirected into flumes, which were built along creek beds to quickly move timber from creeks into larger rivers. These flumes could extend a considerable distance; one flume down a tributary of the Bull River was more than 6 km long (Drushka 1998). Such alterations had enormous consequences for aquatic habitat quality. Blocking water flow to sloughs and backwaters severed the critical connections between the stream system and floodplain vegetation and caused spawning areas to dry up (Sedell et al. 1991). Simplifying watercourses by clearing obstructions accelerated the flow of water and caused enormous quantities of accumulated sediments to be flushed from the streambeds (Cowell 1981). It also dramatically reduced habitat complexity, and altered or eliminated valuable salmonid rearing areas (Sedell et al. 1991).

The construction of splash dams was a common technique used to temporarily prevent water from flowing down river. When this water was released, an "artificial freshet" was created, which carried timber rapidly downstream (Sedell et al. 1991). Splash dams were constructed from stacks of log cribbing and could be many metres high and wide (Sedell et al. 1991). These structures formed impenetrable barriers to fish migration and their repeated use during spawning season decimated local salmon populations. The violent torrent of water that was released when splash dams were opened only intensified the scouring of streambeds, channel erosion, and displacement of gravel spawning areas that were already caused by floating timber; in addition, eggs that had been deposited in streambeds were often lost to scouring and silting (Sedell et al. 1991). The use of splash dams also intensified the deposition of tree bark, which was knocked off as timber moved downstream. This material tended to sink to the bottom of watercourses, where it smothered developing spawn, clogged the gills of young fish, and destroyed shelter for salmon fry, making them more vulnerable to predation (Cowell 1981; Sedell et al. 1991).

Logjams were also frequent during log drives. In the early years, smaller jams were dislodged using pike poles and peaveys (McCaig 1989; Drushka 1998). Larger jams were released with the use of blasting powder or dynamite (Drushka 1998); for example, 30 tons of dynamite was used to clear a single jam on the Bull River, scattering debris for hundreds of metres in every direction (Drushka 1998). Later, bulldozers were used to push logs back into waterways, degrading the stability of streambanks and causing significant erosion (Sedell et al. 1991).

Although the potential for drastic losses to fish habitat caused by log driving was a very early concern to some people (Sedell et al. 1991), this issue was more widely recognized by the 1950s. During the second Sloan Royal Commission on Forest Resources, the Canada Department of Fisheries concluded that log driving in shallow rivers was a serious threat to the salmon fishery (Whitmore 1955, in Sedell et al. 1991). Nevertheless, log driving continued in the Shuswap and Stellako Rivers until at least the late 1960s (Sedell et al. 1991; McCaig 1989) and in the Quesnel and Cariboo Rivers until 1981 (Cowell 1981). The ecological damage to lakes, rivers, and streams that was caused by log driving is still evident in many aquatic systems. Numerous abandoned splash dams were left in place, altering water flow for decades after they were last used (Sedell et al. 1991; Committee on Protection and Management of Pacific Northwest Anadromous Salmonids 1996). Harvesting of large, late-successional conifers in riparian areas also altered the size and species composition of new woody material entering streams and rivers, slowing the rate at which the quality and complexity of these aquatic habitats have recovered (Committee on Protection and Management of Pacific Northwest Anadromous Salmonids 1996). In northern Sweden, where log driving followed a similar timeline and used similar techniques to improve water flow for floating timber, a recent study concluded that log driving has left an indelible imprint on the Vindelälven River (Törnlund and Östlund 2002). It is virtually impossible to determine whether rivers in British Columbia can and will recover from the impacts of log driving because little baseline information exists about river ecology prior to log driving (Cowell 1981).

The Adams River log drives

A classic example of the impacts of log driving on river ecology is that of the Adams River (Figure 5.5). The Adams River Lumber Company began operations in the early 1900s (Drushka 1998). Timber that

FIGURE 5.5 *River transportation of logs on the Adams River: the Adams River splash dam (top), Brennan Creek flume (bottom left), and Adams Lake sorting pond filled with logs to be driven down the Adams River (bottom right). (Photos: Royal BC Museum Archives, I-62756, F-09177, F-09179)*

had been cut in the winter was skidded to the banks of the Upper Adams River, driven down it into Adams Lake, and then towed by sternwheeler to the mouth of the Lower Adams River (Drushka 1998). In 1907, the company built a splash dam at the head of the Lower Adams River that blocked the passage of spawning salmon through the lake and into the Upper Adams River (Hume 1994). Logs that were carried from higher elevations by a system of flumes were dumped into the Adams River below the dam. This dam retained spring meltwater to "flash-float" logs from the Lower Adams River into Shuswap Lake (Hume 1994; Drushka 1998). Once in Shuswap Lake, the logs were boomed and floated to the big mill at Chase. When the water was held back by the dam, river levels were often so low in the Lower Adams River that spawning beds were exposed, causing incubating salmon eggs to freeze in the winter or be suffocated because of a lack of oxygen (Hume 1994). When the water was released, it washed salmon from the river, scoured gravel beds, and destroyed developing eggs in surviving nests (Hume 1994). The use of splash dams on the Adams River lasted from 1907 to 1922 (Hume 1994). The run was also severely affected by the blockage of the Fraser River at Hell's Gate during construction of the second Canadian National Railway line in the Fraser Canyon in 1913. Although reduced numbers of spawning salmon were observed in the Upper Adams River by 1913, the dam was not completely removed until 1945 (Sedell et al. 1991; Hume 1994). By then, the massive run of sockeye that had spawned in the Upper Adams had been driven to extinction (Hume 1994; Drushka 1998). Even though the International Pacific Salmon Fisheries Commission has worked since the 1950s to restore a run on the Upper Adams River, the unique genetic code of the salmon native to this run has been permanently lost (Hume 1994). The run on the Lower Adams River has rebounded, however, and is an important tourist draw in the area.

Rail and road transport of logs

As easily accessible timber close to mills decreased, transportation links became increasingly scarce. Thus, the forest industry needed to develop an intermediate means of transporting logs out of a forest operation. Steam-powered logging railroads (Figure 5.6) appeared on the Coast within a decade of the completion of the Canadian Pacific Railroad in 1885 and by 1917 there were 62 in operation (Gould 1975).

FIGURE 5.6 *"Old Curly" on Thurlow Island in 1894. (Photo: Royal BC Museum Archives, B-06967)*

Because they were labour-intensive and expensive operations, railroads expanded into the most easily accessible areas, which usually also had the highest-value timber. The necessity to operate at a large scale meant that all timber of any value within physical reach of the track was logged as soon as possible after the track was laid. Continuous clearcutting was considered the appropriate harvesting system to match the transportation mode and financial investment.

The use of trucks for hauling logs was well established in the 1920s. Initially equipped with hard, solid rubber tires (Figure 5.7), early trucks often operated on plank roads, which "floated" on the ground surface (Bendickson 2009; Figure 5.8). Power limitations and a lack of good braking hindered the use of trucks as an intermediate transportation system, but trucks did complement the railroads as a feeder system or for the development of small areas.

The use of railroads peaked through the 1920s and 1930s. Steel rails assisted the development of forest resources for almost a half-century before several significant developments took forest operations into a new era.

The earliest road and railroad construction was very labour-intensive. Picks, shovels, axes, and saws were the tools needed to excavate and reshape a roadbed. Animals provided extra power sources for transportation and grading surfaces. One of the earliest forms of excavating equipment was the steam shovel (Figure 5.9). Even with this production marvel, most excavated materials were placed only within the reach of the machine.

In the era of early road construction, roadbeds generally did not create a significant alteration to the landscape. Large excavations and fills were beyond the economic and physical means of most forest operations, so the excavations were avoided and the gullies and rivers were spanned with large, elaborately constructed trestle bridges. Low-gradient roads on lower slopes or valley bottoms did not tend to produce many conflicts in water management.

World War II was a period of significant technological development in the areas of earth-moving equipment and vehicle transportation. Machines such as bulldozers, mechanical shovels, and large-capacity trucks—all powered by internal combustion engines—were capable of developing access beyond the physical or economic limits of the railroad. In addition to the new technology, equipment was suddenly affordable and available after the war effort concluded.

FIGURE 5.7 *Solid-rubber-tire truck in 1924. (Photo: Royal BC Museum Archives, F-08719)*

FIGURE 5.8 *Excavating deep organic soils and hauling ballast for roads could be prohibitively expensive compared to building plank roads. Some of these roads were built over railroad grades by replacing the steel rails with planks. (Photo: Royal BC Museum Archives, NA-08441)*

FIGURE 5.9 *Wood-burning steam shovel near Cowichan Lake in 1935. (Photo: Royal BC Museum Archives, H-05397)*

A primary road construction machine by the late 1940s was the "line" shovel (Figure 5.10), a carry-over from the railroad era, which operated on wooden pads laid down by the operator. Where conditions permitted, the tracked bulldozer (Figure 5.11) was also a popular machine for both constructing subgrade and for spreading the stabilizing ballast over the subgrade. Except for quarrying, rock was avoided as much as possible at this time. Rock drills were most often air-driven units adapted from mining operations. These drills were tied to a separate compressor and were not suited for moving on subgrade.

The greatest limitation of both the front-scooping line shovel and the bulldozer was their inability to adequately sort materials. Given the materials of a 20 m right-of-way and a production budget, these machines constructed a road base from a mixture of mineral and organic soils, stumps, brush, and logs. The result was a road with a component of decomposing material and materials of mixed bearing capacity. Over time, these roads would have problems of settling and pothole formation, or would even collapse entirely.

Line shovels and bulldozers (Figure 5.11) were not ideal machines for grooming cut-and-fill slopes to a natural and stable angle. This resulted in sloughing of constructed slopes until stable angles were achieved. Through this period, road construction was therefore concentrated on lower hillslopes where earth movement was minimal. Because large excavations and fills were uncommon, the machines' deficiencies were revealed only through the significant construction problems that appeared later.

The line shovels also created troughs with spoil piles on both sides of road, which changed the natural water flow. If water was present in any form, line shovels and bulldozers often ended up with a "soup," which was shifted to the side of the intended roadbed. Thus, these machines were the cause of many landslides as operators placed fill on oversteepened slopes. The current method involves the endhaul or movement of excavated material from one section of the road to another or to a disposal site.

In addition to the physical and economic ability to develop new territory, a significant political development provided a complementary incentive. The Royal Commission of 1943–1945 (Sloan 1945) led to changes in the forest tenure system, the concept of forest (or more accurately timber) sustainability, and the availability of an increased land base for forest operations. The new tenure system fit perfectly

FIGURE 5.10 *Road construction with a line shovel in 1936. (Photo: Royal BC Museum Archives, NA-06671)*

FIGURE 5.11 *Bulldozers became popular in the 1940s, but their ability to sort materials on subgrade construction was not a strong feature. Bulldozers were one of the first road-building machines to add hydraulic components (the blade and winch). (Photo: Royal BC Museum Archives, NA-06355)*

with the new technology and by the early 1950s forest operations shifted dramatically from rail to road as the basis of intermediate transportation.

Although the sustainability of the timber resource began to be seriously addressed with the tenure changes of the 1950s, it was not until the early 1970s that foresters and other resource scientists began to apply sustainability principles to the non-timber resources sharing the same land base. Government agencies responsible for wildlife and fisheries resources became increasingly involved in reviewing harvesting plans. Forest companies such as MacMillan Bloedel and British Columbia Forest Products assembled teams of specialists in fisheries, wildlife, hydrology, geomorphology, terrain stability, and other fields to conduct inventories of non-timber resources and to assist planning foresters by developing management guidelines for these resources. Several high-profile land use conflicts such as the "Nitinat Triangle" (west of Nitinat Lake) and the "Tsitika-Schoen" accelerated and increased the work of these professionals. The attempt to have some objectivity in any land use decision, whether single use or multiple use, required inventories and objectives for identified resources, plus a plan and assessment of impact if the timber resource was to be developed. By the end of the decade, development plans for most new areas included multiple resource assessments and objectives.

By the mid-1970s, forest development planning focussed on the rate, pattern, and sequence of forest harvesting to address the integration of resource sustainability. A visible, immediate change was a shift from continuous clearcutting to patch clearcutting with reserves. Tied to this shift in forest operations was an attempt to increase the spatial dispersion of harvesting and rate of harvesting throughout the tenure. Although the tendency to harvest the closest and highest-quality stands first still persisted, the changes to harvest planning significantly increased road construction. The drive to penetrate deeper into watersheds and up into increasingly steep-sloped hillsides produced unprecedented problems of road and bridge washouts and associated terrain failures.

Road construction in forest operations has never enjoyed the luxury of limiting all activity to dry weather conditions. To meet the demands of harvest plans, construction had to be extended for longer periods through the year. This exacerbated the conflicts of quantity, cost, quality, and environmental impact.

The direct costs associated with water management problems were immediately obvious to the forest industry. At the same time, resource managers were becoming increasingly aware of adverse impacts to fish habitat (Figure 5.12). Foresters quickly began to respond to the increased information available to them in their planning. The industry was also encouraged to change practices through increasingly demanding legislation and regulation such as Sections 34–36 of the federal *Fisheries Act*. Beginning in the late 1970s, several legal proceedings firmly established the responsibility of resource professionals and resource companies to exercise the science and technology available to them in the management of forested lands.

A radical change in road construction techniques occurred in the mid-1970s with the introduction of the hydraulic excavator (Figure 5.13) or backhoe (previously used primarily as a trenching machine). Although initially poorly adapted in structural design, the excavator swiftly proved its ability to construct roads quickly and cost efficiently. Just as impressive was the quality of the work that could be produced using these machines. The excavator could effectively sort materials; place logs for optimal value recovery; place stumps, brush, and organic soil in positions where they would not degrade the quality of the road subgrade; and place the best available structural materials in the road. The addition of a hydraulic thumb to the excavator in the mid-1980s further improved the usefulness of this machine. The ability to precisely remove, place, and grade materials has made the excavator the most versatile piece of equipment in the forest industry. Advances and modifications in the design and use of support equipment complemented the excavator.

Excavator construction, coupled with appropriate maintenance, can eliminate many of the causes of road failures such as settling, slumps, and washouts because the road can be constructed without weak or decomposing material and can be appropriately secured against moving or accumulating water. The design of the excavator also allows the machine to effectively deactivate or deconstruct roads deemed environmentally hazardous, temporary, or otherwise not needed.

FIGURE 5.12 *Moving water (in any volume) is an inconvenience to road construction because it requires some form of a special structure to span it. Before stream morphology was commonly understood, moving water was generally viewed as "self cleaning." (Photo: Gould 1975, reproduced with permission of Hancock House Publishers)*

FIGURE 5.13 *The hydraulic excavator brought significant change to the quality of constructed roads. (Photo: D. Bendickson)*

Water management and roads

Forest roads are usually constructed to develop a timber resource for harvesting. Since a road location is a long-term commitment, harvesting and silvicultural systems for both the first and subsequent harvests will dictate that roads are placed for optimal harvest utilization. Yet, construction feasibility and cost are obviously affected by topography and ground composition. Not so obvious is the influence of non-timber resources on road location. Among these values are wildlife and visual quality objectives, which may influence where or how a road will be constructed.

The fact that the direction of a road is often perpendicular to the movement of water is a guarantee of conflict. Without due consideration, roads will intercept, block, restrict, or channel water. Water, in response, will find a way under, over, around, or through any obstruction. Surface water movement, especially over unprotected surfaces, will produce erosion proportional to its velocity and volume. If allowed to saturate soils, any eventual movement may take the soils along with it in the form of slumping or flows (for more information see Chapters 8 and 9).

Water-related road problems generally involve surface erosion or cut/fill slope failures (see Chapters 9 and 10). These effects of water on roads have been somewhat understood since the earliest roads were built but, arguably, the effects of roads on water were not well appreciated until the 1970s when multiple use resource planning was incorporated into forest operational planning.

Long-term water management plans consider the optimal accommodation of surface and subsurface flow. These plans include prescriptions for deactivating (or even deconstructing) inactive roads to minimize risk of erosion or mass wasting. Thus, good engineering practices in the design and construction of roads are necessary.

Bridges and culverts

Logs and timbers were obvious material choices for any structures associated with forest transportation systems. Simple log-stringer/gravel-decked culverts have been popular since the days of early railroads for spans of up to 6 m. Where suitable species and log diameters were available, plank-decked spans of up to 21 m were common (Nagy et al. 1980).[1]

1 Span lengths can vary significantly based on the design load, timber species, and size.

Early forest roads inherited the log and timber truss construction of railroads, but this labour-intensive structure was short-lived. The ability of logging trucks and equipment to handle steeper grades and curves gave engineers the flexibility to locate crossings at sites that offered the option of shorter and less complex spans.

To keep spans as short as possible, it was common for crib abutments to encroach on the wetted perimeter of the watercourse (Figure 5.14). For wider rivers, piers were constructed to create multiple span crossings. The limited span lengths of the wood construction often meant that crib footings and piers presented an obstacle to water flow. This created scouring problems and the structures also tended to trap debris moving downstream.

In the 1960s, glue-laminated timber beams became a preferred girder for longer spans (Figure 5.15). Manufactured in Vancouver, "glulam" beams of up to 27 m could be transported and erected with reasonable efficiency. Longer glulam spans were not common because of the excessive weight and handling difficulty. Where longer spans could not be avoided, glulam structures were reinforced from below with tension trusses.

Steel and concrete became financially viable alternatives to wood in the late 1970s. The versatility and durability of these materials quickly made them the preferred option in many applications. The ability to custom design major spans and to transport those spans in sections allowed for longer clear spans, which minimized structures in the streambed (Figure 5.16). Furthermore, the ease of moving a portable bridge to a new location once a short-term road was deactivated was another milestone. Additional refinements of steel and concrete bridge components, including prefabricated decks, footings, and retaining walls, now allow the construction of high-quality temporary structures that completely span a watercourse from bank to bank.

FIGURE 5.14 *Because of span limitations for log-stringer bridges, the footing structures often intruded into the wetted perimeter of the creek. Scouring associated with such construction was an ongoing problem. Moakwa Creek (Upper White River tributary) circa 1972. (Photo: D. Bendickson)*

FIGURE 5.15 *The bridge over the lower Klinaklini River illustrates a "glulam" bridge with simple spans and a longer inverted truss span. (Photo: D. Bendickson)*

FIGURE 5.16 *The use of steel girders has allowed for longer spans without obstructions to water flow (Klanawa River). (Photo: D. Bendickson)*

Forestry in British Columbia has a long and rich history and the strands of political, economic, social, and ecological change are woven together in complex ways that defy easy description. In recent years, numerous books have been written on various aspects of forest history, including theoretical works and personal memoirs. Novels and poems involving the province's forest history also add a special perspective. A succinct summary of this literature is found in the Pearse Royal Commission on Forest Resources (Pearse 1976).

Although British Columbians share their forest heritage with other parts of Canada and northern European nations, the province stands out in several unique ways. First, experience with the primeval forest is very recent; consequently, substantial portions of the forest remain largely untouched by visible human impacts. Second, the forest estate is very large, very important in economic, social, and political terms, and makes up a considerable proportion of the Earth's remaining old-growth conifer forests, many of which are destined for future harvest. Third, the forests remain in public ownership. Forest policy developments over the last century have been strongly influenced by all three.

Kimmins (1999) proposed that forestry (defined as "the art and science, and practice, of managing forested landscapes to provide a sustained production of a variety of goods and services for society") progresses through four stages.

1. The initial stage is unregulated exploitation, the negative effects of which lead to stage two.
2. In stage two, legal and political mechanisms are put in place to regulate the rate and location of exploitation and to determine who reaps the benefits and who pays the costs, usually in a very uneven fashion.
3. Stage three involves modification of these mechanisms to improve their ecological, economic, and social sensitivity, but the focus of management is on marketable products such as timber.
4. The final stage sees the development of a forestry that is responsive to a much wider range of demands from society, including values and goods having no obvious market and price. As a consequence, the benefits and costs of forestry activities are shared more equitably among all sectors of society.

Forestry policies and activities in British Columbia (and in the rest of Canada) have passed through the first two stages into the third stage, and are now struggling to move to the final social stage in which forestry activities are sanctioned by broad public support. Along the way, legislative tools to guide management have become increasingly complex, and sometimes bewildering.

In 1865, a *Land Ordinance* was passed that permitted colonial governors to make timber available via leases, but left the land in government hands. In 1912, the first *Forest Act* was put into effect. This legislation led to the creation of the Forest Branch (later the Forest Service and Ministry of Forests) with a mandate to:

- administer timber leases, value timber, and collect royalties
- protect forests against damage, especially fire
- undertake reforestation[2]

The *Forest Act* has received a series of revisions and other statutes were also added to deal with various contemporary issues. However, it took another 60 years before water and water issues were dealt with specifically in forest legislation. The 1976 Report of the Royal Commissioner on Forest Resources (Pearse 1976) made recommendations for forest resources planning, including the need to address water, fisheries, recreation, wildlife, and other concerns. The 1978 *Ministry of Forests Act* charged the Ministry, as the agency responsible for administering forest management activities, with planning for all forest resources, recognizing that the non-timber resources were usually not inventoried or well understood. Therefore, the 1970s became a period of information gathering, distilling, and dissemination.

Further reforms that drew attention to water and water-based resources took place in the 1980s and 1990s. For example, the objective of "no net loss" of fish habitat was clearly communicated to forest planners in 1986 by Fisheries and Oceans Canada. The following year, in a joint provincial/federal effort, the *British Columbia Coastal Fisheries/Forestry Guide-*

2 Based on the 1912 *Forest Act*, the Forest Branch had jurisdiction over reforestation, but none was undertaken until 1930.

lines became one of the early publications to take a set of objectives and provide some communication of expectations in achieving those objectives. In 1995, the *Forest Practices Code of British Columbia Act* was supported by a series of guidebooks that provided forest planners and practitioners with condensed and practical reference material. Among these publications, the *Riparian Management Area*, *Green-up*, *Forest Road Engineering*, and *Fish-stream Crossing* guidebooks were significant aids. In addition, the Act created the Forest Practices Board as a "watch dog" agency to report on adherence to the Code and to provide citizens with an avenue of complaint. The Board has issued a number of water-related reports since its creation.

The Code was revised in 2002 when the *Forest and Range Practices Act* was enacted. The new government of the day believed that the Code was too bureaucratic, and representatives often made this point by standing beside a large stack of Code-related documents. It sought to reduce the hand of government from forest management decisions and place more responsibility in the hands of forest professionals. Whether this approach is more or less effective than other possible approaches is open to debate, but since the new legislation was introduced, its relative importance has diminished in face of many changes in the forestry environment. The softwood lumber dispute, a crash in lumber prices, company mergers, mill closures, the forest fires of 2003, and the mountain pine beetle outbreak have combined to change the political and economic landscape in a way that was never anticipated.

SUMMARY

Forest harvesting and the transportation of logs to mills has evolved from primitive manual means to the use of powerful, high-tech, heavy equipment. The physical ability to significantly modify forested landscapes may potentially affect other natural resources and systems if impacts to these resources are not considered. In the first half of the 1900s, British Columbia's forests were considered primarily as a commercial timber resource, and logging activities reflected a single-use objective of the land base. Beginning with the recognition of timber resource sustainability in the mid-20th century, other forest-based resources began to receive recognition, and were officially included in planning decisions by the early 1970s, creating a demand for new information.

Legislation has complemented the changes. Although many problems emerged during its implementation, the 1995 *Forest Practices Code of British Columbia Act* provided dramatic pressures for change. At the same time, it defined forest resources and many objectives.

In the current decade, new legislation has moved the emphasis of forest land management to sustainable forest management (SFM), which sets objectives and measures results. The Canadian Forest Service has defined SFM as management that maintains and enhances the long-term health of forest ecosystems for the benefit of all living things while providing environmental, economic, social, and cultural opportunities for present and future generations. This places expectation and accountability directly on industry, technologists, and professionals. The process is reinforced with penalties for non-compliance.

Although planning and implementing forestry activities is maturing to current expectations, the interaction of forest roads with forest resources is dynamic and the understanding of this interaction is undergoing continual refinement. Perhaps the greatest variable and the greatest challenge is the continuing education of the people that do the work.

Forest policy changes are still in a state of flux; further changes can be expected in response to the current economic, social, and political realities. The value of the water resource has grown enormously in the last 50 years, and is likely to become the pre-eminent resource in the future. Human health concerns will continue to keep water issues at the forefront of forest management in the 21st century. These concerns will likely result in increasing restrictions of logging practices throughout the province.

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Hydrologic Processes and Watershed Response

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INTRODUCTION

Streamflow reflects the complex interactions between the weather and the biophysical environment as water flows through the hydrologic cycle. It represents the balance of water that remains after all losses back to the atmosphere and storage opportunities within a watershed have been satisfied. This chapter describes the hydrologic processes that affect the generation of streamflow in British Columbia's watersheds. These processes include precipitation, interception, evaporation, infiltration, soil moisture storage and hillslope flow, overland flow, and groundwater. The processes and their spatial and temporal variability are described at both the stand and watershed scales. Understanding how streamflow is generated is vital in evaluating the effects of forest disturbance on hydrologic response and in identifying best management practices in a watershed. Chapter 7 ("The Effects of Forest Disturbance on Hydrologic Processes and Watershed Response") describes how forest disturbances, such as insects, fire, logging, or silviculture, alter both stand-scale processes and watershed response as streamflow. The discussion in the current chapter is organized by surface and subsurface processes, beginning with precipitation and ending with streamflow.

SURFACE PROCESSES

Forest vegetation directly affects the amount of water available for streamflow through the interception of rain and snow, the evaporation of intercepted water, and through transpiration. Altering forest vegetation can have an important influence on water balances at the site and watershed scale. In the hydrologic cycle (Figure 6.1), the forest canopy is the first surface that intercepts and stores precipitation, resulting in some amount (depending on canopy characteristics) returning to the atmosphere through sublimation and evaporation. Water also returns to the atmosphere through transpiration and by evaporation from the forest floor, soil surface, and open water bodies, such as lakes and wetlands (Figure 6.1). Forest vegetation reduces the rate at which snow melts and hence the rate of soil moisture recharge and downslope flow (Spittlehouse and Winkler 2002).

A water balance equation is often used to express the contributions of these processes to streamflow. At the watershed scale, the water balance equation

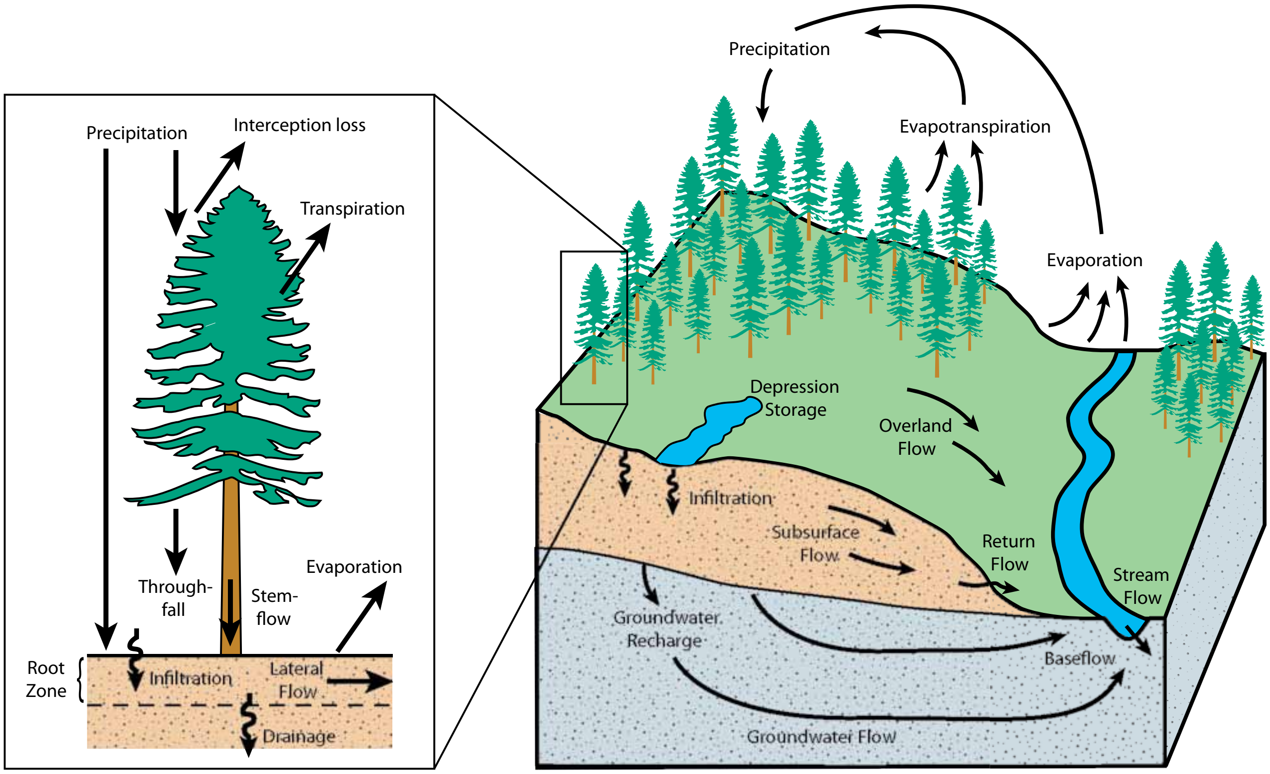

FIGURE 6.1 *The hillslope hydrologic cycle and stand water balance.*

links streamflow (R), precipitation (P), evaporation (E), losses to regional groundwater (G), and changes in watershed storage (ΔS):

$$R = P - E - G \pm \Delta S \tag{1}$$

where the variables are measured in water depth equivalent (millimetres) over time. The quantity G represents water that leaves the watershed as groundwater and thus does not contribute to streamflow.

Watershed storage includes water stored as snow, surface water (lakes, ponds, depression storage), glacier ice, soil moisture, and groundwater. As water flows through a watershed, it often moves between these storage compartments before discharging to a stream or being lost through evaporation and transpiration. For example, meltwater draining from a snowpack and infiltrating the soil represents a loss of snowpack storage and a gain of soil moisture, but with no net effect on total watershed storage.

At smaller spatial scales, the site or stand water balance equation links soil water storage (W), precipitation (P), interception (I), overland flow (O), infiltration ($F = P - I - O$), plant transpiration (E_t), evaporation from the soil surface (E_s), and drainage from the root zone (D):

$$\Delta W = F - E_t - E_s - D \tag{2}$$

where ΔW is the change in soil water storage (millimetres) over time. During winter in areas where precipitation falls as snow, there is a delay in infiltration until snowmelt: E_s is zero when snow covers the ground, and water content change depends only on drainage from the root zone except where tree transpiration occurs during snowmelt (Spittlehouse 2002). These processes are discussed in detail in the following sections. Equation 2 does not account for lateral downslope flow of soil water, and therefore may not be applicable on hillslopes.

Precipitation

Precipitation is defined as liquid (rain) or frozen (sleet, hail, graupel, snow) water, or a combination of both, falling from the sky. Water is also deposited at the land surface through condensation (e.g., dew, fog, rime, and hoar frost) (UNESCO and World Meteorological Association, International Hydrology Program 1998). British Columbia's precipitation regimes are a result of weather systems that are typical of mid-latitudes on the east side of the Pacific Ocean and that interact with the mountain ranges and valleys paralleling the coast. Alternating sequences of eastward-flowing high- and low-pressure systems, incursions of arctic air, and flows from the midwestern United States interact with topography to produce precipitation regimes including wet coastal, dry interior, moist northern, and snow-dominated high elevations (Hare and Thomas 1974; Phillips 1990; also see Chapter 3, "Weather and Climate").

Precipitation and other climatic elements vary seasonally because of the combined effects of solar radiation and weather patterns. For example, most coastal locations are dominated by a wet-winter/dry-summer pattern. In the lee of the Coast Mountains and throughout much of the interior, seasonal differences in precipitation are smaller than on the coast (see Chapter 3, "Weather and Climate"). Variations in sea surface temperatures in the Pacific Ocean (e.g., El Niño/La Niña and the Pacific Decadal Oscillation [PDO]) also influence precipitation on annual and decadal cycles, producing large between-year variations superimposed on long-term trends (Rodenhuis et al. 2007).

In general, rainfall volume and intensity are the highest during the winter in coastal British Columbia. Storms that produce over 100 mm of rain in a 24-hour period are common, as are intensities that reach 30 mm/h. Coastal winter rainstorms can last for many days with long periods of light drizzle intermixed with periods of high rainfall intensities. Summer storms on the coast are usually much shorter in duration and lower in total precipitation than winter storms. In contrast, rainstorms in the province's interior and north are usually of shorter duration (less than 1 day) and occur in late spring through early fall when temperatures are above 0°C. Storms with higher rainfall volumes and intensities are often convective, with durations of a few hours. In the province's interior and north, infrequent warm frontal storms in the late winter can also deliver large amounts of rain. Further information on precipitation and intensity-duration-frequency data is provided in Chapter 17 ("Watershed Measurement Methods and Data Limitations").

Dew forms on surfaces above 0°C, whereas hoar frost forms at or below this temperature (Glickman [editor] 2000). These condensation events occur overnight and are facilitated by clear skies and low wind speeds that cool the condensation surfaces. Rime is formed by the rapid cooling of water blown onto exposed frozen surfaces. Condensation precipitation events can be important biologically. For example, these events can have a significant effect on vegetation (e.g., dew facilitates the growth and spread of disease); however, the amount of water deposited during these events makes a negligible contribution to the annual water balance.

Ice storms are more damaging than rime. These storms occur when liquid precipitation freezes on contact with a surface and forms a glaze (layer of ice) (Glickman [editor] 2000). The glaze has a much higher density than rime and hoar frost, and its weight can break branches or whole trees and cause widespread damage (Irland 1998, 2000).

Fog drip is another form of condensation precipitation. It occurs when wind moves clouds at ground level (fog) so that water droplets collide with and adhere to foliage, branches, stems, and other surfaces. Subsequently, individual droplets coalesce and drip from the tree canopy or flow down the branches and stems to the ground. Forests on windward slopes and ridge tops are most prone to fog drip because of the large canopy area that is exposed to the cloud. On flat terrain and leeward slopes, most of the fog drip occurs at the edges of stands (clearcut edges, lake edges, etc.). Fog drip can be a significant component of the annual water balance in some ecosystems, predominantly in coastal areas (Harr 1982; Schemenauer 1986; Ingraham and Matthews 1995). For example, Harr (1982) reported that fog drip in a windward-facing watershed in coastal Oregon resulted in an approximate 30% increase in water reaching the soil surface, which makes an important contribution to summer flows. Studies in coastal British Columbia (Beaudry and Sagar 1995; Spittlehouse 1998a, 1998b) show that fog drip is not a significant component of the water balance. Fog drip generally results in a net gain in water to the land base. Fog drip can have negative impacts where it is combined with acidic deposition (Schemenauer 1986; Schemenauer et al. 1995). This has not been reported as a problem in the province, although localized effects may occur near cities and certain industrial facilities.

Snow forms in saturated air at subzero temperatures when a water droplet containing a condensation nucleus (e.g., dust, or inorganic or organic matter) drops below some particle-specific temperature and freezes. The temperature at which the water droplet freezes depends on its size and chemical composition, and on the ice nucleation mechanism by which it freezes (Wallace and Hobbs 1977; Schemenauer et al. 1981). Initially, all ice crystals are very small (< 75 µm) and have simple shapes; however, these small crystals continue to enlarge through the condensation and freezing of liquid water molecules onto the ice crystals or through the collision of large ice crystals and water droplets as the ice crystals fall. Rimed crystals, graupel, or snow pellets form when supercooled water droplets freeze onto large snow crystals as they fall (Schemenauer et al. 1981).

Snowfall requires sufficient cloud height to permit snow crystal growth and temperatures less than 0°C in most of the area through which the snow falls (Schemenauer et al. 1981). Snow can remain suspended in the air column for a considerable length of time because its surface area is large in relation to its mass. During this time, snow may be transported by wind, or may melt or sublimate. Wind redistribution of falling snow occurs at various scales. The movement of air masses over topographic barriers and large water bodies influences snowfall at the macro-, or regional, scale. At the mesoscale, snow may be redistributed over distances of 100–1000 m through the combined effect of wind, terrain, and vegetation. At the microscale of 10–100 m, snowfall is affected by surface roughness and airflow patterns (Schemenauer et al. 1981; Gray and Prowse 1993).

While snow is falling, its crystal mass changes through vapour exchange with the surrounding air. If the air temperature is above 0°C, or if the vapour pressure is less than the saturated vapour pressure, then falling snow may melt or sublimate before it reaches the ground (Satterlund and Adams 1992). The amount of energy required to sublimate snow at 0°C is 2.83×10^6 J/kg, which is the amount of energy required to melt snow (0.33×10^6 J/kg) plus the amount of energy required to evaporate water (2.50×10^6 J/kg at 0°C) (Oke 1987). This energy comes from several sources, including solar and longwave radiation, and convective heat transfer between the air and the snow.

As snow falls, air temperature controls its dryness, hardness, and crystalline form. Depending on the temperature, snow crystals can vary in shape from plates to prismlike crystals to dendrites and hollow columns (Wallace and Hobbs 1977; Schemenauer et al. 1981; McClung and Schaerer 2006). Generally, snow crystals that fall through a cold atmosphere are smaller than those that fall through warmer air (McClung and Schaerer 2006). Snow falling at low temperatures is also drier and less dense than that falling at warmer temperatures (Geiger et al. 1995). Detailed discussions of snow formation and crystalline structure can be found in McClung and Schaerer (2006) and Gray and Male (editors, 1981).

Net Precipitation

Only a portion of the precipitation falling during a given period reaches the ground. Some falls directly to the ground through gaps in the forest canopy; the rest is caught (intercepted) by vegetation or other surfaces and subsequently drips or slides from these surfaces to the ground or is lost through evaporation and sublimation. The precipitation that reaches the ground is referred to as *net precipitation*. Net precipitation depends on the weather (e.g., solar radiation, humidity, temperature, and wind speed), the time between precipitation events, the size and duration of events, and the type of vegetation cover present.

Rain interception

The fraction of rain intercepted by a forest depends on storm intensity and duration, weather conditions (wind speed, air temperature, humidity), and amount and type of vegetation present. Rain either falls directly to the forest floor through gaps in the canopy (throughfall) or hits foliage, branches, and stems. Rain that hits the vegetation may bounce off and fall to the forest floor (throughfall), run down branches and stems (stemflow), be absorbed by bark, lichens, and moss in the canopy, or remain on the surface of the foliage and branches and evaporate after the storm (interception loss) (Calder 1990). Interception loss increases with an increase in rainfall and the volume of material in the canopy (e.g., leaves, branches, stems, mosses, epiphytes) (McMinn 1960; Rothacher 1963; Plamondon et al. 1984; Giles et al. 1985; Spittlehouse 1998a, 1998b; Pypker et al. 2005).

The interception storage capacity of a forest is the maximum amount of water that can be absorbed by and reside on the canopy at any one time, and varies from 0.5 to 2 mm (Shuttleworth 1989). The total interception loss during storms is often much greater because of increased energy exchange and efficient mixing of the air, and moisture evaporated from the canopy is constantly replenished (Calder 1990). Most of the rain from small storms (e.g., < 10 mm) is intercepted and lost, particularly from the dense canopies of old coastal forests (Figure 6.2). In these forests, interception (volume) increases as storm rainfalls increase up to 100 mm (point of maximum interception loss), whereas the fraction of rainfall lost decreases. The less dense canopies of lodgepole

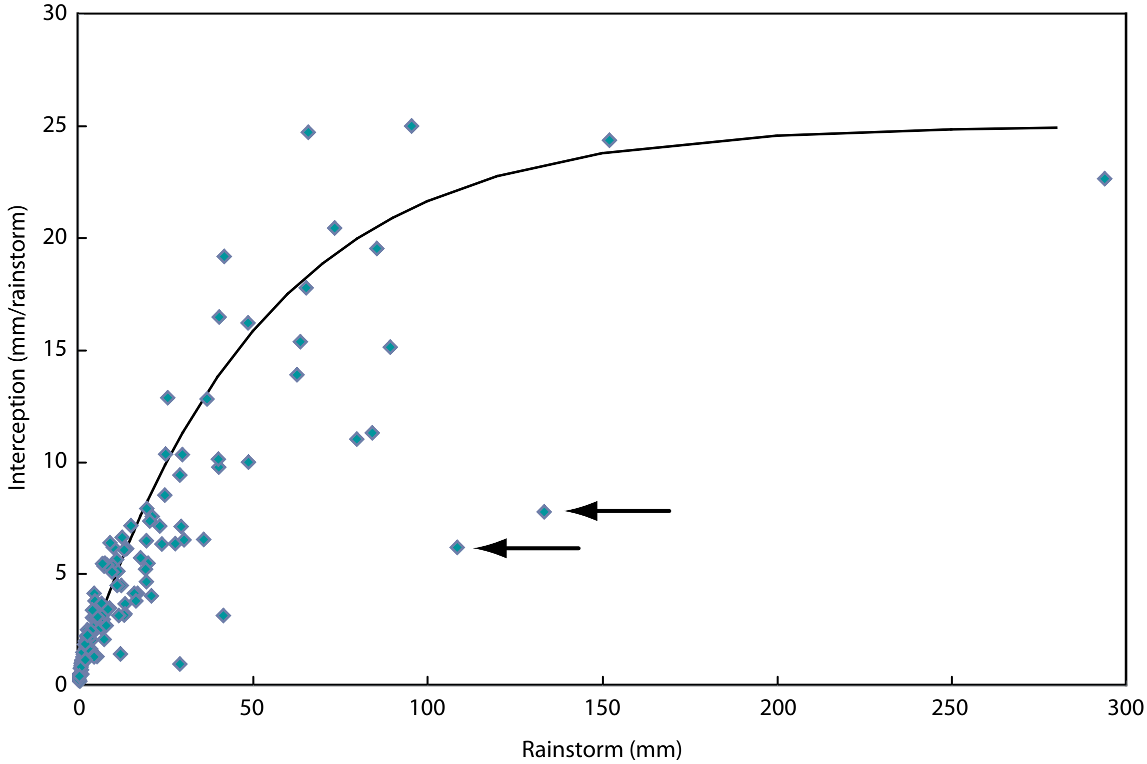

FIGURE 6.2 *Interception loss as a function of the amount of rainfall in an individual rainstorm for a mature coastal hemlock forest at Carnation Creek on the west coast of Vancouver Island. The arrows indicate two high-intensity winter storms. Line fitted by eye. (Adapted from Spittlehouse 1998a)*

pine and Engelmann spruce in the British Columbia interior have the same general effect; however, peak interception losses occur with storm rainfalls of approximately 25 mm (Figure 6.3). Young coastal stands are intermediate between these two situations (Figure 6.4) (Spittlehouse 1998a, 1998b). Maximum interception losses range from 3 to 30% of the storm precipitation depth depending on forest age, canopy density, and climate (Table 6.1).

Storms of long duration and low intensity have greater interception loss than higher-intensity, shorter-duration storms with the same total rainfall. This is because storms of short duration and high intensity have less favourable weather conditions for evaporation (Calder 1990; Spittlehouse 1998a, 1998b; Crockford and Richardson 2000; Carlyle-Moses 2004). Antecedent conditions, such as the time since the last storm, are also important because these conditions influence the remaining interception storage capacity of the vegetation. Figures 6.2 and 6.3 provide examples of the influence of antecedent conditions.

Points highlighted by the arrow symbols show interception losses during a wide range of storm sizes in coastal and interior forests. The arrows in Figure 6.2 and the arrows in Figure 6.3 highlight storms that occurred when the foliage was wet from previous storms. Interception losses were also measured over a series of storms at Upper Penticton Creek. The first storm with 6.2 mm rainfall occurred in the late afternoon, 2 days after the last rain, and had a 2.3 mm (37%) interception loss. A second storm began just after midnight and was of higher intensity; however, with little time for the canopy to dry out, the resulting interception loss for the higher-intensity storm (29 mm rainfall) was only 1.4 mm (5%) (Spittlehouse 1998b).

Understorey vegetation, slash, downed wood, rocks, and the forest floor also intercept rainfall (Black and Kelliher 1989; Kelliher et al. 1992; Putahena and Cordery 1996). The lower amount of leaf area in the understorey and less favourable conditions for evaporation because of low wind speed means that

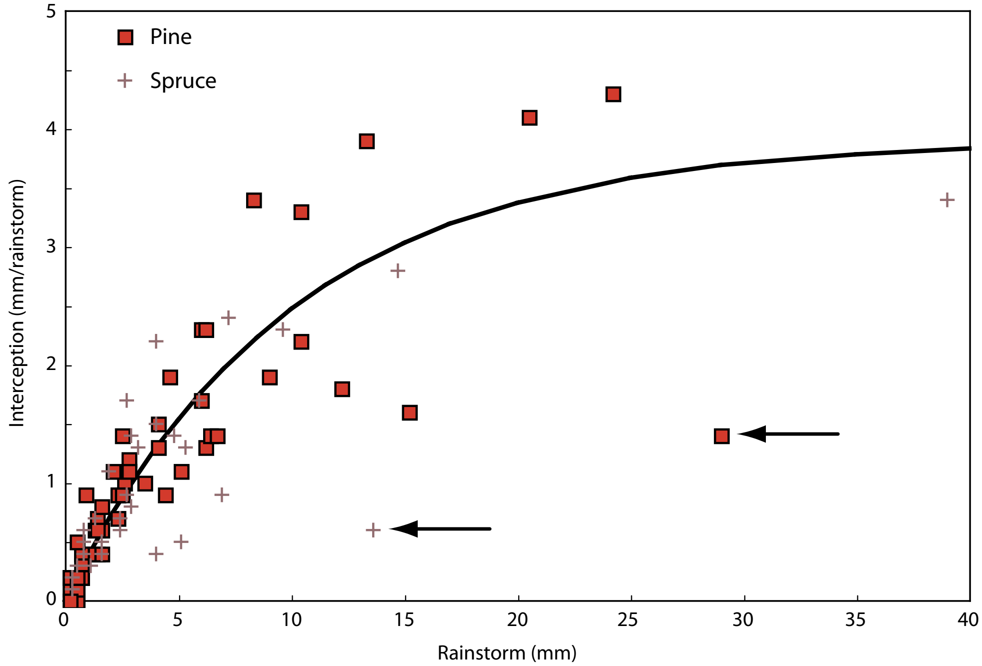

FIGURE 6.3 *Interception loss as a function of the amount of rainfall in an individual rainstorm for lodgepole pine (■) and Engelmann spruce–subalpine fir (+) forests at Upper Penticton Creek. The arrows indicate interception for storms that occurred when the canopy was wet from a preceding storm. Line fitted by eye. (Adapted from Spittlehouse 1998b)*

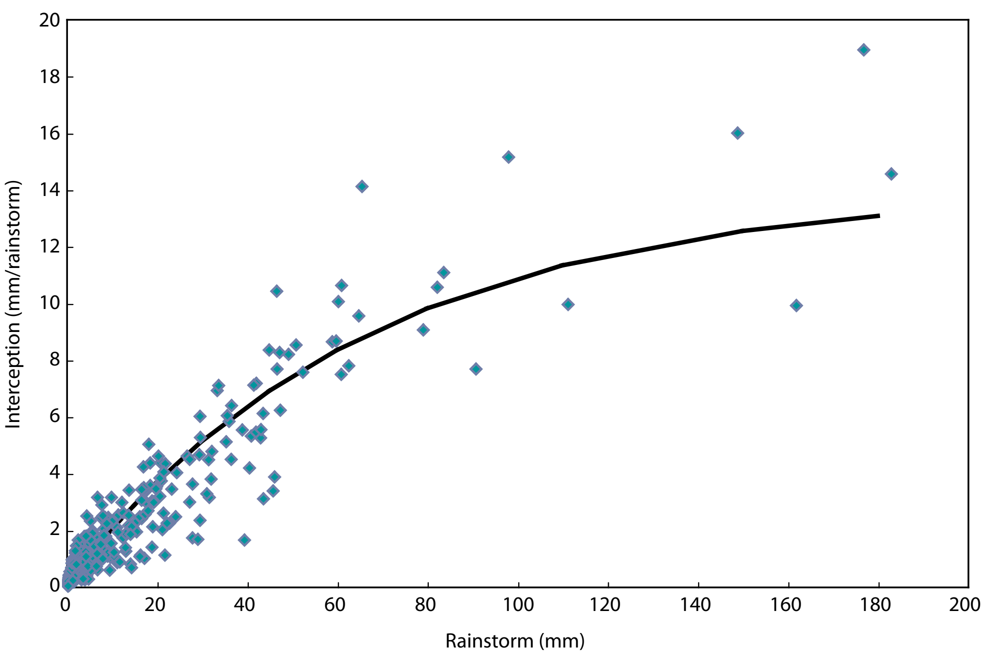

FIGURE 6.4 *Interception loss as a function of the amount of rainfall in an individual rainstorm for a young coastal Sitka spruce forest at Carnation Creek on the west coast of Vancouver Island. Line fitted by eye. (Adapted from Spittlehouse 1998b)*

TABLE 6.1 *Throughfall (T), stemflow (S), and maximum storm interception (Imax) percentages of season-long rainfall in various forest types in British Columbia (Spittlehouse 1998a, 1998b, 2004, unpublished data; D. Maloney, unpublished data)*

Forest type	Age (yr)	Location	Elevation (m)	Canopy cover (%)	T (%)	S (%)	Imax (mm)
Western hemlock–western redcedar	> 250	Prince Rupert area	50	75–80	73–78	1	30
Western hemlock	> 125	Carnation Creek	450	85	69	1	30
Sitka spruce, western redcedar, western hemlock	approx. 20	Carnation Creek	5	75	77	9	14
Douglas-fir	55	Campbell River	300	85	75	3	11
Douglas-fir # 1	25	Cowichan Lake	175	70	70	9	15
Douglas-fir # 2	25	Cowichan Lake	175	40	85	4	9
Lodgepole pine	> 125	Upper Penticton Creek	1650	45	71	< 0.5	5
Engelmann spruce–subalpine fir	> 125	Upper Penticton Creek	1800	45	71	< 0.5	3.5
Lodgepole pine	25	Upper Penticton Creek	1750	40	71	5	4

maximum interception is similar to the interception storage capacity—about 0.2 mm per unit leaf area (Black and Kelliher 1989).

Throughfall and stemflow may be negligible for rainfalls that are less than 3 mm and during the initial stages of larger events because much of the rainfall is intercepted and stored by the canopy. As rainfall increases, a greater proportion of it becomes throughfall (Beaudry and Sagar 1995; Price et al. 1997; Spittlehouse 1998a, 1998b). Stemflow is usually not produced until the canopy interception storage capacity has been exceeded. In an old coastal hemlock forest (Figure 6.2) and in an interior lodgepole pine forest (Figure 6.3), a rainfall of 15 mm or more on a dry canopy was required for stemflow to occur. In younger forests, stemflow commenced with as little as 2 mm of rainfall (Spittlehouse 1998b). In older forests, stemflow is often negligible (Table 6.1).

On an annual basis, interception loss from conifer forests typically ranges from 10 to 40% of total rainfall depending on canopy characteristics and weather conditions (e.g., McMinn 1960; Rothacher 1963; Sollins et al. 1980; Beaudry and Sagar 1995; Spittlehouse 1998a, 1998b; Link et al. 2004; Levia and Frost 2006). At lower elevations in coastal British Columbia where most of the precipitation falls as rain, the percentage of rainfall lost to interception is greater in summer than in winter because precipitation events are generally smaller and weather conditions are more suitable for evaporation. Up to 50% of rainfall can be intercepted in the summer in mature coastal conifer stands. Annual variability in the rainfall regime also affects interception losses. Annual rainfall interception losses from the old interior lodgepole pine site reported in Table 6.1 and Figure 6.3 varied from 23 to 31% over a 10-year period. If this stand was subjected to the rainfall regime of the mature coastal western hemlock stand, then interception loss would be about 12%. No published interception data exist for deciduous trees in British Columbia. In a review of the interception loss literature, Carlyle-Moses (2004) reported that, on average, canopy interception loss from coniferous forests was double that of deciduous stands, with typical values of approximately 26% and 13% of incident rainfall, respectively, during the growing season.

Snow interception

The amount of snow intercepted by forest canopies is affected by weather variables and stand characteristics, including existing snow load, air temperature, wind speed, time since last snowfall, and snowfall amount, as well as tree species, leaf area, stand density, and stem distribution (Schmidt and Troendle 1992; Gray and Prowse 1993; Pomeroy and Gray 1995; Hedstrom and Pomeroy 1998). The lowest interception rates of falling snow occur at wind speeds greater than 2 m/s (Gray and Prowse 1993); the highest rates occur at temperatures of −3 to 0°C (Pomeroy and Goodison 1997). At these temperatures, snow crystals are more cohesive and less likely to rebound from branches, needles, and previously intercepted snow. At lower temperatures, the less cohesive snow is more likely to be redistributed by wind (Schmidt and Troendle 1992). The exact proportion of the total snowfall that is redistributed once it has been intercepted is the subject of much debate. Estimates range from a minor proportion of the total snowfall to as high as 90% during individual storms. Wheeler (1987) suggested that only 2% of the difference in the water equivalent of the snowpack between an opening and an Engelmann spruce–subalpine fir–lodgepole pine forest in Colorado could be explained by the redistribution of snow by wind. As snow ages, the ability of wind to transport intercepted snow decreases (Miller 1962; Schmidt and Troendle 1992).

Schmidt and Troendle (1992) suggested that when winter storms are small (e.g., 5–25 mm of water equivalent), 50% of the snowfall may be intercepted when conifer crown closure exceeds 50%. Pomeroy et al. (1998) found that 56% of the total snowfall was intercepted in a 19 m tall jack pine stand in Saskatchewan with a crown closure of 82%. At a high-elevation site in the Kootenays, Schmidt and Gluns (1991) found that snow interception was 45–50% on individual Engelmann spruce, subalpine fir, and lodgepole pine branches during a 10 mm water equivalent, low-density snowfall event. These authors also found that snow catch decreased to about 10% as snow density increased to 13%, highlighting the importance of meteorological conditions during snowfall events. Snow intercepted by the canopy gradually builds up, forming bridges between the needles and branches. This process increases the surface area on which any additional snow can be intercepted until the weight of the snow can no longer be supported and some or all of it slides off the branches. Intercepted snow that remains on the canopy can be lost through sublimation. In spring, snow on the canopy may also melt, slide off, or wash off during rainfall and warm weather.

Snow accumulation

Snow accumulation on the ground is usually described by its depth, density, and water equivalent. Snow water equivalent (SWE) is the depth of water that would result from melting a given depth of snow. Snow water equivalent is calculated as the product of the snow depth and snow density. Snow accumulation is affected by snowfall, topography, and vegetation. In a given climatic region, snow accumulation generally increases with increasing elevation as a result of greater storm frequency, decreased evaporation, and decreased melt throughout the winter (Gray and Prowse 1993). The relationship between elevation and snow accumulation varies considerably from year to year. Snow accumulation also varies with slope position and orientation, decreasing along a slope oriented parallel to the prevailing winds and increasing in depressions and on lee slopes (Gray and Prowse 1993). Aspect influences the amount of energy that reaches the snow surface, and as a consequence, the magnitude of melt and sublimation losses in late winter before the main melt period. Topographic variability in SWE is reduced by tall vegetation, an effect that generally increases with increasing vegetation density (Pomeroy et al. 1998). Snow accumulation under tall vegetation is generally less than in the open. The loss of forest cover caused by logging, fire, insects, or disease generally results in increased snow accumulation on the ground (see Chapter 7, "The Effects of Forest Disturbance on Hydrologic Processes and Watershed Response").

In forests, snow accumulation varies between stands of different species, canopy density, and stem distribution. Within the same stand, snow accumulation varies significantly from year to year; however, within-stand variability has been found to decrease with increasing mean accumulation (Winkler and Moore 2006). Snow accumulation also varies with distance from individual trees, increasing up to a distance of approximately 3 m from the trunk of a coniferous tree (Pomeroy and Goodison 1997; Faria et al. 2000). The amount of forest canopy, often described by estimates of canopy closure, gap fraction, or leaf area, is inversely related to SWE on the ground, although the exact nature of the relationships between these variables changes with climatic conditions (Metcalfe and Buttle 1998).

Pomeroy and Goodison (1997) reported that snow accumulation in the boreal forest was greater in stands of aspen than of jack pine and was least in stands of black spruce. The increases in snow accumulation corresponded to the lack of foliage during winter in aspen stands and to lower leaf area in pine relative to spruce stands. Pomeroy et al. (1998) stated that up to 70% of the spatial variation in SWE under forest cover is related to winter leaf area. At high-elevation sites in central British Columbia, Teti (2003) found that canopy density, measured in a cone 60° wide, explained 22–73% of the variability in mean peak SWE, with the least variability occurring in the smallest openings, likely as a result of edge effects. In the southern interior of British Columbia, crown closure explained 68% of the variability in pre-melt SWE across a range of typical forest types (Winkler et al. 2004). In a survey of lodgepole pine stands of varying ages on the Thompson-Okanagan plateau, April 1 SWE was reduced by approximately 6% for every 10% increase in crown closure (Winkler and Roach 2005).

Snow metamorphism

Newly fallen snow generally has a complex crystal structure and low density, ranging from 50 to 120 kg/m^3 (Pomeroy et al. 1998). On the ground, snow undergoes continual change in crystal form, surface conditions, temperature, water content, permeability, and density. Snowpack metamorphism refers to the change in the shape of snow grains through time. The nature of metamorphism depends on the presence or absence of liquid water, snowpack temperature, and vertical temperature gradient through the snowpack. The vertical temperature gradient depends on the temperature at the ground–snow interface, as well as on air temperature and snowpack depth.

In dry snow with weak vertical temperature gradients (less than 10°C change over each metre of snow depth), metamorphism is initially caused by the migration of water molecules from the convex areas of a snow crystal (which have a high equilibrium vapour pressure) to the concave areas (which have a lower equilibrium vapour pressure). As a result, snow crystals evolve from original complex forms toward rounded grains, typically about 0.5 mm in diameter. This process is referred to as *equilibrium metamorphism*. It dominates in regions with consistently below-freezing air temperatures and relatively deep snowpacks, such as the Columbia Mountains and inland, and at higher elevations in the Coast Mountains. The rate at which equilibrium metamorphism proceeds decreases with decreasing temperature.

Strong vertical temperature gradients (greater than 10°C per metre of snow depth) can develop where extremely low air temperatures coincide with relatively shallow snow because the base of the snowpack is maintained within a few degrees of freezing by heat conduction from the soil. In these situations, the strong temperature gradient is accompanied by a strong vapour pressure gradient: the lower, warmer portions of the snowpack have higher vapour pressures in the pore spaces than does the overlying snow. In a process called *kinetic metamorphism*, these gradients drive the diffusion of water vapour upward, where it condenses and refreezes to form faceted crystals, which are often called *depth hoar*. It is a common occurrence in the Rocky Mountains, but also occurs in other regions.

In the Coast and Cascade Mountains, and at lower elevations in the interior of the province, snowpacks experience periodic mid-winter melt and rainfall followed by percolation of liquid water into the snowpack. In the presence of liquid water, snow grains rapidly evolve into relatively coarse, rounded grains up to 1 mm or more in diameter. If the snowpack temperature is below freezing, some or all of the percolating water freezes, releasing its latent heat of fusion and warming the snowpack. Repeated melt–freeze cycles produce bonded clusters of coarse grains. If a snow surface exposed to melt and refreezing is buried by additional snowfalls, it can be a barrier to vertical percolation through the snowpack during subsequent melt–freeze cycles. Percolating melt and rain may pond on such a surface and refreeze, sometimes producing thick ice layers. The resulting stratification in the snowpack is not uniform across the landscape, which produces large spatial variability in snow density and meltwater percolation (Langham 1981).

As a result of snow metamorphism, the density of the snowpack tends to increase through the winter, reaching densities of 200–300 kg/m^3 or more once the snowpack has settled and been exposed to wind (McKay and Gray 1981). During the seasonal transition from winter to the onset of spring melt, snowpack densities typically increase to 250–500 kg/m^3 because of ongoing metamorphism and retention of liquid water (Pomeroy and Gray 1995; Pomeroy et al. 1998). Figure 6.5 shows an example of this general increase in snow density over the ablation season at Mayson Lake on the Thompson Plateau. The density of the snowpack measured on day 92 was less than that on the previous date as a result of snowfall between days 84 and 92, as shown by the increase in SWE.

Snowpack metamorphism is of great interest in relation to avalanche hazard, but is also relevant to hydrological processes. For example, the albedo of freshly fallen snow is high due in part to the complex grain shapes. As snow metamorphoses into rounded grains, the albedo declines. Depth hoar has a low thermal conductivity, which restricts the conduction of heat from the soil into upper layers of the snowpack. Small snow grains have a higher water retention capacity than coarser grains, and also have a lower hydraulic conductivity, which results in lower rates of water percolation through the snowpack. Thick ice layers within a snowpack can impede vertical water percolation and encourage downslope flow. See Langham (1981) and McClung and Schaerer (2006) for further discussions of snowpack processes

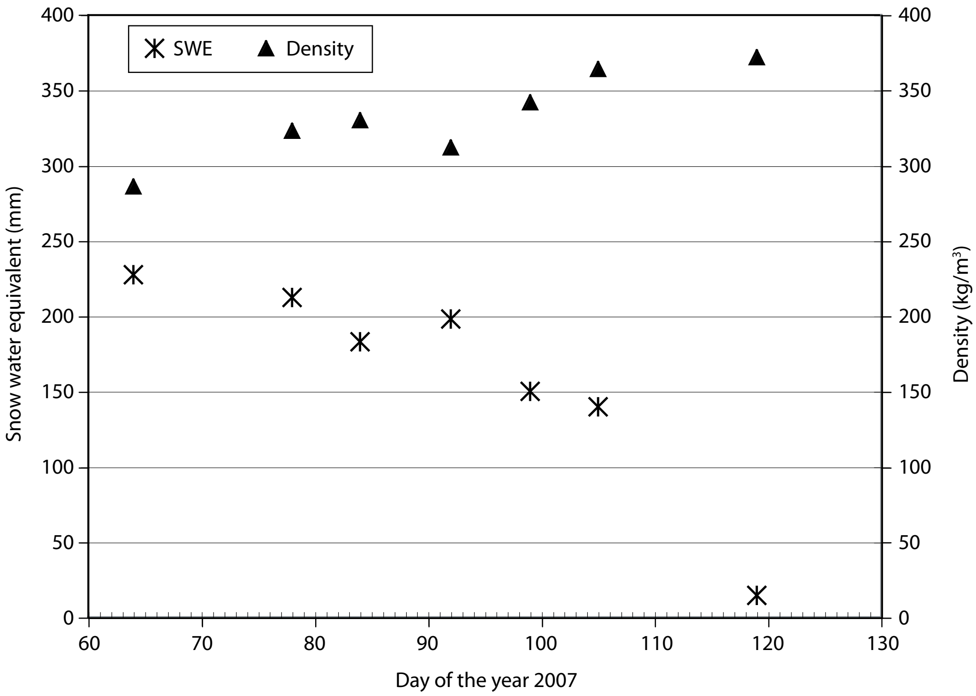

FIGURE 6.5 *Changes in snow water equivalent (SWE) and snow density in a clearcut at Mayson Lake. (R.D. Winkler, unpublished data)*

and properties, and Haegeli and McClung (2007) for a treatment of the geographic variability of these processes in southern British Columbia.

Snowmelt

The rate at which snow melts depends on the amount of energy available to change snow to liquid water. The energy fluxes to and from a snowpack include: radiant energy (Q_R, shortwave and longwave radiation); latent heat flux (Q_E, energy released through the condensation of water onto the snowpack or lost through evaporation); sensible heat flux (Q_H, energy transferred to or from the snow surface by turbulent exchange); conduction/ground heat flux (Q_G, energy conducted to the bottom of the snowpack from the ground); advection (Q_P, transfer of heat to the snowpack through rain); and the rate of change of internal energy (stored) in the snowpack (Q_S). The amount of energy available for melt (Q_M) can be expressed using the energy balance equation:

$$Q_M = Q_R + Q_E + Q_H + Q_G + Q_P - Q_S \qquad (3)$$

All streams of energy are expressed in units of megajoules per square metre per day (MJ/m^2 per day) or watts per square metre (W/m^2). During snowmelt periods, the term Q_S is often omitted (especially for daily totals) because its assumed value is small relative to the other energy balance components and because the snowpack is close to 0°C during melt (Male and Gray 1981). Net radiation is the difference between incoming and outgoing shortwave and longwave radiation:

$$Q_n = S\downarrow - S\uparrow + L\downarrow - L\uparrow \qquad (4)$$

where: $S\downarrow$ is the shortwave (global) radiation arriving at the snow surface; $S\uparrow$ is the shortwave radiation reflected by the snow surface, which is a function of the surface albedo (reflectivity of the snow); $L\downarrow$ is the longwave radiation arriving at the snow surface from the sky and forest canopy; and $L\uparrow$ is the longwave radiation emitted by the snow surface, which is determined by the temperature of the snow surface plus a typically small contribution attributed to the reflection of incident longwave radiation.

The amount of meltwater produced (M), expressed as a snow water equivalent (millimetres), can be calculated from:

$$M = (Q_m / [\rho_w \lambda_f B])\ 1000 \qquad (5)$$

where: Q_m is in MJ/m^2 per day; ρ_w is the density of water (approximately 1000 kg/m^3); λ_f is the latent heat of fusion (0.334 MJ/kg); and *B* is the thermal quality of snow (generally between 0.95 and 0.97) (Pomeroy and Goodison 1997).

Shortwave radiation reaching the Earth's surface is a function of slope, aspect, cloud cover, time of year, and time of day. The amount of shortwave radiation reaching a snow surface under forest cover is reduced relative to that in an open site. The amount of reduction is not easily related to simple measures, such as canopy closure, because this reduction also depends strongly on canopy volume, stem distribution within the canopy, and density and arrangement of foliage elements, as well as on the relative proportions of direct and diffuse (direct radiation scattered by particles in the atmosphere) solar radiation reaching the top of the canopy, and the Sun's location within the sky dome. In stands with sparsely dispersed trees and at the edges of openings, shading effects can extend up to a distance of two to three times the crown height from the tree trunk (bole) (Bohren and Thorud 1973; Spittlehouse et al. 2004).

The fraction of incident shortwave radiation that is reflected is termed the *albedo*. As discussed above, albedo depends on the snow age and the nature of metamorphism. Snow albedo also depends on the fractions of diffuse and direct radiation (Pomeroy and Goodison 1997). The albedo can be as high as 0.95 for a fresh snow surface colder than 0°C and as low as 0.2–0.3 for an old, thin snowpack (Geiger et al. 1995). Below a forest canopy, albedo is also affected by the amount of leaves, lichens, and moss that has fallen on the snow surface (Adams et al. 1998; Melloh et al. 2001; Spittlehouse and Winkler 2004). Shortwave radiation can be absorbed and reflected within the snowpack, the amount of which decreases exponentially with depth. As much as 40% of the shortwave radiation received at the snow surface penetrates to a depth of 10 cm, and only 10% or less reaches a depth of 25 cm (Geiger et al. 1995). This energy may be reflected back to the surface or may contribute to snowpack metamorphism.

Longwave radiation depends on the temperature of the radiating surface and its emissivity. The emissivity is a measure of the efficiency with which a body emits longwave radiation relative to a theoretical black body, which has an emissivity of unity. The emissivity of the atmosphere depends on temperature and vapour content (and thus cloud cover). Under clear sky conditions, a typical emissivity for the atmosphere would be 0.75, whereas the emissivity under thick cloud cover can approach unity. Longwave radiation below the forest canopy is composed of sky radiation that penetrates through gaps in the canopy plus radiation from the canopy and tree trunks. It is usually greater than sky radiation because the trees are warmer than the sky and have a higher emissivity than is typical for clear sky conditions. Snow absorbs 95–99% of the incident longwave radiation, reflecting less than 5%. Because snow has a high emissivity approaching unity (Oke 1987), longwave radiation emitted by snow depends primarily on its surface temperature (Gray and Prowse 1993).

The increase in longwave radiation emitted by a forest canopy compared to sky radiation can partially compensate for the reduction in shortwave radiation reaching a snow surface below the canopy. For example, Woo and Giesbrecht (2000) found that shortwave radiation was reduced under a mixed black and white spruce canopy relative to the open areas, and longwave radiation was enhanced, particularly during overcast conditions. Overall, net radiation decreases as canopy density increases up to approximately 60%, after which net radiation increases (Pomeroy and Goodison 1997). Under dense canopies, increased longwave radiation from the canopy and multiple reflections of solar radiation between the canopy and snowpack can compensate for reductions in solar radiation as canopy density increases (Bohren and Thorud 1973). Net radiation can also be greater under a deciduous than a coniferous canopy (Pomeroy and Goodison 1997).

Convective fluxes depend on the intensity of turbulence and thus on wind speed, and the roughness of the snow surface. These fluxes also depend on the differences in temperature (for sensible heat) and vapour pressure (for latent heat) between the overlying air and the snow surface. Convective flux often provides a much smaller source of energy to the snow than radiation. Under forest cover, where wind speeds are low, these fluxes are small (Woo and Giesbrecht 2000; Spittlehouse and Winkler 2004); however, convective fluxes can be important energy sources at open sites during weather conditions that are dominated by warm, moist maritime air masses and strong winds (such as "Pineapple Express" or "Tropical Punch" events, as described in Chapter 3, "Weather and Climate").

Ground heat flux is usually negligible compared to radiation and to latent and sensible heat fluxes, and is often ignored or capped in energy budget predictions of daily snowmelt (Pomeroy and Goodison 1997). Nevertheless, ground heat flux is important for maintaining the heat content of the snowpack during winter and thus for minimizing the amount of energy required to raise the snowpack to the melting point in the initial stages of snowmelt.

Sublimation from the snowpack occurs when the temperature is less than 0°C and the vapour pressure of the air is less than that of the snow surface. Sublimation and evaporation from snow-covered clearcuts is usually less than 1 mm/d, amounting to a total seasonal loss of 10–20 mm SWE (Bengtsson 1980; Bernier 1990; Prévost et al. 1991; Adams et al. 1998). Although water losses from the snowpack are small, the energy lost or added through sublimation or condensation can affect melt rates.

The transition from winter snow accumulation to spring melt begins when net heat exchange at a snow surface becomes dominantly positive, primarily as a result of increasing shortwave radiation and air temperature. At the time of this transition, the snowpack often has temperatures below freezing and thus negligible liquid water content. Before substantial quantities of water can percolate to the base of the snowpack, the snowpack must become "ripe"—that is, it must become isothermal at 0°C and its water-holding capacity must be satisfied. Water often percolates through the snowpack via preferred pathways, and therefore it is not necessary for the entire snowpack to become primed for some amount of drainage to occur in the early stages of a rain-on-snow event or spring melt (Conway and Benedict 1994).

Initially, the upper layers of a snowpack will warm to 0°C and then begin to melt. Warming of deeper layers of the snowpack occurs partly by conduction of heat through the snow grains. In addition, meltwater percolates down through the pack and refreezes, releasing its latent heat of fusion and warming the snowpack until it becomes isothermal at 0°C. After a layer of snow has been warmed to 0°C, some of the meltwater or rain percolating to this layer will be retained and any surplus can percolate deeper. The water retention capacity of a snowpack depends on grain size and structure, and is highly variable, ranging from near zero to more than 10% by volume (Kattelmann 1986). As a result of these internal snowpack processes, there is a delay between the onset of rainfall or melt at or near the snow surface and the arrival of that water at the base of the snowpack. This delay is augmented by refreezing of water within the snowpack, which decreases its permeability by filling in voids, thus reducing percolation rates (Pfeffer and Humphrey 1996). Detention in the snowpack can moderate peak streamflows by delaying the delivery of rain or meltwater to channels compared to areas with no or shallow snowpacks. Figure 6.6 illustrates the warming process at Upper Penticton Creek. It shows the distinct lag between the time that maximum daily air and snow surface temperatures reached zero and the time when the entire snowpack became isothermal and the period of continuous melt began.

In an isothermal snowpack, the rate of water percolation varies with the snowpack's structure (particularly permeability, and the presence or absence of ice layers) and water content. During periods of melt, a snowpack's liquid water content can exceed its water-holding capacity because of the presence of water that is percolating under the force of gravity. In fact, during periods of high melt rates, 20% or more of the snowpack volume may be liquid water (Male and Gray 1981). As liquid water enters or moves through the snowpack, metamorphism occurs, changing the water retention characteristics of the pack, as described earlier. As snow grains become larger and more rounded, pore space increases, making the snowpack more permeable and reducing its water-holding capacity. In addition, ice layers tend to decay in the presence of liquid water. These structural changes, in conjunction with decreasing snowpack depth, result in reduced travel times through the snowpack as the melt season progresses, and thus less time lag between peak snowmelt and peak streamflow (Jordan 1983).

During the main spring-melt period, snowmelt rates tend to be highest during episodes of clear weather and are enhanced significantly by sensible and latent heat fluxes under windy conditions at open sites. Snowmelt rates vary strongly with aspect, with the highest rates occurring on south-facing slopes (in the northern hemisphere) because of the aspect-dependence of incident solar radiation. In contrast, mid-winter rain-on-snow events (see next section) typically involve relatively warm, humid air, high wind speeds, and low solar radiation.

Snowmelt rates reported in the literature are often calculated from repeated measurements of SWE over time, and therefore represent the combined losses of water from the snowpack through sublimation, evaporation of meltwater, and outflow. This combined loss is referred to as ablation. Average ablation rates reported in the literature vary from 4 to 25 mm/d in the open and 3–17 mm/d in the forest (Winkler et al. 2005). In the Thompson-Okanagan region, ablation rates calculated from repeated SWE measurements typically vary from 4 to 5 mm/d in mature Engelmann spruce or lodgepole pine

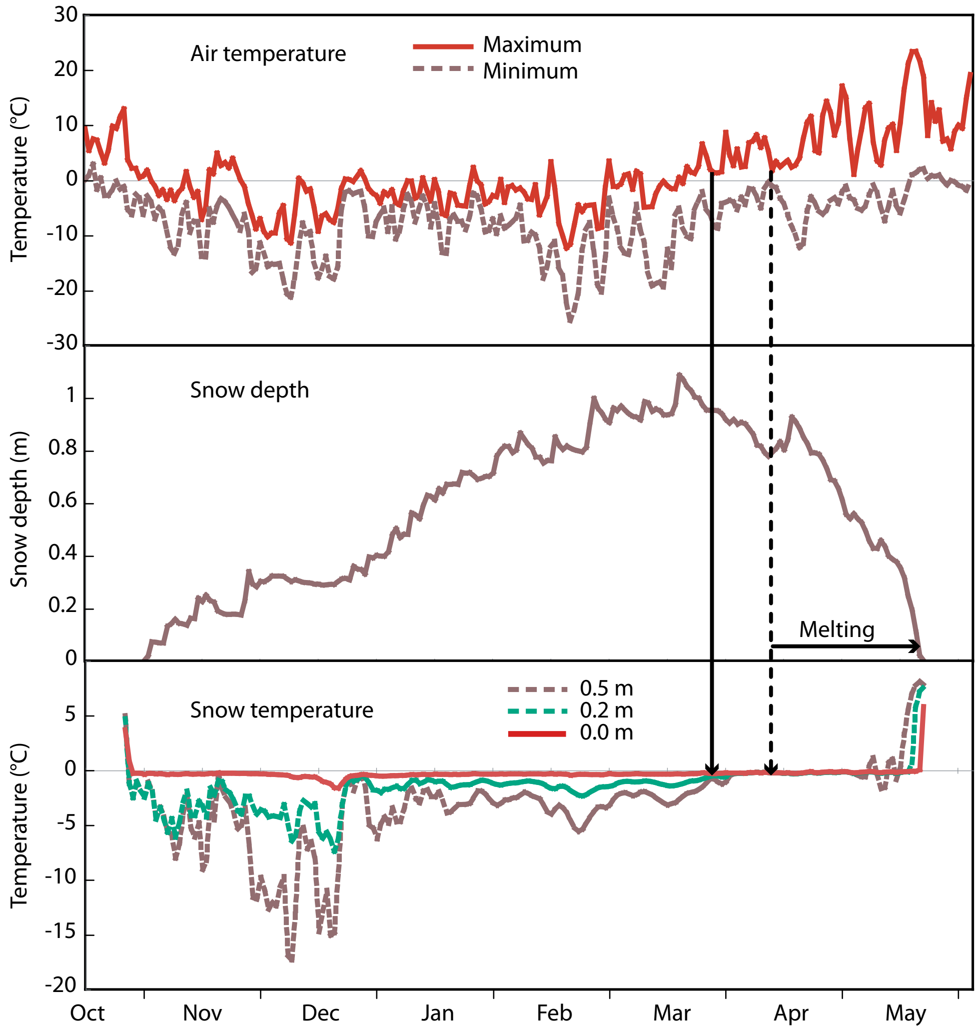

FIGURE 6.6 *Changes in maximum and minimum air temperature (upper panel), snow depth (middle panel), and daily mean snow temperature (lower panel) under a forest during winter 2005/06 at Upper Penticton Creek. When a snow temperature sensor is above the snow (Oct–Dec and May) its readings approximate air temperature. The solid arrow indicates the time when the snowpack has uniformly warmed to about 0°C and the dashed arrow indicates when it has ripened sufficiently to start melting. (D. Spittlehouse, unpublished data)*

stands and from 6 to 10 mm/d in the open (Winkler 2001). At Mayson Lake, maximum daily snow melt rates, measured as lysimeter outflow, were as high as 29 mm/d in the open (Winkler et al. 2005). At 15 sites in the Kootenays, average ablation rates determined from repeated snow surveys were 1.1 and 0.8 mm/d in the open and forest, respectively (Toews and Gluns 1986). Snow ablation, estimated using an energy balance approach, at sites near Vanderhoof ranged from 2 to 5 mm/d in the open and forest, respectively (Boon 2007). In the southern interior of British Columbia, 57% of the variability in standardized (to account for the weather) ablation rates among forest stands was explained by crown closure. Year alone explained 24% of the variability in ablation rates among these sites. These results highlight the importance of both forest structure and the weather in regulating snowmelt (Winkler et al. 2004).

Rain-on-snow events

Mid-winter rain-on-snow events can generate major floods, especially in the Coast and Cascade mountains. During these events, solar radiation is low and melt rates are typically governed by sensible heat transfer from the relatively warm air and from condensation of water vapour onto the snowpack, and in some cases, by the sensible heat of rainfall (Beaudry and Golding 1983; Berris and Harr 1987; Marks et al. 1998). Under these conditions, snowmelt combined with rainfall can result in increased peak streamflow (Harr 1981, 1986). These effects are particularly important in open sites (such as clearcuts) in the transient snow zone between about 300 and 800 m above sea level in south coastal British Columbia, where shallow snowpacks can develop and melt one or more times each winter. The elevational extent of the transient snow zone varies geographically and from year to year. The significance of a shallow snowpack is that it becomes primed and drains quickly during rainfall; however, even for deeper snowpacks, storms that persist for several days (as is typical of "Pineapple Express" events) can successfully prime the snowpack and produce significant floods.

Snow held in a forest canopy usually melts faster than snow on the ground—either under the canopy or in a cutblock—because of its greater exposure to wind and thus convective exchanges of sensible and latent heat (Berris and Harr 1987). If a significant snowpack exists on the ground compared to the amount of snow held in the canopy, then the effect of canopy melt is unlikely to have an important influence on peak flow response except for lower-magnitude events (Harr 1986).

Rain-on-snow events can also generate high peak flows during spring melt when soil moisture levels are high over significant portions of even large watersheds. Under these conditions, a moderate rain event can generate a significant peak flow, especially when augmented by snowmelt.

Evaporation

Evaporation includes all processes by which water returns to the atmosphere as water vapour: evaporation of intercepted rain and snow; evaporation from bare soil and water bodies, such as ponds, lakes, and streams; and transpiration from plant leaves. Evaporation requires the following four conditions: (1) available water; (2) higher humidity at the evaporative surface (i.e., vapour pressure) than in the surrounding air; (3) energy to evaporate the water; and (4) movement, or transfer, of water vapour away from the evaporative surface.

Energy required to evaporate water depends on incoming solar radiation, reflectivity of the evaporative surface, and air and surface temperature. Diffusion and convection move the vapour away from the surface. Increasing solar radiation, air temperature, and wind speed and decreasing atmospheric humidity all create an increase in evaporation rate. Evaporation is enhanced by warm air flowing over a cooler surface (e.g., air moving from dry rangeland over an irrigated crop or a small lake [Oke 1987]), but decreases rapidly with distance from the boundary between dry and wet surfaces.

Intercepted rain or snow, and open water are in direct contact with the air. Both boundary-layer and aerodynamic resistance affect water loss from these surfaces. The boundary layer is a thin layer adjacent to a surface through which vapour moves by diffusion. Aerodynamic resistance describes vapour movement in the rest of the atmosphere. Both resistances depend on the size and shape of the evaporative surface, and both decrease as wind speed increases. Tree needles have a lower boundary-layer resistance than large leaves and a much lower resistance than that of a lake. Trees generate more turbulence to airflow than smooth surfaces, such as a lake; consequently, trees have lower aerodynamic resistances at the same wind speed. The combined resistances for a wet surface are relatively low compared to the resistance to movement of water from inside leaves or from below a dry soil surface.

Soil evaporation

A wet or moist soil surface is similar to a water surface in that the water is essentially in contact with the air. The hydraulic properties of wet or moist soil are such that liquid water can move upward (through capillary action) to the surface to maintain a moist surface layer as evaporation proceeds. As the near-surface soil dries, the water must diffuse through pores to reach the surface, which forms another resistance (a decrease in conductance) in the path of water vapour flow to the air (Hillel 1998). The first stage of evaporation from soil is *demand limited* because the evaporation rate is limited by the *atmospheric demand* imposed by factors such as the intensity of solar radiation, air temperature and humidity, and wind speed. The second stage is *supply limited*, where evaporation is controlled by the rate at which water in the deeper soil layers can move to the evaporating surface.

Transpiration

Plant leaves are analogous to dry soil in that water must pass from the cells within the leaf through stomata to the air. This results in a resistance to evaporation that is at least 10 times that of the combined boundary-layer and aerodynamic resistances. The stomata regulate water loss to maintain an appropriate water status in the leaves by balancing the atmospheric demand for water with the ability of the roots to supply water from the soil. Once evaporative demand reaches a certain level, even though the soil is moist, the stomata begin to close, which maintains evaporation at a constant level. The vapour pressure deficit of the air (see Chapter 17, "Watershed Measurement Methods and Data Limitations") is a good predictor of evaporative demand and its effect on transpiration (Figure 6.7). Transpiration occurs when stomata open to allow carbon dioxide to diffuse in for photosynthesis; thus, stomata usually close at night when there is no light for photosynthesis.

There are species differences in the ability to regulate water loss. Trees usually have higher stomatal resistances to water loss than shrubs and grass (Kelliher et al. 1993, 1995). As with soil evaporation, transpiration has a supply-limited phase caused by soil drying. Stomata open less and start to close earlier in the day, resulting in a substantial increase in stomatal resistance (Tan et al. 1978; Spittlehouse 2003) and subsequently a decrease in transpiration (Figure 6.7). Stomatal resistance and the amount of leaf area of the forest are usually combined in the term *canopy resistance* (Tan and Black 1976). Canopy resistance generally decreases as the canopy of a stand develops, but it tends to plateau at a leaf area index (LAI) of about 4–6 (units of LAI are square metre of leaf area per square metre of ground surface) (Kelliher et al. 1995). Further increases in leaf area produce shading that alters the below-canopy environment, and this combined with the physiological state of what are usually older leaves results in an increase in stomatal resistance. Mature interior lodgepole pine stands typically have an LAI of 2–4, whereas mature coastal Douglas-fir stands have an LAI of 6–20.

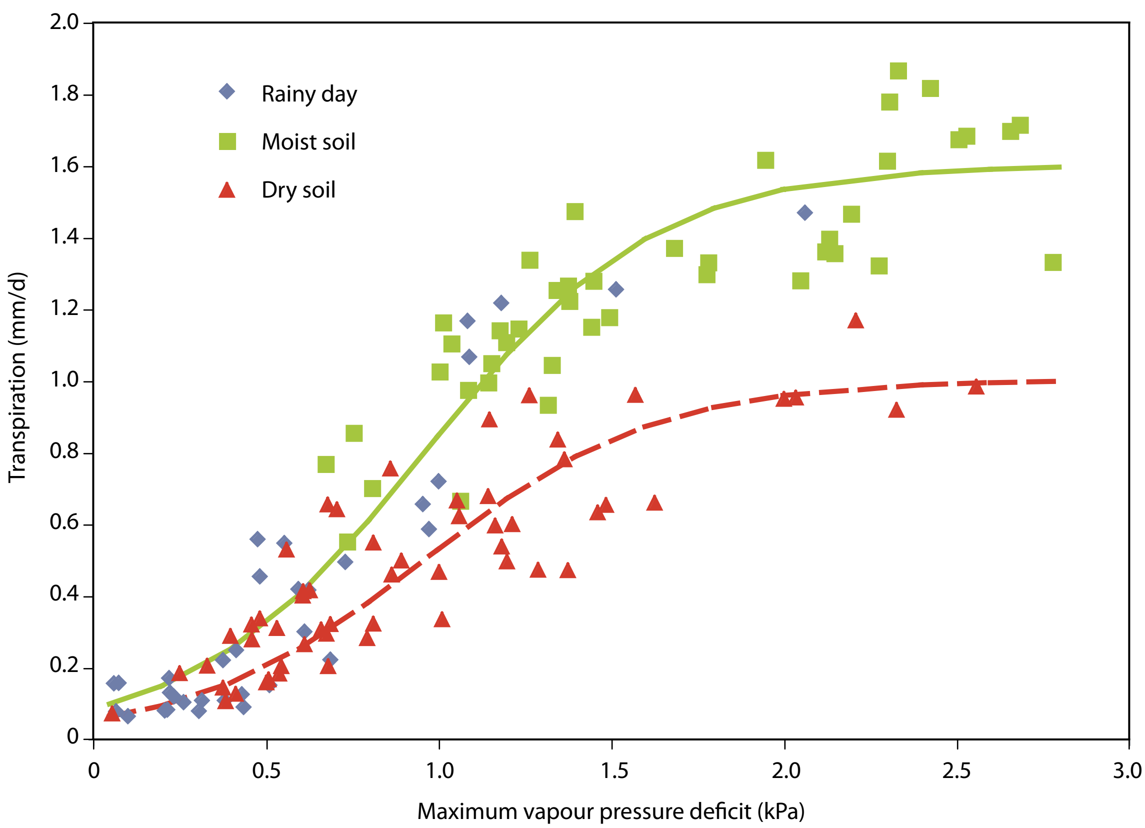

FIGURE 6.7 *Daily transpiration rates of old lodgepole pine trees at Upper Penticton Creek and daily maximum vapour pressure deficit (kPa) on days when the soil was moist (prior to mid-August) when the soil was dry or when nighttime temperatures were below zero (mid-August to late September), and on days with rain (adapted from Spittlehouse 2002).*

Free water evaporation

Seasonal timing and rates of evaporation from water bodies depend on the physical dimensions of the water body (depth, width, and length) and its geographical location (Phillips 1990; Schertzer 1997). As air passes over the water body, evaporation increases the water content of the air and cools it, thereby reducing evaporation rates. Thus, ponds and small lakes are more affected by the air around them than large lakes and usually maintain higher evaporation rates under the same weather conditions. Summer evaporation rates for lakes vary from 50 to 150 mm per month (Phillips 1990; Schertzer 1997). The length of the ice-free season affects annual evaporation, which varies from 800 mm/yr in southern British Columbia to under 400 mm/yr in the northern part of the province. Wetlands, ranging from those with open water to those with full vegetation cover, have evaporation rates of 2–5 mm/d, which translates to an annual evaporation of between 200 and 500 mm/yr (Roulet et al. 1997).

Evaporation of intercepted precipitation

Evaporation rates from forest canopies wetted by rain can exceed transpiration rates from dry canopies under the same environmental conditions. Evaporation rates from wet canopies vary from 0.07 to 0.7 mm/h (De Villiers 1982; Dykes 1997; Humphreys et al. 2003; Price and Carlyle-Moses 2003) and can be similar during daytime and nighttime conditions (Pearce et al. 1980). These high rates of evaporation are sustained by the sensible heat flux to the canopy from the planetary boundary layer and the low aerodynamic resistance of the canopy. For grasses and similar short vegetation, such as understorey plants, the aerodynamic resistance is similar in magnitude to that of the canopy resistance. In this case, the evaporation rates of intercepted rain range from 0.07 to 0.4 mm/h depending on whether the canopy is partially wet or saturated (Calder 1991). Transpiration rates are less than this during rainfall because the vapour pressure gradient is small.

In winter, sublimation losses of intercepted snow depend on wind speed, air temperature, humidity, and radiation, and on canopy characteristics, including canopy surface area. Most of the energy used in sublimation comes from warm, dry air advected into the canopy (Schmidt and Troendle 1992). Maximum rates of sublimation occur during clear periods between small frequent storms, which deposit snow on cold, stiff branches, and are facilitated by the high surface area of snow exposed to the air on the tree crowns. In dense conifer canopies, all of the intercepted snow (up to 30% of the total snowfall on average) can be lost to sublimation and evaporation over the winter (Schmidt and Troendle 1992). Maximum sublimation and evaporative losses from boreal and subalpine forest canopies were found to vary from 3 to 5 mm/d (Woo et al. 2000; Molotch et al. 2007), and average 0.1–0.2 mm/d over the winter (Arain et al. 2003).

Forest evaporation rates

In British Columbia, evaporation rates from a wet soil surface in a clearcut can be greater than 3 mm/d, but would be 10–20% of this under a closed forest canopy. This rate is rarely maintained for more than a day or two, and as the soil dries, the rate rapidly decreases to a steady loss of 0.1–0.2 mm/d after about 10 days with no rain (Kelliher et al. 1986). On an annual basis, evaporation from a high-elevation clearcut in the southern interior of British Columbia varied from 175 mm in a dry year to 300 mm in a wet year (Spittlehouse 2006a). The evaporation characteristics of a clearcut slowly change as the vegetation regrows (Adams et al. 1991; Vertessy et al. 2001; Delzon and Loustau 2005; Spittlehouse 2006b).

Increasing the proportion of forest cover does not correspond to an equal increase in evaporation; there is often a threshold amount of forest cover at which evaporation peaks. In stands with low canopy cover, evaporation from the soil and understorey vegetation contributes substantially to stand evaporation. As tree cover increases, shading reduces these losses. Also, individual trees tend to transpire more water under conditions of lower stand density (e.g., Tang et al. 2003; Bladon et al. 2006; Simonin et al. 2006). As noted earlier, above a certain leaf area, canopy resistance tends to remain constant (Kelliher et al. 1995), offsetting the effects of increased leaf area. Younger trees with efficient water transport mechanisms can maintain higher transpiration rates than tall, old trees under the same weather conditions (Hubbard et al. 1999; Delzon and Loustau 2005) because of the higher water conductivity of their stems.

On sunny days, typical evaporation rates from provincial forests with a dry canopy are 2–4.5 mm/d (Kelliher et al. 1986; Spittlehouse 1989, 2002; McCaughey et al. 1997; Humphreys et al. 2003). These rates decrease as the soil dries and soil water potential drops below about −0.2 MPa. Shallow and (or) coarse-textured soils have a relatively low water storage capacity. Consequently, transpiration and evaporation can be substantially reduced during periods of low rain (Giles et al. 1985; Fleming et al. 1996; Spittlehouse 2003). In an old lodgepole pine stand at Penticton Creek, the mean daily areal tree transpiration rate was between 1 and 1.5 mm/d in mid-summer (Figure 6.8). This decreased to about 0.5 mm/d in late August and September when soil water potential dropped below –0.5 MPa, and remained low in the cool, late September and October weather. Average below-canopy evaporation rates varied from 0.7 to 0.2 mm/d. Mean daily evaporation rates varied from 2.4 mm/d in June through August to 1 mm/d in late September. The trees and the below-canopy vegetation plus soil contributed 42% and 25%, respectively, of the total evaporation (239 mm) during the snow-free season. The remaining 33% was contributed through the evaporation of intercepted rainfall (78 mm, or 28% of the rainfall).

In British Columbia, few annual or seasonal measurements of evaporation are published for land surfaces. In most cases, seasonal and annual totals must be determined using process-based models of evaporation based on measured characteristics of the evaporative surface (e.g., Spittlehouse and Black 1981; Kelliher et al. 1986; Spittlehouse 2004, 2006a). The only long-term measurements are for second-growth Douglas-fir stands on the coast (Jassel et al. 2009). From 1998 to 2007, annual evaporation varied from 410 to 480 mm for a stand that was 59 years old in 2007. Monthly evaporation rates varied from as low as 10 mm in winter to over 70 mm in mid-summer. Although soils are about 0.8 m deep at the site, evaporation was reduced by about 20% in July and August in a summer with low precipitation. An 18-year-old stand had similar annual evaporation rates, illustrating that once plant canopy leaf area exceeds a certain level, evaporation remains relatively constant. Similar numbers were obtained for the 59-year-old stand using a water balance model (Spittlehouse 2004). The modelling study showed that, on average, about 25% of the evaporation was from intercepted water. In contrast, annual evaporation from high-elevation, old lodgepole pine stands varied from 335 mm in a dry year to 430 mm in a wet

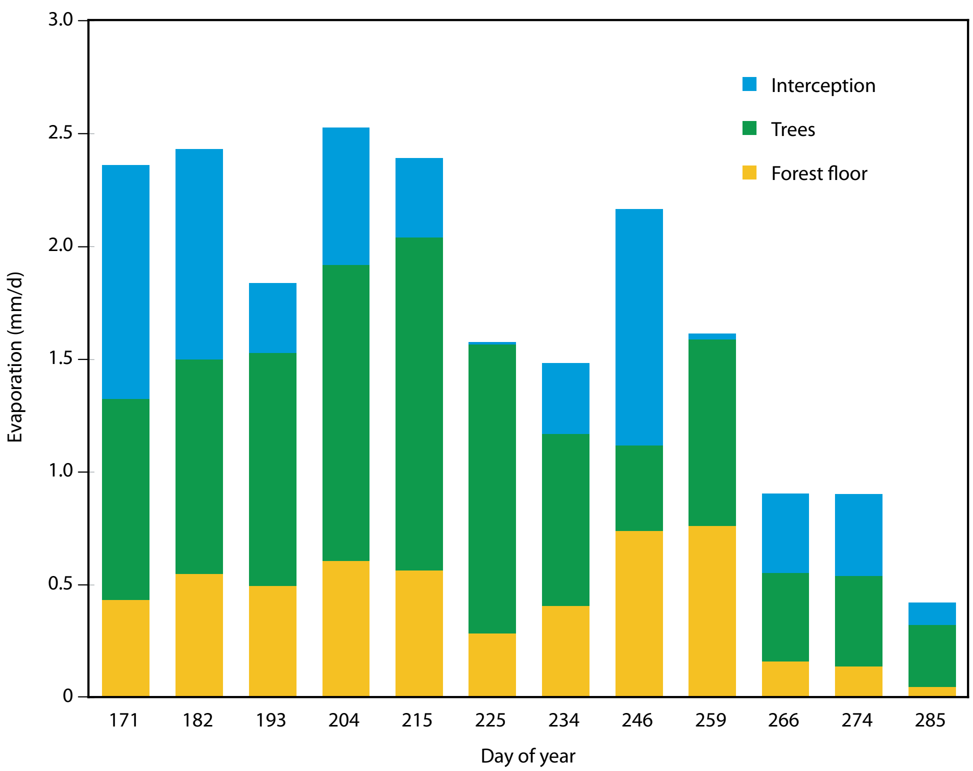

FIGURE 6.8 *Average daily evaporation of intercepted water (blue), tree transpiration (green), and below-canopy evaporation (yellow) (transpiration from the understorey and trees less than 3 m tall, plus evaporation from the soil surface) at Upper Penticton Creek. Uncertainty in interception is 0.2 mm/d, in tree transpiration is 0.2 mm/d, and below-canopy evaporation is 0.4 mm/d. (Adapted from Spittlehouse 2002)*

year (Spittlehouse 2006a). In this forest type, about 40% of the evaporation was through sublimation and evaporation of intercepted precipitation (Figure 6.8).

Measurements of annual evaporation from the boreal forest in Saskatchewan may be suggestive of rates in northern British Columbia ecosystems. For example, annual evaporation from a mature aspen stand ranged from 270 mm in a year with substantial drought to 400 mm in a wet year. The growing season was only 4 months long but accounted for 75% of the annual evaporation (Kljun et al. 2006; Krishnan et al. 2006). Conversely, during the same period in the same area, annual evaporation rates in an old black spruce stand on a wet site ranged from 300 to 330 mm, whereas rates in an old jack pine stand on a dry site ranged from 220 to 260 mm (Arain et al. 2003; Kljun et al. 2006). Consequently, maximum daily evaporation from the aspen stand during summer was more than 4 mm compared to 3 mm for the black spruce and jack pine stands. Average evaporation rates during summer for the black spruce and jack pine stands were 3 and 2 mm/d, respectively (Kljun et al. 2006), and during winter were 0.1 to 0.2 mm/d, respectively.

Calculation of evaporation for water balance analysis is best done by separating interception from transpiration. Allen et al. (1998) showed how to calculate a reference evaporation rate that can be adjusted to different surfaces. This method can be used on a daily or monthly time step. Chapter 3 ("Weather and Climate") provides examples of calculated reference evaporation rates for selected locations in British Columbia (see Figure 3.8 and Table 3.2).

Water Storage and Movement on Hillslopes

Water arriving at the ground surface can accumulate on the surface, infiltrate into the forest floor and soil, and (or) flow over the surface. The flow path that water takes (Figure 6.1) is primarily determined by surface and soil properties, antecedent moisture conditions, and the characteristics of the precipitation (rain or snowmelt) delivered to the soil surface.

Depression storage

If the soil surface has a low infiltration capacity and low hydraulic conductivity, and if the topography allows for surface storage, then water may be stored at the surface in small pools or depressions. These water-filled depressions, called *vernal pools*, are often seasonal features (Rains et al. 2006) that form because of perched water tables. These depression storage areas may become hydrologically connected during high water conditions and develop a flow network to deliver water to streams or other surface water bodies (Rains et al. 2006).

Overland flow

If the rate of rainfall and (or) snowmelt exceeds the infiltration capacity, then some of the surplus water will flow downslope over the soil surface as infiltration-excess overland flow (often termed *Hortonian* overland flow) (Tarboton 2003). In British Columbia, overland flow seldom occurs on undisturbed forest soils, which typically have sufficiently high infiltration capacities and hydraulic conductivities (Cheng 1988). In most soils, infiltration is aided by the presence of vertical macropores fed by flow concentration within the overlying organic horizons (deVries and Chow 1978).

In near-stream zones (e.g., riparian zones or floodplains), the water table may rise to the soil surface during storm or snowmelt events because of the direct input of rain and (or) snowmelt plus contributions of subsurface stormflow from surrounding hillslopes. These zones of saturated soil can generate saturation overland flow, which flows directly to the stream channel. Saturation overland flow includes return flow (water flowing to the surface from below) as well as snowmelt or precipitation falling directly onto the saturated zone (Figure 6.1). Depending on the topography, these areas may expand during storm events and during the wet season in response to a rising water table, and subsequently contract between events and during the dry season. Saturation overland flow may be significant in watersheds with relatively wide riparian zones or floodplains, particularly in relatively low-gradient terrain (e.g., Taylor 1982). In steeper headwater areas, riparian corridors constitute only a small portion of the watershed and may be the primary area for runoff generation during dry periods (e.g., Sidle et al. 2000) when most of the rain falling onto hillslopes (outside of the riparian zone) is retained within the soil and is unavailable to generate streamflow.

Water repellency (also know as hydrophobicity) and restricted infiltration can occur in situations where mineral soil grains become coated with organic compounds. Many forest soils naturally exhibit water-repellent characteristics when dry, but this effect diminishes once soil moisture increases. Barrett and Slaymaker (1989) found that such water repellency occurred naturally in shallow soil layers in subalpine forests at sites across southern British

Columbia. Wildfire may create or enhance water-repellent soil conditions. It can create a more severe and thicker water-repellent layer by partially volatilizing soil organic compounds, which subsequently condense onto cooler soil particles deeper in the profile (Letey 2001; Wondzell and King 2003). Such water-repellent conditions may lead to localized overland flow, slowed water movement through the soil, and hence altered subsurface recharge, quickened streamflow delivery, and increased potential for surface erosion, especially during storms in dry watersheds (Scott and Pike 2003; Curran et al. 2006; also see Chapter 8, "Hillslope Processes").

Two studies showed that soils in slashburned clearcuts had a higher tendency to water repellency than soils in old-growth stands or in clearcuts that had not been burned. McNabb et al. (1989), working in Oregon, showed that infiltration rates recovered quickly after burning. Henderson and Golding (1983), working in south coastal British Columbia, reported little evidence of widespread overland flow at the hillslope scale that could be attributed to fire-induced repellency, although erosion was observed on one steep slope that had been slashburned 2 years earlier.

SUBSURFACE PROCESSES

The relationship between precipitation received at the land surface (rain and melting snow) and the hydrologic response of surface waters (streams, wetlands, lakes) is strongly controlled by the properties of the surface soil layers, surficial geology, and bedrock (for more information on British Columbia's geology, see Chapter 2, "Physiography of British Columbia"). A knowledge of the hydrological connections between upland portions of a watershed and surface waters is critical to understand the cycling of nutrients and movement of pollutants through a watershed (Stieglitz et al. 2003). The rate at which water can move, as well as the pathways followed between deposition of precipitation and appearance in surface waters, affects peak flows, low flows, and water quality (McDonnell 2003). The study and understanding of subsurface hydrologic processes is complicated by many factors that influence the paths and rates of water movement and by an inability to directly (visually) observe subsurface processes. Therefore, subsurface processes are the major source of uncertainty in hydrologic understanding and our ability to model hydrologic systems (Beven 2001).

Hydrologic Properties of Soils and Porous Media

Subsurface processes in hydrology are controlled by the properties of porous media (e.g., soil, unconsolidated sediment, and rock), which store and transmit water. The primary properties affecting the storage and transmission of water are soil composition (mineral vs. organic), texture, structure, coarse fragment content, and density. Soil properties can be modified by biotic processes (e.g., plant root growth), abiotic processes (e.g., frost activity), and disturbance (e.g., fire, compaction), which results in changes in storage and transmission properties. Soils and other porous media (herein referred to as "soil") are often analyzed as a three-phase system consisting of solids, liquids (water), and gas (air). The proportions of each phase are variable between soils and with depth in most soils. The key points regarding the hydrologic role and properties of soils and porous media are presented below. Detailed information on the hydrologic properties of soil can be found in textbooks on soil science (e.g., Brady and Weil 2007) and soil physics (e.g., Hillel 1998; Rose 2004; Brady and Weil 2007), and in many reference texts on hydrology and hydrogeology.

Soil physical properties
Soil texture describes the relative proportions of different size classes of mineral constituents. Sand, silt, and clay fractions (soil particles ≤ 2 mm diameter) are used to classify soil texture. Sand is the largest size fraction, clay the smallest, and silt is intermediate. Soil texture is estimated in the field using various diagnostic tests and keys, or quantified in the laboratory using a range of analytical methods. Typically, soils with a greater proportion of finer materials (silts and clays) hold more water than soils with coarser textures (sands). Coarse fragments are solids that range from more than 2 mm to over 200 mm in diameter. These fragments reduce the volume of soil that can hold water or air and can influence its hydraulic properties. Soil structure describes the

arrangement and organization of soil particles (Hillel 1998). Structural voids can result in large pore spaces that allow for enhanced water flow (preferential flow) through soils. Bulk density is the mass of soil solids within a given volume of soil, and increases as the proportion of solids in the three-phase system increases. The amount of soil volume that is occupied by pore spaces affects water storage and transmission. Bulk density is also related to the type of soil solids (mineral vs. organic). The bulk density of forest soils ranges from 1000 to 1600 kg/m^3 for mineral soils and from 100 to 1000 kg/m^3 for organic soils, and is typically greater for soils that have been compacted (Fisher and Binkley 2000). Particle density is the ratio of the mass of soil solids to the volume of solids, and is greater than bulk density because it does not include pore spaces. The particle density of organic materials is typically around 1500 kg/m^3 (Redding et al. 2005), whereas mineral materials are typically assumed to have a particle density of 2650 kg/m^3 (Hillel 1998).

The proportion of the soil that is occupied by void spaces is termed the *soil porosity*. These void spaces can be filled with either water or air. The porosity of a soil (f, unitless) is calculated as follows:

$$f = 1 - (\rho_b / \rho_s) \tag{6}$$

where: ρ_b is soil bulk density (kilograms per cubic metre), and ρ_s is particle density (kilograms per cubic metre). When all the pore space is filled with water, the soil is considered to be saturated. The distribution of pore sizes is a function of soil texture, density, and structure. It controls how much water can be held in unsaturated soil and the rate at which water moves in soil.

Water storage in soil

Water is held in the soil by cohesive forces between water molecules and adhesive forces between water molecules and soil particles. Films of water cover the solid particles, and as the soil becomes wetter, water sequentially fills the smallest pores through to the largest pores. The amount of water stored within the soil is typically expressed in one of four ways: (1) gravimetric, (2) volumetric, (3) relative saturation, and (4) depth of water. Gravimetric water content (θ_g) is the ratio of the mass of water (kilograms) to the mass of dry soil (kilograms). Volumetric water content (θ_v) is the ratio of the volume of water to the total volume of soil, and is related to θ_g by soil bulk density (ρ_b):

$$\theta_v = \theta_g / \rho_b \tag{7}$$

where: θ_v is expressed as a ratio (cubic metres water per cubic metres soil) or a percentage (ratio per 100%).

The relative saturation (S) of a soil is:

$$S = \theta_v / f \tag{8}$$

where: S is expressed as a ratio or as percentage.

Knowing the equivalent depth of water (D) stored within a soil profile (e.g., root zone depth), is useful for calculating water balances:

$$D = \theta_v / z \tag{9}$$

where: z is the depth range (metres) over which the storage of water is to be calculated. For example, if the mean θ_v in the top 1 m of soil is 0.3, then the depth of water stored is 0.3 m.

To quantify water movement through the soil, it is necessary to understand the energy state of water in the soil. The energy state of soil water is related to how water is held in the soil (surface tension). As soils dry, larger pores empty, the films of water in fine pores and on the surfaces of the soil particles become thinner, and the energy required to remove the water increases. The energy state, or potential, of water depends on the sum of three forces that act on water in the soil: (1) gravitational, (2) matric, and (3) osmotic. Matric forces are caused by the attraction of the water to the soil; osmotic forces are caused by the attraction of ions (i.e., solutes) in the water. Osmotic forces are small, except in saline soils, and are usually neglected. The gravitational potential depends on the vertical distance between a point in the soil and some reference level. In most cases, as the soil dries, the gravitational potential becomes small compared to the matric potential. When soils are unsaturated, the water potential is negative (less than atmospheric). Other terms used to describe the energy state of water include soil water tension, soil water suction, and pore pressure. In the unsaturated zone, suction and tension are typically considered to be positive values, whereas pore pressure is negative.

The relationship between water content (or relative saturation) and water potential is termed the *soil moisture characteristic curve* or the *water retention curve* (Figure 6.9). Retention curves are generally presented for soil fractions that are less than 2 mm. Correction of the volumetric water content for the coarse fragment content (gravel, stones, and rocks) is done by multiplying the volumetric water content by 1 minus the coarse fraction proportion.

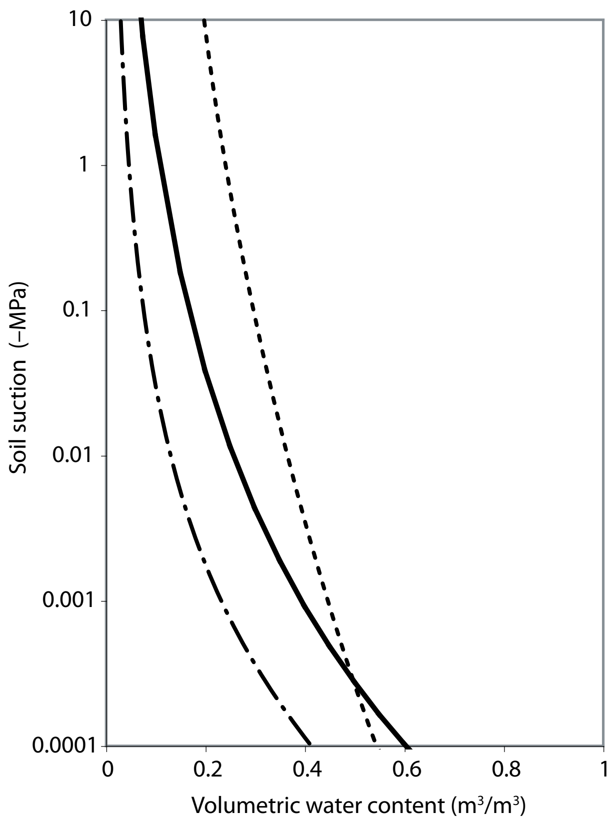

FIGURE 6.9 *Water retention curves for sand (dash-dot), loam (solid), and clay (dotted) soils.*

Two points on the water retention curve are commonly used to understand water storage and transport: (1) field capacity, and (2) permanent wilting point. *Field capacity* (approximately −0.01 MPa or −0.1 bars) is used to denote the water content at which the drainage rate is minimal. If soils are wetter than field capacity, then water will rapidly drain under the force of gravity. *Permanent wilting point* (approximately −1.50 MPa or −15 bars) refers to the water content at which most plants can no longer remove water from the soil. The water content at field capacity and wilting point varies between soils depending on texture and other soil properties. The permanent wilting point is also determined by the type of plant—some plants can access water under drier soil conditions. The soil water available to plants is often described as that between field capacity and the permanent wilting point. Water between −0.01 and −0.20 MPa is readily available to plants. Below this, the soil starts to strongly restrict the water supply, and plants reduce transpiration accordingly (i.e., stomata close). The portion of soil porosity that occurs at a volumetric water content between saturation and field capacity is termed the *drainable porosity*. This is the portion of the soil volume that contributes most strongly to stormflow generation (Weiler et al. 2005). Drainable porosity typically decreases with depth in soils as organic matter content decreases and bulk density increases.

Water movement in soils

Water movement in soil is controlled by several factors, including texture, bulk density, occurrence of preferential flow pathways, and water content. Water flow under unsaturated conditions is controlled by the gradient in water potential between two points (difference in total potential [sum of matric and gravitational potentials] divided by the distance between the points of interest) and the ability of the soil to conduct water. Water movement occurs from zones of higher water potential (wetter, or less negative) to zones of lower water potential (drier, or more negative). It is important to note that water may flow upward (i.e., against gravity) when soils are drier at the surface and wetter with depth. The ability of the soil to conduct water is expressed quantitatively as hydraulic conductivity (K), which is related to soil texture, structure, density, and water content. Hydraulic conductivity is higher in wetter or coarser-textured soils and decreases non-linearly with decreasing water content (or water potential) and in finer-textured soils. The highest hydraulic conductivity occurs under saturated conditions (K_s).

Preferential flow pathways are an important characteristic of water transmission in forest soils (Beven and Germann 1982; Hendrickx and Flury 2001; Weiler et al. 2005). These large pores (also termed *macropores*) can transmit large quantities of water at rates greater than the saturated hydraulic conductivity (K_s) of the surrounding soil matrix. This flow bypasses much of the soil matrix and has major implications for processes such as lateral and vertical flow from hillslopes, and for the transport of dissolved materials (e.g., nutrients or pollutants) to groundwater and surface water bodies. Preferential flow pathways can develop from root channels, animal burrows, structural voids, cracks or fractures in the soil, or subsurface pipes developed over time through subsurface erosion of these pathways (Beven and Germann 1982; Hendrickx and Flury 2001). Preferential pathways may also develop within the soil matrix at transitions between soil layers of significantly differing texture, in zones of greater permeability, such as at the soil–bedrock interface (e.g., Hutchinson and Moore 2000), or in lenses of higher permeability (e.g., Buttle and McDonald 2002). Preferential flow pathways typically constitute only a small portion of the total soil volume (often less than 1%), but, near saturation, pathways can contribute a large proportion of the water flow (Weiler et al. 2005). For example, at Russell Creek on Vancouver Island, Anderson (2008) found that for 12 events where measurements were available, preferential flow constituted approximately 80% of the total hillslope lateral flow.

Infiltration

Infiltration is the process by which water enters the soil. The maximum rate at which water can flow into the soil, or infiltrate, is called the *infiltration capacity* or *infiltrability*. Detailed descriptions of the infiltration process and models of infiltration are available in soil physics and hydrology textbooks (e.g., Ward and Robinson 1989; Hillel 1998; Dingman 2002). Just as with water movement, described previously, the ability of water to infiltrate soil is controlled by soil properties and moisture content, and the presence of frost.

Most forest soils are covered by an organic forest floor layer, which has a high porosity and abundance of preferential flow pathways. Connectivity between large pores across the forest floor–mineral soil interface provides conditions for rapid infiltration (deVries and Chow 1978). In areas without an organic forest floor layer, rapid infiltration can occur in coarse-textured soils or soils with preferential flow pathways described previously (e.g., roots, fractures). Infiltration capacity is reduced in areas where the organic forest floor layer has been removed, where there is no forest floor (e.g., grasslands, agricultural fields), where soil textures are fine, or where soils have been compacted. Soils that have been compacted typically have lower infiltration capacities, owing to the loss of larger pores and macropores (Startsev and McNabb 2000). Compaction effects are more pronounced on fine- than coarse-textured soils. In the absence of an organic forest floor or dense vegetation, the surface may be altered by the impact of raindrops, which can cause larger pores to become clogged by finer materials, thereby reducing the infiltration capacity of the soil.

Soil moisture content prior to the initiation of rainfall (or snowmelt) also influences the infiltration capacity of the soil. Infiltrating water is subject to two major driving forces moving it into the soil. The first is gravity, which pulls water downward from the surface into the subsurface; the second is the water potential gradient created between wet soils above the wetting front and drier soils below it. The latter depends greatly on soil moisture conditions before infiltration (i.e., antecedent soil moisture). The water potential gradient increases when a large difference in moisture content exists between the soil around the wetting front and the soil below it. For soils that are initially wet, this driving force will be smaller; for soils that are initially dry, the infiltrating water near the surface creates a large driving force (water potential gradient) across the wetting front. As the wetting front moves deeper into the soil, the driving force becomes smaller as the water potential gradient decreases. The process of infiltration commonly follows a trend of high initial infiltration rates (infiltration capacity) followed by a decrease in rate through time until a steady state is reached (Figure 6.10). This steady state rate approximates the water movement rate for soils that are completely saturated (i.e., the saturated hydraulic conductivity, K_s).

In areas where soils experience freezing, the infiltration pattern is altered. Under frozen conditions, a fraction of the soil porosity is filled with ice (instead of water), which impedes the flow of liquid water until the ice has melted. Therefore, frozen soil has low initial infiltration rates, which increase with time as ice in the soil melts and water is able to flow into all pores (Figure 6.10). Soils can be frozen at a range of water contents; the higher the water content (which translates into ice content), the lower the infiltration rate. In addition, ice tends to form preferentially in larger pores (Stahli 2005), which are also the most highly conductive to water flow. Dry soils in fall and early winter promote greater infiltration capacity during snowmelt (Stahli 2005). The development of concrete frost in the upper portions of the soil can result in reduced infiltration capacity (Proulx and Stein 1997). The development of concrete frost layers is controlled by soil properties, meteorological conditions prior to melt (cycles of melt and refreezing), and snowpack depth (Proulx and Stein 1997).

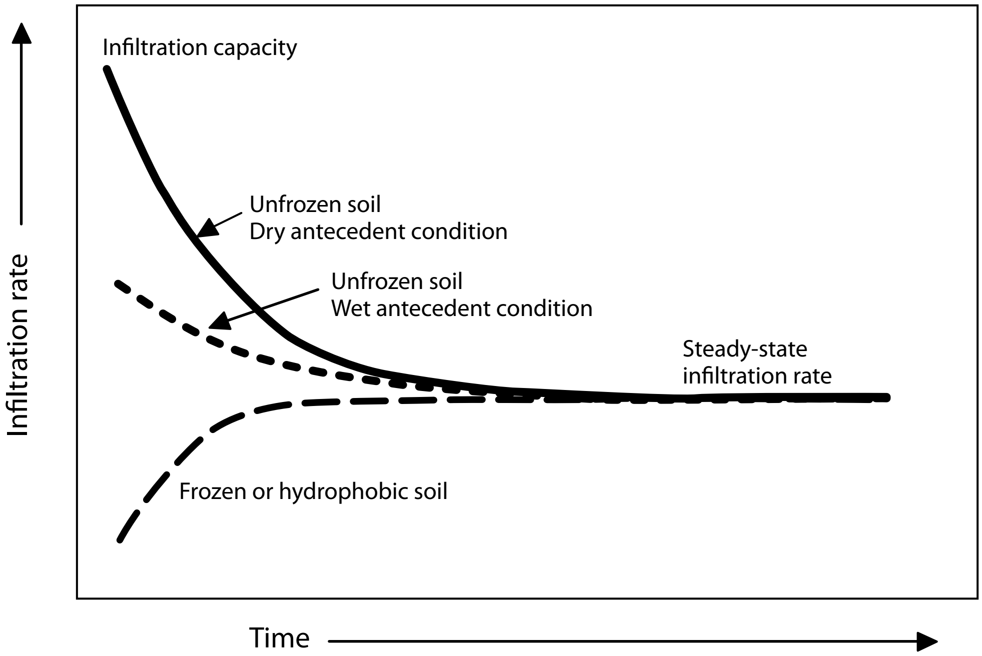

FIGURE 6.10 *Hypothetical example of relative infiltration rates in unfrozen, frozen, and hydrophobic soils.*

Lateral and Vertical Flow from Hillslopes

Following infiltration, water can move vertically, laterally, or some combination of both, depending on subsurface properties, antecedent moisture content, and precipitation characteristics. The flow pathway(s) of water and flow mechanisms that result from these combined factors are the subject of hillslope hydrology. Overviews of hillslope hydrology are available in the compilations edited by Kirkby (1978), Anderson and Burt (1990), and Beven (2006). Review papers on the subject have been published by Bonnell (1993, 1998) and Weiler et al. (2005).

If the available water storage capacity of hillslope soils is large relative to the precipitation amount, then most of the water will be stored within the unsaturated zone of the hillslope (e.g., Redding and Devito 2008). If the available storage capacity of the soil is small relative to the precipitation, then water will flow either laterally downslope (as surface runoff or shallow subsurface flow, referred to as *subsurface storm flow*) or vertically downward to recharge the groundwater (or deeper unsaturated zones). The available storage capacity is a function of soil type (soil texture) and depth to a restricting layer. If soils are shallow, less water is required to fill available storage than in deeper soils. Antecedent soil moisture storage can be a strong control on lateral flow generation. A wetter soil will have less available storage than a drier soil to a given depth. The amount of water that is stored in the soil also depends on plant water use.

The interactions between storage capacity and subsurface storm flow are threshold relationships, with a minimum precipitation amount necessary to result in subsurface storm flow. Studies from steep, humid watersheds around the world indicate that storm total precipitation amounts of 20–55 mm are required to generate any subsurface storm flow (Weiler et al. 2005). Therefore, in many areas of the world, a large proportion of precipitation events will not result in appreciable subsurface storm flow from hillslopes, depending on soil properties and geology. For example, at aspen-forested boreal plain sites with deep, fine-textured soils, Redding and Devito (2008) found that a threshold event rainfall amount of more than 20 mm was required to generate more than 1 mm of lateral flow.

The generation of lateral flow depends on the presence of a restricting layer that impedes vertical water flow and directs it laterally (e.g., Hutchinson and Moore 2000). A restricting layer will have low hydraulic conductivity, and may be formed by bedrock, compacted sediments, a hardpan layer within the soil, or a change in soil texture. For soils and sediments without an impermeable bedrock layer, the restricting layer may be defined relative to the rate of water input to the soil. If the hydraulic conductivity of a layer is greater than the input rate, it will not restrict flow, whereas if the input rate is greater than the saturated hydraulic conductivity of the layer, it will result in lateral deflection of the flow.

Two primary lateral flow–producing mechanisms are associated with a restricting layer. The first involves preferential flow in the saturated layer along the interface between the restricting layer and the overlying sediments, where the materials at the interface have a higher permeability than materials above or below. The preferential flow pathways may be related to the presence of tree roots (e.g., Hutchinson and Moore 2000) or erosional soil pipes (e.g., Buttle and MacDonald 2002). The second lateral flow mechanism involves the downward wetting of the soil column above the restricting layer, followed by the development of a perched water table. As the water table rises towards the surface, the saturated hydraulic conductivity of the soil typically increases, allowing for greater lateral flow as the water table intersects higher-permeability layers or preferential flow paths (e.g., Stieglitz et al. 2003; Devito et al. 2005).

The combination of rapid vertical flow to a restricting layer followed by lateral flow along the restricting layer is common for hillslopes in humid and steep environments (Hetherington 1982; Anderson et al. 1997; Hutchinson and Moore 2000). Depending on local geology, water exchanges between the saturated soil layer and the underlying bedrock may occur. For example, Anderson et al. (1997) showed that a significant amount of water infiltrated from the soil into fractures in the underlying sedimentary bedrock at a site in the Oregon Coast Range, then re-emerged into the soil some distance downslope. Shallow saturated layers typically respond rapidly (i.e., within minutes to hours) to stormwater inputs (e.g., Pierson 1980; Sidle 1984; Jackson and Cundy 1992; Fannin et al. 2000). Regions with abundant subsurface stormflow typically have a shallow restricting layer and a precipitation surplus relative to available soil moisture storage capacity. The control of available soil moisture storage on hillslope-surface water connectivity was shown to be important in shallow soil systems underlain by relatively impermeable bedrock (Buttle et al. 2005; McNamara et al. 2005; Tromp-van Meerveld and McDonnell 2006), and should be even more critical in systems without a well-defined restricting layer and (or) deeper unsaturated sediments (Redding and Devito 2008). Unsaturated lateral flow may occur on forested hillslopes, and is likely more important outside of the growing season, when plant demand is at a minimum.

Riparian groundwater plays an especially important role during periods of low flow. In watersheds with shallow soils and steep hillslopes, base flow is normally supplied by drainage of water stored in the riparian zone, which may be fed by a slow movement of water from the hillslopes (Hewlett and Hibbert 1967). Transpiration by riparian vegetation can extract riparian groundwater that would otherwise discharge into the stream, producing a diurnal decrease in streamflow followed by recovery at night (Hewlett 1982; Bond et al. 2002).

Groundwater Hydrology

Groundwater refers to water that completely fills pore spaces within the zone of saturation beneath the Earth's surface (Meinzer 1923). All geologic materials are composed of solids (i.e., actual grains, sediment, or rock matrix) and pore space (i.e., voids). The amount of available pore space and the interconnectivity of pores govern the storage and transmission of groundwater. If all pore spaces are filled with liquid, then a porous medium is considered to be saturated. If air fills some pores, the material is considered to be unsaturated (Figure 6.11).

The distinction between saturated and unsaturated subsurface zones is based on location of the water table, which is found at the top of the saturated zone (Figure 6.11), where the pore water pressure is equal to atmospheric pressure. Above the water table, water and air occupy pore spaces (Figure 6.11), and the water is held under tension by capillary forces at less than atmospheric pressure. Below the water table, the pore water pressure is greater than atmospheric pressure, and spatial variation in pore water pressure governs groundwater flow.

Groundwater contributes to the generation and regulation of streamflow in watersheds, and the sustainability of many wetlands, ponds, and lakes. The position of the water table (and distribution of pore water pressure) can also strongly influence slope stability (Sidle and Ochiai 2006). In cases where the water table rises rapidly because of a large storm event, many soils have reduced strength, and the potential for slope failure increases (Fannin et al. 2000). In streams, groundwater inflow and exchange with surface water in the hyporheic zone (the region below and adjacent to the streambed where surface

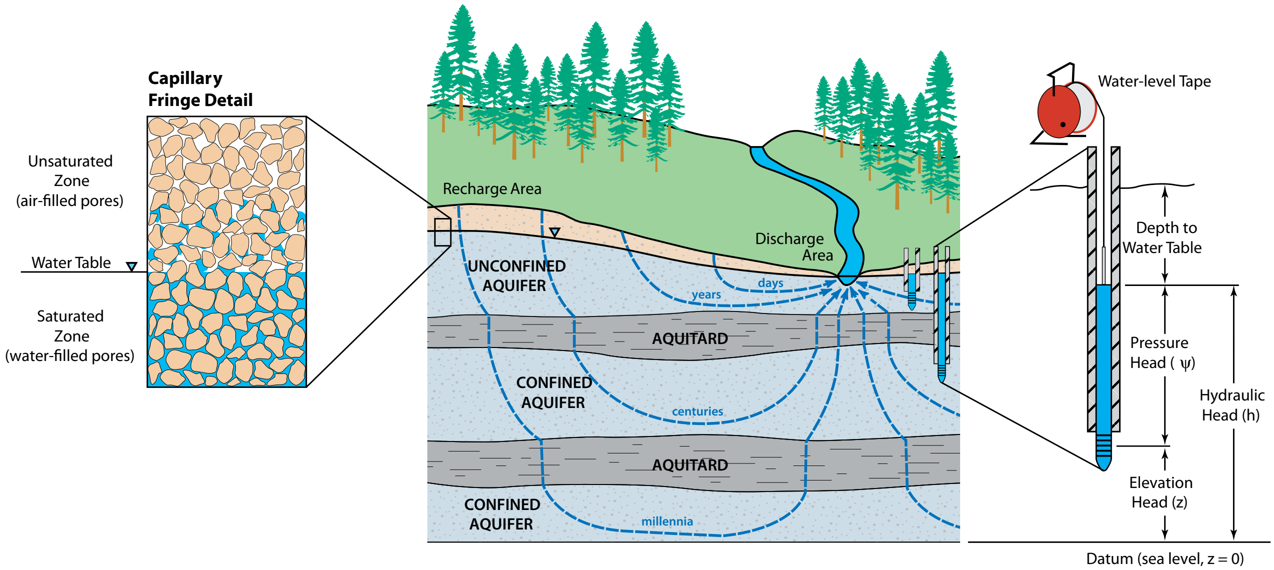

FIGURE 6.11 *Groundwater flow systems (Smerdon and Redding 2007).*

water and groundwater mix) have been shown to regulate stream temperature and aquatic health (Mellina et al. 2002; Moore et al. 2005).

Groundwater flow through porous media

Geologic units can be defined by their ability to store and transmit water. An aquifer is a permeable material that can transmit significant quantities of water to a well, spring, or surface water body (generally, "significant" is defined based on human need rather than on an absolute standard). Often, aquifers are composed of unconsolidated sand and (or) gravel deposits, consolidated deposits that are permeable (e.g., sandstone, limestone), or consolidated formations that are generally less permeable (e.g., granitic and metamorphic rocks) but which have become fractured. An aquitard is a saturated geologic unit that is less permeable than an aquifer and is incapable of transmitting useful quantities of water. Typically, aquitards are composed of clay, silt, shale, or other dense geologic materials. Aquifers may be unconfined (those permeable geologic units open to the atmosphere where the water table forms the upper boundary) or confined (those covered by an aquitard) (Figure 6.11).

Movement of water in the saturated zone is driven by spatial differences in the potential energy supplied by elevation and pore water pressure. This energy potential, termed *hydraulic head*, incorporates the driving forces caused by gravity and differences in pore pressure from point to point in the subsurface. Similar to unsaturated flow, the driving force is based on a gradient (difference in hydraulic head) across a given distance. The amount of water transmitted (flux) depends on the size of the hydraulic head gradient and the properties of the porous medium (i.e., an aquifer or aquitard). In the field, hydraulic head is measured using a piezometer or a water well. As shown in Figure 6.11, hydraulic head is measured at a known point, which means that the exact intake depth of a particular well (or piezometer) must be known. By measuring the depth to water, the hydraulic head—the combination of elevation head and pressure head—can be determined. Spatially distributed measurements of hydraulic head throughout a watershed can be compiled to infer the directions of groundwater flow, as water moves from areas of high hydraulic head to areas of lower hydraulic head (e.g., toward lowland areas [Figure 6.11]).

The relationship between the volume of groundwater flow to the driving force and properties of the porous medium (Darcy's Law) is:

$$Q = K_s A \frac{dh}{dL} \tag{10}$$

where: Q is the volumetric flow rate (cubic metres per second), K_s is the (saturated) hydraulic conductivity (metres per second), A is a representative cross-sectional seepage area (square metres), and dh/dL is the hydraulic head gradient (difference in hydraulic head divided by difference in distance; this term is dimensionless). Hydraulic conductivity (K_s)

is an empirical proportionality constant describing the ease with which water passes through porous media. It is a specific version of permeability (*k*), which is an empirical constant describing the ease with which any fluid (water, oil, etc.) passes through porous media. These empirical constants range over many orders of magnitude, with higher values corresponding to aquifers (i.e., highly permeable) and lower values corresponding to aquitards (i.e., less permeable).

Groundwater flow is greatest with larger gradients of hydraulic head (i.e., steeply sloping water table) and higher hydraulic conductivity values. The geology (surficial and bedrock) of the area of interest strongly affects both gradient and hydraulic conductivity. The pore-size distribution and interconnectivity of pores govern hydraulic conductivity. Larger pores typically conduct more water than smaller pores. Many geologic materials also possess preferential flow pathways, such as macropores or fractures in bedrock, which allow water to be transported at high rates relative to that in the surrounding substrate.

Role of groundwater in watersheds

Groundwater flow systems establish below the water table. These organized systems are comprised of recharge areas (high hydraulic head) that drive water to discharge areas (of low hydraulic head) (Figure 6.11). As a result, stream base flow can originate some distance from a stream. The travel time from recharge to discharge areas may be as short as days or longer than centuries depending on flow system depth and whether the flow path is at a "local" or "regional" scale (Toth 1962). Shallow, local-scale flow systems exhibit seasonal variability in flow rates because they respond to variations in infiltration rates and can be greatly affected in the short term by land use or climate change. Deeper, regional-scale flow systems tend to buffer short-term variability but integrate a multitude of changes over the long term, making deleterious impacts more difficult to reverse. Local and regional scales of groundwater flow systems can create different surface and subsurface watersheds (Winter et al. 2003). A well-defined surface watershed may not correspond to that for groundwater, which is controlled by geology.

Groundwater discharge to streams is commonly expressed as base flow, which is streamflow that occurs during dry times of the year (not caused by specific storm events or seasonal phenomena, such as snowmelt). Figure 6.11 illustrates a typical groundwater flow system and groundwater that supplies a stream. This is an example of a gaining stream, which gains discharging groundwater along its length. Surface water bodies, however, are both sources of and sinks for groundwater. Losing streams are those in which stream discharge decreases downstream because of water losses through the streambed to groundwater. Flow-through streams simultaneously gain and lose groundwater along their length. Flow-through stream reaches typically occur where a stream meanders through a low-gradient area in which the groundwater flow is parallel to the general valley alignment and tends to "short-cut" the meander bends. When the groundwater table lies below the stream channel, the stream is perched and loses flow in the downstream direction. A single stream can have reaches that are gaining, losing, or flow-through. This is determined largely by the physical properties of the underlying sediments (Winter 1999) and the topography of the watershed. Other surface water bodies, such as wetlands or lakes, exhibit the same gaining/losing/flow-through characteristics depending on the groundwater flow system and their position within the landscape. Many wetland types are therefore classified by their interaction with groundwater (Rydin and Jeglum 2006). For example, bogs form in locations where groundwater inflow to the wetland is minimal, whereas fens have groundwater flowing into and through the wetland. The wetland classification system for the province includes the role of groundwater (MacKenzie and Moran 2004).

In addition to the complexity of flow systems described so far, some saturated zones may become perched above a deeper, regional water table. For these isolated zones of saturation, which meet the definition of groundwater, the word "perched" is added to note their disconnection from most of the groundwater regime. Perched conditions often develop above a layer of low permeability (perching layer), which creates a saturated zone in a generally unsaturated subsurface zone (Sidle and Ochiai 2006). Perched water tables may be relatively long term or transient (seasonal or an event), and can be common in environments with high rainfall and shallow soils that lie over a suitable perching layer. The development of these transient perched systems is a common driver of subsurface stormflow (Weiler et al. 2005).

The role of groundwater in a watershed is complex, given that groundwater flow systems may have cycling times that range from days to centuries, and that the interaction with surface water can create gaining, losing, or flow-through systems (Winter et al. 2003). Besides contributing to stream base flow, groundwater adjacent to streams can buffer peak flows if bank sediments are sufficiently permeable. A rise in stream stage may temporarily exceed the level of the adjacent water table and cause gaining stream reaches to become losing streams until the stream stage declines (known as a *groundwater flow reversal*). Also, transpiration by vegetation in a riparian zone may extract local groundwater and cause a diurnal decrease in streamflow followed by recovery at night (Dunford and Fletcher 1947; Hewlett 1982, Bond et al. 2002). Studies conducted in Australia and South Africa demonstrated that riparian vegetation is a more liberal user of water than vegetation in other parts of a watershed (Bren 1997; Scott 1999), and can cause sections of small streams to dry up in periods of low flow. Stream drying may occur frequently in the headmost portions of the channel network, interrupting connectivity. For example, in the central interior of British Columbia, Story et al. (2003) found that dewatering of an intermediate segment of stream channel effectively decoupled a lower reach from the warming effects associated with harvesting and road construction on an upper reach of a small stream. The broader biogeochemical and ecological implications of changing connectivity along the length of streams in forested areas do not appear to have been examined.

Groundwater flows are critical to maintaining aquatic health (Douglas 2006, 2008) since these flows buffer nutrients and temperature fluctuations (Story et al. 2003), especially in riparian and hyporheic zones (Dahm et al. 1998; Hayashi and Rosenberry 2001). In northern climates, where many surface water bodies freeze in winter, groundwater inflows or seepage can maintain open water, thus providing temperature refugia for fish (Power et al. 1999). Groundwater inflows can also help maintain healthy temperatures for overwintering eggs of sockeye salmon (Leman 1993). In summer, groundwater inflows to streams may reduce stream temperatures, which is critical for fish survival.

STREAMFLOW

Streamflow is water flowing in, or discharging from, a natural surface stream. Streamflow is often referred to as *discharge* or *runoff*, which can be confusing. Discharge is the volume of water flowing past a reference point in a stream, canal, pipe, or other structure. In addition to being used synonymously with discharge, runoff is also often used to refer to water flowing downslope along or from any surface. Definitions of these terms are provided in Appendix 1 ("Glossary of Hydrologic and Geomorphic Terms").

Streamflow regimes, including the timing and duration of high and low flows, vary depending on the predominant form of precipitation (rain, snow, or mixed rain and snow), on whether streamflow is augmented by glacial melt, and on how precipitation is stored and released within the watershed. The hydrologic regimes of British Columbia are detailed in Chapter 4 ("Regional Hydrology"). If water in a stream flows throughout the year, then the stream is referred to as *perennial*. The total length of perennial streams per unit watershed area in a watershed, referred to as the *drainage density* (km/km^2), provides an indication of how quickly streamflow responds to rainfall or snowmelt in one watershed relative to another. Streams with flow only in response to storm events and no flow during dry seasons or in particularly dry years are referred to as *ephemeral*. Streams in which flow occurs only seasonally in direct response to precipitation or to flow from intermittent springs are referred to as *intermittent* (UNESCO and World Meteorological Association, International Hydrology Programme 1998). The stream network and adjacent contributing areas expand and contract in response to rainfall and snowmelt depending on antecedent conditions and on the amount and duration of rain or snow input (Figure 6.12). These contributing areas are often referred to as the *variable source area* (Hewlett 1982). Streamflow reflects the cumulative effects of the biophysical environment on the balance of precipitation inputs to a watershed, plus losses through evaporation and transpiration, and storage within the watershed.

Streamflow is expressed by the continuity equation as the volume of water passing a given channel cross-section area per unit time:

$$Q = wdv \tag{11}$$

where: Q is streamflow, discharge, or the volumetric flow rate (cubic metres per second); w is the channel width (metres); d is mean depth (metres); and

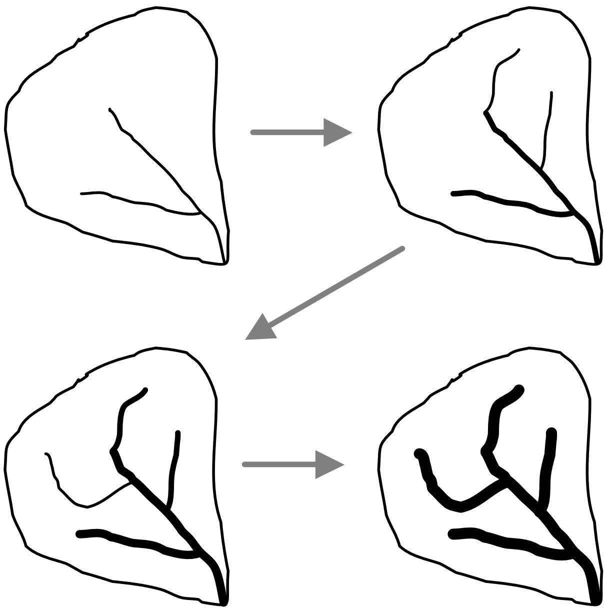

FIGURE 6.12 *Hypothetical example of expanding variable source areas within a headwater watershed during a runoff event. Thicker lines indicate greater runoff contribution. (Adapted from Hewlett 1982)*

v is mean stream velocity (metres per second). The equation above is a simplified version of the cross-section equation applied by most hydrometric monitoring agencies. (For a more detailed explanation of discharge measurements, see Chapter 17, "Watershed Measurement Methods and Data Limitations." Appendix 2, "Acronyms, Initialisms, Symbols, and Conversion Factors," contains details on calculating unit discharge.) Since total streamflow at any point along a channel is affected by the size of the contributing area, it is often also useful to express flow as a depth equivalent over the watershed (millimetres per unit of time). This metric, known as the unit area discharge, is useful when comparing streamflow to precipitation and determining water budgets (expressed in millimetres), and when comparing water yield between watersheds or geographic areas.

At most gauging stations, streamflow is recorded at intervals as frequent as 15 minutes. (For a more detailed explanation of discharge measurements, see Chapter 17, "Watershed Measurement Methods and Data Limitations.") The 24-hour average of these values is published as mean daily discharge. In Canada, a national network of streamflow gauging stations is maintained by the Water Survey of Canada through agreements between Environment Canada, Indian and Northern Affairs Canada, and the provinces. Streamflow data, both archived long-term and real-time, are available electronically through the Environment Canada website (www.wateroffice.ec.gc.ca/index_e.html). In British Columbia, 451 stations were active in May 2010. Of these, 313 recorded natural (unregulated) flows from watersheds without diversions, storage, or water extraction. These gauging stations are located in watersheds ranging in size from less than 3 km^2 (Jamieson Creek at the mouth) to 104 000 km^2 (Liard River at lower crossing). Several stations, such as those located in the Lillooet, South Thompson, and Slocan River watersheds, provide more than 80 years of record. The most common variables used to describe streamflow are seasonal or annual water yield, peak flow, low flow, frequency at which streamflows of a certain magnitude occur, and duration of streamflows of a certain magnitude.

The Hydrograph

Streamflow variations over any time period can be represented by a hydrograph (Figure 6.13), which shows peak flows, low flows, and water yield over the course of a year, plus individual storm events or flows over a day, several days, or a season. The hydrograph can also show how streamflow responds to rainfall or snowmelt (Figure 6.13). Hydrographs typical of streams across British Columbia were described in Chapter 4, "Regional Hydrology." Note that these hydrological classifications are somewhat arbitrary and depend on the conceptual model of streamflow generation assumed by the analyst.

Streamflow begins to increase some time after the onset of a rainstorm or snowmelt. This increase is shown as the rising limb of the hydrograph (Figure 6.14). The time from the centroid (or centremost point in time) of a rainstorm to the peak in streamflow generated by that storm is referred to as *basin lag*. The time required for water to flow from the farthest point on the watershed to a point of interest, such as a gauging station or culvert, is referred to as the *time of concentration*. The maximum streamflow or discharge in response to a storm or snowmelt is referred to as *peak flow*. The definition of peak flow depends on the time period of interest (e.g., event, seasonal, annual). The magnitude of a peak flow is a function of rainstorm duration or snowmelt volume; storm intensity or melt rate; antecedent conditions affecting storage opportunities, such as soil moisture, soil characteristics, forest floor condi-

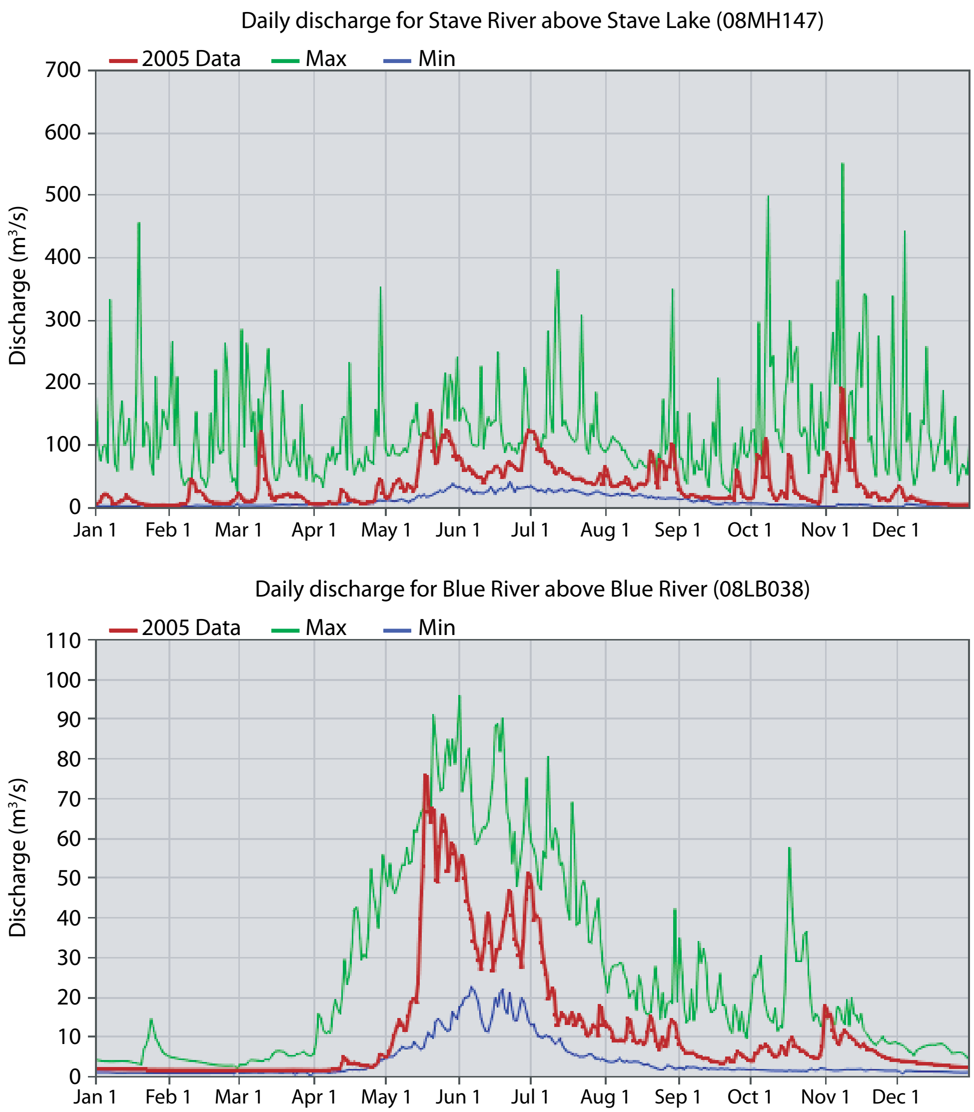

FIGURE 6.13 *Typical hydrographs for coastal (Stave River above Stave Lake [282 km^2]) and interior (Blue River near Blue River [280 km^2]) watersheds in British Columbia. The hydrographs show the 2005 daily discharge rates, and the maximum and minimum daily flows for Stave River (1983–2005) and Blue River (1926–2006), respectively. (Source: Water Survey of Canada, www.wateroffice.ec.gc.ca/index_e.html)*

tions, snowpack conditions, and canopy wetness; and watershed characteristics, including drainage area, topography, and physiography. In watersheds with large floodplains, streamflow may be stored or detained in the floodplain alluvium. From there, it is later released back into the channel as base flow, thereby reducing peak flow (Ritter et al. 2002). Similarly, lakes, reservoirs, and wetlands in a watershed may also reduce the magnitude of peak flows.

After a storm or snowmelt, contributions to the channel decrease, as shown by the recession (falling) limb on the hydrograph (Figure 6.14). The shape of the recession limb is influenced by the physical characteristics of the watershed. Once rain- and snowmelt-generated surface and subsurface contributions to the channel are depleted, the remaining streamflow is referred to as *base flow*. It is the proportion of the flow in a stream that is contributed by groundwater, or by flow from lakes, wetlands, or glaciers during prolonged periods without precipitation. Base flow also forms part of the hydrograph peak, but is small relative to the total streamflow during a rainstorm or snowmelt.

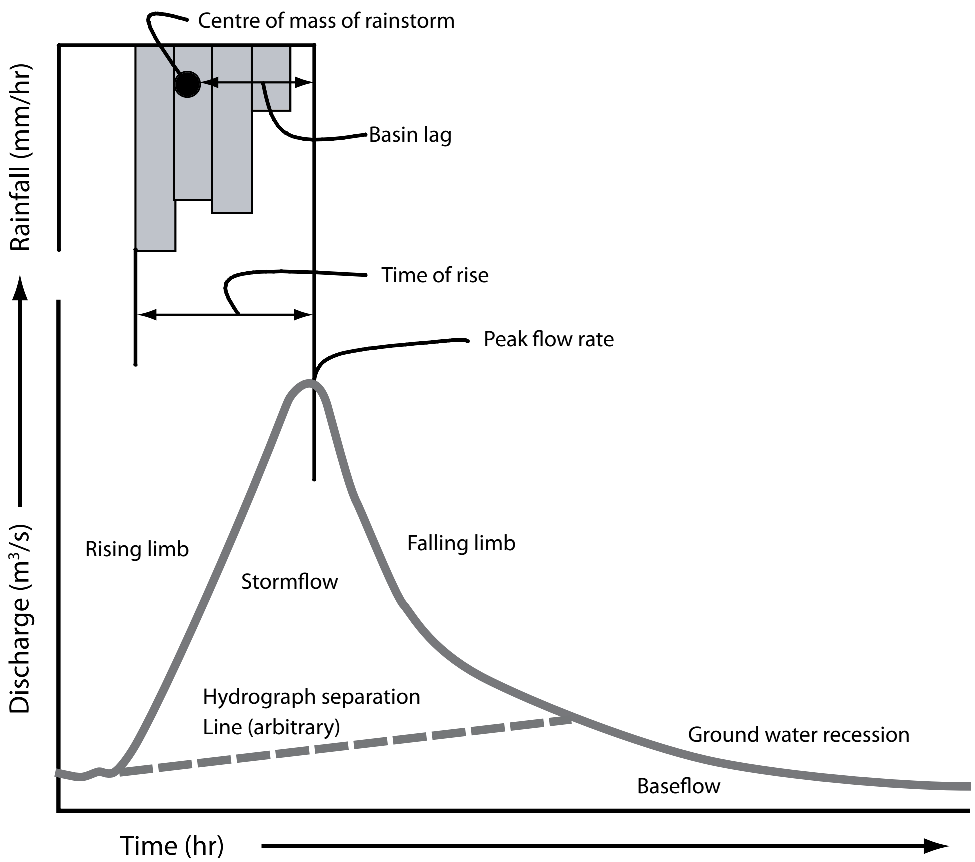

FIGURE 6.14 *Components of the hydrograph. (Adapted from Hewlett 1982)*

In many snow-dominated watersheds, streamflows are typically low during late summer or early fall through the winter until spring freshet. Low flows are the minimum flow or absence of flow in a stream, and are often characterized by the lowest average flow over a defined time interval (i.e., 7-day period each year) (Pike and Scherer 2004). These flows are referred to as *low flow* and are maintained by base flow and occasional rainfall. The sources of base flow in a watershed can be thought of as storage reservoirs. The amount of precipitation ultimately defines the amount of recharge that occurs in these storage reservoirs, whereas the release of water is more a function of physiographic characteristics, such as climate, topography, soils, and geology (Smakhtin 2001). The natural variability of low flows is caused by the complex interaction and heterogeneity of watershed characteristics. During the dry season, many natural processes can affect low flows, including "the distribution and infiltration characteristics of the soils, the hydraulic characteristics and extent of aquifers, the rate, frequency and amount of recharge by periodic rainstorms, the evapotranspiration rates from the basin, distribution of vegetation types, topography and climate" (Smakhtin 2001:149).

Annual water yield is the total streamflow (measured in millimetres, cubic metres, or cubic decametres [1 dam^3 = 1000 m^3]) from a watershed over a water year. A water year can be defined as January through December, or any other 12-month period that best captures the precipitation that generates the described streamflow. In snowmelt-dominated regimes, the water year is often set at October 1 through September 30. The streamflow generated during this period is a result of the total precipitation that occurs during this time, including snow that accumulates over the winter. The advantage of using the water year in many parts of the province is that it begins and ends at the time of year when flows and watershed storage are at their lowest, which allows for greater precision in water balance calculations. In British Columbia, annual water yield varies from 50 mm to more than 3000 mm depending on location. Examples of annual water yields from streams

TABLE 6.2 *Total annual streamflow volume and water yield per unit area from selected research watersheds throughout British Columbia*

General location	Watershed	Area (km^2)	Years of record	Annual streamflow volume (dam^3)[a]	Annual water yield per unit area(mm)
Coast	Jamieson Creek	3	2002–2004	8 750	2917
	Carnation Creek	10	1973–2006	25 700	2570
	Tsitika River	365	1975–2006	709 000	1942
Southern Interior	240 Creek	5	1984–2006	1 880	376
	Redfish Creek	26	1968 ; 2006	27 200	1046
	Fishtrap Creek	135	1972–2002 ; 2005–2006	24 900	184
Northern Interior	Chuchinka Creek	310	1976–2006	157 000	506
	Baker Creek	1550	1964–2006	151 000	97
	Bowron River	3420	1977–2006	2 030 000	594

a dam³: a cubic decametre equals 1000 m³.

throughout the province are provided in Table 6.2.

Hydrograph shape is affected by both climate and the biophysical characteristics of the watershed, with climatic factors dominating the rising limb and watershed characteristics the recession limb. Key climatic factors that influence the hydrograph include rainfall intensity, duration and distribution, storm direction and type, and form of precipitation (Gray 1970). Rainfall intensity and duration determine the volume and duration of stormflow. For a given duration, increased rainfall intensity increases both peak flow and stormflow volume once soil storage and (or) infiltration capacity has been exceeded. For a given intensity of storm, its duration will affect the duration of stormflow and, in part, peak flow. These effects are most noticeable in small watersheds but are rarely measurable in large watersheds. The areal distribution of rainfall in a watershed will affect hydrograph shape. A storm near the outlet of a watershed will result in a sharp-peaked hydrograph, whereas if the storm occurs in the upper reaches, the hydrograph will generally be broader and have a lower peak. Storms that move down a valley tend to produce more rapid and higher streamflow peaks than storms moving up a watershed. Hydrographs generated by melting snow tend to be lower and broader than those generated by rain. Thunderstorms produce the sharpest peaks in small watersheds, whereas large frontal storms produce the most significant peaks in large watersheds (Gray 1970). How quickly streamflow responds to precipitation or snowmelt is often referred to as *flashiness* or *hydrologic response*.

The topographic and physiographic characteristics of a watershed influence the amount and rate at which excess rainfall and snowmelt are delivered to the channel. These characteristics include: watershed size and shape; drainage density and distribution; slope of the watershed and channel; and presence of natural storage reservoirs, such as lakes, ponds, and wetlands (Gray 1970). Generally, the smaller and steeper the watershed, the greater the drainage density, and the more uniform the distribution of channels, the steeper and narrower the hydrograph. When there is greater opportunity for detention and storage of flow in a watershed, the hydrograph becomes broader. Each factor influencing the hydrograph may be compounded or obscured by the effect of another; therefore, the hydrograph for individual watersheds will depend on the cumulative effect of all factors specific to the watershed (Gray 1970).

Extreme Events

Streamflow records covering a sufficient time frame to describe the frequency and duration of specific flow conditions provide an understanding of extreme streamflow conditions such as flooding and low flow, which are of central importance to the physical and ecological integrity of streams in forested watersheds (Poff et al. 1997). Streamflow frequency and duration are described only briefly here. Thorough discussions of flow regime analysis are provided in engineering and flood forecasting texts, such as Maidment (editor, 1993) and references therein.

Streamflow frequency (return interval) refers to how often a flow above a given magnitude (the amount of water moving past a fixed location per unit time) recurs over some specified interval (e.g., a 10-year flood is equalled or exceeded once every 10 years or has a 10% chance of occurrence in a given year). Flow duration is the period of time associated with a specified flow condition (e.g., a floodplain may be inundated for a specific number of days by a 10-year flood).

One way of quantifying the frequency and duration of flows over time is through the use of flow duration curves. A flow duration curve represents the relation between the magnitude and frequency of daily, weekly, monthly, (or some other time interval) streamflow for a particular watershed. A flow duration curve, such as that shown for Fishtrap Creek near Kamloops (Figure 6.15), provides an estimate of the percentage of time a given streamflow was equalled or exceeded over a historical period.

Flow duration curves have been used to evaluate the effects of forest harvesting and other land use disturbances on streamflows (e.g., Troendle 1970; Fallas 1982; Burt and Swank 1992; Austin 1999). Flow duration curve analysis evaluates the effects of forest harvesting on the magnitude and frequency of flows simultaneously; therefore, it accounts for the highly non-linear relation between the two attributes—a signature of the streamflow regime of a watershed. As is the case for any other statistical tool, its usefulness hinges heavily on the length of pre- and post-treatment flow records. Differences in flow duration curves before and after disturbance will also be influenced by climatic variability, which complicates their use for interpreting the influence of forest disturbance.

The term "flood" has different meanings to different professionals, scientists, and the public (Eisenbies et al. 2007). For instance, to a geomorphologist, a flood is an event that moves bedload and scours streambanks. To an ecologist, a flood is an event that inundates the banks and affects habitat. To an engineer, a flood is an event that washes out a stream crossing (bridge or culvert). To the public, a flood is an event that causes structural damage to property, and at times, the loss of lives. The scientific literature and regulatory setting are consistent in emphasizing the importance of floods that have recurrence intervals of at least 1 year, and most often much greater.

The magnitude and frequency of floods is characterized through a frequency analysis of a partial duration (peak-over-threshold) or annual maxima time series (Figure 6.16). The partial-duration approach is often intuitively attractive as it can focus on peak flows that exceed some relevant threshold (e.g., flow at which bedload movement begins or above which

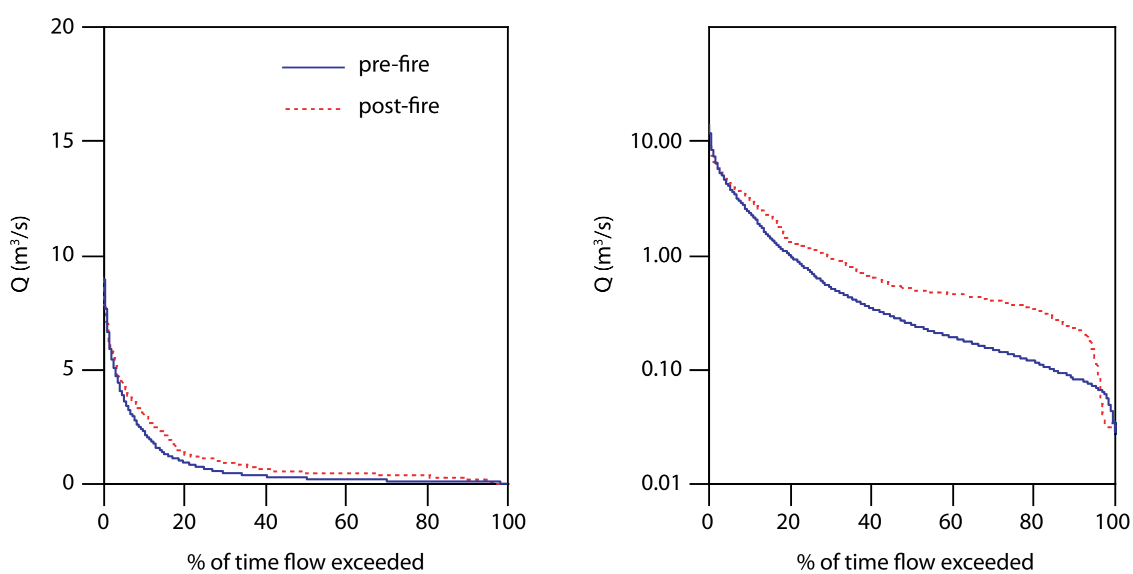

FIGURE 6.15 *Flow duration curves for Fishtrap Creek, with arithmetic scale (left) and logarithmic scale (right). Separate curves are shown for pre-fire (1972–2002) and post-fire (2004–2006) years. Note that the differences between the pre- and post-fire curves reflect the effects of climatic variability and the effects of the fire and post-fire salvage harvesting (R.D. Moore, unpublished data).*

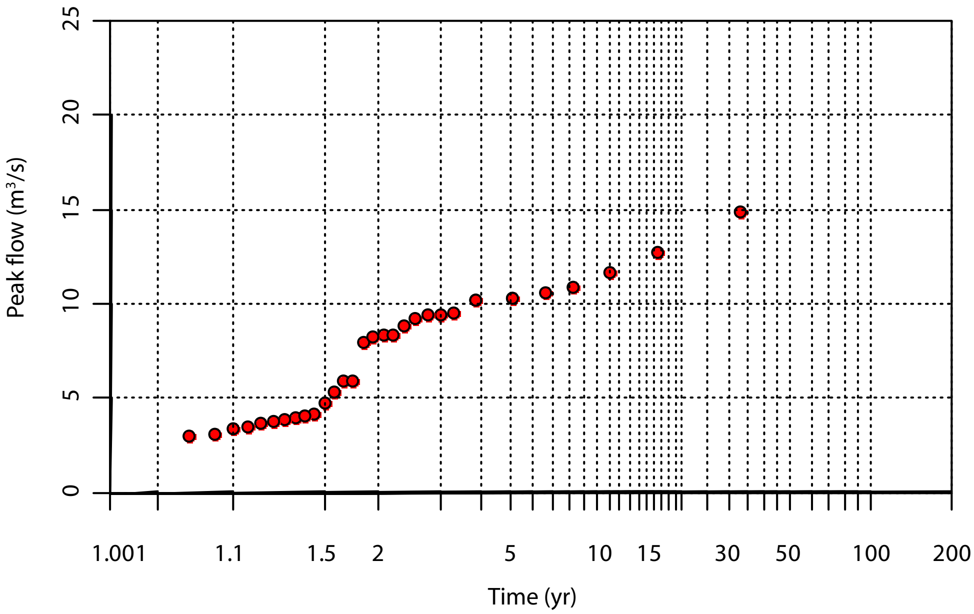

FIGURE 6.16 *Annual flood frequency plot for Fishtrap Creek for pre-fire years (R.D. Moore, unpublished data).*

the stream overflows its banks); however, the theoretical basis for interpreting partial-duration series is weaker than that for analyzing annual extreme series.

At a gauged watershed with a long continuous record of flows, a time series of annual extremes is fitted to a theoretical frequency distribution by the method of moments, L-moments, or maximum likelihood. The fitted distribution is then used to estimate the magnitude of extremes for specific return periods of interest (Coulson 1991). Three major sources of error are associated with such at-site frequency analyses.

The first source of error is caused by the misidentification of the underlying frequency distribution, the magnitude of which depends on how much the assumed model differs from the actual parent distribution. For the relatively short record length of streamflow data that is typically available, several distributions with extremely different upper tail characteristics often give an equally good fit in the range of observed floods. This can have major implications for the predictions of extreme flood events.

The second source of error in at-site frequency analysis is caused by sampling deficiencies, where the length of record (i.e., sample size) is not long enough for a reliable statistical analysis (especially for long return periods). This problem is particularly relevant in British Columbia, where a relatively small number of hydrometric stations have more than 30 years of record.

The third source of error is caused by lack of stationarity. Conventional flood frequency analysis is based on the fundamental assumption that the time series of observed floods is stationary, and as such, has not been affected by either climate or land use. Newly emerging flood frequency analysis techniques that are not constrained by such assumptions should be used to accommodate time series that may have been affected by climate change or forest regrowth (El Adlouni et al. 2007).

For a watershed with short or non-existent flow records, various regional flood frequency analysis techniques can be used to estimate the magnitude and frequency of floods. Regional models are based on the premise that if a set of gauged watersheds can be used to identify a hydrologically "homogeneous" region, flood frequency characteristics could be transferred from the gauged to ungauged basins within such a region. Pooling of data from a set of nearby gauged watersheds within a homogeneous region is equivalent to increasing the flow record length; therefore, it provides a more reliable way of identifying a regional flood frequency distribution and estimating its higher-order parameters. Several regional flood frequency models are available for British Columbia (e.g., Obedkoff 1998; Eaton et al. 2002). Model development was based on a sparse network of gauged watersheds, particularly in northern British Columbia; therefore, these models are appropriate only for generating first-order approximations of the magnitude and frequency of floods at ungauged watersheds. In addition, the historical hydrometric database in the province consists largely of medium and large basins (i.e., drainage area > 100 km^2), and so all British Columbia regional models are limited in application to basins within this size range. Few data are available to indicate how well these regional models can be extrapolated to small forested watersheds of less than 100 km^2. Successful application of these models requires considerable judgement, experience, and understanding of hydrologic and hydro-climatic principles, particularly when estimating floods for small ungauged watersheds.

Low-flow frequency analysis is used to evaluate the ability of a stream to meet specified flow requirements. Such analysis can provide an indication of the adequacy of the flow to meet a given demand with a stated probability of experiencing a shortage. Low-flow frequency analysis is used for applications such as designing hydroelectric power plants, determining minimum flow requirements for fish or wildlife, and designing water storage projects. Annual low flows are usually computed for several durations (in days), with flow rates expressed as the mean flow for the period. The number of days over which the average low flow is calculated depends on the operational problem at hand. The estimation of low flow magnitude and frequency follows the same procedures and is subject to the same limitations as that for floods, even though the governing physical processes are different. In the case of low flows, the return period is the average length of time between events in which the flow index is as low as or lower than a specific value. Site and regional low-flow analysis models are based on a set of different frequency distributions developed for time series of minima.

Hydrologic Response

How quickly streamflow increases at the beginning of a rainfall event or as snow begins to melt is referred to as *hydrologic response*. This depends on the form of precipitation, antecedent conditions, watershed characteristics, and the flow paths along which water is delivered through the watershed to the stream channel. For example, during an intense, wet-season rainstorm and at the end of the snowmelt season, hydrologic response may be almost immediate. Conversely, hydrologic response may be slow, lagging precipitation events by months during the winter in snow-dominated environments or after periods of drought. When a precipitation event occurs over an entire watershed, and the antecedent conditions plus rates of flow through the area are similar, contributions to streamflow from all portions of the watershed may be synchronized. This generally occurs in small watersheds and results in a "flashy" (rapid) response to storm events. In a large watershed, greater variability is likely to occur across the landscape, resulting in desynchronization of flow to the outlet and a more moderated storm response.

Water that contributes to streamflow arrives at the channel along one or more of four generalized flow paths: (1) direct precipitation onto the stream channel, (2) overland flow, (3) shallow subsurface flow or throughflow, and (4) deep subsurface flow or groundwater.

Rain intercepted directly by the channel becomes streamflow immediately. This is usually a minor contribution to the total flow in a stream unless the surface area of the stream is large relative to the area of its watershed (> 1%) (Satterlund and Adams 1992). Stream surface area varies with time of year and storm size. It is small to intermittent during dry periods and larger towards the end of a significant storm, during the wet season, and during snowmelt when the stream network connects to intermittent and ephemeral channels.

Precipitation that reaches the watershed area outside a stream channel can be stored at the surface, or can infiltrate into the soil or flow over an impermeable surface. Overland flow does not occur unless the precipitation rate exceeds the infiltration rate, an impermeable layer (e.g., road or hydrophobic layer) is present, or all surface storage and infiltration capacities have been filled. Where these conditions occur and are continuous down the slope to the stream channel, overland flow contributes to streamflow relatively quickly after the onset of rain or snowmelt. Overland flow is most common in areas of rock outcrop, along roads, in areas of shallow and (or) waterlogged soil, on soils with reduced surface permeability caused by logging, grazing, or fire, and over densely frozen soil. The rate at which overland flow reaches a stream channel depends on the distance to the channel, vegetation density, surface roughness, and soil permeability downslope (Satterlund and Adams 1992). Overland flow in undisturbed forest soils with a well-developed litter layer is not common, especially on upper slopes that have a high infiltration capacity and greater soil moisture storage capacity (i.e., soil water deficits are common).

Water that has infiltrated into the soil flows towards channels, adding to streamflow more slowly than water contributed at the surface. The most rapid delivery of water along subsurface flow paths occurs near the surface in areas where surficial materials are highly porous or where interconnected macropores or preferential flow pathways occur through the soil. This is also referred to as *interflow*. Subsurface flows on lower slopes and in riparian areas can also be rapid because of less available storage (i.e., higher antecedent soil moisture storage) and proximity to stream channels.

Channel interception, surface runoff, and interflow are often referred to together as *quickflow* or *stormflow*. Deep subsurface flow or base flow, which is also referred to as *delayed runoff*, is water that drains from the landscape slowly and sustains streamflow during dry periods. This may include contributions from groundwater, saturated soil, and unsaturated soil. The soil moisture or groundwater contributions to base flow are not always replaced during every storm or during snowmelt, as is the case with water moving along other flow paths.

Water can move from one flow pathway to another on its way downslope to a stream channel. Therefore, small headwater streams, where flow pathways are short and storms deliver rainfall uniformly across the watershed, are most responsive to rainfall events. These streams are also most affected by changes in flow pathways caused by disturbance, even though the proportion of precipitation that appears as streamflow differs from storm to storm. In larger watersheds, lag effects and differences in flow contributions from tributary streams complicate hydrologic response to precipitation, providing opportunities for desynchronization of flows from different parts of the watershed and potentially moderating hydrologic response and snowmelt-generated peak flows (Ward and Robinson 1989; Satterlund and Adams 1992).

SUMMARY

Hydrologic processes, including precipitation, interception, evaporation, and subsurface storage and flow, all affect water delivery to, storage in, and flow from forested watersheds. Each process, as affected by the weather, influences the quantity, timing, and variability of streamflow events. These events, in turn, affect slope and stream channel integrity, stream ecology, water supplies, infrastructure, and the public. The consequences of forest disturbance, through natural events or human activity, on hydrologic processes and streamflow are described in the next chapter ("The Effects of Forest Disturbance on Hydrologic Processes and Watershed Response").

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The Effects of Forest Disturbance on Hydrologic Processes and Watershed Response

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INTRODUCTION

Hydrologic processes that affect the generation of streamflow were described in the previous chapter ("Hydrologic Processes and Watershed Response"). This chapter builds on the understanding of these fundamental processes by describing how changes in forest cover, brought about by logging, insects, disease, fire, or forest regrowth, affect these processes at both the stand and watershed scale. The discussion in this chapter focusses on the water balance, groundwater, water yield, peak and low streamflows, and hydrologic recovery.

STAND- AND HILLSLOPE-SCALE EFFECTS

Surface Processes

Net precipitation

Forest cover significantly influences the amount of precipitation that reaches the ground, as described in Chapter 6 ("Hydrologic Processes and Watershed Response"). The loss of forest cover results in an increase in net precipitation (i.e., gross precipitation minus the amount intercepted and subsequently evaporated or sublimated). The amount of net precipitation depends on stand characteristics (tree species, leaf area, canopy density) and the weather (time since the last rainstorm, existing snow load, air temperature, wind speed, and the size of storm events). Increases in net precipitation may occur immediately after forest disturbance, such as following clearcutting or severe fires, or may occur over a period of years when foliage loss is gradual, as can occur following attack by insects such as the mountain pine beetle.

In British Columbia and similar forested environments, 5–70% more water can accumulate as snow in clearcuts than in the forest, depending on the winter precipitation in a given year and forest cover type (Toews and Gluns 1986; Hudson 2000; Winkler 2001; Pomeroy et al. 2002; Winkler and Moore 2006). At Mayson Lake and Upper Penticton Creek in south-central British Columbia, 47% and 29% higher snow water equivalent (SWE) was measured in a clearcut than in a mature, mixed-species stand, respectively. In the clearcut at Upper Penticton Creek, SWE was only 12% higher in the open than in a mature lodgepole pine stand (Winkler 2001). At these same locations, the largest relative differences in SWE between mature forest and open snow survey sites did not occur in the year of heaviest snowfall, likely because of an upper limit to the interception capacities of these stands and to annual variations in snowfall pattern (few large snowfall events vs. many small events) (Winkler and Moore 2006).

Studies in Colorado indicated that most of the increased snow accumulation in openings occurs during storms, and that the redistribution of intercepted snow by wind between storms is not significant (Wheeler 1987). Comparisons of snow accumulation under forest cover and in clearings, at both the plot and watershed scales, indicated that the increased accumulation in openings was due mainly to a decrease in interception loss rather than preferential deposition by wind patterns (Troendle and Meiman 1984, 1986; Troendle and King 1985; Wheeler 1987). In Alberta, maximum increases in snow accumulation were measured in openings 2–5 tree heights wide (Golding and Swanson 1986).

Losses to evaporation, sublimation, or early melt can be important influences on peak SWE within cutblocks. For example, although more snow accumulated in cutblocks than in the forest on the North Fork of Deadhorse Creek, Colorado, peak SWE averaged over the watershed did not increase significantly following harvesting (Troendle and King 1987). The researchers hypothesized that early ablation in south-facing openings offset the increase in snow accumulation that occurred from reduced interception loss. This effect was also observed in a high-elevation, south-facing clearcut at Upper Penticton Creek, where at the start of the spring melt, less snow remained in the clearcut than in the adjacent forest (R.D. Winkler, unpublished data).

Tree mortality caused by insects or disease can result in reduced interception losses and consequent increases in net precipitation, though not as large as those associated with harvesting. Research into the effects of beetle-killed stands on snow accumulation is under way throughout the interior of British Columbia (Winkler and Boon 2010). Initial results of this research show that averaged over all research sites, maximum SWE is reduced by 24% under mature, green-to-red attacked lodgepole pine relative to the open, and by 13% under grey pine (Table 7.1); however, the wide range in results for both stand conditions highlights the large variability in interception among stands, sites, and years.

Reductions in snow interception and increases in net precipitation are generally proportional to reductions in canopy cover or to percent basal area removal (Harestad and Bunnell 1982; Moore and McCaughey 1997). Maximum snow accumulation, measured as SWE, increased by 16% following 40% basal area removal in lodgepole pine stands in Montana (Woods et al. 2006), and by 21% following 50% removal in similar stands in Colorado (Troendle and King 1987).

Net precipitation may also increase after a forest fire. If only a few small openings are created, the effects on net precipitation may be similar to those following thinning or patch cutting; however, in cases where most of the forest cover is lost, changes in net precipitation can be expected to be similar to those following clearcutting (Neary and Ffolliott 2005). Few studies have quantified the effects of fire on snow accumulation. In a lodgepole pine stand in southern Montana, where fire reduced canopy cover by 90% (from 42 to 4% cover), SWE increased by 9% relative to the adjacent similarly structured unburned forest, the same difference as was observed in the clearcut (Skidmore et al. 1994). Although this increase in SWE was small, the authors suggested

TABLE 7.1 *Percent reduction in maximum snow water equivalent (MSWE) and average ablation rate (AAR) in the forest relative to the open and the number of days difference in timing of snow disappearance in stands affected by mountain pine beetle in the British Columbia Interior (Winkler and Boon 2010, based on data collated in Winkler and Boon 2009)*

		Reduction in MSWE (%)			Reduction in AAR (%)			Difference (days) in snow depletion date (Forest–Open)	
Forest age class (years)	Attack class	No. sites	Average	Range	No. sites	Average	Range	Average	Range
Old (120+)	Green/red	3	22	31 to 6	2	39	42 to 36	7	4 to 10
Mature (40–120)	Green/red	12	26	57 to +9	3	38	48 to 27	2	0 to 3
Intermediate (10–40)	Green/red	12	16	72 to +7	5	22	49 to +7	2	–5 to 9
Old (120+)	Grey	7	11	21 to +9	2	22	29 to 14	9	–1 to 3
Mature (40–120)	Grey	9	16	58 to +28	3	37	57 to 25	3	0 to 6
Intermediate (10–40)	Grey	1	21		1	38		12	12

that larger increases should be expected where stand densities are higher. At Border Lake, south of Keremeos, SWE was 4% higher in a burned stand than in a nearby clearcut (Dobson 2008). At Mayson Lake, near Kamloops, maximum SWE was reduced by 4–11% in a burned stand relative to a clearcut (R.D. Winkler, unpublished data).

Changes in rainfall interception follow a similar pattern to that of snow interception, generally decreasing with increasing loss of forest cover; however, the magnitude of change also depends on the type of forest cover loss. For example, interception in stands where trees are killed by insects and disease may not change until the trees start to lose their needles (Héile et al. 2005; Spittlehouse 2007; Winkler et al. 2008a), whereas disturbance by fire immediately removes much of a stand's rainfall-intercepting capacity (Moore et al. 2008). After harvest, some understorey remains, although interception losses from shrubs, herbs, mosses, slash, and large woody debris tend to be smaller than those from forest canopies (Kelliher et al. 1992). At Mayson Lake, Carlyle-Moses (2007) found no significant difference in rainfall interception loss during the growing season between an unlogged mature pine-spruce-fir stand (20–25 m tall; basal area = 52 m^2/ha) and a portion of the stand where beetle-attacked trees had been removed (remaining basal area = 19 m^2/ha). Similar results have been observed in other forests. For example, Knoche (2005) found that in a 66-year-old Scots pine stand in Germany, rainfall interception losses declined by only 7% (from 38 to 31%) when basal area was reduced by 47% (from 38 to 20 m^2/ha). Thinning of an 11-year-old radiata pine stand in New Zealand reduced rainfall interception by 27% (Whitehead and Kelliher 1991). Both rain and snow interception losses increase as the forest regenerates (see Chapter 6, Table 6.1; see also "Watershed-scale Effects" section below).

Evaporation

Changes in the amount and (or) type of vegetation caused by fires, insects, disease, harvesting, and silvicultural treatments can alter evaporation from the forest (i.e., evaporation or sublimation of intercepted precipitation, transpiration from the vegetation, and evaporation from the soil surface) (Bonan 2008). Understanding how changes in forest cover affect evaporation is critical to comprehend the effects of disturbance on water yield, and on peak and low flows.

Evaporation rates from a wet, bare, exposed soil surface, which often occurs immediately after clearcutting or a fire, may be as high or higher than evaporation rates from a forest (3–4 mm/d) (Novak and Black 1982; Spittlehouse 1989); however, bare soil surfaces usually dry within 1–2 days after a rainfall, and evaporation decreases rapidly (Novak and Black 1982). This results in a dry 0.05–0.10 m surface layer and a moist lower soil profile; consequently, soil moisture stays relatively high throughout the summer compared to the forest (Figure 7.1). Average evaporation rates depend on the frequency of rainfall and on weather conditions. At high-elevation sites in the province's southern interior, bare soil evaporation during intermittent, short, wet and dry periods averaged 1–2 mm/d, similar to that of forests, but rapidly decreased to less than 0.5 mm/d during extended dry periods (Figure 7.2).

Silvicultural practices such as partial retention and thinning remove only part of the forest canopy. If water is not a limiting factor for the site, the relationship between partial vegetation removal and decreases in evaporation is generally considered as linear (Hibbert 1983). When transpiration is water-limited, however, the relationship is not linear. In some situations, where trees are able to use additional water, a threshold level of forest cover must be removed before a change in evaporation loss can be detected (Tang et al. 2003; Bladon et al. 2006; Simonin et al. 2006; Li et al. 2007; Stednick 2008). Canopy openings also allow for increased evaporation from the soil and increased transpiration from understorey vegetation, which compensates somewhat for the reduction in stand-level transpiration caused by the removal of trees (Black 1979; Spittlehouse and Black 1982; Knoche 2005).

Soil moisture levels increase following harvesting (Ziemer 1964; Hart and Lomas 1979; Adams et al. 1991; Elliott et al. 1998; Bhatti et al. 2000). By using a physically based stand water balance model, Spittlehouse (2006) found that summer evaporation from a high-elevation clearcut in the province's southern interior was about 30% less than that from the forest. This lower evaporation rate combined with a reduction in interception losses increased soil moisture in the clearcut and increased the water available for streamflow (Figure 7.3). Similar results were obtained in a clearcut and Douglas-fir stand in coastal British Columbia (Jassal et al. 2009). Liu et al. (2005) found a similar reduction in seasonal and daily evaporation at burned sites in Alaska.

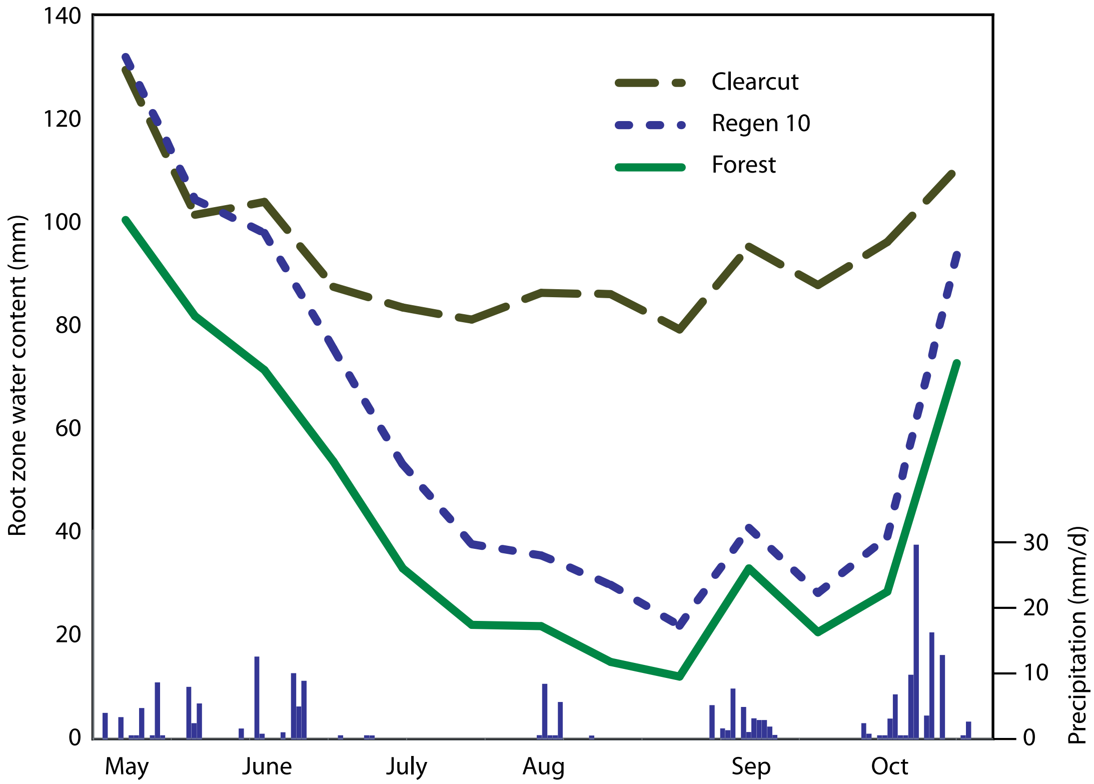


FIGURE 7.1 *Measured root zone soil water storage (mm) and daily precipitation (mm/d) for a lodgepole pine forest (solid line), a clearcut (long dashed line), and a regenerating 10-year-old lodgepole pine stand (short dashed line) at Upper Penticton Creek. Differences in water content between sites in May are the result of differences in water storage capacity (depth and stone content) of the three sites (D.L. Spittlehouse, unpublished data).*

In stands killed by mountain pine beetle, evaporation from the shaded understorey and the forest floor is low, and soils remain moist for most of the summer (Spittlehouse 2007). Numerical modelling indicates that if tree mortality is less than 40%, the water balance (evaporation, soil moisture storage, and drainage) in an attacked stand is similar to an unattacked stand; however, if most of the trees are killed, the stand may have a water balance similar to that of a clearcut (Spittlehouse 2007; Moore et al. 2008).

Changes in vegetation with succession and reforestation may also influence evaporation. Grasses and deciduous plants tend to have a lower canopy resistance to transpiration than do conifers (Kelliher et al. 1993; Bonan 2008) (see Chapter 6, "Hydrologic Processes and Watershed Response"); thus, these plants have a higher transpiration rate than conifers under the same weather conditions. In Australia, Putuhena and Cordery (2000) studied the hydrological effects of clearcutting a watershed covered by a native, dry, sclerophyll eucalypt forest and replacing it with radiata pine. The study results showed that evaporation decreased after clearcutting and then increased with increasing pine age (especially between the 4th and 16th year of growth). The equilibrium evaporation rate reached after approximately 16 years was lower than the rate recorded in the pre-disturbance eucalypt forest. These results are consistent with data for the province's southern interior (Figures 7.2 and 7.3).

Forest practices that alter or remove vegetation affect the amount of evaporation by changing interception losses, transpiration, and direct evaporation from the mineral soil. Decreased evaporative losses result in increased water storage (e.g., soil moisture, groundwater) within watersheds. Higher antecedent soil moisture increases the potential for greater water yield and for more rapid and higher peak flows (Bosch and Hewlett 1982; Iida et al. 2005). The magnitude of such increases varies from small to large depending on available watershed storage capacity relative to the volume of increased water input. In addition, the total area and spatial distribution of disturbance are important to consider because the areas of a watershed contributing to streamflow at any given time will vary in both space and time (i.e., variable source area) (Hewlett and Hibbert 1967).

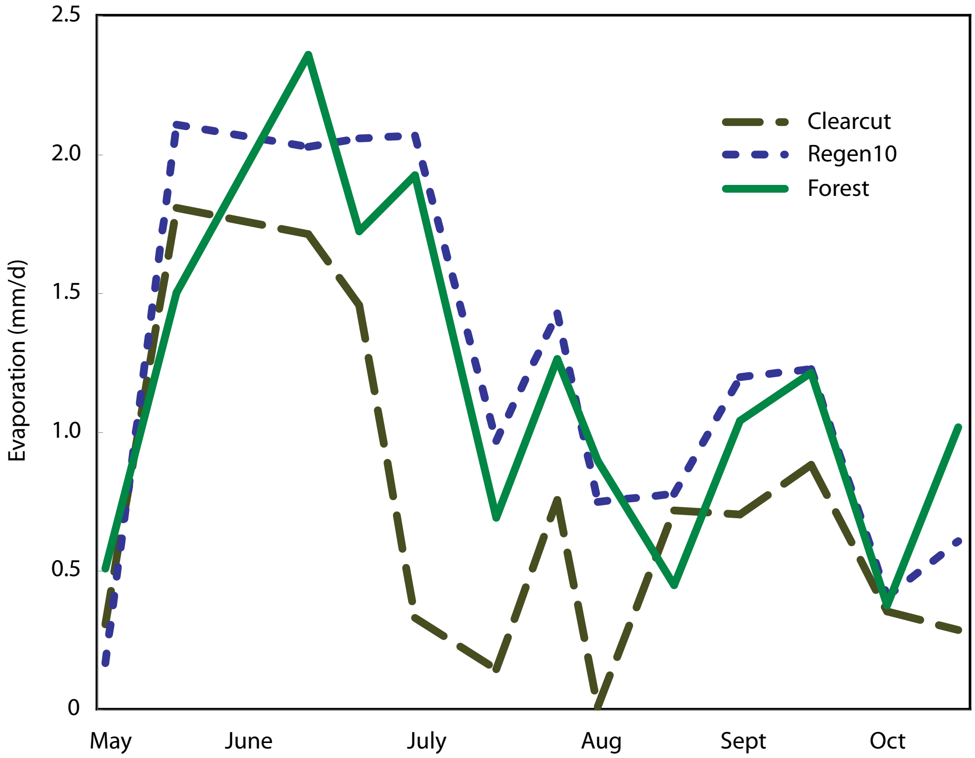


FIGURE 7.2 *Daily evaporation (mm/d) averaged over 10- to 20-day periods for a lodgepole pine forest (solid line), a clearcut (dashed line), and a regenerating 10-year-old lodgepole pine stand (dotted line) at Upper Penticton Creek. Data are based on measurements of water content and precipitation (Figure 7.1), and a water balance model estimate of drainage. Evaporation is the sum of plant transpiration, evaporation of intercepted water, and evaporation from the soil surface (Modified from Spittlehouse 2006).*

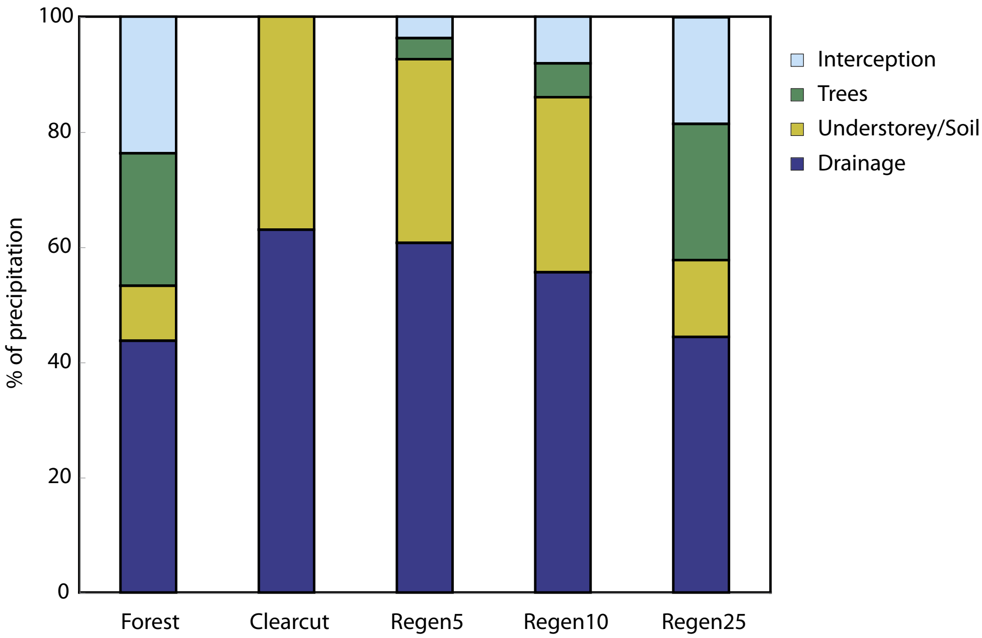


FIGURE 7.3 *Mean annual water balance for an old lodgepole pine forest, clearcut, and regenerating 5-, 10-, and 25-year-old lodgepole pine stands at Upper Penticton Creek from October 2002 to September 2005. Data are from a water balance model calibrated using measured root zone water content, precipitation interception, and tree transpiration data. Shown are interception of precipitation (light blue), tree transpiration (green), evaporation from the soil plus understorey transpiration (yellow), and drainage below the root zone (blue), which are expressed as a percentage of the mean October to September precipitation of 490, 840, and 645 mm for the 3 years (D.L. Spittlehouse, unpublished data).*

Snow ablation

Ablation refers to the loss or disappearance of a snowpack caused by snowmelt (drainage of meltwater at the base of the pack) and evaporation (including sublimation). The separation of ablation into melt and evaporation is necessary to understand snowmelt contributions to spring streamflow. In most situations, melt dominates over evaporation in snowpack ablation as the energy required for evaporation and sublimation is approximately 7.5 and 8.5 times greater than for melt, respectively (see Chapter 6, "Hydrologic Processes and Watershed Response").

In Finland, Kuusisto (1986) found that snow evaporation losses are generally low (less than 0.6 mm/d). At Mayson Lake, B.C., latent heat flux densities measured under forest cover and in a clearcut averaged less than 2 W/m^2 (Adams et al. 1998). These fluxes indicate negligible (less than 0.5 mm/d) evaporation losses from, or condensation on, the snowpack during seasonal snowmelt. Bernier and Swanson (1992) found that daily snow evaporation rates were low in both openings and in a lodgepole pine forest in Alberta, and ranged from an average of 1.1 to 2.5 mm/d, respectively. Bernier (1990) studied snow evaporation rates in the open and in an artificial stand of 2.5 m tall lodgepole pine trees in Alberta. At a stand density of 1650 stem per hectare, the evaporation rate was equal to that in the open. At 2500 stems per hectare, the rate decreased to one-third of that in the open. The consensus from available research is that vapour losses are minor relative to the rates of snowmelt, and that measured ablation provides a good approximation of meltwater draining from a snowpack. The relative importance of sublimation is greater for thinner snowpacks, as it becomes a larger proportion of the water balance.

Forest cover affects snow ablation by changing the surface energy balance. The amount of shortwave radiation reaching the snow surface is reduced, longwave radiation is increased, and the exchange of latent and sensible heat is attenuated through the reduction of wind speed, near-surface temperature, and humidity gradients relative to the open. Removal of the forest canopy exposes the snow surface to greater incident solar radiation and higher wind speeds, which can increase sensible and latent heat inputs (Berris and Harr 1987; Adams et al. 1998). Snow ablation rates are typically 30% to more than 100% higher in the open than in the forest (e.g., Toews and Gluns 1986; Berris and Harr 1987; Spittlehouse and Winkler 2004; Winkler et al. 2005).

Snow ablation rates reported in the literature are usually average values over a 1- to 4-week period. The rates vary from 4 to 25 mm/d in open areas and 3 to 17 mm/d under forest canopies similar to those found in south-central British Columbia (Winkler 2001). Metcalfe and Buttle (1998) found that in boreal stands, snowmelt rates increased exponentially with increasing canopy gap fraction (the fraction of sky visible to a sensor under the canopy). Hardy and Hansen-Bristow (1990) found that average snowmelt rates in a mature stand were 0.6 and 0.8 times those in 4 and 14 m tall juvenile stands, respectively. At Mayson Lake average melt season ablation rates measured over a 3-year period were 4 mm/d in a spruce-fir stand and 10 mm/d in a clearcut (Winkler et al. 2005). At long-term snow research sites on the Thompson-Okanagan Plateau, snow ablation averaged 15% lower and snow persisted up to 8 days longer in mature lodgepole pine stands than in the open. Snow ablation rates in mixed-species stands varied by up to 70% of those in the open (Winkler 2007).

Maximum snowmelt rates indicate how quickly snow cover can generate meltwater under extreme conditions. These rates are more difficult to measure and are less frequently reported than seasonal melt rates, but often exceed seasonal average rates by two to three times. Kattelmann et al. (1998) summarized snowmelt rates reported in the literature and suggested that under typical mountain conditions, maximum rates in the open are generally less than 50 mm/d. The authors concluded that the following conditions are most conducive to rapid snowmelt in the absence of rain: long periods of daylight around the time of the summer solstice; clear days with cloudy nights; warm humid air combined with high wind conditions; low snow albedo; discontinuous and shallow snowpacks so that light can penetrate to the ground surface; and local sources of longwave radiation (e.g., burned stems, logs, or rocks). In a clearcut at Mayson Lake, the maximum snowmelt rate measured as lysimeter outflow was 29 mm/d, which was more than three times the average of 8 mm/d measured as SWE loss over the melt season (Winkler et al. 2005). Haupt (1969) reported a maximum daily lysimeter outflow rate of 52 mm during clear weather at a high-elevation, south-facing site in Idaho. At Murphy Creek near Castlegar, B.C., Nassey reported[1] a maximum lysimeter outflow of 23 mm/d compared to the average SWE loss of 12 mm/d over the melt period.

Changes in snow ablation rates also occur following natural disturbance. After a forest fire in southern Montana reduced lodgepole pine canopy cover by 90%, average snow ablation rates increased by 57%, the same increase as was observed following clearcutting (Skidmore et al. 1994). In lodgepole pine stands near Fraser Lake, Boon (2009) found that average ablation rates were 14–17% higher in dead stands than in live stands, whereas differences between a live forest and open areas varied from 15 to 34%, depending on year. The average snow ablation rate in a dead stand near Vanderhoof was similar to that in a green stand and about one-half the rate of that in the open (Boon 2007). Initial results from other studies in British Columbia, summarized in Table 7.1, show that average ablation rates are 27–48% lower in green-to-red attacked pine stands, except at Border Lake where ablation rates were slightly higher in the forest than in the clearcut, and 14–29% lower in grey stands compared to rates in the open. Snow disappeared 3–10 days later in the green-to-red attacked stands and 0–6 days later in the grey stands than in the open. In a comparison of maximum ablation rates, Teti (unpublished data) found reductions of 17–48% in green-to-red attacked stands and rates 3% higher to 47% lower in grey stands.

Hillslope Runoff Generation

Because undisturbed forest soils in the Pacific Northwest, including British Columbia, generally have sufficiently high infiltration capacities, infiltration-excess overland flow (also known as *Hortonian* overland flow) is generally not an important process for streamflow generation (Cheng 1988; Wondzell and King 2003). Therefore, runoff generation in undisturbed forest catchments tends to be dominated by subsurface flow. In montane catchments in

1 Nassey, J.M. 1994. Measurement and modelling of snowmelt on a clear-cut site in the West Kootenays, British Columbia. Directed Stud. Rep., Simon Fraser Univ., Vancouver, B.C.

coastal British Columbia, which are typically dominated by shallow soils overlying bedrock or relatively impermeable glacial till, downslope flow occurs through the development of a transient saturated zone overlying the till or bedrock, with rapid flow facilitated by root channels and other preferred pathways (Hetherington 1982; Hutchinson and Moore 2000; Anderson 2008).

Forest harvesting can influence subsurface hydrologic response through decreased interception and evaporation of water, alteration of soil physical properties that control infiltration and transmission of water, and the construction of roads that reroute water. Decreased interception loss can lead to an increase in the amount of water that infiltrates the soil, which can result in higher water-table levels during storms (Dhakal and Sidle 2004). Forest canopy removal reduces interception losses. This increases rainfall intensity at the soil surface, which may cause more rapid subsurface flow and larger peak flows (Keim et al. 2006). Logging can also affect subsurface stormflow by compacting surface soils (i.e., reducing macropore space), which slows the transmission of water through the soil (deVries and Chow 1978).

Forest harvesting commonly leads to higher soil moisture content because removal of forest cover leads to a reduction in interception and evaporation of water. This trend has been observed in several different ecosystems (e.g., Megahan 1983; Adams et al. 1991; Keppeler et al. 1994; Troendle and Ruess 1997; Elliott et al. 1998; Hetherington 1998). Increases in subsurface stormflow after harvest are typically related to higher antecedent soil moisture content, resulting in a smaller soil moisture deficit that must be satisfied prior to initiation of lateral flow (Keppeler et al. 1994). This effect is most pronounced in summer and early autumn when soil moisture deficits are high (e.g., Ziemer 1981).

Very few published studies focus on changes in hillslope runoff generation following harvesting. At Carnation Creek on Vancouver Island, Hetherington (1998) noted greater post-harvest hillslope flow inputs to the valley bottom, but did not provide details. In Idaho, Megahan (1983) measured a large, post-disturbance (harvest and wildfire) increase in both subsurface stormflow volume (+ 96%) and flow rate (+ 27%) during the snowmelt period. The increase in volume and rate was attributed to a post-disturbance increase in both SWE and snowmelt rates. Similarly, by using a combination of hydrometric and geochemical methods, Monteith et al. (2006a, 2006b) found that a higher proportion of snowmelt-generated streamflow originated from surface and near-surface soil horizons in harvested than in unharvested watersheds on the boreal shield in Ontario. At the Fraser Experimental Forest in Colorado, Troendle and Ruess (1997) found that annual plot outflows increased from 15% of annual precipitation under mature forest cover to 60% in a clearcut plot. Keppeler and Brown (1998) measured a 400% increase in peak pipeflow after harvesting in the Caspar Creek watershed in northern California.

At many sites, particularly in British Columbia's interior, harvesting is conducted with skidders that can compact soil surfaces and cause overland flow. This can lead to an increase in the flashiness of streamflow response and the magnitude of surface erosion, and a decrease in the chemical interactions of water with the subsurface environment. The significance of this soil compaction and resulting overland flow depends on the degree of compaction and how much of the watershed area is disturbed (Putz et al. 2003), as well as whether the skid trails direct water to the natural drainage network. In south coastal British Columbia, Cheng et al. (1975) found that infiltration rates remained sufficiently high, even following compaction, and that infiltration-excess overland flow was not observed. Tracked machinery, such as hoe-forwarders and feller-bunchers, can also cause soil compaction, particularly if used when soil moisture levels are high (Greacen and Sands 1980). Excavated trails and constructed haul roads typically have compacted surfaces with lower permeability than forest floors, and can generate overland flow even in moderate rainstorms (Luce and Cundy 1994). The significance of this overland flow at the watershed scale depends on the proportion of the watershed that is covered by logging roads and landings, and on the connectivity of these surfaces to streams via ditches, road surfaces, and culverts.

Harvesting in riparian zones can have a significant effect on riparian zone hydrology. Changes in transpiration and water table drawdown caused by harvesting can decrease the fluctuations in discharge through the day and lead to an increase in low flows (Dunford and Fletcher 1947). These effects should lessen over time as riparian forests regrow. Nevertheless, changes in species composition during forest succession can affect low flows. For example, in the Oregon Cascades, a change in riparian vegetation from conifers to deciduous species following clearcut logging resulted in increased transpiration by streamside vegetation and reduced dry weather streamflow (Hicks et al. 1991).

Groundwater

The effects of forest harvesting on groundwater are generally driven by the same processes that result in greater hillslope flow and water yield (Smerdon et al. 2009a, 2009b). Very few published studies have focussed on the direct link between forest management activities (both harvesting and road construction) and groundwater (i.e., subsurface water in the saturated zone). Some studies have reported on the effects of tree removal (i.e., clearcutting, partial retention, or selective cutting) and have variously observed changes in the position of the water table, estimated changes in catchment water yield, or changes in streamflow. Most often, reductions in evaporation and interception lead to an increase in groundwater recharge, which results in elevated water tables. Changes in water-table elevations after harvesting are important because groundwater is the source of most base flow in streams, which has many economic and ecological values (Douglas 2008).

This section draws from two recent publications that summarize the available literature on forest management effects on groundwater hydrology, specifically water-table position, groundwater recharge, and the effects of road construction on groundwater flow (Smerdon et al. 2009a, 2009b). These authors also developed a hydrogeological classification for British Columbia and used it to place the potential effects of forest management on groundwater within a landscape context.

Effect of forest harvesting on water-table position

Forest harvesting generally leads to a rise in the elevation of the water table towards the ground surface. This is primarily attributed to the reduced interception and evapotranspiration that results from the loss of forest cover. The effect of changes in the interception of rain and snow by the canopy has been relatively well documented; however, changes in plant transpiration before and after harvest are complex and depend on vegetation type. The net effect is that wetter soil more readily conducts water to the saturated zone, which in turn may increase the elevation of the water table (Table 7.2).

The degree of water-table rise depends on specific watershed characteristics, including bedrock geology, surficial geology, soil type, and landform topography. At Carnation Creek on west Vancouver Island, late summer water-level rises on the floodplain of 30–50 cm persisted for 10 years after harvesting (Hetherington 1998), and contributed to the process of triggering preferential flow in humid, steep watersheds (e.g., Beckers and Alila 2004). Rex and Dubé (2006) noted elevated water-table levels in lowland areas (e.g., toe slopes, wetlands) after harvesting in the Vanderhoof Forest District (Table 7.2). On the boreal plains, the water table in low-relief watersheds with dry climates increased by 26 cm following harvesting (Evans et al. 2000).

An increase in water-table elevation may have different effects on forest management operations (e.g., trafficability) depending on soil characteristics and slope gradient. Reduced surface-water drainage density and understorey vegetation, and increased area of poorly drained soils, promote wetter ground conditions (Dubé and Rex 2008). In the Vanderhoof Forest District, summer logging on wet ground has been stopped in favour of logging when the ground is frozen and generally more stable (Rex and Dubé 2006). In steep mountainous watersheds, a rise in the water table may lead to a greater risk of slope failure (Sidle and Ochiai 2006). Each of these scenarios imposes constraints on logging operations, which illustrates the potential feedback between forest harvesting and trafficability.

A rise in the water table and wetter ground can also affect the selection of appropriate silvicultural systems (Pothier et al. 2003) and post-harvest species selection. Forest regeneration may be affected by higher water tables because some tree species do not tolerate saturated conditions in the rooting zone, which may decrease productivity or result in regeneration failure (e.g., Landhäusser et al. 2003). Some species, however, promote higher evapotranspiration rates, thereby helping to lower the water table following harvesting. These "nurse crops" could contribute to water-balance recovery in recently harvested and replanted sites (Landhäusser et al. 2003). Restoration of a vegetation cover that has similar evapotranspiration characteristics to the original species is an important step in minimizing the long-term effects of harvesting on groundwater systems and in maintaining forest productivity.

Effect of forest harvesting on groundwater recharge

Water entering the groundwater flow system (i.e., recharge) is difficult to quantify (deVries and Simmers 2002); however, inferences are possible from changes in water-table position, watershed water yield, and

TABLE 7.2 *Effects of forest harvesting on water-table position (from Smerdon et al. 2009b)*

Source	**Study site**	**Location**	**Annual precipitation (mm)**	**Forest management practice**[a]	**Change in water table**
Hetherington (1998)	Carnation Creek	Vancouver Island, British Columbia	2100–4800	CC	30–50 cm rise that persisted for 10 years following harvest
Fannin et al. (2000)	Carnation Creek	Vancouver Island, British Columbia	2100–4800	CC	50–150 cm rise (approx.) following individual storm events. Large spatial variability because of soil conditions, but all water-table response was rapid. An upper limit to pressure-head increase was observed, above which preferential flow pathways activated.
Rex and Dubé (2006)	Vanderhoof Forest District	Central British Columbia	496	CC + MPB	10 cm (approx.) higher water table in toe-slope of cut area compared to MPB-kill area; 30 cm (approx.) higher water table in upland of cut area compared to MPB-kill area
Evans et al. (2000)	TROLS	Central Alberta	468	PC	26 cm higher in cut area compared to uncut area
Dubé et al. (1995)	Beaurivage Forest	St. Lawrence lowlands, Quebec	957	CC	7–52 cm rise, depending on soil texture
Pothier et al. (2003)	Villroy	St. Lawrence lowlands, Quebec	510	PC + CC	Up to 22 cm rise in cut areas. Water-table rise increased linearly with percentage of cut area in the first year following harvest. Five years after harvest, water tables remained elevated but were less dependent on the percentage of area cut.
Megahan (1983)	Pine Creek	Central Idaho	890	CC	90 cm rise in water table, decreasing to approximately 40 cm after 2 years
Bliss and Comerford (2002)	-	Gainesville, Florida	1150	CC	21–49 cm rise after 900 days. Larger seasonal fluctuations observed for 4 years following harvest.
Peck and Williamson (1987)	Collie River Basin	Western Australia	820–1120	CC + PC	100–400 cm rise following wet season. Water table increased by 260 cm/yr in clearcut areas and 90 cm/yr in partially cleared areas.

a CC: clearcut; PC: partial cut; MPB: mountain pine beetle

base flow (in some cases). Similar to water-table rises following harvesting, wetter soil conditions may lead to increases in groundwater recharge rates because of the wetter antecedent conditions and the lower available storage in the unsaturated zone. Appreciable amounts of groundwater recharge (compared to runoff) are not expected in watersheds in steep terrain (e.g., Hudson and Anderson 2006). Although no published studies directly measured increased rates of groundwater recharge following harvesting, a few have shown increases in watershed yield, which may be a result of higher groundwater recharge (Table 7.3). For example, in the northeastern United States, a harvested headwater watershed supplied an

TABLE 7.3 *Effects of forest harvesting on groundwater recharge (from Smerdon et al. 2009b)*

Source	Study site	Location	Annual precipitation (mm)	Forest management practice[a]	Inferred change in groundwater recharge
Bates (2000)	Fernow Experimental Forest	West Virginia	1470	PC	Harvested watershed had greater low flows (base flow) to headwater streams caused by higher soil moisture in the years following harvest. Minor amount of (event) stormflow noticed, compared to deeper subsurface flow.
Bent (2001)	Cadwell Creek	Massachusetts	1174	PC	Groundwater recharge increased by 68 mm/yr for six seasons following harvest
Cornish (1993)	Karuah	Australia	1450–1750	PC	Yield increased 150–250 mm/yr following harvesting, depending on percentage of area cut. Increased recharge and overall water yield remained higher for 3 years following harvesting.
Bren (1997)	Cropper Creek	Southeast Australia	660	CC	Increase in amplitude of diurnal streamflow fluctuations attributed to increased subsurface flow, after removal of vegetation
Cook et al. (1989)	Western Murray Basin	Australia	340	CC	Recharge increased by 20 mm/yr very gradually following harvest (approx. 200 years based on simulation modelling)

a CC: clearcut; PC: partial cut

increased base flow for a few years that was apparently not generated from shallow stormflow (Bates 2000). In another case, recharge increased by 68 mm/yr for 6 years following harvesting (Bent 2001). Pike and Scherer (2003) summarized similar results for snowmelt-dominated hydrologic regimes. Results of stand water-balance models for the Upper Penticton Creek Experimental Watershed in the Okanagan Basin showed an increase in water drainage from the soil rooting zone after harvesting and natural disturbance (e.g., mountain pine beetle) compared to undisturbed mature forest stands (Spittlehouse 2007, Figure 7.3). Similarly, simulations of fire effects on pine stands at Mayson Lake indicated that greater root zone drainage is a result of high antecedent soil moisture conditions (Moore et al. 2008).

The implications of changes in groundwater recharge vary according to watershed characteristics (e.g., bedrock geology, surficial geology, soil type, and topography). Further study of the relationship between groundwater recharge and streamflow in forest management areas is needed to quantify the hydrologic mechanisms that control flow regimes. The magnitude of increase in water yield, and potentially the increase in groundwater recharge, may be linearly related to the percentage of area cut (partial harvesting or clearcutting), especially in the first few years following harvesting (e.g., Stednick 1996). Groundwater flow systems adjust to increased water input, and effects may be short- or long-term depending on the scale of the flow system. Recharge that is part of a local-scale system may discharge relatively quickly to nearby headwaters (e.g., Bren 1997), in which case the effects of harvesting could be detectable and (or) ecologically significant. Conversely, the effects of relatively short-term forest disturbances (i.e., decades) may not be detectable in larger-scale flow regimes (e.g., regional-scale flow systems), which tend to respond over long time scales.

Forestry roads and groundwater flow

The effects of logging roads on hillslope hydrology and watershed response are a major focus of concern and debate (Luce and Wemple 2001). Roads that cut into a hillside may intersect shallow groundwater. This creates a seepage face along the road cut (Megahan and Clayton 1983) and more rapidly redirects surface water flow in ditches and culverts into the stream network, which potentially increases peak flows (Jones and Grant 1996; Wemple and Jones 2003). Forestry roads that intercept and re-direct shallow groundwater may also reduce groundwater flow to downslope environments (e.g., springs and seepage areas).

The effect of flow interception by roads depends on how much subsurface flow is intercepted and how much is conveyed directly to the stream network. The proportion of intercepted subsurface flow is a function of many factors, including soil depth, permeability of the bedrock underlying the soil, depth of the road cut/ditch surface, permeability of the roadbed material, and location of the road on the slope. These factors combine with specific watershed characteristics (e.g., bedrock geology, surficial geology, soil type, and topography) that vary across the province. On the south coast of British Columbia, Hutchinson and Moore (2000) found that most of the net rainfall input to a hillslope segment was intercepted at a road cut that was more than 1 m deep with soils averaging 1 m deep and underlain by a compacted basal till. At Carnation Creek on Vancouver Island, road construction at different locations either increased, decreased, or resulted in no change to downslope water-table elevations (Hetherington 1998). The differences in downslope effects were related to road location and construction practices. The effect of road interception on groundwater flow may have significant effects on aquatic ecosystems that rely on specific rates and timing of groundwater discharge or duration of base flow. Subsurface flow intercepted at a road cut during snowmelt in the Idaho Batholith was about 30–40% of the total upslope water input, suggesting that over one-half of the meltwater flowed downslope below the road cut, through the bedrock (Megahan 1972).

Runoff from a road network can flow into stream channels via two pathways: (1) roadside ditches that drain directly to streams, and (2) roadside ditches that drain to culverts that feed water into incised gullies (Wemple et al. 1996). In other cases, ditch flow may be diverted back onto the slope below a road by a cross-drain or culvert, where it will re-infiltrate and flow downslope as subsurface flow. In this case, the road may not increase the rate of transmission of water to the stream channel, but rather redistribute subsurface flow laterally across the slope. For steep valley-side streams without well-defined watershed boundaries, this redistribution could transfer water from one small stream to the next. The redistribution of flow from its initial path into gullies and onto slopes can increase the potential for slope failures (see Chapter 9, "Forest Management Effects on Hillslope Processes").

In more gently sloped terrain, the potential for road cuts to intersect groundwater flow systems is typically lower than in steep terrain, except near groundwater discharge areas (streams, wetlands) where the water table is shallow. Under such conditions, the physical attributes of the road may be more important than those of the surrounding landscape. For example, compacted road surfaces can limit infiltration and pre-existing lateral flow (if present). Whether this effect is significant depends on the area of the watershed that is covered by compacted surfaces (Putz et al. 2003), the location of roads, and the type and quality of road construction.

Forest management effects on regional groundwater resources

Groundwater flow systems at the regional scale tend to buffer short-term variability in climate and land use changes (including forest management activities), but also tend to integrate long-term changes, which makes deleterious impacts more difficult to reverse. As a result, widespread changes in upland recharge areas caused by forest harvesting could go unnoticed for decades in adjacent valley-bottom aquifers. The effects may also be masked or magnified by climate variation and change. Widespread forest clearing in Western Australia throughout the 1950s and 1960s provided an example of the time involved for changes to propagate through groundwater regimes. Low-rainfall areas (850 mm/yr) and high-rainfall areas (1120 mm/yr) were clearcut, and groundwater response was monitored over the succeeding decades (Hookey 1987). Water budget studies and simulation modelling for the flow systems in Western Australia have shown that groundwater equilibrium for the basin re-establishes 25–30 years after cutting (Hookey 1987).

To date, no studies on the potential large-scale effects of forest disturbance on regional groundwater resources have been conducted in British Columbia. Therefore, studies that focus on the dynamics of complete groundwater flow systems and the mechanisms that govern groundwater recharge and discharge in forest management areas are warranted, particularly in the context of the current mountain pine beetle infestation and associated salvage harvesting in central British Columbia. Rex and Dubé (2006) found that low-relief watersheds with fine-textured soils and dead pine stands have wet soils and a raised water table. Cutblocks may have higher water-table elevation than beetle-kill areas, and the difference between beetle-kill and cutblock areas could be more pronounced at upper slope locations (Rex and Dubé 2006). Such findings suggest that the effects of forest disturbance on regional groundwater resources may not manifest for decades. Clearly, further research on the effects of forest disturbance on regional groundwater resources in British Columbia is needed, especially as the use of groundwater resources and urban interface forest management increases in the province.

WATERSHED-SCALE EFFECTS

Although many of the hydrologic effects of forest operations are reasonably well understood at the site or stand scale, it is more difficult to make quantitative predictions at the watershed scale (Committee on Hydrologic Impacts of Forest Management 2008; also see Chapter 16, "Detecting and Predicting Changes in Watersheds"). Three streamflow variables of primary interest are: (1) water yield, (2) peak flows, and (3) low flows (see Chapter 6, "Hydrologic Processes and Watershed Response). Understanding the effects of forest disturbance and regrowth on total annual and seasonal water yield, peak flows, and low flows is important to sustain water supplies, protect aquatic habitat, design infrastructure, and mitigate risks to lives and property.

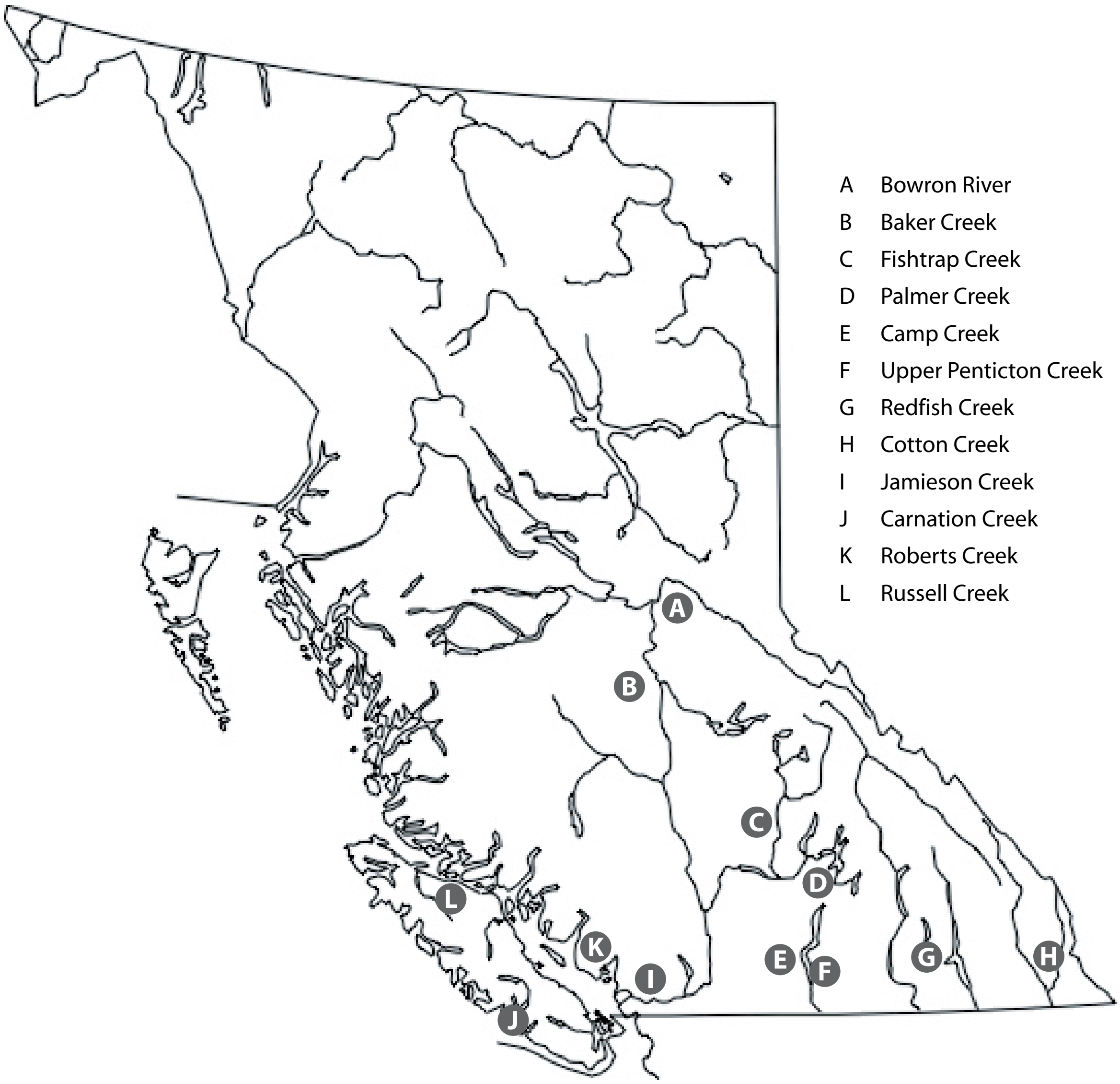
Detecting the effects of disturbance on streamflow variables at the watershed scale is difficult because of the natural variability in driving factors, such as climate, geology, topography, and forest cover within and between watersheds. The most statistically rigorous method of quantifying the effects of forest operations on streamflow involves paired-watershed experiments, which include pre- and post-harvest data and at least one untreated control (Hewlett 1982). Alternative approaches to paired-watershed experiments include: retrospective studies that use existing operational streamflow data; pre- and post- treatment studies without a control; chronosequence analysis in which trends in streamflow in a suite of watersheds with different management histories are correlated with changes in land use and (or) climatic variables; and computer simulation models. Most recently, Alila et al. (2009) suggested pairing peak-flow events based on frequency rather than chronology to evaluate changes after logging. Table 7.4 summarizes information about watershed experiments within British Columbia that quantify disturbance effects on streamflow. Figure 7.4 shows the locations of these watersheds. Table 7.5 summarizes similar information for watershed experiments outside British Columbia.

TABLE 7.4 *Watershed experiments that quantify forest cover effects on streamflow in (a) coastal, (b) northern interior, and (c) southern interior regions of British Columbia*

Forest region /watershed (Reference)	Status[a]	Leading tree species[b]	Sub-basins	Area (km^2)	Elevation range (m)	Hydrologic regime[c]	Duration[d]	Average annual precipitation (mm/yr)	Average annual water yield (mm/yr)	Design[e]	Treatment[df] (years): total of area cut (%)	Changes[g] measured (%) or under investigation (√): Annual yield	Changes[g] measured (%) or under investigation (√): Peak flow
a) Watershed experiments in coastal British Columbia													
Carnation Creek													
(Hetherington 1982, 1998; Hartmann and Scrivener 1990)	O	Hw, Cw, Ba, Fd	B	10.0	8–884	R	1970–p	2100-4800	2544	PP	CC (1976–1981): 41 CC (1987–1995): 25; (66% total)		
	O		C	1.5	46–700	R	1970–p				Control: 0%	√	√
	O		E	2.6	150–884	R	1970–p				Control:	√	√
											CC (1987–1993): 38	√	√
	O		H	0.1	152–305	R	1970–p				CC (1977–1978): 90	14	20
	C		J	0.2	30–300	R	1970–1996				CC (1977–1978): 94	√	√
Flume Creek													
(Roberts)	C	Hw, Fd	F4	< 1	505–850	ROS	1995–2005	1640	1416	BACI	VR (1998): 44		80
(Hudson 2001)			F5	< 1	505–850				1405		SC (1998): 32		85
			F6	< 1	395–560				1486		Control: 0		
Jamieson Creek													
(Golding 1987)	C	Fd, Hw, Cw	Jamieson	3	305–1310	ROS	1972–1984	3525	2995	BACI	CC (1978, 1982–1984): 19		13.5
			Elbow	1	275–1065				1525		Control: 0		
Russell Creek													
(Anderson 2008; Floyd and Weiler 2008)	O	Hw, Fd	Stephanie	32	275–1700	ROS	1991–p	2395	1993	M, ISS	CC (1978–p): 35		√
b) Watershed experiments in northern interior British Columbia													
Baker Creek													
(Forest Practices Board 2007)	O	Pl		1570	470–1530	SM	1963–2005, 2006–p	400	80	M, ISS	CC (2005*): 34; MPB (2005*): 53; CC (< 1970–p)	31*	61*
Bowron River													
(Lin and Wei 2008)	C	Se, Bl		3590	602–2447	SM	1954–1996	1149 (upper)	594	PP	CC (1966–1996): 25	NS	NS

c) Watershed experiments in southern interior British Columbia													
Camp Creek (Cheng 1989; Moore and Scott 2005)	C	Pl	Camp	34	1070–1920	SM	1970–2000	600	140	CI	CC (1976–1978): 30	NS	30 (April)
			Greata	41	880–1620						Control: 0		
Cotton Creek (Jost et al. 2007)	O	Pl, Se, Bl		17	1100–2100	SM	2004–p	650	284	ISS	CC (1975–p): 34	√	√
Fishtrap Creek (Moore et al. 2008)	O	Pl, Se, Bl, Fd		135	370–1620	SM	2004–p	471	180	ISS	B (2003): 75; CC(?–p)	√	√
Palmer Creek (Cheng and Bondar 1984)	C	Fd, Pl, Se, Bl	Palmer	18	960–1950	SM	1967–1977	750	350	BACI	B(1973): 50	24(Apr–Aug), 37(Aug–Nov)	50
			Upper Salmon	143							Control: 0		
Redfish Creek (Whitaker et al. 2002)	O	Fd, Lw, Pl, Se, Bl		26	700–2370	SM	1992–p	1582	1018	M, ISS	CC (*): 22max CC (1979p)		22*
Upper Penticton Creek (Winkler et al. 2008b)	O	Pl	240	5	1620–1920	SM	1984–p	700	325	BACI	Control: 0		
		Pl	241	5	1600–2015				326		CC (1995): 10 CC (1998): 18 CC (2002): 28 CC (2006): 47	NS, NS, √ √	NS, NS, √ √
		Se, Bl, Pl	Upper Dennis	4	1780–2065				389		CC (1995): 10 CC (1998): 21 CC (2000): 52	NS, NS, √	NS, NS, √

a O: ongoing; C: complete
b Ba: amabilis fir; Bl: subalpine fir; Cw: western redcedar; Fd: Douglas-fir; Hw: western hemlock; Lw: western larch; Pl: lodgepole pine; Py: ponderosa pine; Se: Engelmann spruce
c R: rain; ROS: rain on snow; SM: snowmelt
d p: present
e BACI: before after control intervention (paired watersheds all with pre- and post-disturbance data); CI: control intervention (paired watersheds with no pre-disturbance data); PP: pre- and post- (single watershed time series); ISS: intensive study site for detailed modelling; M: modelling
f CC: clearcut; B: burned; VR: variable retention (% of canopy removed); SC: shelterwood (% of canopy removed); PC: partial cut (% basal area removed); MPB: attacked by mountain pine beetle (% of watershed area attacked)
g * modelled; NS: non-significant or non-detectable



A Bowron River
B Baker Creek
C Fishtrap Creek
D Palmer Creek
E Camp Creek
F Upper Penticton Creek
G Redfish Creek
H Cotton Creek
I Jamieson Creek
J Carnation Creek
K Roberts Creek
L Russell Creek

FIGURE 7.4 *Locations of watershed experiments in British Columbia.*

TABLE 7.5 *Watershed experiments that quantify forest cover effects on streamflow in hydrologic regimes similar to those in British Columbia*

Study site	Hydrologic regime[a]	References
Cabin Creek (Alberta)	SM	Swanson et al. 1986
Alsea Watershed Study (Oregon Coast Range)	R	Harr et al. 1975; Harris 1977; Harr 1983; Stednick 2008
Hinkle Creek (Oregon Coast Range)	R, ROS	n/a
H.J. Andrews (Oregon Cascades)	R, ROS	Rothacher 1973; Harr 1983; Harr 1986; Hicks et al. 1991; Jones and Grant 1996; Thomas and Megahan 1998; Beschta et al. 2000; Jones 2000
Fox Creek (Oregon Cascades)	R	Harr 1980, 1983
Coyote Creek (Oregon Cascades)	R	Harr et al. 1979; Harr 1983
Mica Creek Experimental Watershed (Idaho)	SM	Hubbart et al. 2007
Horse Creek (Idaho)	SM	King and Tennyson 1984
Fraser Experimental Forest (Colorado)	SM	Alexander et al. 1985; Troendle and King 1985
Deadhorse Creek (Colorado)	SM	Troendle and King 1987
Wagon Wheel Gap (Colorado)	SM	Bates and Henry 1928; Van Haveren 1988

a R: rain; ROS: rain on snow ; SM: snowmelt

Annual Water Yield

Studies in temperate forest environments around the world have demonstrated that annual water yield generally increases following removal of forest cover (Bosch and Hewlett 1982; Trimble et al. 1987; Stednick 1996). Moore and Wondzell (2005) reviewed results of paired-watershed studies in the Pacific Northwest and found that, in rain-dominated watersheds, water yield increased by up to 6 mm for each percentage of the watershed area clearcut or patch cut, whereas selective cutting increased yields by up to 3 mm for each percentage of basal area removed. Most of the increased yield occurred in the wet autumn–winter period (Harr 1983; Keppeler and Ziemer 1990). Water yield decreased slightly following patch cutting in two watersheds in the northern Oregon Cascades (Fox Creek 1 and 3), likely because of a decrease in fog drip (Harr 1982).

In snow-dominated watersheds, post-logging water yields increased from 0.25 mm to more than 3 mm per percentage of watershed area harvested (Van Haveren 1988; King 1989; Stednick 1996). At the Fraser Experimental Forest, Colorado, water yield from Fool Creek was 40% higher than the long-term expected flow and 45% higher for the first 5 years post-logging, even though the SWE averaged over the watershed increased by only 9% (Troendle and King 1985). This disproportionate increase in water yield relative to snowpack likely resulted from the combined effect of increased snow accumulation and decreased evaporation (and thus reduced soil moisture deficits) within the harvested areas. At the Horse Creek watersheds in Idaho, annual yields increased by about 3.6 mm per percentage of watershed cleared (King 1989). Small increases in yield were reported at Cabin Creek, Alberta (Swanson et al. 1986) and Deadhorse Creek, Colorado (Troendle and King 1987), but these increases were not statistically significant.

Few studies have been conducted on the effects of roads on water yield because harvesting usually occurs either at the same time as, or immediately following, road building. At the Horse Creek watersheds in Idaho, water yield changes for road-only treatments were not statistically significant (King and Tennyson 1984); however, the short study period (i.e., small sample size) likely limited the ability of the analysis to detect a change.

Peak Flow

The effect of forest disturbance on peak flows has been a source of controversy for decades (Food and Agriculture Organization 2005; Committee on Hydrologic Impacts of Forest Management 2008; Grant et al. 2008). Data from numerous studies (notably those from the H.J. Andrews Experimental Forest in the Oregon Cascades) have been analyzed repeatedly using different data-processing approaches and statistical analyses to assess the magnitude of peak flow increases following forest harvesting (Jones and Grant 1996; Thomas and Megahan 1998; Beschta et al. 2000; Jones 2000; Grant et al. 2008; Alila et al. 2009). The term "peak flow" has been used in various ways to refer to the single largest flow event recorded in a year, the largest flow over a 7-day period, and the total volume of water delivered over the high-flow season and other time periods depending on the flows of interest. Peak flow events are described in terms of magnitude, frequency of occurrence, and variability. Often, the data will limit the type of analysis that can be undertaken. The most common limitation is a short data record, which may not include extreme events. The type of analysis can further limit interpretation of the data. For example, the average of a series of high flow events does not provide insights about flow extremes or the frequency at which events of various magnitudes occur.

The effects of disturbance on peak flows can differ between rainfall, rain-on-snow, and spring snowmelt events. Reported changes in mean peak flows following logging in rain-dominated watersheds range from increases of 20–30% (Harr 1983) to a decrease of 22% (Cheng et al. 1975). Cheng et al. (1975) also reported a several-hour delay in peak flow following harvesting in south coastal British Columbia. The authors suggested that extensive skid road development and soil disturbance (affecting 50% of the watershed area) had reduced the efficiency of macropore flow, which resulted in slower delivery of water to the stream channel (Cheng et al. 1975).

Lewis et al. (2001) analyzed peak flow responses to forest harvesting in 10 watersheds in the North Fork of Caspar Creek, northern California. In this study, roads occupied less than 7% of the area in each watershed. In most watersheds, maximum peak flow increased by less than 100%, but in some, maximum flows increased by up to 300%. The size of the treatment effect increased with the proportion of watershed harvested and decreased with an index of antecedent wetness; it therefore tended to be greatest in early autumn. The mean percentage increase in peak flows decreased with storm size. For a storm event with a 2-year return period, the average increase in peak flows was 27% in the watersheds that were 100% clearcut.

Peak flow responses to forest harvesting during rain-on-snow events can be highly variable, depending on the magnitude of rainfall, wind speed, temperature, water equivalent of the remaining snowpack, and the extent of snow cover (Kattelmann 1987; McCabe et al. 2007). In some studies, peak flows during rain-on-snow events increased following harvesting because of increased melt rates and reduced rainfall interception (Harr 1986). At temperatures close to zero, Beaudry and Golding (1983) found that snow intercepted by forest canopies melted, whereas snow on the ground remained frozen and resulted in increased runoff from the forest relative to the open. Hudson (2001) found highly variable peak flow responses in two watersheds in south coastal British Columbia that could be related to the role of transient snow cover and the percentage of canopy removal. Jones (2000) re-analyzed peak flow data for the H.J. Andrews Experimental Forest (Oregon Cascades) using analysis of variance, and found that winter rain-on-snow peak flows increased by 25–31% in four of five watersheds.

Increased peak flows have been reported in snow-dominated watersheds where more than 20% of the watershed was harvested; however, there was no direct correlation between the extent of harvesting and peak flow change. For example, in Colorado, mean peak flows increased by 20% to almost 90% in research watersheds where 20–40% of the watershed had been harvested (Troendle and King 1985, 1987; King 1989). Van Haveren (1988) found that 100% clearcutting produced a 50% increase in mean peak flow. Alila et al. (2009) re-analyzed the 48-year streamflow record for Fool Creek, Colorado, by pairing peak flow events according to estimated frequency rather than chronology. This re-analysis allowed an assessment of logging effects on the magnitude of peak flow associated with a given return period. They found that peak flows increased after logging across the full range of event frequencies and that the frequency of peak flow events larger than the mean also increased. Because of the non-linear relationship between peak flow magnitude and frequency, a small change in peak flow translated into a large change in the frequency of the event (i.e., a given peak flow was equalled or exceeded more frequently following forest harvesting). Results of forest harvest modelling for the 25-km^2 Redfish watershed near Nelson suggest that removing 100% of the forest cover in 60% of the watershed (without the simulation of roads) would cause a 7-day annual maximum discharge event (i.e., an event that would normally be expected once every 30 years) to occur every 8 years (Schnorbus and Alila 2004). These modelling results also suggest that an event previously expected to occur once every 8 years could be expected every 3 years (Schnorbus and Alila 2004). The magnitude of the snowmelt-generated peak in this study increased by 15%. This is a small increase compared to other studies, and was attributed to desynchronization of the snowmelt in the upper-elevation alpine zone from that under forest cover.

Peak flows can also be affected by forest cover losses caused by fire, insects, and disease. The effects of fire on peak flow vary with burn severity (spatial extent and intensity). High-intensity fires can alter infiltration and overland flow processes, which can result in increases in peak flow that may be as large as one or two orders of magnitude during intense rainstorms (Scott 1993; Neary et al. 2005). A summary of the effects of fire on peak flow indicated that increases in peak flow ranged from negligible in a snow-dominated fir forest, to 1.4 times in a coastal Douglas-fir forest, to 20–2000 times in ponderosa pine forests (Committee on Hydrologic Impacts of Forest Management 2008). Moore et al. (2008) found that after a fire burned throughout the Fishtrap Creek watershed near Kamloops, flows early in the freshet season increased and high flows were sustained longer than before the fire. In an Engelmann spruce beetle epidemic in Colorado, defoliation of up to 80% of the trees in 30% of the area in a north-facing and a west-facing watershed resulted in 4% and 27% increases in maximum annual instantaneous flow, respectively (Bethlahmy 1975). These increases were expected to persist for 25 or more years after the infestation.

In general, salvage harvesting is expected to have a greater hydrologic impact than mortality related to the mountain pine beetle (Redding et al. 2008). Modelling of four scenarios involving beetle-related tree mortality and subsequent salvage harvesting at Baker Creek in British Columbia's central interior suggested that with extreme disturbance, peak flows would occur more than 2 weeks earlier than normal, and average peak flows would increase by 60–90% (Forest Practices Board 2007). These results indicate the potential for a major shift in flood frequency to occur in the watershed. With extensive salvage harvesting (80% of the watershed area), the pre-beetle disturbance flood with a return period of 20 years may increase to a return period of 3 years. This shift would have major implications on the design of infrastructure on the floodplain. It also highlights the need to plan the extent of clearcut salvage harvesting in infested watersheds, and the need to designate reserve areas and carefully design stream crossings (Forest Practice Board 2007).

An important issue in understanding the effects of forestry operations and natural disturbance on snowmelt-driven peak flows is synchronization of snowmelt. In some situations, partial logging of a watershed could reduce peak flows through desynchronization of snowmelt in openings and under forest cover. For example, clearcutting 50% of a low-relief peatland watershed in north-central Minnesota resulted in two relatively low-magnitude peak flows instead of a single, higher peak flow (Verry et al. 1983). The first peak resulted from snowmelt in the cut area; the second occurred several days later in response to melting of the forest snowpack. After the remaining upland area was clearcut, snowmelt was uniform over most of the watershed, and peak flows increased compared to flows that would have occurred in the absence of clearcutting. At Fishtrap Creek, harvesting also appears to have desynchronized melt runoff over the watershed, resulting in lower peak flow magnitude but longer duration of high flow periods (Moore et al. 2008); however, this inference is based on a relatively short post-disturbance record and requires further study. Logging of north-facing slopes or high-elevation areas could advance the timing of melt from those areas and synchronize it with melt on forested south-facing or lower-elevation slopes (Toews and Gluns 1986).

Logging roads did not appear to have a measurable effect on peak flows in most studies that included a road-only treatment (Harr et al. 1975; Ziemer 1981; King and Tennyson 1984; Jones 2000). In these studies, however, the road-only condition was typically short (1–2 years), so the analyses would have low power (i.e., low ability to detect an effect). One exception was the Deer Creek 3 watershed (Alsea Watershed Study, Oregon), where peak flow increases averaged 18% following road construction that affected 12% of the area (Harr et al. 1975).

A particular point of controversy is the hydrologic and geomorphic significance of peak flow changes. Some studies have concluded that changes in peak flows are largest for small events that have "no hydraulic consequence" (e.g., less than a 2-year return period) (Ziemer 1981). Several studies have shown that the mean effect of harvesting decreases with increasing event magnitude (as indexed by the return period for the control watershed peak flow response) (Beschta et al. 2000; Moore and Wondzell 2005; Grant et al. 2008). However, this type of analysis does not provide a basis for interpreting the effect of forest harvesting on flood frequencies, or for assessing the change in peak flows of a given return period or the change in recurrence interval for a specified magnitude of peak flow. For example, the change between return-period events is not linear, so a 5% change in a 25-year flood can result in a peak flow that approximates a 100-year event.

One difficulty in assessing the effect of forest harvesting on larger peak flow events is that few large storm events have been sampled in experimental studies. This is due, in part, to the relatively short periods of pre- and post-treatment monitoring and the rarity of high-magnitude events. The ability to draw generalizations about the effect of forest harvesting on peak flows is further confounded by the wide range of observed response, which is linked to the nature of the events generating peak flows (e.g., rain vs. rain-on-snow vs. spring melt), watershed characteristics, details of the road and harvesting systems, and the time since harvesting. Separating the effects of road building from forest harvesting is particularly difficult because, in most studies, road building and harvesting occur either simultaneously or in close succession.

Even if forest harvesting does not have an effect on major channel-disturbing events, changes in small to medium events can potentially affect channel morphology by changing the frequency of events that can move substantial amounts of sediment. A more detailed discussion of channel morphology is provided in Chapter 10 ("Channel Geomorphology: Fluvial Forms, Processes, and Forest Management Effects").

Development of guidelines for managing the risks associated with forestry-related peak flow increases is hampered by uncertainties about how logging changes peak flows and is compounded by uncertainties regarding a channel's sensitivity to those changes. Grant et al. (2008) generated some tentative guidelines for managing these risks in coastal catchments in Oregon. For snowmelt-dominated catchments, the current knowledge base suggests that risks are low when up to about 20% of the catchment is clearcut. Risks increase, but become highly uncertain, as the area clearcut approaches and exceeds 30%. Further research that combines field studies and modelling with long-term monitoring is needed to quantify the hazards associated with forestry-related peak flow increases in British Columbia watersheds of varying sizes, biophysical characteristics, and different hydrologic regimes.

Low Flow

Low flows are minimum flows that are sustained by groundwater during the dry season or prolonged periods without rain. They may occur in summer or during the time of year when precipitation falls as snow. Reductions in low flows can have a significant effect on aquatic ecology and on water supplies for irrigation, domestic use, and industry. Various metrics have been used to describe low flows in experimental studies. These include the number of low flow days each year (the number of days that daily discharge was less than some arbitrary threshold, which varies among studies), water yield during August (typically a period of low flows), and the total flow for months when the discharge from the control watershed was below an arbitrary threshold. It is difficult, therefore, to make quantitative comparisons across studies without re-analyzing the original data.

In a review of 28 studies on low flows conducted in the United States, Austin (1999) found that 16 cited a statistically significant increase in streamflow, 10 showed no change, and two studies noted less streamflow following logging. Pike and Scherer (2003) reviewed eight forest harvesting studies in snow-dominated environments and found that four reported an increase in low flow volumes, whereas four showed no significant change.

In coastal rain-dominated watersheds in northern California, the Oregon Cascades and Coast Range, and Vancouver Island, more summer streamflow was noted in the first few years after harvesting (Harris 1977; Harr et al. 1982; Hetherington 1982; Keppeler and Ziemer 1990; Hicks et al. 1991). An exception occurred in the Bull Run watershed in the northern Oregon Cascades, where the number of days with low flows increased significantly following patch cutting of 25% in Fox Creek 1 and 3 watersheds.

The decrease in streamflow during the low flow period was attributed to a reduction in fog drip (Harr 1982).

Greater streamflow during the low flow period has been shown to diminish over time with the establishment of vegetation after logging (i.e., because of increased evaporative losses), and in some cases, can be reduced below pre-harvest levels. For example, in the H.J. Andrews Watershed 1 (HJA-1) in Oregon, August flows were higher than expected in the first 8 years after logging, but then dropped below expected flows during the next 18 years (Hicks et al. 1991). This response was attributed to the post-harvest colonization of the watershed's relatively wide valley floor by hardwoods, including red alder, which has lower stomatal resistance and thus a higher transpiration rate per unit of leaf area (which was also higher) than conifers (Moore et al. 2004). The high transpiration rate from the riparian zone reduced streamflow. At the South Fork of Caspar Creek in northern California, summer flow volume increased and the number of low flow days decreased after harvesting, but these effects persisted for only about 5–10 years (Keppeler and Ziemer 1990). In fact, there was some indication that summer flows 10 years after harvest decreased to levels less than expected for pre-harvest conditions. Similar, though not statistically significant, trends were recorded at Needle Branch in the Alsea watershed study (Harris 1977).

Low flow responses to forest harvesting in snow-dominated watersheds are not well documented, as the lowest annual flows often occur in mid- to late winter when it is difficult to measure flows accurately because of ice formation in channels and on weirs. At Wagon Wheel Gap, Colorado, 100% clearcutting increased base flows by 17%, but had no statistically significant effect on 30-day low flows (Van Haveren 1988). At Cabin Creek, Alberta, clearcutting 20% of the watershed area increased flows in August, September, and October by 10–15%, but these changes were not statistically significant (Swanson et al. 1986).

Roads may influence low flows in small, upland watersheds by diverting subsurface flow laterally across hillslopes, particularly where streams are weakly incised. The net effect could be an increase in flows in some streams at the expense of others; however, this effect is not well documented by field measurements.

HYDROLOGIC RECOVERY

Hydrologic recovery, or the return to the pre-disturbance hydrologic regime, involves complex processes that are difficult to measure or estimate at stand and watershed scales. Recovery is driven primarily by the return of the canopy (i.e., leaf surface area) to pre-disturbance levels. As the forest canopy re-establishes, increases in water yield and peak flows following harvesting decrease with time. The timing and rate of recovery depends on site- and season-specific hydrologic processes, the species that occupy the site, stocking density, site productivity, the rate of forest regrowth, and the cumulative effects of varying levels of recovery across the watershed. Where streamflow is dominated by radiation snowmelt that produces a single annual peak flow event, the reduction in snow accumulation and melt caused by forest regrowth can be the largest single influence on hydrologic recovery. In coastal watersheds, where multiple streamflow peaks occur as a result of rain or a combination of rain, rain-on-snow, and snowmelt events, rainfall interception and snow interception influence hydrologic recovery. In Canada, most hydrologic recovery research has focussed on stand-scale processes, primarily interception, which represents the recovery of the canopy or leaf area. At the watershed scale, cumulative recovery at the watershed outlet or other point of interest along the stream channel (such as a community water system intake) depends on the integrated effect of changes in all components of the water balance, both surface and subsurface, over time and space. Watershed-scale recovery has been studied at several long-term research sites in the United States. In British Columbia, information at the watershed scale will become available as harvested areas in long-term watershed experiments regrow and as areas are monitored for mountain pine beetle recovery. Most long-term field experiments are limited in area, so the results are extrapolated to larger watersheds through hydrologic modelling.

At the stand scale, changes in snow accumulation and melt caused by forest regrowth were investigated in coastal British Columbia by Hudson (2000) and in the southern interior of the province by Winkler

(2001) and Winkler et al. (2005). Hudson (2000) used tree height as a predictor variable and space-for-time substitution to develop snow recovery curves for the coast, whereas Winkler (2001) and Winkler et al. (2005) compared snow accumulation and melt in clearcuts and in young and mature stands to infer recovery rates in the southern interior. Hudson (2000) found that maximum SWE recovery was 50% in 4 m tall mixed subalpine fir, western hemlock, and cedar stands with 20% crown closure, and 75% in 8 m tall stands with 50% crown closure. The author suggested that complete recovery would occur when trees were 20 m tall and crown closure exceeded 95%. Hudson and Horel (2007) subsequently incorporated changes in rainfall interception in young stands into their estimates of recovery, which also considered differences in elevation, precipitation regimes, and storm types. The authors suggested that for a stand at 550 m elevation, recovery during a 35-mm rain-on-snow event was approximately 65% in 10 m tall stands and 90% in 20 m tall stands. Using a mature mixed pine, spruce, and subalpine fir stand to represent the "undisturbed" hydrologic condition, Winkler (2001) and Winkler et al. (2005) found that maximum SWE in a 4.5 m tall lodgepole pine stand with 28% crown closure was 43% recovered and that average ablation rates over the melt season were 29% recovered. Snow accumulation recovery (SAR) was calculated using measured SWE as follows, and ablation recovery was measured by replacing SWE with ablation rate.

$$\mathrm{SAR}\ (\%) = \left\{1 - \left[\frac{\mathrm{SWE}_{\mathrm{open}} - \mathrm{SWE}_{\mathrm{regen}}}{\mathrm{SWE}_{\mathrm{open}} - \mathrm{SWE}_{\mathrm{mature}}}\right]\right\} \times 100 \qquad (1)$$

The recovery of snow ablation rates is generally slower than that of snow accumulation rates. In the boreal forest of northern Ontario, maximum SWE recovered to 80% of that in the mature forest within approximately 15 years after clearcutting, when the regenerated stand was about 7 m tall and had a canopy density of 40% (Buttle et al. 2005). Average ablation rate recovery was not expected to reach 50% until trees were 16 m tall or the same height as the unlogged stands. Melt rates in very young regenerating stands exceeded those in clearcuts, possibly as a result of increased longwave radiation and sensible and latent heat fluxes to the snow surface as young trees and other vegetation began to extend above the snowpack (Buttle et al. 2005). Talbot and Plamondon (2002) found that ablation recovery in a clearcut balsam fir stand in southern Quebec was 50% at 15 years after clearcutting, when young trees were 4 m tall, stand crown closure was 50%, light interception was 55%, and basal area was 16 m^2/ha. In Montana, a lodgepole pine stand of 10–14 m tall trees with a basal area of 17 m^2/ha achieved a 41% and 48% recovery of maximum SWE and ablation, respectively, compared to a mature stand of predominantly subalpine fir (Hardy and Hansen-Bristow 1990).

In general, results of snow surveys in pine stands in the southern interior of British Columbia suggest that a 6% reduction in maximum SWE can be expected for approximately every 10% increase in crown closure (Winkler and Roach 2005). Moore and McCaughey (1997) found a 6.4% decrease in maximum SWE per 10% increase in canopy density. The authors also noted that the relationship between canopy density and SWE was stronger for spruce-fir than pine stands, which suggests that other forest structure variables (e.g., crown shape) also influence SWE.

Regenerating stands also slowly increase rainfall interception losses at a site. At Upper Penticton Creek, rainfall interception losses in a 25-year-old, well-stocked, 4 m tall lodgepole pine stand were similar to those in the adjacent mature forest. In coastal ecosystems, however, where mature canopy biomass is high, interception loss in 20- to 25-year-old juvenile stands was about one-half that of a mature stand (Spittlehouse 1998a, 1998b; see also Table 6.1 in Chapter 6, "Hydrologic Processes and Watershed Response").

Within 5 years of disturbance, the root systems of young trees, grasses, and shrubs can effectively withdraw moisture from most of the soil profile. Consequently, during the summer, soil water content and evaporation on these sites is similar to that of a mature forest (Adams et al. 1991; Elliott et al. 1998; Spittlehouse 2002). Similar results were found in boreal forest sites in Saskatchewan (Amiro et al. 2006) and black spruce stands in Alaska (Chambers and Chapin 2003). Further regrowth increases interception losses and increases transpiration to levels where water losses can exceed those of some types of mature forests (Vertessy et al. 1993). In the Oregon Cascades, summer soil moisture content in a regenerated clearcut was 20 mm less than in a forested site 5 years after harvest (Adams et al. 1991). The declining surplus and shift to a slight deficit was attributed to the rapid increase in plant growth, primarily by fireweed, vine maple, and snowbrush.

Figure 7.3 shows model estimates of the mean annual water balance over 3 years in a mature lodgepole pine stand, a recent clearcut, and juvenile lodgepole pine stands after 5, 10, and 25 years of regeneration within the Upper Penticton Creek Experimental Watershed in south-central British Columbia. For this forest type and region, the water balances for a well-stocked 25-year-old stand and the mature uncut forest are similar, whereas those for 5- and 10-year-old stands resemble a clearcut. Figure 7.3 also illustrates an interesting pattern in the water balance throughout the various stages of recovery. The sum of evaporation from the soil and transpiration from understorey plants and overstorey trees remains relatively constant under all forest scenarios, and the decrease in drainage from clearcut to mature stand conditions appears to be a consequence of increased interception loss with increasing stand age (i.e., increasing canopy cover). As noted earlier in this chapter (and in Chapter 6, Table 6.1), the recovery of rainfall interception in coastal forests is slower than in interior forests because of the additional time needed to develop the high canopy biomass characteristic of coastal old-growth forests. Jassal et al. (2009) found that annual evaporation from a 15-year-old and a 60-year-old stand of coastal Douglas-fir stand was similar, which suggests that the higher transpiration rates of younger trees and brush somewhat compensated for lower rainfall interception.

Estimates of post-treatment streamflow recovery rates vary among studies and depend partly on the method of analysis. Thomas and Megahan (1998) found that in the western Cascades of Oregon, treatment effects on peak flows decreased through time, persisting for more than 20 years on the clearcut HJA-1 watershed, but for only 10 years on the patch cut and roaded HJA-3 watershed. Jones and Grant (1996) suggested that the slower recovery of HJA-1 compared to HJA-3 was likely caused by the slower regeneration of conifers. Jones (2000) reported recovery times of at least 10 years in all cases where a significant treatment effect was evident and at least 30 years in two cases (HJA-1 and HJA-3). Hydrologic recovery for annual water yield in Needle Branch, a clearcut watershed in the Alsea watershed, took 31 years to return to pre-treatment water yields (Stednick 2008). At the Coyote Creek watersheds in the southern Oregon Cascades, effects of clearcutting appeared to decline over the 5-year, post-harvest period (Harr et al. 1979). An exponential model fitted to the data suggested that the effect would decrease by about 60% over the first 10 years and by 95% after 30 years. For practical applications, recovery curves can be developed that relate the percent recovery toward pre-logging conditions to some index of stand development, typically canopy height or basal area. These are surrogates for leaf area, which is more difficult to measure and not routinely available.

In British Columbia, estimates of recovery are used to calculate equivalent clearcut area (ECA), which is a commonly used index of the extent of forest disturbance and regrowth in a watershed. Equivalent clearcut area was originally used in provincial watershed assessment procedures as an indicator of potential hydrologic change due to forest harvesting (B.C. Ministry of Forests 1999). In these procedures, data from government and industrial forest inventory databases (e.g., the date of harvest, extent, and distribution of logging and roads in a watershed, and forest regrowth) were used to infer the potential for past and new developments to affect peak flows, channels, aquatic habitat, water supplies, and communities. It was assumed that the greater the proportion of the total watershed area disturbed, the greater the potential for hydrologic change. Additional weight was given to roads and logging near stream channels as well as harvesting in zones that contributed snowmelt at the time of peak flow.

The potential for harvesting-related changes in streamflow, particularly snowmelt-dominated spring peak flows, was one of the key hydrologic changes of concern. Information on quantitative relationships between forest cover removal or regrowth and streamflow at the watershed scale in British Columbia is not widely available; therefore, changes in streamflow generation processes at the stand scale were assumed to result in changes at the watershed scale if the area affected was of sufficient size. In completely forested watersheds, the potential for hydrologic change was assumed to be large if much of the forest was removed. In a watershed with limited distribution of forest cover—for example, where a large portion of the watershed was alpine—even if most of the forest was altered, little change in streamflow would be expected since flows would originate mainly in the non-forested area. It was further assumed that forest regrowth, or recovery, would mitigate the effects of past development. Deactivation of roads was also considered.

The ECA of a clearcut is derived by reducing the total area cut by recovery, which is estimated from relationships between snow accumulation and melt and crown closure (Winkler and Roach 2005) or tree height (Hudson and Horel 2007). The cumulative ECAs for all openings are summed to provide an ECA for the entire watershed. High ECAs would trigger a detailed field assessment of watershed condition. Lewis and Huggard (2010) combined snow and stand information from the coast and the interior, and from research in similar forest types in Montana, to derive a single relationship between ECA and tree height.

Several sources of uncertainty complicate the application of recovery curves and simple indices of hydrologic change, such as ECA. The changing hydrologic function of a growing forest stand depends on the tree and understorey species present, tree spacing, climatic characteristics, and site topography. Consequently, estimates of recovery at one site may not apply to another. Standard forest inventory metrics, such as tree height or stocking, may not best represent the influence of forests on hydrologic processes, including interception, evaporation, soil moisture, and water yield. These processes are generally better correlated with leaf area, canopy density, or basal area (Pomeroy et al. 2002; Talbot and Plamondon 2002; Teti 2003). The complex interactions between forests and hydrologic processes are highly dependent on the weather. For example, in the southern interior of British Columbia, attributes of the forest canopy account for some of the variability in hydrologic processes, such as snow accumulation and melt, among stands, but year-to-year differences in snowfall patterns account for the largest proportion of the variability (Winkler et al. 2005; Winkler and Moore 2006). Therefore, estimates of recovery derived from measurements made over a relatively small number of seasons may not adequately represent the range of weather conditions expected within a given watershed, and at best, represent average conditions. The greatest uncertainty lies in the application of stand-scale recovery estimates and ECA indices to the evaluation of hydrologic change at the watershed scale. Linkages between stands, hillslopes, and entire watersheds are complex and vary with the weather and the watershed; consequently, these linkages have not been quantified. A 50% recovery in snow accumulation or melt may not represent a 50% recovery in peak flow magnitudes. The ECA index provides only an indication of potential hydrologic change based on the extent of disturbance. It should not be used as a substitute for professional analyses and field assessments.

SUMMARY

The effects of forest disturbance on hydrologic processes and the generation of streamflow are highly variable and are influenced not only by the disturbance, but also by the weather and the biophysical characteristics, of the watershed. This chapter has described how forest harvesting and natural disturbance-related changes in fundamental hydrologic processes, including precipitation, interception, evaporation, and soil moisture, can influence hillslope flow, groundwater, and streamflow. The details of past and ongoing research in British Columbia and elsewhere can be found in the references cited. New initiatives are building on this foundation by linking stand-scale processes to watershed response; by conducting modelling over greater temporal and spatial scales, with scenarios of extensive harvesting and changing climates; and by conducting field research to quantify the effects of natural disturbance on hydrology and hydrologic recovery.

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CHAPTER 8

Hillslope Processes

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INTRODUCTION

Understanding and managing landslides and soil erosion can be challenging in British Columbia. Hillslope processes occur on a complex template, composed of a wide array of topographies, climates, geologies, and ecosystems (Chapter 2, "Physiography of British Columbia").

British Columbia is a diverse province with mountain ranges and incised plateaus. The mountainous topography is responsible for creating a variety of climates. On the Coast, mountain slopes support rainforests under a maritime climate, and at higher elevations have an extensive cover of snow and ice. The interior mountains are drier and colder in winter. In the southern Interior, some valleys are semi-arid grasslands. In the northern Interior, mountain ranges and plateaus have sporadic permafrost.

Underlying these climates and topographies is a diverse geology. Bedrock types and surficial materials vary greatly throughout the province (Chapter 2, "Physiography of British Columbia"). In general though, bedrock types range from flat-lying sedimentary rock in the northeast, to faulted and folded sedimentary rocks in the Rocky Mountains, extrusive volcanics in the central Interior, and igneous intrusive rock on the Coast. Surficial sediments are eroded from this bedrock and re-deposited by glaciers, water, wind, and mass movement. Some sediments are deposited directly by glaciers (till), some settle out in running water (fluvial), in lakes (lacustrine/glaciolacustrine), or in seas (marine/glaciomarine), and others are deposited by wind (eolian) or by landslides (colluvium).

This chapter is divided into three parts. The first describes landslides and landslide processes, the second explains soil erosion and related processes, and the third examines the reading and interpretation of the landscape. However, several important geomorphic processes are not included in this chapter. Although snow avalanches are important hazards in western Canada, with thousands occurring each year and claiming more than 10 lives annually, a treatment of these avalanche processes is beyond the scope of this chapter. Snow avalanches are further discussed in Chapter 9 ("Forest Management Effects on Hillslope Processes"), and an excellent discussion of snow avalanche processes is given in the Ministry of Forests and Range's Land Management Handbook 55 (Weir 2002). Periglacial processes, such as nivation and solifluction, are also ubiquitous in British Columbia's mountains. Although these processes may play a role in the priming of alpine areas for debris flows and rock slides, a large body of literature already exists on periglacial geomorphology.

Landslides

WHAT IS A LANDSLIDE?

Landslide is a generic term used to describe the downward movement of soil, rock, or other earth material under the influence of gravity. Landslides occur in several material types (earth, debris, rock, organic materials), move at varying rates (millimetres per year to tens of metres per second), and can involve different styles of movement (fall, flow, slide, spread). Landslides can evolve through various stages of activity ranging from relict to dormant to active. They can be retrogressive or progressive, advancing or enlarging, move along planar or curved surfaces, and be shallow or deep. In addition, landslides are often complex, involving more than one type of material and style of movement.

The various types of landslides all behave differently and thus have different associated hazards. The management of landslides and landslide-prone terrain necessitated a classification scheme to enable intelligent and efficient communication. Several classifications are in use today. The main classifications used in British Columbia are those of Cruden and Varnes (1996) and Hungr et al. (2001). It is useful to indicate which classification system is used when describing a landslide. British Columbia landslide types are also discussed in Chatwin et al. (1991) and in B.C. Ministry of Transportation and Highways (1996).

Material Types

The terms used for landslide materials should describe the displaced material as it was before the landslide occurred. For example, the term “rock” in “rock slide” describes the in-place, intact bedrock before displacement and not the rubble or debris in the landslide’s accumulation (deposition) zone.

Rock describes landslides resulting from the failure of intact, in-place bedrock.

Debris refers to generally loose, unsorted material, typically derived from colluvium, till, glaciofluvial sands and gravels, and anthropogenic materials such as mine waste and embankment fills. Debris typically consists of a mix of pebbles, cobbles, and boulders in a matrix of sand, silt, and clay. Debris can contain significant volumes of woody material, including branches, tree stems, stumps, and finely ground or decomposed organic soil. Debris is usually non-plastic to weakly plastic (Hungr et al. 2001).

Earth refers to generally unsorted, cohesive, plastic materials with a low percentage of sand-sized and larger fragments. Cruden and Varnes (1996) stated that 80% of the particles in a material referred to as “earth” should be smaller than 2 mm. Hungr et al. (2001) proposed that the term “earth” should apply to materials with a consistency closer to the plastic limit than the liquid limit but did not specify limits for coarse fragment content.

Mud, a term favoured by Hungr et al. (2001), refers to liquid or semi-liquid, clay-rich materials (or “earth” with high water content). Such materials include clay-rich, sensitive, marine and glaciomarine sediments occurring in some coastal areas, lake sediments, and some volcanic soils.

Organic materials include both saturated lowland peats and thick upland forest humus forms known as “Folisols” (Soil Classification Working Group 1998). These are mainly non-mineral materials resulting from biogenic accumulation.

Movement Type

Topple refers to the forward rotational movement of a mass of soil or rock outward from a slope about a point or axis below the centre of gravity of the displaced mass. Toppling can result from forces exerted by surficial materials or bedrock upslope of the displaced mass and sometimes by ice or water in cracks in the mass (Cruden and Varnes 1996). Because the moving mass is still attached to its base, toppling is considered by some to be a precursor to a landslide, rather than a true landslide movement. Topples are usually precursors to falls (Figure 8.1).

Fall refers to detachment of soil or rock with little or no shear displacement (frictional movement between two surfaces) and descent mainly through the air by falling, bouncing, and rolling (Cruden and Varnes 1996) (Figure 8.2).

Slide involves movement of a relatively intact soil or rock mass along a surface of rupture or along one or several discrete shear surfaces. These surfaces often form characteristic slickensides (Figure 8.3), or polished shear surfaces akin to those in fault zones.

FIGURE 8.1 *Flexural topple in limestone in the Rocky Mountains northeast of Prince George. (Photo: M. Geertsema)*

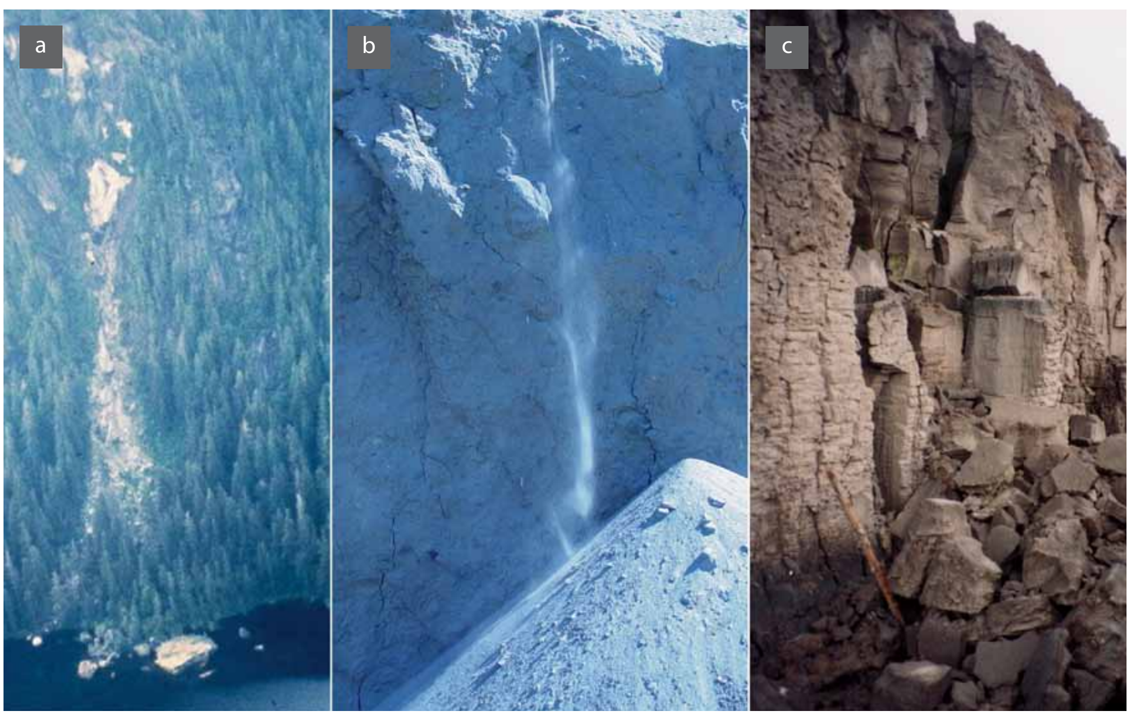

FIGURE 8.2 *Examples of (a) rock fall, (b) sand fall, and (c) earth topple and fall near the British Columbia towns of Prince Rupert, Williams Lake, and Terrace, respectively. (Photos: M. Geertsema)*

FIGURE 8.3 *Slides, such as this earth slide near Fort St. John, have discrete shear surfaces. The rupture surface here is polished and is known as a slickenside. (Photos: M. Geertsema)*

Sliding can be translational (the surface of rupture is sub-planar or undulating) or rotational (the surface of rupture is curved and concave and is sometimes referred to as a "slump"), or intermediate or a combination of the two (rotational-translational). The displaced mass initially remains intact, but as the detached material slides further, the displaced mass tends to break up. When the disrupted mass begins to flow, the landslide is no longer termed a "slide" (Cruden and Varnes 1996) but is instead referred to as a "complex slide-flow."

Flow involves the movement of a mass with significant internal distortion or disruption (Figure 8.4). Flows have multiple dissipated shear surfaces distributed throughout the mass. A *flow-like* landslide will often begin as a slide moving along a rupture surface, but then continues as a flow down unconfined surfaces (e.g., debris avalanche) and (or) confined channels (e.g., debris flow), and can travel for long distances. In granular materials, the initial sliding movement leads very quickly to complete disintegration, producing flow-like motion characterized by nearly complete remoulding of the moving mass (Cruden and Varnes 1996; Hungr et al. 2001). Flow-like behaviour often depends on water entrainment after the initial failure. Hungr et al. (2001) restricted use of the term "flow" to channelized movement.

Avalanche is often used to describe an unconfined flow, and is especially used for rapid landslides that have long run-out distances (Figure 8.5). Debris avalanches and rock avalanches are commonly occurring types.

Spread is extension of a cohesive soil or rock mass and subsidence of the fractured mass of cohesive material into a softer, underlying material (Cruden and Varnes 1996). Spreads may result from the liquefaction or flow and extrusion of soft or weak materials underlying competent materials. An example of bedrock spread is shown in Figure 8.6.

Complex landslides involve more than one type of material, and consequently involve more than one type of movement. Large rock slides can change behaviour when the displaced rock mass moves downslope and impacts other materials. Three common scenarios include rock slides that transform into earth flows, debris flows, or debris avalanches (Figure 8.7).

FIGURE 8.4 *Mud flows following a rainstorm in the Peace River area. Sliding in the source areas gives way to flowing in the narrow transport zones. The landslides shown are about 100 m long. (Photo: M. Geertsema)*

FIGURE 8.5 *Rock avalanche near Chisca River west of Fort Nelson. (Photo: M. Geertsema)*

FIGURE 8.6 *Rock spread west of Fort Nelson. The ridges are transverse to the direction of movement. (Photo: M. Geertsema)*

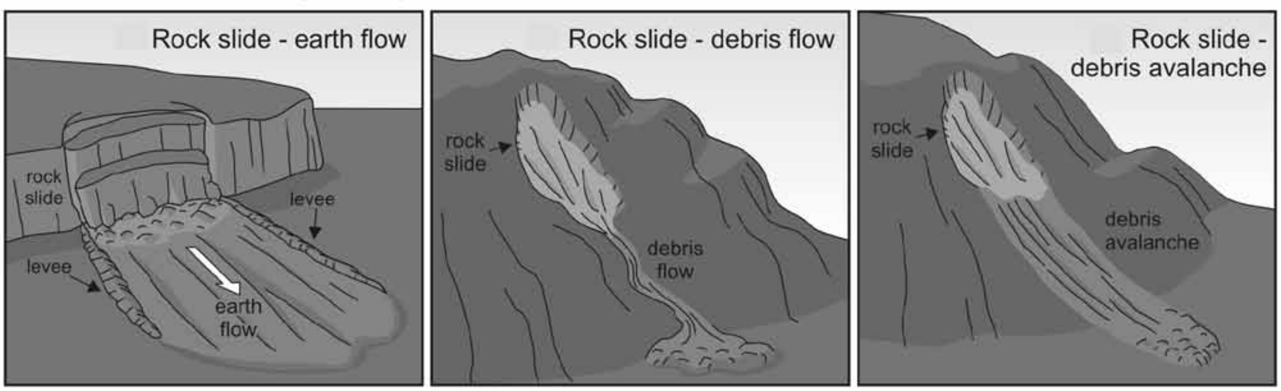

FIGURE 8.7 *Illustrations showing various rock slide interactions upon impact with soil. (Modified from Geertsema et al. 2006a)*

WHAT CAUSES LANDSLIDES?

It is important to distinguish between the *causes* and *triggers* of landslides. In the broad sense, *cause* is the sum of factors that renders a slope unstable; *trigger* is the final, possibly small, event that leads to the ultimate point of slope failure. Causal factors of instability include preconditions such as weathering, glacial erosion, river erosion, geologic structure, tectonics, changes in climate, wildfire, deforestation, and road construction. The trigger may be an earthquake, increase in surface loading (surcharge), intense or prolonged rainfall, or river undercutting of a slope.

If an increase in landslides in a specific area is evident, it is useful to look for changes in potential causal factors. Some disturbances result in large-scale changes to forested watersheds. Wildfires (Cannon 2001; Sanborn et al. 2006) and logging practices

(Schwab 1983; Rood 1984; Jakob 2000; Guthrie 2002) are known to cause debris slides and flows. Changes in climate can also be important (Chapter 19, "Climate Change Effects on Watershed Processes in British Columbia"). Global climate warming may be affecting landslide rates in mountain ranges around the world (see "The Influence of Climate on Slope Stability" below).

Landslide Triggers

Numerous inherent (internal) and variable (external) forces interact to control or modify hillslope stability. Principal among these factors are changes in strength–stress relationships on a slope over geologic time. The importance or effectiveness of a specific factor is determined by the local geologic and hydrologic conditions. The factors modifying slope stability are discussed in numerous publications. Jakob and Hungr (editors, 2005) have provided an in-depth presentation on debris-flow phenomena. *Landslide Investigation and Mitigation* (Turner and Schuster [editors] 1996) provides a detailed discussion on large and complex landslides. In British Columbia, Chatwin et al. (1991) have discussed landslides in a forest management context. Sidle and Ochiai (2006) have also provided an overview of management effects on landslides within a global context.

A landslide trigger is the external force that acts on a slope to increase shear stress, which results in a landslide. The trigger attributed to the landslide is the driving force that has set off the landslide on a predisposed unstable slope. The triggers discussed within a British Columbia context are earthquakes and climate (meteorological conditions, intense rainfall and rapid snowmelt, long-term climate trends, frost wedging, and thawing permafrost). Vallance (2005) offered an in-depth presentation of direct and indirect effects of volcanic eruptions, a trigger of large debris flows (lahars), and hence these are not discussed here. However, massive landslides have occurred in the vicinity of eruptive centres, particularly within the Garibaldi volcanic belt (e.g., Devastation Glacier landslide [Patton 1976]; Rubble Creek landslide [Moore and Mathews 1978]; and Dusty Creek, Mount Cayley [Clague and Souther 1982]).

Earthquakes

Ground shaking caused by earthquakes has triggered catastrophic landslides around the world. Earthquakes increase stress, increase temporary pore-water pressures, and decrease soil strength. The overall effect of ground motion depends on the topographic and geological setting. Generally, an earthquake greater than Magnitude 6 is required to trigger landslides (G. Rogers, Pacific Geoscience Centre, Sidney B.C., pers. comm., 2001). The Magnitude 9.2 earthquake that occurred March 28, 1964, in Alaska's Prince William Sound caused landslides in marine sediments and rock slides–avalanches on mountain slopes. Near Fairbanks, Alaska, the large inland Magnitude 7.9 Denali Fault earthquake of November 3, 2002, preceded by a Magnitude 6.7 event on October 23, 2002, triggered many rock falls, snow avalanches, and a large rock avalanche. British Columbia's short recorded history and sparsely populated area provide only a few anecdotal reports of earthquake-triggered landslides. Vancouver Island's largest historic earthquake (Magnitude 7.3) occurred on June 23, 1946, in the vicinity of Forbidden Plateau, central Vancouver Island, and caused numerous landslides including submarine landslides (Hodgson 1946; Mathews 1979; Rogers 1980). The largest historic earthquake recorded in Canada occurred on August 22, 1949, off the west coast of Graham Island, Haida Gwaii (formerly known as the Queen Charlotte Islands). This Magnitude 8.1 earthquake was believed to have caused extensive rock fall, debris slides, and debris avalanches. Geologists with the Geological Survey of Canada working on the northern end of Graham Island recorded that they could not stand up during the event, which lasted 5 minutes. The *Prince Rupert Daily News*, on August 28 and 30, 1949, provided several eyewitness accounts: a fisheries patrol boat anchored at Lockeport, Kulunkwoi Bay, reported the crashing of rocks and timber downslope, and the next day many new slide scars were visible; a logging camp on the north shore of Cumshewa Inlet received heavy damage, and new slides were reported all along Cumshewa Inlet. Meteorological conditions before the 1949 earthquake likely aggravated the occurrence of slope failures. The field observations of Alley and Thomson (1978) noted that many large landslide areas on the northern part of Graham Island were characterized by forest cover of similar age and species, suggesting that rock slides and debris avalanches may have been triggered by the 1949 earthquake or a large prehistoric earthquake. Schwab (1997, 1998) and Martin et al. (2002) indicated that the majority of the debris avalanches and debris flows on the north coast and Haida Gwaii were triggered by large precipitation events. Nevertheless, large 100- to 300-year-old landslides are visible on Haida Gwaii. Likewise, many pre-historic landslides were identified on Vancouver Island (VanDine and Evans 1992). These large landslides on Haida Gwaii, Vancouver Island, and the mainland were likely caused by a combination of factors with a probable earthquake trigger.

The influence of climate on slope stability

Precipitation

In a simple sense, the occurrence of landslides will increase in relation to the amount and duration of precipitation; however, slope factors may be more complicated and thus preclude a simple analysis. Some landslides respond rapidly to rainfall, others have delayed responses. In any case, antecedent conditions have been shown to be very important. In general, soils must become saturated, allowing the build-up of pore pressures. Threshold pore pressures develop more rapidly in shallow materials than in deeper materials. Landslides that respond rapidly to changes in precipitation include shallow debris slides, debris flows, and some rock falls. Landslides that have delayed responses to changes in precipitation and temperature are usually the larger, deeper-seated rock slides, earth slides, and earth flows.

Rapid response landslides

In mountainous areas of British Columbia, shallow debris slides, debris avalanches, and debris flows are typically associated with heavy rainfall or rapid snowmelt, provided threshold conditions are met. Precipitation attributes that influence pore water pressure are antecedent storm precipitation, storm duration, total rainfall, and precipitation intensity. Studies from many parts of the world document the response of rainstorms of different intensity and duration for various geologic settings (Wieczorek and Glade 2005). In British Columbia, only a few studies attempt to link type of storms, antecedent meteorological conditions, and precipitation intensity and duration as triggers of shallow hillslope failures. However, most landside investigations use probable climatic triggers to describe the event. For example, the 1978 storm in Rennell Sound, Haida Gwaii, triggered 264 debris avalanches and debris flows (Schwab 1983). A meteorological station located at sea level in the centre of a study area recorded the cumulative precipitation over a 3-day period with two intense 12-hour periods of 120 and 110 mm of rainfall. Unfortunately, rainfall at higher elevation was probably underestimated and the time of the many slope failures can only be approximated. The climatological station located at Rennell Sound was relatively close to the debris flow activity; however, the general problem with landslide studies throughout British Columbia is a sparse climatological network located at valley-bottom sites that are not representative of weather at the location of landslide activity. Satellite images, combined with weather radar and synoptic radiosonde,[1] provide a record of upper air masses and synoptic climatology and may aid future studies in areas of sparse climatic stations (Egginton 2005).

Following devastating debris flows in the early 1980s, Church and Miles (1987) undertook a detailed analysis of meteorological antecedents for case studies in the Howe Sound area and eastern Fraser Valley. A remarkably wet period in the early 1980s was the best correlate found for the increased debris flow activity. The described meteorological conditions for debris-flow–triggering storms based on regional climatological stations show no exceptional events. Eyewitness reports, however, suggest that locally intense cells of precipitation, not recorded in the regional rain gauge network, are likely important. Such locally intense precipitation is sometimes generated by convective cells embedded in a frontal storm. Free convection of warm air is enhanced by the impingement of the air stream and rapid forced lifting on mountain slopes. Church and Miles (1987) concluded that the traditional hydrometeorological indices, based on routine meteorological measurements, are unlikely to provide consistent indications of the likelihood for debris flows to occur.

On Haida Gwaii, Hogan and Schwab (1991) studied rainfall characteristics before and during verified debris flow events. They compared the temporal frequency of slope failures to antecedent precipitation trends for time scales ranging from years to days. Over the longer time scale, a positive correlation was found between annual moisture conditions and the reported frequency of hillslope failure; however, annual moisture was less significant than precipitation over shorter periods (months or days) preceding a slope failure. On the shorter time scale, only the months immediately preceding slope failure were important in conditioning the hillslope to failure. The most important situation leading to a high

1 Upper Air Observing System measures temperature, pressure, humidity, wind speed, and wind direction.

frequency of slope failures was continuously wet weather before the landslide-triggering storm. Precipitation values required to trigger slope failures are regularly exceeded during fall and winter months on Haida Gwaii and British Columbia's north coast.

Jakob and Weatherly (2003) studied landslide-triggering storms on the North Shore Mountains of British Columbia's south coast. Data were collected for 18 storms that triggered landslides and 18 storms that did not. Data included antecedent rainfall, rainfall intensity-duration during storm events, and discharge exceedance of a small watershed. Through discriminant function analysis, the authors separated landslide-triggering storms from storms that did not trigger landslides. This analysis indicated that antecedent conditions are more important in developing a threshold for landslide initiation than short-term rainfall intensities alone, which fail to include hydrologic effects such as snowmelt and antecedent rainfall. The 28-day antecedent precipitation amounts are critical in bringing soil moisture to saturation levels conducive to landslides. Jakob and Weatherly (2003) concluded that considerable amounts of precipitation are required to saturate forest soils to the point where a landslide is likely, once an intensity threshold is exceeded. They also found the 6-hour rainfall intensity during a storm to be significant because it causes already-wet soils to generate positive pore-water pressures, which further decrease the effective strength of the soil.

Schwab (1997, 1998) used dendrochronology techniques to determine the ages (back to the early 1800s) of large debris slides, debris avalanches, and debris flows in selected areas along British Columbia's north coast and on Haida Gwaii. A catalogue of storm information compiled by Septer and Schwab (1995) as part of the study was used to verify dates determined through tree-ring analysis of increment cores and impact scars on trees. A conclusion from this work is that most shallow landslides occur during a few major storms. Data from Graham Island, Haida Gwaii, and from the Prince Rupert area on the British Columbia north coast (Figure 8.8), indicate that six storms over the past 150 years transported 76% of the volume of landslide debris: 9.5% (1875), 14% (1891), 30.9% (1917), 6.5% (1935), 6.4% (1957), and 9.1% (1978). Interestingly, different watersheds (study areas) did not experience the same magnitude of landslide activity for the various storms; for example, at Beresford Creek on the northwest corner of Graham Island, 36.2% of the volume transported by landslides occurred in 1917, but no landslides were

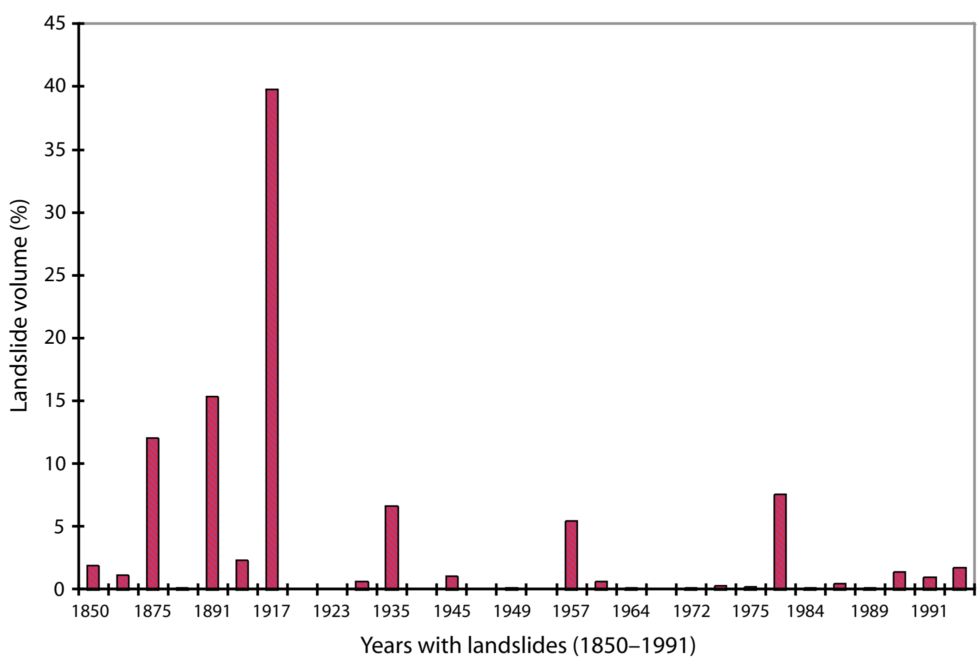

FIGURE 8.8 *Percent volume transported by debris slides, debris avalanches, and debris flows on forested terrain at Rennell Sound and Pivot Mountain, Haida Gwaii. Most of the sediment was mobilized during a major storm in 1917. Landslides occurring before 1875 are probably under-represented because only large events were identified on aerial photographs.*

experienced in the watershed in 1978. Overall, the data suggest that British Columbia's north coast has yet to re-experience meteorological conditions similar to the event of October 28 to November 19, 1917.

Most precipitation on the British Columbia north coast is associated with frontal systems that pass across the region from the Pacific Ocean (Chapter 3, "Weather and Climate"). Loci of warm-wet conditions shift north or south depending on the prevailing storm track in the Pacific (Karanka 1986). Most landslide-triggering storms involve a warm front followed by a cold front (Jakob et al. 2006). The events appear to involve a flow in the lower part of the atmosphere, extending south to incorporate subtropical moisture. Jakob et al. (2006) found that the following combination of meteorological conditions is associated with the occurrence of debris slides, debris avalanches, and debris flows:

- warm fronts with strong SE–SW winds at the 850-mb level
- warm fronts that extend south over the Pacific Ocean, bringing much additional moisture to north coastal British Columbia
- warm fronts with high freezing levels
- a strong jet stream with airflow arcing north of the region and W–SW winds exceeding 90 knots at the 250-mb level

These meteorological conditions prevailed during landslide-triggering storms in 2003 and again in 2004. The event on October 25, 2003 triggered many large landslides, with 12 reported in the vicinity of Prince Rupert alone (see Chapter 3, "Weather and Climate," Figure 3.16, for satellite photograph showing moisture transport during the October 2003 event). Landslides on November 4, 2004 severed a natural gas pipeline serving Prince Rupert and closed Highway 16, the only highway link to the city (Geertsema et al. 2006a).

Delayed response landslides

Larger, and especially deeper, landslides tend to have delayed responses to precipitation. It takes time for water to infiltrate and saturate potential slide masses. Geertsema et al. (2006a, 2007) studied large landslides in northern British Columbia. They found that some large landslides in soil occurred after long periods of above-average precipitation. The large earth flow spread at Mink Creek (see "Flows and Slides in Sensitive Clays" below) occurred after a 10-year wet period (Figure 8.9). Many flow slides in glaciomarine sediments occurred in the Terrace area 2000–3000 years ago when the climate was wetter (Geertsema and Schwab 1997). This suggests that long-term climate trends may influence these deeper landslides.

Individual rainstorms, rain-snowmelt events, and outburst floods can indirectly trigger landslides by increasing peak flows in streams. High water flows are known to increase bank erosion. Many landslides are triggered by bank erosion (Figure 8.10). An example is the June 2002 rain-on-snow event that resulted in a flood flow from the Gillis Mountain area in central British Columbia. The flood triggered

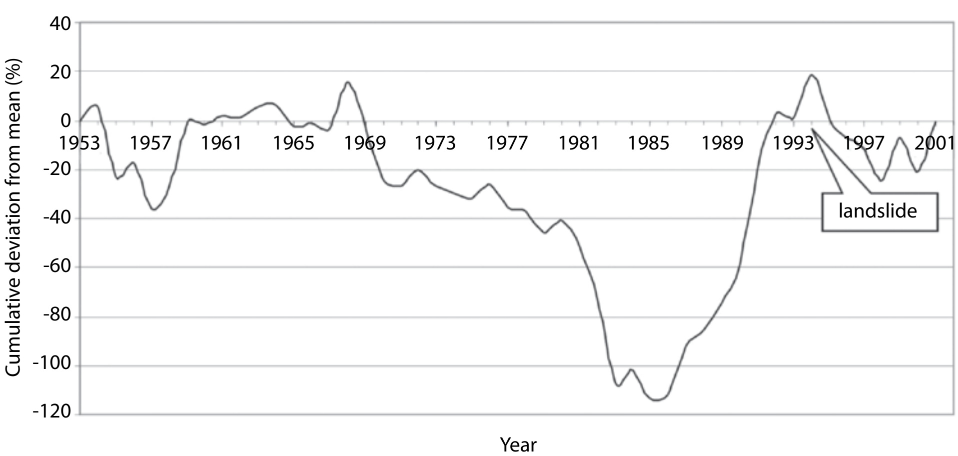

FIGURE 8.9 *Percent cumulative deviation from mean precipitation at Terrace airport from 1953 to 2002. (Modified from Geertsema et al. 2006b). The Mink Creek landslide occurred after a decade of increased precipitation.*

FIGURE 8.10 *Small flows within a larger landslide along the Prophet River shows partial reactivation of a landslide. The movements were triggered during a rainstorm in July 2001. The larger (deeper) landslide appears to be influenced by bank erosion. (Photo: M. Geertsema)*

numerous shallow debris slides, destabilized an alluvial fan, and washed out a bridge.

Temperature Landslides in mountainous regions may be especially responsive to increases in temperature. Two important factors may be the debuttressing of valley walls as a result of glacier melting and retreat, and the thawing of permafrost. Recent research has shown that 20th-century retreat of glaciers in British Columbia has debuttressed rock slopes, causing deep-seated slope deformation, joint expansion, and catastrophic failure (Evans and Clague 1994; Holm et al. 2004; Geertsema et al. 2006a). Mountain permafrost appears to be degrading, especially in the last decade, decreasing the stability of slopes (Davies et al. 2001; Harris et al. 2001: Gruber and Haeberli 2007). Recent large rock slides in the European Alps have been attributed to thaw of mountain permafrost (Dramis et al. 1995; Bottino et al. 2002; Noetzli et al. 2003), and permafrost thaw may also play a role in initiating large rock slides in northern British Columbia.

In addition to debuttressing and permafrost degradation, changes to the dynamics, amount, and seasonal distribution of snow and rain can have a significant impact on the stability of slopes. For example, the sudden and delayed melt of above-average snowpacks can lead to increased occurrence of landslide events caused by increases in pore pressure.

Whatever the cause, large natural landslides appear to be increasing in frequency in northern British Columbia. When comparing the 1970s and 1980s to the last two decades, Geertsema et al. (2007) reported a near tripling of rock slides and a 2.5-fold increase in large soil landslides. Rock slides in particular have occurred in years of above-average temperature and at least one of the rock slides was confirmed to have permafrost in its main scarp. Thawing permafrost could contribute to the destabilization of slopes.

Large rock slides have happened in recently deglaciated areas as well as in areas that have been ice-free for most of the Holocene. Half of the rock slides in northern British Columbia reported by Geertsema et al. (2006a) occurred on steep rock walls above glaciers. They suggested that glacier thinning under a warming climate following the Little Ice Age had destabilized many slopes through the debuttressing of steep rock walls.

In British Columbia, the Pacific Decadal Oscillation (PDO) strongly influences temperature and, to a lesser extent, precipitation patterns in 20–30 year cycles (Egginton 2005; Chapter 3, "Weather and Climate"). The last decade of the current warm phase of the PDO, which started in 1977, was a period of frequent rock slides. We need to extend our comparison of landslide frequency and the PDO beyond the current phase into the past and also to investigate the link, if any, between landslides events and El Niño–Southern Oscillation.

COMMON LANDSLIDE TYPES IN BRITISH COLUMBIA

Landslides are common in mountain regions around the world, and British Columbia is no exception. The varied geology, topography, and climate result in several landslide types. Shallow debris slides and flows are especially common on the coast. Rapid flows and spreads (flow slides) are common in some lake sediments in northeastern British Columbia and in glaciomarine sediments near the coast. Large rotational rock slides (slumps) are common in glacial sediments and volcanic bedrock in the Interior. Rock avalanches occur in various settings in mountainous regions of the province. Many of the larger landslides are complex, involving more than one style of movement and often more than one type of material.

Debris Slides

Debris slides are the most simple and perhaps most common landslide type in British Columbia. The landslides are shallow, occur on steep slopes, and move rapidly (Figure 8.11). Shallow debris slides typically have rapid response times to rainfall, and it is possible to establish hydrometeorological thresholds for their initiation.

Shallow debris slides occur in various earth materials including till, colluvium, and water-deposited sediments. On the north and central coast, debris slides also initiate in upland organic soils called Folisols (see "Debris slides and debris avalanches in organic soils," below). Rupture surfaces may develop along a bedrock interface, and along textural change boundaries or soil horizons, including cemented horizons, common in Podzols, and clay-enriched Bt horizons in Luvisols caused by soil-forming processes.

Natural debris slides tend to initiate at a point on a slope and flare distally downslope. Debris slides are often the feeders for other types of landslides. In some cases, debris slides will transform into debris avalanches, which are essentially open-slope debris flows. If debris slides enter gullies, they tend to become channelized debris flows.

On natural slopes, the initiation points are often seepage zones (see "Indicators of Instability" in Part III) or concave hollows where seepage water and hydrostatic pressures can build up. Debris slides also initiate where water is concentrated by improper road drainage, and in unstable fillslopes.

Case study: Debris slides and debris avalanches in organic soils

The Wet Hypermaritime and Maritime subzones of the Coastal Western Hemlock biogeoclimatic zone extend from northern Vancouver Island along the mainland of British Columbia's mid- and north

FIGURE 8.11 *Shallow debris slides near Prince Rupert. (Photo: J.W. Schwab)*

coast. The cool, wet climate results in extensive forested landscapes with folic soils that develop on the nutrient-poor plutonic rock. In the south, Folisols (Soil Classification Working Group 1988) have also developed on deep, compact, impermeable basal till. Debris slides and debris avalanches occur in these shallow organic soils (Figure 8.12).

Nagle (2000) described the physical and chemical characteristics of Folisols and the terrain in which they occur in the Prince Rupert area (Figure 8.13). Steep slopes (30–60°) with an irregular slope configuration had the greatest influence on slope stability. This geologic/hydrologic condition creates many seepage exit points or locations of high hydrostatic pressure that in turn create the potential for slope failure through a reduction in soil strength. These sites are often concave in shape and distinctively recognized by water-tolerant vegetation and wet organic soil.

Slope stability problems with organic soils appear during road construction at slope angles of 20–30°. Problems arise when vibrating heavy machinery both overloads and liquefies the saturated organic materials, triggering a landslide as the machine passes over the site. Landslides of this type were documented at Kennedy Island and Kumealon Inlet on the province's north coast and at Security Inlet and Mackenzie Sound on the mid-coast.

These sites generally appear relatively stable with no apparent landslides evident on the landscape. At some sites, signs of periodic natural creep, single-tree slump blocks, or small, scallop-shaped depressions of an old head scarp may or may not be present. The organic soils are 10–100 cm deep over bedrock (sometimes over compact till), and are commonly found in isolated concave pockets where surface and subsurface seepage concentrates (i.e., irregular or convergent slopes), but they can also occur on convex slopes and across larger, more homogeneous slopes. The most obvious sites will be the poor to very poorly drained areas where seepage or water flow is visible at the surface. Less obvious are the moderately well to imperfectly drained sites where subsurface seepage dominates (Figure 8.14). Poor to very poorly drained sites can be identified by the presence of skunk cabbage, sphagnum moss, goldthread, and devil's club.

FIGURE 8.12 *Close-up of a moderately well-drained Folisol at a landslide head scarp (organic soil situated on bedrock). (Photo: J.W. Schwab)*

FIGURE 8.13 *Debris avalanche in a Folisol near Prince Rupert. (Photo: J.W Schwab)*

FIGURE 8.14 *Debris flow initiated at a low slope angle in a poorly drained folic soil. (Photo: J.W. Schwab)*

Debris Flows

Debris flow is a form of mass movement in which a saturated slurry of soil, stones, and vegetation flows rapidly down a slope, usually in a confined channel. The terminology used to describe debris flows and related phenomena in North America has varied over the past several decades, and has therefore caused some confusion. In British Columbia and the U.S. Pacific Northwest, the term "debris torrent" was widely used (Swanston and Swanson 1976; VanDine 1985; Chatwin et al. 1991) to describe coarse-textured debris flow in steep mountain channels, and was also sometimes loosely used to refer to debris floods. The term "debris torrent" has fallen out of use and is no longer recommended. Other sources have applied the term "debris flow" to all flow-like mass movement processes, regardless of speed or degree of confinement (after Varnes 1978), including unconfined, open-slope events that are more properly classified as "debris avalanches" (Fannin and Rollerson 1993). We recommend the use of Hungr et al.'s (2001) definition: "Debris flow is a very rapid to extremely rapid flow of saturated non-plastic debris in a steep channel." The distinguishing attributes of debris flows, according to this definition, are: rapid speed (usually greater than 1 m/s), saturation, confinement in a channel, and relatively coarse texture (low clay content). The term "mud flow" is used for rapid flows of fine-textured, cohesive (clay-rich) material (defined by Hungr et al. 2001 as having a plasticity index > 5%).

Debris flows are probably one of the most important mass movement processes in the amount of geomorphic work done. In some areas of the province, rock avalanches and earth slides overwhelm sediment budgets, but overall it is the debris flow that moves most of the sediment produced by weathering, rock fall, and small landslides in steep terrain into valley bottoms. Debris flows occur at all scales, ranging from tiny flows a few centimetres deep that can be observed on any road cut during a spring thaw, to rare events that can carry a cubic kilometre or more of debris from a collapsing volcano over a hundred kilometres down a low-gradient river valley (Costa 1984; such volcanic-source debris flows are often called "lahars"). Most of the debris flows of interest in a forestry context range from a few hundred to a few tens of thousands of cubic metres.

Debris flows have a number of distinctive properties (e.g., VanDine 1985; Pierson and Costa 1987). The debris is very poorly sorted, ranging from silt and clay to large boulders, and it often includes a substantial proportion of organic material, from finely shredded wood to entire trees. A debris flow in motion resembles, and has similar fluid properties to, wet concrete pouring down a flume from a cement truck. The debris is saturated with water and flows under a fairly limited range of water contents (sediment concentration is typically greater than 80% by weight). If the water content is too low, the debris flow will stop; if it is too high, it will separate into two phases of muddy water and bedload. Large stones tend to accumulate on top of the flow and at the front (Figure 8.15). Where the channel widens, levees of boulders are often deposited along the edges of the flow (Figure 8.16). Debris flows frequently occur as a series of surges, possibly caused by the coarse front jamming behind obstructions and then releasing. In many steep mountain channels, the peak discharge of debris flows is typically 10–100 times the peak discharge of normal streamflow floods (Jakob and Jordan 2001).

In steep channels, large debris flows can reach high speeds—5–20 m/s is typical—and have great erosive power (Figure 8.17); however, on losing confinement on a fan, they move much more slowly, covering or engulfing objects, often without damaging them. They can surround automobiles without breaking windows or scratching the paint (Figure 8.18).

Coarse-textured debris flows tend to slow down and stop when the slope reaches about 15% (9°), although this critical slope varies somewhat, depending on the magnitude of the event and its water content. Fans that are formed primarily by debris flows tend to approximate this gradient (Figures 8.19, 8.20). Fine-textured debris flows (those containing more clay) can traverse more gentle slopes, as low as 2% for large volcanic mud flows (Figure 8.21). The deposits of debris flows on alluvial fans can often be recognized by levees and irregular lobes of bouldery debris (Figures 8.22, 8.23), and by unsorted layers of debris that are matrix-supported (large clasts not touching), and that contain more silt, fine sand, and wood fragments than streamflow deposits (Costa 1984; Figure 8.24).

Debris flows can originate in a number of ways (Hungr et al. 2005). The most common in the forested mountain landscapes of British Columbia is for a small landslide to enter a steep stream channel (Millard 1999). The resulting debris flow can be many times larger than the originating landslide, as debris is eroded from the channel bed and banks

FIGURE 8.15 *Bouldery debris flow deposits at Lower Arrow Lake near Burton. (Photo: P. Jordan)*

FIGURE 8.16 *Steep debris flow fan at Salal Creek near Pemberton. (Photo: P. Jordan)*

when the debris flow moves down the channel. Most open-slope debris slides lack sufficient water content to flow under their own weight. A landslide into a channel may dam the stream, and the dam soon breaks, producing a slurry with the proper water content to flow as a debris flow. Alternatively, a rapidly moving landslide may overtake and consume water flowing in the channel, transforming into a debris flow. Another common mechanism that enhances transport is that unusually high streamflow in a small, steep channel may undercut the streambanks, causing a temporary dam, which when failure occurs, results in a debris flow. Yet another mechanism (well described in the literature from Japan, where higher rainfall intensities occur than in Canada) is that unusually high streamflow in a small, steep channel may cause the channel bed itself to fail, reaching a sufficiently high sediment concentration through entrainment of streambed and bank material to transform into a debris flow (Takahashi 1981).

In western Canada, debris flows tend to be restricted to channels with drainage basins of about 1–10 km^2, which contain channels steeper than about 50% in their upper reaches, and which are nowhere less steep than about 15% (VanDine 1985). A simple

FIGURE 8.17 *Channel stripped of vegetation and alluvial sediment by a large debris flow in the Monashee Mountains near Edgewood. (Photo: P. Jordan)*

FIGURE 8.18 *Vehicle engulfed by slow-moving, distal deposit of a debris flow, which damaged a logging camp at Meager Creek near Pemberton. (Photo: P. Jordan)*

FIGURE 8.19 *Aggrading alluvial fan formed by debris flows and fluvial sedimentation in the Monashee Mountains near Edgewood; note former channel and avulsion caused by recent debris flow. (Photo: P. Jordan)*

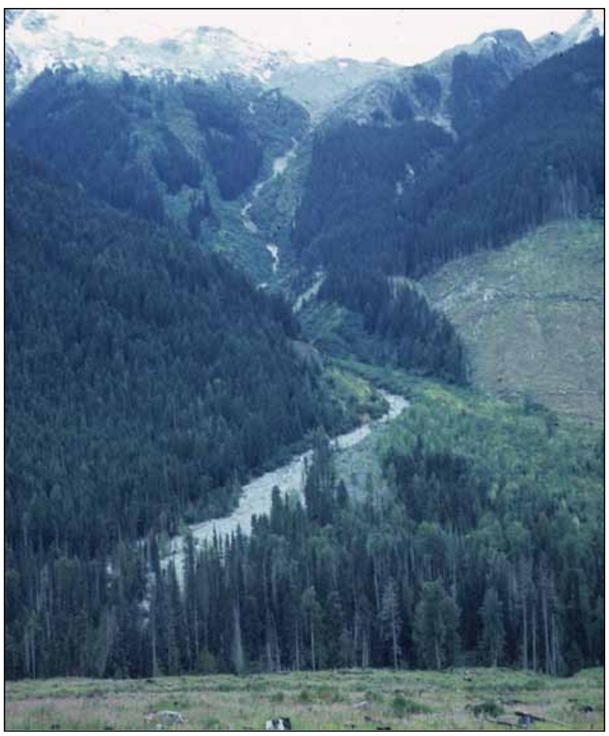

FIGURE 8.20 *Alluvial fan formed by frequent debris flows into the Ryan River near Pemberton. (Photo: P. Jordan)*

index of debris flow susceptibility is that the "Melton ruggedness number" (the ratio of elevation range to the square root of watershed area, in consistent units) be greater than 0.6 (Wilford et al. 2005b).

The time of occurrence of debris flows is quite unpredictable, although the susceptibility of a channel to debris flow hazard can be assessed fairly reliably from evidence on the fan and in the channel, and from geomorphic conditions in the watershed (Jakob and Jordan 2001). For a debris flow to be initiated, enough sediment and enough water must arrive at the same place and time in a channel that is sufficiently steep and confined to support debris flow. Although heavy rainstorms and peak snowmelt events are responsible for triggering most debris flows, many debris flow events are triggered by rainfall or snowmelt that is not unusually heavy; extreme rainstorms fail to trigger debris flows most of the time in most susceptible channels. Perhaps some random occurrence, such as a small debris slide or bank failure caused by the peak flow, is necessary to add enough sediment to trigger a debris flow. Alternatively, when released, a transient flow blockage (a fallen tree, woody debris, or snow) may increase the peak discharge to a level sufficient to entrain sediment by scouring the channel. The volume of loose, entrain-

FIGURE 8.21 *Repeated debris flows have deposited sediments near the Zymoetz River. (Photo: M. Geertsema)*

FIGURE 8.22 *Levees formed by debris flows originating in fractured sedimentary rocks at Fountain Ridge near Lillooet. (Photo: P. Jordan)*

FIGURE 8.23 *Levee formed by debris flow originating in granitic rocks at the Upper Lillooet River near Pemberton. Largest boulders are 5 m in diameter. (Photo: P. Jordan)*

FIGURE 8.24 *Debris flow deposit showing inverse grading (larger stones toward top) and matrix support (stones not touching, surrounded by fine material) at Meager Creek near Pemberton. (Photo: P. Jordan)*

able material stored in a steep channel provides some indication of the possible volume of a future debris flow, assuming all the material is mobilized (Hungr et al. 2005).

An almost unlimited supply of debris is possible in watersheds dominated by active slope processes such as rock fall (Bovis and Jakob 1999). In such watersheds, debris flows may occur annually, or more often if rainstorm events generate enough streamflow; however, in most forested watersheds or alpine areas with competent rock, debris flows are infrequent. On many steep stream channels, debris flows occur at average intervals of several decades on the Coast, to several centuries in the Interior. Following debris flow events, stream channels go through a cycle in which debris in the channel is slowly replenished by rock fall, tree fall, and soil creep (Jakob et al. 2005). After several decades or centuries, a critical hydrologic event occurs that triggers a debris flow, and the accumulated debris in the channel is carried away. Disturbances in a watershed, such as logging, road construction, or wildfire, can cause hydrologic changes that lead to greater peak streamflow, or they can cause small slope failures that enter the channel. Both these changes can greatly increase the probability of a debris flow in a channel.

Flows and Spreads in Sensitive Clays

Rapid landslides may occur on extremely low gradients in some low-lying areas of the British Columbia coast. They occur in fine-textured sediments that may be prone to sudden liquefaction. The sediments can be relatively strong in the undisturbed state, but a small vibration or load can make them flow like wet porridge. Such materials are referred to as "sensitive clays." In the extreme states, when the undisturbed strength is more than 30 times greater than the remoulded strength, such sediments are called "quick clays."

The British Columbia coastline was submerged by the weight of a large ice sheet during the last glaciation (Clague 1989). As the ice retreated, the sea migrated inland with the retreating ice fronts. Glaciomarine sediments composed of rock flour, silt, and clay minerals were deposited in the sea. In freshwater, clay particles tend to settle much more slowly than the larger silt particles. In salt water, clays and silts aggregate together forming floccules and settle in a random orientation. Negative, repulsive charges on the clay particles are neutralized by cations such as Na^{+} and Ca^{2+} in seawater. The resulting sediment has an open structure with high water content. The positive charges of the salts maintain the interparticle bonds.

As the glaciers melted, the land began to rebound isostatically, rising as much as 230 m above present-day sea level (McCuaig 2000). This exposed the glaciomarine sediments to rainfall and groundwater. Salts in the clays were gradually leached out of the sediments. Salt content would decrease from an initial 30 g/L to below 1 g/L. With a lower salt content, repulsive forces between particles increased, leaving the saturated, porous sediment prone to collapse.

An imposed load, vibration, or bank erosion can trigger collapse of the sedimentary structure in sensitive glaciomarine deposits, causing liquefaction. During liquefaction, the weight of the soil is transferred from the solids to the pore water.

Examples of landslides in sensitive clays

In preparation for widening the British Columbia highway between Kitimat and Terrace in 1962, earth fills were placed beside the highway on sensitive clays soils. When the loads failed, two enormous landslides occurred May and June. These landslides, which both flowed into Lakelse Lake, destroyed sections of the highway (Figure 8.25).

In winter 1993/94, a similar landslide flowed into Mink Creek (Figure 8.26), a tributary of Lakelse River, moving 2.5 million m^3 of mud. The landslide failed on a nearly flat slope of just over 2°. The landslide likely began with a small failure along the creek, which exposed sensitive clays further in the slope. Once the sensitive materials were exposed, the landslide retrogressed rapidly, eating its way into the slope.

Landslides of this nature often move as spreads or flows. Flows tend to occur in materials with lower shear strengths than spreads. Flows may also occur where a thick zone is sensitive. Spreads are more likely to occur where a thin zone is sensitive. In flows, ridges (if they occur) are oriented parallel to movement direction. In spreads, ridges are transverse to movement. The Mink Creek landslide was complex, displaying both spreading and flowing behaviour (Figure 8.27).

In 2003 a large flow in sensitive clays ruptured a natural gas pipeline between Terrace and Prince Rupert (Schwab et al. 2004). The landslide cut off the natural gas supply to Prince Rupert for 10 days. The landslide also dammed Khyex River.

FIGURE 8.25 *One of two destructive, sensitive clay landslides at Lakelse Lake, 1962 showing a) fluid mud in the zone of depletion, b) debris accumulation in Lakelse Lake, and c) damage to vehicles. (Photos: Terrace Library)*

FIGURE 8.26 *The Mink Creek earth flow spread illustrates a rapid failure on a nearly flat gradient. (Photo: M. Geertsema)*

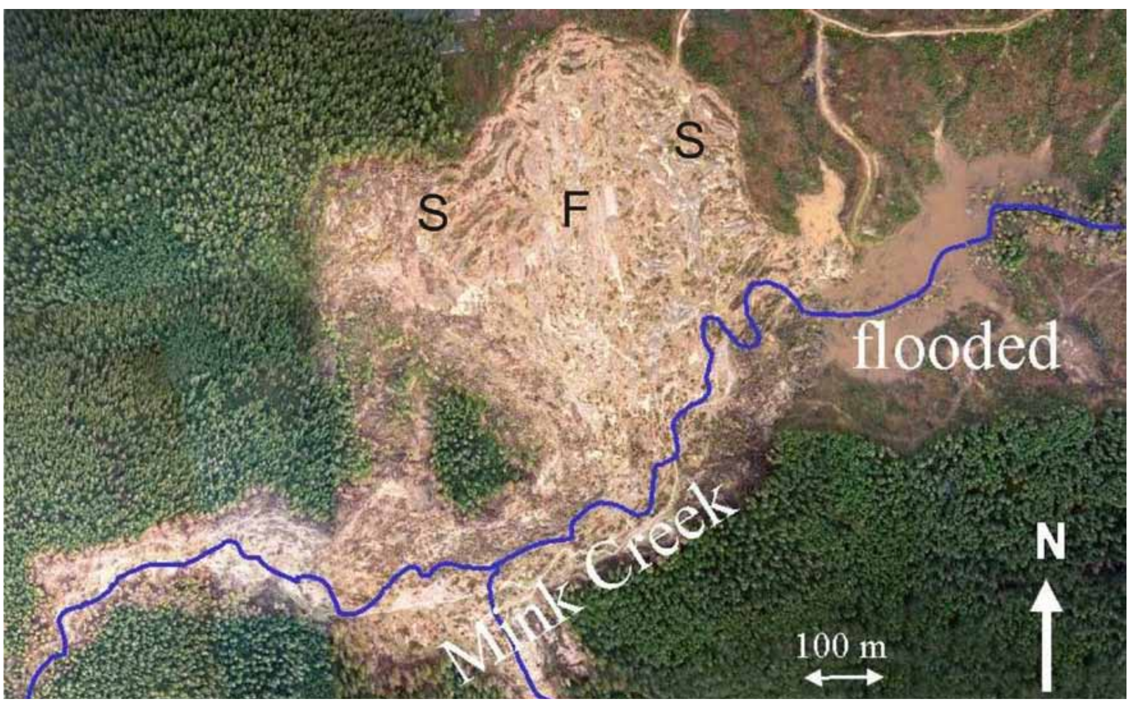

FIGURE 8.27 *Transverse ridges of spreading (S) give portions of the Mink Creek landslide a ribbed appearance; the central part of the slide experienced more complete liquefaction and became a flow (F). (Modified from Geertsema and Torrance 2005)*

Large Earth Flows

Large, slow-moving earth flows, involving millions of cubic metres at individual sites, are common along the incised valleys of interior British Columbia's dissected plateaus (Bovis 1985). Some gigantic landslides, such as the Sheslay slide in northwestern British Columbia (Souther 1971), cover tens of square kilometres and may involve billions of cubic metres of material. These large landslides occur in weak, fine-textured sedimentary and volcanic bedrock, or in the softened materials derived from the bedrock. Typically, the landslides have lateral ridges and lobate tongues and often have transverse ridges (Figure 8.28). Although these landslides resemble flows, movement generally occurs along discrete shear surfaces, with the body of the landslide moving as a rigid plug. For this reason, the movements may more accurately be described as "slides" rather than flows, and some researchers refer to these types of landslides as "mudslides" (e.g., Hutchinson 2004; Picarelli et al. 2005). Most of these landslides display very slow seasonal movement in response to groundwater fluctuations, and have been more active in the past, when climate was wetter, than at present.

Rock Avalanches

Rock avalanches involve the initial failure and subsequent disintegration of large rock masses on mountain slopes and the rapid downslope movement of this debris to a lower slope. Documented cases involve volumes of more than 10 million m^3. These landslides often attain velocities in excess of 100 km/hr, sometimes exceeding 300 km/hr. The velocity of rock avalanches depends not only on slope gradient, but also on the thickness of the debris wave; thus, greater velocities may be attained where the debris is channelized, and lower velocities where it is able to spread and thin.

Catastrophic movements in rock avalanches last for a matter of minutes. The conditions that weaken a rock mass for catastrophic sudden failure can take thousands of years. Globally, a common triggering mechanism for rock avalanches is seismic activity (Keefer 1984). The historic rock avalanches in

FIGURE 8.28 *Pavilion earth flow; note the transverse ridges in the zone of accumulation. (Photo: J. Ryder)*

northern British Columbia have not been associated with earthquakes (Geertsema et al. 2006a). For some, there have been no apparent triggers. Other movements have been associated with intense and prolonged rainfall.

An important consideration is not only the trigger, but also the preconditions that make the slopes susceptible to movement. These preconditions include:

- development of fractures or joints dipping downslope;
- progressive reduction of cohesion along shear planes;
- development of low friction angles and low residual strength along weak planes; and
- vertical relief and steep slopes.

In British Columbia, rock avalanches have occurred in the following major settings (Figure 8.29):

- cirque walls above glaciers
- on sedimentary dip slopes
- in association with slow mountain slope deformation

Cirque wall failure
Cirque walls tend to be steep and are often fractured. Fracturing probably occurs as a result of stress release such as the removal of glacial ice buttressing, or from exposure to modern weathering processes such as freeze–thaw cycling. A 1999 rock avalanche involving rock fall from a cirque wall occurred in the Rocky Mountains near McBride (Figure 8.30). The landslide was associated with heavy, prolonged rainfall (Geertsema et al. 2006a).

Downslope dipping bedrock
Long run-out rock slides are sometimes initiated on dipping sedimentary rock slopes, particularly in fault zones. West of Fort Nelson, two rock avalanches had detachment zones on 27–36° west-dipping strata (Geertsema et al. 2006a). Exposed rock on these dip slopes tended to be weak shale but occasionally sandstone would be exposed (Figure 8.31). On steeper slopes, failures would occur more continually, preventing the conditions for a single movement of a large volume of rock. If the slopes are too shallow, the rock masses will not gain enough energy to travel at a high velocity.

Deforming mountain slopes
Mountain slope deformation is common in northern British Columbia. These sagging slopes, or *sackungen*, are often the result of glacial debuttressing following deglaciation (see "Sackungen" below). These features are especially common in weak volcanic and sedimentary bedrock. Slope sagging may also be attributed to differential zones of weakness in bedrock, weathering, and tectonic activity.

During the last glaciation, glaciers gouged out valleys, putting stress on rock masses. When the glaciers melted, isostatic adjustments of the Earth's crust and lateral adjustments from valley walls caused some mountain slopes to sag. Such sagging is often a precursor to rock avalanching (Figure 8.32).

Most rock avalanches are complex landslides that frequently start out as frictional rock slides, but then transform into debris avalanches or debris flows when impacting and entraining soil lower on the slopes. Run-out models that incorporate these various phases of rock slides have been developed in British Columbia by Oldrich Hungr and his PhD

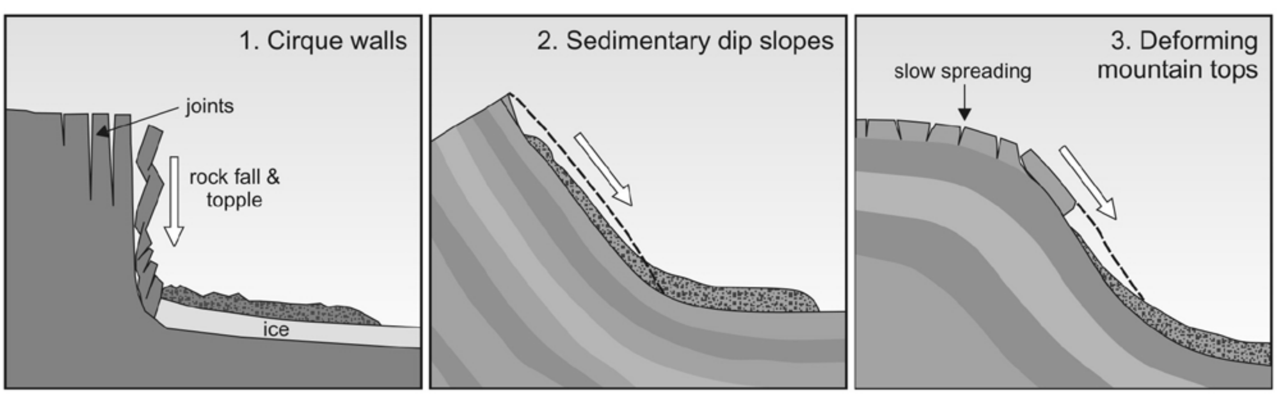

FIGURE 8.29 *Illustration showing the settings for rock avalanches in British Columbia. (Modified from Geertsema et al. 2006a)*

FIGURE 8.30 *Rock fall from this cirque wall above the Kendall Glacier, 45 km northwest of McBride, disintegrated and transformed into a rock avalanche with more than 1 km of run-out in July 1999. The landslide was associated with prolonged, heavy rainfall. (Photo: C. Erickson)*

FIGURE 8.31 *The Tetsa rock avalanche was triggered by disintegrating rock masses on dip slopes of approximately 30° in the Rocky Mountain foothills; note the run-up on the valley wall opposite the landslide. (Photo: M. Geertsema)*

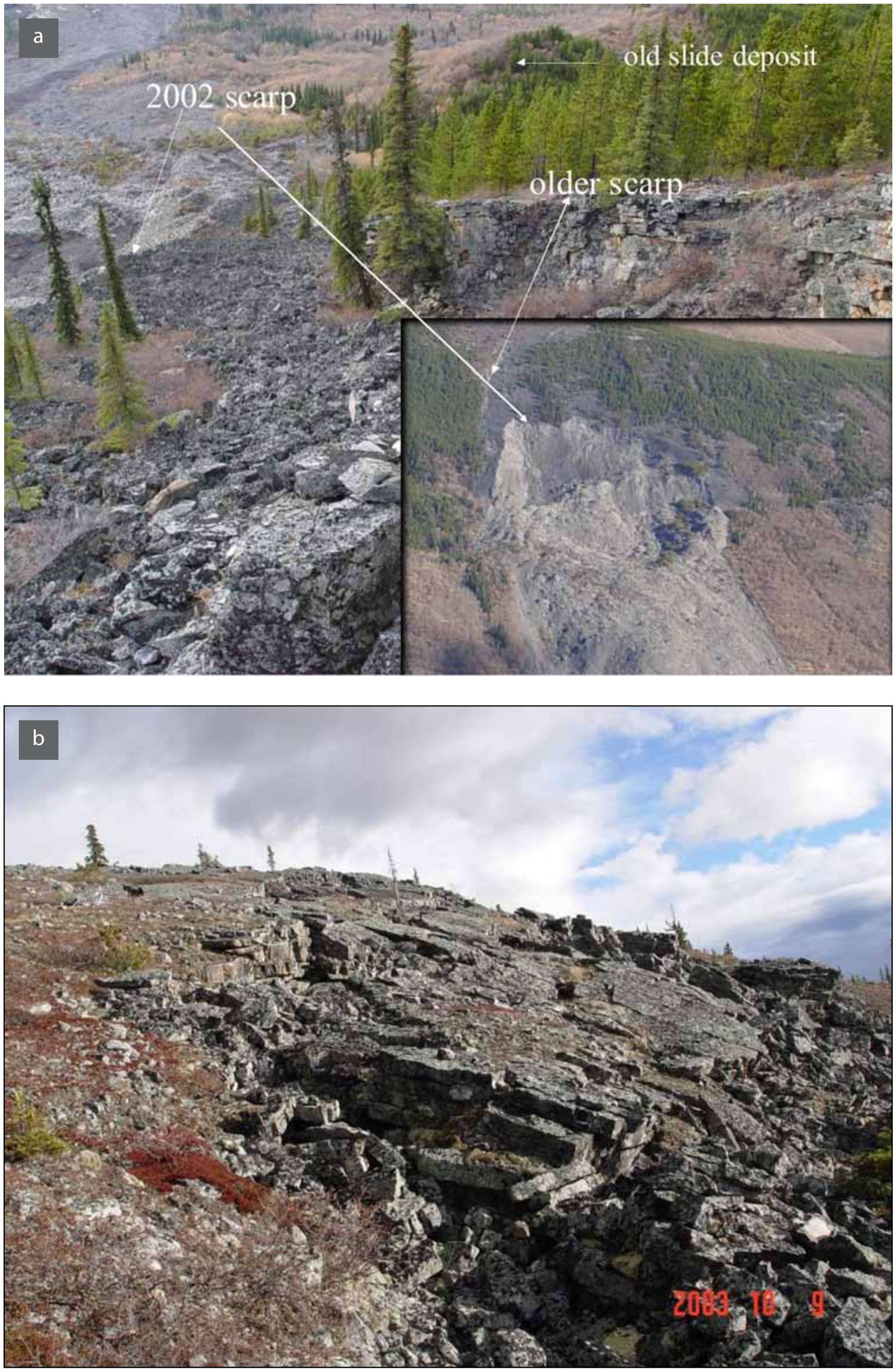

FIGURE 8.32 *The 2002 Pink Mountain rock slide–debris avalanche was associated with mountain slope deformation (Geertsema et al. 2006c). (a) Note ongoing deformation on the mountain, and (b) an older scarp and tension fractures above the 2002 landslide. (Photos: M. Geertsema)*

students, such as Scott McDougall (McDougall and Hungr 2004).

Sackungen *Sackungen* is a German term defined as a deep-seated rock creep that produces ridge-top trenches through the settling of a stable mass into an adjacent valley (Zischinsky 1969). It denotes "slope sagging," "ridge spreading," gravitational spreading, or deep-seated gravitational slope deformation. It is used to describe anomalous linear features, including upslope-facing (antislope) scarps and troughs trending parallel to slope contours. These linear features are found in mountainous landscapes and are often associated with the retreat of glacial ice from valleys leaving slopes over-steepened and unsupported (Bovis 1982, 1990; Bovis and Evans 1995, 1996). Sackungen may also be the precursors of catastrophic rock avalanches (Geertsema et al. 2006a).

Sackungen features were identified on a forested slope in the Kitnayakwa River drainage of west-central British Columbia (Schwab and Kirk 2003). The sackungen features at Kitnayakwa create giant steps in the landscape, with moss-covered downslope scarps up to 30 m high dropping down to meet antislope scarps up to 10 m high (Figure 8.33).

The scarps and troughs between scarps trend roughly parallel to slope contours. A distinct fracture line runs along the trough in most features. Larger ridges are situated approximately 150 m apart with a trough up to 25 m wide. Trees are growing in the trough as well as on the steeper ridge crests. Elongate pools of standing water are also found in some troughs. Smaller ridges in-between the larger ridges are spaced roughly 25 m apart. Trenches up to 7 m wide and 3 m deep cut across the hillslope at regular intervals. Vegetation on the smaller, antislope scarps shows signs of movement, and little growth occurs at the base of the troughs along the fractures.

Dead, fallen trees stretch across the troughs, some showing signs of excessive bow before falling (Figure 8.33). Converging and diverging ridges are found where troughs meet or separate, creating a maze on the hillslope. Along ridges, bowed trees are leaning both ways in an effort to grow upright (Figure 8.33). Split trees and trees with healed cracks occur in some troughs, indicating recent, abrupt movement.

Discerning sackungen features in a forested landscape from aerial photographs or by flying over the terrain is difficult because many of the features are visible only on the ground. Nevertheless, the following larger features are evident on aerial photographs.

- Lineations in the forest cover extend parallel to the contours of the slope (Figure 8.34). These are breaks in the forest cover where tree growth has been prevented by the fractures in the ground, or where scarp slopes are too steep for tree growth.
- Sharp elevation breaks in the hillslope are sometimes accompanied by a rise in elevation at the bottom of the break. These breaks indicate larger sackungen, and the accompanying antislope scarp.
- An over-steepened slope at the base of the hill breaks sharply onto the valley flat where no evidence of this type of slope is seen in the surrounding terrain.
- A small pond present at the base of a slope may show a dissimilarity from neighbouring slopes.
- Ridge crest depressions are often parallel to the slope.

FIGURE 8.33 *Antislope scarp and ridges indicative of slope sagging near the Kitnayakwa River. (Photos: J.W. Schwab)*

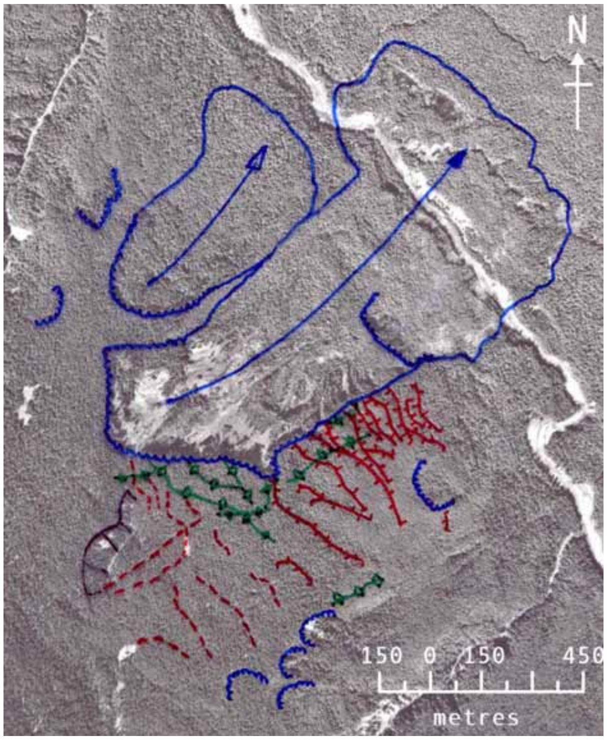

FIGURE 8.34 *Old landslides (blue outline) and transverse sackungen ridges (red) at the Kitnayakwa River. The landslides are about 220+ and 460+ years old. The largest landslide involved about 9.5 million m^3 of material. The bedrock is a highly fractured and crumbling, well-bedded, fine-grained, crystal-lithic tuff. (Photo interpretation: J.W. Schwab)*

The sackungen features at Kitnayakwa signify an unstable, sagging slope. Another landslide could occur at any time in this deeply weathered, over-steepened, unsupported glacial debuttressed slope through seasonal and climatic cycle effects on groundwater and (or) from earthquake shock.

Complex Landslides

If one examines any landslide closely enough, complexity will be found; however, some landslides clearly involve distinct materials and separate movement types. For example, rock slides will often trigger much larger secondary movements in soil, which would not have been possible without the initial slide. Two examples are presented below.

Rock slide – earth flow
Sometime around 1979, a rotational rock slide (slump) involving an estimated 2–3 million m^3 of material triggered an earth flow of 15 million m^3 in a clay-rich diamict near Muskwa River, west of Fort Nelson (Figure 8.35). The landslide travelled more than 3 km on a 3.5° slope (Geertsema and Cruden 2008). Such a long, low-gradient movement in this type of soil is impossible without invoking some type of dynamic loading (Geertsema and Schwab 2006). Hutchinson and Bhandari (1971) explained such behaviour by a process called "undrained loading." In this case, the rock slide became a dynamic load for the clay soil. The load transfer caused pore-water

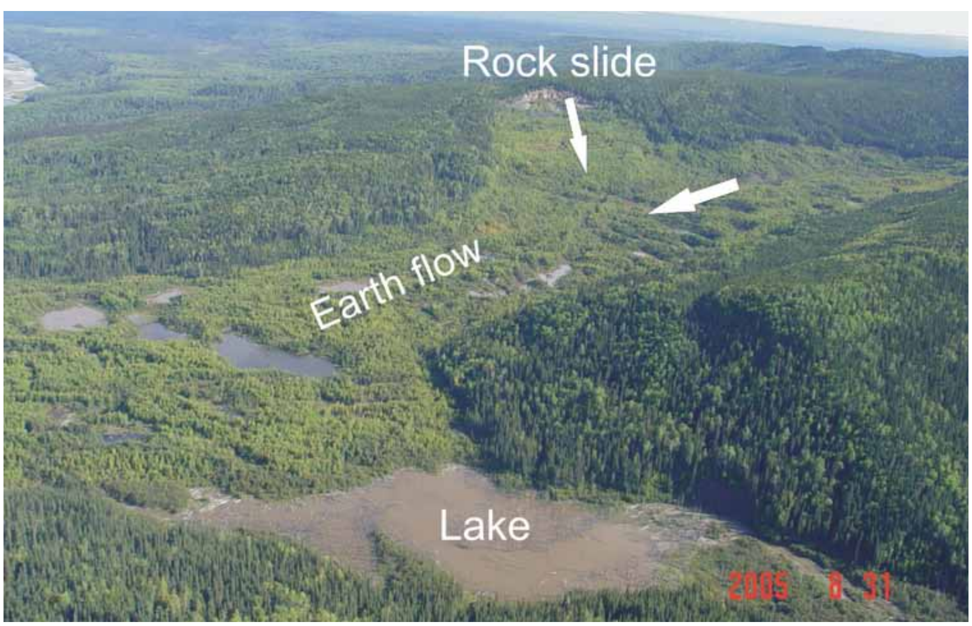

FIGURE 8.35 *Muskwa rock slide–earth flow, west of Fort Nelson. (Photo: M. Geertsema)*

pressures in the soil to increase, and because the groundwater could not readily escape from clay-rich soil, liquefaction within the clay occurred, mobilizing the mass to instability in what would otherwise have been relatively stable ground.

Rock slide – debris avalanche – debris flow
In late June 2002, a rock slide west of Smithers impacted glacial till above Harold Price Creek (Figure 8.36). After travelling 1.3 km, the rock slide transformed into a debris avalanche, and after 2.2 km the landslide entered a gully and changed into a debris flow, travelling a total of 4 km in three different phases. The initial frictional rock slide impacted, partially liquefied, and entrained glacial sediments, transforming the movement into a debris avalanche.

FIGURE 8.36 *A complex landslide (2002) at Harold Price Creek, near Smithers: (a) a rock slide transformed into a debris avalanche, which in turn transformed into (b) a debris flow, travelling a total distance of 4 km. (Photos: M. Geertsema)*

Most of the rock rubble was left upslope of this. The thinly spread debris avalanche entered a gully, which confined the moving mass into a debris flow.

Landslide-generated Tsunamis

In British Columbia, landslide-generated tsunamis are not well recorded and the majority likely remain unknown. Two main types have been documented: (1) displacement waves generated by subaqueous (underwater) landslides, and (2) those created by subaerial (on land) landslides.

Tsunamis from subaqueous landslides
Active fans and deltas are high-energy fluvial systems subject to continuous and progressive sediment loading. Where these deposits are underlain by soft muds, they may be prone to large-scale slumps, earth flows, or flow slides, as a result of site loading (natural or human-caused), vibration from heavy machinery, undercutting, interruption of intertidal drainage, and high artesian pressures.

Fan-deltas are common along the sidewalls of coastal fjords and also on steep-sided, long, narrow, fjord-type lakes in mountainous areas of British Columbia. The geomorphology of fan-deltas has been studied and described in coastal fjords (Prior and Bornhold 1988, 1990; Bornhold and Harper 1998). The processes are similar in freshwater lakes. Fan-deltas have developed over the last 8 000–12 000 years, commencing with the high-energy, sediment-laden meltwater during deglaciation of fjord-side drainage basins. Rivers continue to supply sediment during spring freshets and floods with the aggradation of sediments over the delta through multiple distributary channels on its surface with deposition in various layers of cobbles, gravels, and sands. Materials at depth generally include complex layers of interstratified muds, sands, and gravels. A fan-delta at first glance often appears as a relatively flat, unassumingly stable landform, but evidence of a historic failure is sometimes revealed as a steep, exposed head scarp along the fan-delta face. Large portions of fan-delta surfaces are submerged. Submerged materials perched on the steeply sloping faces are generally highly unstable. The evidence of landslide activity on fan-deltas is often concealed because landslide features are under water; however, the delta front can be viewed through side-scan sonar imagery. Debris slides and debris flows are displayed as a slight trough and debris lobe. The presence of crescent-shaped slide scarps and blocky debris ridges suggests the occurrence of a retrogressive rotational type of landslide movement (Figure 8.37).

Large fan-deltas are tempting sites for waterfront development because of their relatively flat terrain. The possibility of landslide activity and hazard is often not considered because the evidence is covered by water. However, working with equipment or the placement of structures close to the fan-delta water edge may cause disastrous slope failures. Before development, a detailed investigation of the fan or fan-delta must be carried out to ensure stability and safety of the sites.

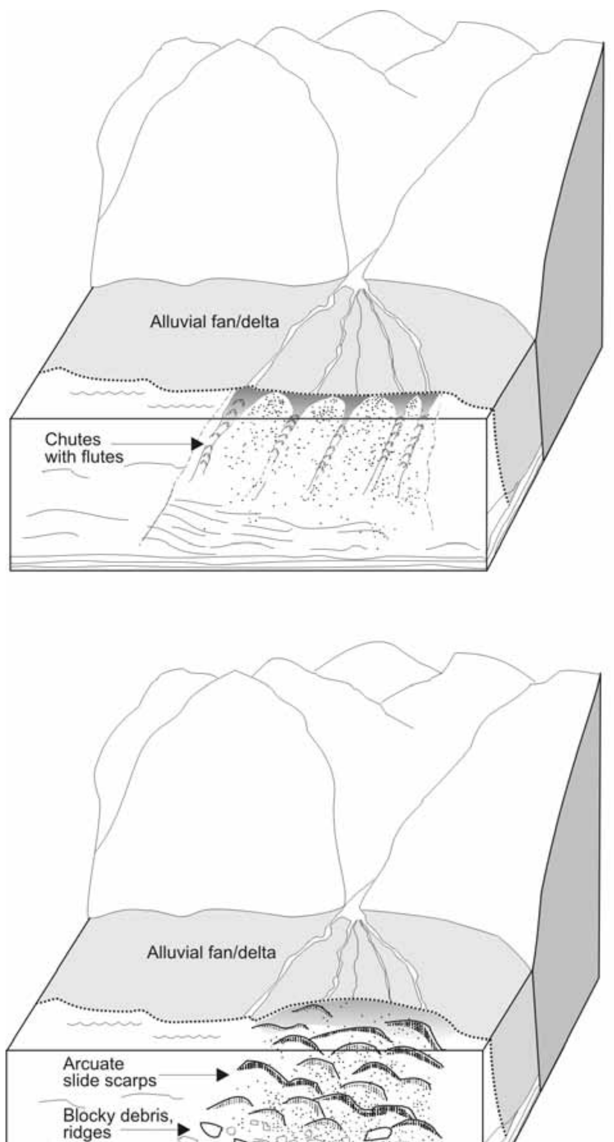

FIGURE 8.37 *Illustration showing subaqueous landslide associated with fan-delta collapse. (Modified from Schwab 1999)*

In part because of the remote location, little documentation exists of landslides that have occurred on fan-deltas; however, a few destructive landslide-tsunami events within British Columbia coastal fjords and on fjord-type freshwater lakes associated with fan-deltas have been recorded. For example, in 1975, a large submarine flow slide occurred on the front of a fjord-head delta in Kitimat Arm. The failure (volume estimated at up to 55 million m^3) happened about 1 hour after an extreme low tide (Prior et al. 1984). According to Murty (1979), the slide generated a tsunami up to 8.2 m high, causing considerable property damage. Site loading (construction activities) under conditions of high pore-water pressures, created by the extreme low tide, was thought to have triggered the slide.

Damaging landslides are not restricted to marine coastlines. On October 6, 1998, a fan-delta collapse occurred in Troitsa Lake, near Houston. Schwab (1999) estimated the volume of the largely underwater slide to be 3 million m^3. The resultant displacement wave was 1.5 m high and it damaged dock facilities. It travelled 1 km to the opposite side of the lake where it caused trees to topple. A backwash wave about 2 m high travelled back to the head scarp area carrying debris up to 150 m inland. Waves oscillated on the lake for 4 hours.

The cause of sediment failure along the delta front is unknown; however, human activity and seismicity were ruled out. Interestingly, a row of horizontally lying trees, approximately 650 years old, were exposed at the head scarp, indicating a past subaqueous slope failure event that was likely a landslide-generated tsunami.

Tsunamis from subaerial landslides

Some landslides discharge into water bodies and generate displacement waves (a type of tsunami). The most spectacular events occurred in Lituya Bay, Alaska, in the last century. The most recent event (1958) occurred when a large rock avalanche entered the bay. The landslide generated a displacement wave of more than 500 m high that stripped the forest from the opposite mountain slope and the lateral shorelines as it travelled down the bay (Pararas-Carayannis 1999).

A large rock slide (4 million m^3) entered Knight Inlet sometime in the mid-1500s, dropping more than 800 m vertically (Bornhold et al. 2007). The resulting tsunami destroyed the village of Kwalate, and about 100 (or possibly more) of its First Nations inhabitants. The wave height was estimated at 2–6 m.

In 1946, a Magnitude 7.2 earthquake triggered a 1.5 million m^3 rock avalanche from the north face of Mount Colonel Foster on Vancouver Island that entered Landslide Lake (Evans 1989). The resultant tsunami wave ran up a maximum height of 51 m, washing away the forest in its path.

The most recent landslide-generated tsunami occurred in 2007 at Chehalis Lake, about 40 km north of Chilliwack (Figure 8.38). The event results were first observed December 6, 2007, with the event likely occurring December 3–4 during heavy rains. A large, jointed mass of quartz diorite estimated at 3 million m^3, slid about 450–600 vertical metres into the lake. When the rock mass entered Chehalis Lake, it caused at least one wave that eroded the shoreline in some places to a height of more than 18 m. At the south end of the lake, 8 km from the slide location, the maximum wave height was 6.5 m. Trim lines were observed in the Chehalis River downstream of Chehalis Lake, indicating that a wave continued down the river. The displaced water continued downriver at least 14 km to the Chehalis alluvial fan, where large volumes of fresh woody debris were evident. Landslides entering rivers can also create displacement waves.

FIGURE 8.38 *The 2007 Chehalis rock slide triggered a large displacement wave. (Photo: B.C. Ministry of Forests and Range, Recreation, Sites and Trails Branch)*

TYPES OF EROSION

Erosion is generally considered the gradual wearing away of the Earth's surface by water, wind, glaciers, and frost action. Erosion is a generic term for a group of processes in which materials are loosened or worn away and simultaneously moved from one place to another. Erosion is intimately related to chemical and physical weathering of rock and other materials on and below the surface. Forest fires, logging, grazing, tillage, mining, road building, and other phenomena that remove surface vegetation can increase erosion rates. Wave action is an important component of shoreline erosion, and both ice jams and high streamflow erode riverbanks. Erosion is accompanied by transport and sedimentation.

The geomorphic processes of soil erosion and landsliding exist along a continuum. In this section, we consider various forms of water erosion ranging from splash erosion to rill erosion. We also consider piping (a type of subsurface erosion), washouts, gullies, and fire-generated erosion leading to landslides.

Rain splash erosion occurs when soil particles are dislodged and lifted into the air upon impact by raindrops (Figure 8.39). The level of erosion depends on the intensity and kinetic energy of the rainfall and the cohesion of the soil. Splash erosion results in crusting, detachment, and transport of soil. Once dislodged, the soil particles may be easily washed downslope by sheet erosion.

Sheet erosion involves the removal of thin layers of surficial material by broad, continuous, uniform flow over a slope. *Rill erosion* is the development of small, closely spaced channels resulting from surface soil removal by running water. Rill erosion is intermediate between sheet and gully erosion (Figure 8.40). Erosion is concentrated at these sites and may result in both downward and headward erosion. When a non-erodible base is encountered, lateral erosion tends to occur. With time, rill erosion may develop into gully erosion (Figure 8.40) (see "Gully Processes," below).

The erosion of an area depends on several factors relating to soil, slope, vegetation, climate, and erosion control practices. The Universal Soil Loss Equation (Wischmeier and Smith 1960) may be used to calculate erosion (soil loss):

$$X = R \times K \times L \times S \times C \times P$$

where: X = soil loss, R = rainfall factor, K = soil erodibility index, L = slope length factor, S = slope gradient factor, C = cropping or vegetation factor, and P = erosion control practices factor.

Rainfall factor (R) is based on the sum of storm

FIGURE 8.39 *Splash erosion: the raindrop is a few millimetres wide. (Photo: M. van Hees)*

FIGURE 8.40 *Rill and gully erosion along the Chilcotin River. Gullies are eroded into weakly cemented glaciolacustrine deposits (right) and into shallow rills in less cohesive colluvial deposits (left). Both gullies and rills on the lower 20 m of the slope were washed away after inundation from a landslide dam. (Photo: M. Geertsema)*

energy and intensity and will generally be specific to a given region or area. Soil erodibility index (*K*) provides an estimate of how erodible the local soil is and, therefore, will be site-specific. In general, soils with low cohesion (e.g., sandy soils derived from fluvial or glaciofluvial deposits) will have a high index, and soils with high cohesion (e.g., silty or clayey soils derived from basal tills) will have a lower index. Slope length factor (*L*) is determined by the length of the slope at a specific slope gradient. This slope gradient defines the slope gradient factor (*S*). Together, these factors (*LS*) may be considered the "topographic" factor. Long, steep slopes will be more susceptible to erosion than short, gentle slopes because of the greater volume of water running off the slope and a higher flow energy. The topographic factor will depend on site conditions; however, it may be possible to alter the topographic factor through treatments such as re-sloping or terracing. The cropping or vegetation factor (*C*) reflects the cover provided by vegetation or other material (e.g., mulch, erosion control matting), with higher cover yielding a lower cropping factor. Appropriate revegetation treatment can lead to a reduced cropping factor. The erosion control practices factor (*P*) depends on the application of non-vegetative erosion control practices, such as installation of check dams or silt fencing and surface roughening. Any such treatments can reduce this factor.

Hydraulic connectivity refers to the "connection" of hillslope areas to larger creeks or streams beyond the toe of the hillslope. In areas of direct hydraulic connectivity, any sediment that reaches a hillslope creek is transported directly downstream at a significant gradient (i.e., greater than 5%) to locations where it may result in adverse effects to water quality or aquatic resources. For areas with no connectivity, the hillslope stream must flow into a swamp or lake and trap sediment. Indirect connectivity may occur where the hillslope stream flows through a lower-gradient reach (typically less than 5% gradient for a minimum length of 100 m) before connecting with any stream reach with water quality or resource values.

Surface erosion from rain splashing and sheet wash delivers fine sediment to streams. Landslides such as debris flows tend to deliver coarser materials.

Gully Processes

Gullies are small, steep, and incised drainages on hillslopes. Gullied terrain may indicate a geomorphically active area that is sensitive to forest management activities. Landslides common in gullies include debris slides and debris flows. With increasing water concentration, debris floods and floods also occur.

Gully Morphology

In general, two types of gullies occur in British Columbia. The first type is incised into deep surficial materials, typically glacial till, and glaciofluvial or glaciolacustrine sediments. The second type is incised into bedrock. Often gullies formed in bedrock are the result of differing bedrock types, competency and structure, and groundwater conditions. In some locations, smaller gullies lie parallel to each other across an open slope (dissected terrain), and other gullies are tributary to one another, forming dendritic gully systems.

Gullies are divided into a number of zones: the headwall, sidewall, transport zone (or channel base), and mouth. Often a depositional fan is present downslope of the gully mouth. The headwall is the uppermost part of the gully system, and is typically a concave-shaped area shallowly incised into the hillslope. Headwalls in unconsolidated materials frequently exhibit soils with moderate or poorer drainage and are often the site of slope failures. Headwalls can be subtle features and are not always recognized.

The transport zone of the gully is what many people recognize as "the gully." It is typically V- or U-shaped, with steep sidewalls forming the sides of the V or U. The transport zone usually carries surface discharge along the channel at the base of the V or U. Sidewalls can range from a metre to tens of metres in height. The Forest Practices Code (B.C. Ministry of Forests 2001) defined gullies as those with sidewalls greater than 3 m in height. Similarly, sidewall gradients are generally steeper than 50% (27°) but in some cases are less steep. The size of the channel in the gully will depend on the amount of surface water draining into the gully as well as any groundwater that emerges from the gully sidewalls or from the base of the gully. Defining drainage area for gullies can be difficult, since groundwater movement on adjacent open slopes may or may not deliver subsurface flows towards the gully.

A fan is often located at the mouth of the gully unless the gully discharges into a larger stream or lake. The fan is a conical-shaped deposit formed by sediment and woody debris discharged from the gully. Where the gully channel discharges onto the fan, it may be incised into the fan deposits or it may flow across the fan surface. Channels that are incised into the fan surface may emerge onto the fan surface at a location further downslope. Where a channel flows across the surface of a fan with little incision, it may avulse (a complete or partial change in channel location). On many fans, multiple channel locations are the result of avulsions. Many fans below gullies are subject to debris flow deposition. Many gullies are relict features of deglaciation, particularly in the Interior, and are no longer active. Relict fans ("paraglacial fans") below these gullies may have had little or no sedimentation for millennia; however, development can re-activate gully processes. In some cases, fans are a combination of relict surfaces and active surfaces. Additional information on fans is available in Wilford et al. (2005b) and Millard et al. (2006).

Landslides in Gullies

Within gullies, the steep slopes and drainage concentrations result in a range of possible hazards. Steep headwalls and sidewalls are often sites of slope failures (usually debris slides). Debris slides that enter the gully channel may continue down the channel as a debris flow, a result of the contribution of water from the channel.

Almost all gully debris flows in British Columbia initiate from a debris slide or other slope failure from a headwall or sidewall. In one study of debris flows in coastal British Columbia, only 2% of debris flows initiated from high water discharges in the channel (Millard 1999). A debris slide from headwalls is most likely to initiate channelized debris flows because it maintains a directional momentum on entering a gully channel to which it is already closely aligned. In contrast, most debris slides from gully sidewalls need to turn almost 90° to continue down the gully channel. This loss of momentum results in more debris slides depositing in the gully channel. In the Interior of British Columbia, in-channel initiation of debris flows as a result of high-discharge events appears to be more common.

Since confined debris flows travel much farther than unconfined debris flows, most gully debris flows reach the fan or valley bottom. Debris floods and flows from gullies can result in significant aggradation of streams and fans.

High water discharge rates in gullies can typically result in water floods or debris floods. Water floods are the typical type of flood, transporting coarse sediment as bedload and with woody debris floating on the surface of the flow. Large discharge events supplied with large volumes of sediment can achieve sediment concentrations sufficient to support smaller clasts in a slurry, resulting in a debris flood. Debris floods may be four times larger than water floods expected in the same channel and are more destructive.

Piping

Piping is subsurface soil erosion (Parker and Higgins 1990). Pipes may be exposed as cavernous openings in eroding banks (Figure 8.41). Terrain undergoing subsurface erosion often exhibits hollows or collapsed depressions aligned along routes of subsurface erosion. Classic examples are strings of collapse features forming beaded gullies common in silt fans derived from glaciolacustrine deposits in the Thompson-Okanagan area (Evans and Buchanan 1976).

FIGURE 8.41 *Pipe exposed on a bank of the Chilcotin River. (Photo: M. Geertsema)*

These features, combined with an accompanying cave or pipe, are generally observed in arid climates in landforms composed of silt and clay. The term "piping" is also used to describe soil pipes and (or) water movement in forest soils, particularly along root channels and small erosion conduits in surficial materials. In some cases, piping may contribute to landslides and washouts.

Washouts (rapid erosion of seepage faces) have been observed in glaciofluvial sands, silts, and gravels throughout British Columbia. Some examples include: the grand campus washout of 1935 at the University of British Columbia, Vancouver (Williams 1966); the Maryhill gravel pit washout, Coquitlam River valley (Allen 1957); caving erosion in the Lower Mainland (Hungr and Smith 1985); gully erosion / caving erosion found in the Coquitlam River valley (Siebert 1987); and seepage-face erosion (Parker and Higgins 1990). Catastrophic seepage-face erosion was also observed at Bowser River in 1995, a result of the diversion of a small stream by beaver onto a glaciofluvial terrace of sands and silts—no surface water flow reached the catastrophically formed gully. The most spectacular is the Donna Creek wash-out flow of 1992 described by Schwab (1997, 2000, 2001).

A schematic representation of catastrophic seepage-face erosion is presented in Figure 8.42. Under normal conditions, seepage exits the slope without causing erosion as the ground slope of the seepage face has adjusted to the highest recurrent seepage exit gradient[2] and resulting seepage forces. A change in the discharge conditions results in an increase in seepage exit gradient above a critical value. This increase in exit gradient may result from a rapid increase in groundwater discharge or sudden removal of material from the seepage face by a landslide or by running water. If the resultant erosion is of sufficient intensity, the retreating seepage face uncovers material with greater groundwater pressure, the exit gradient increases, and a continuing chain reaction results (O. Hungr, University of British Columbia, Earth and Ocean Science, Vancouver, B.C., pers. comm., 1997).

2 Seepage exit gradient is the slope of the piezometric surface adjacent to the drainage discharge point.

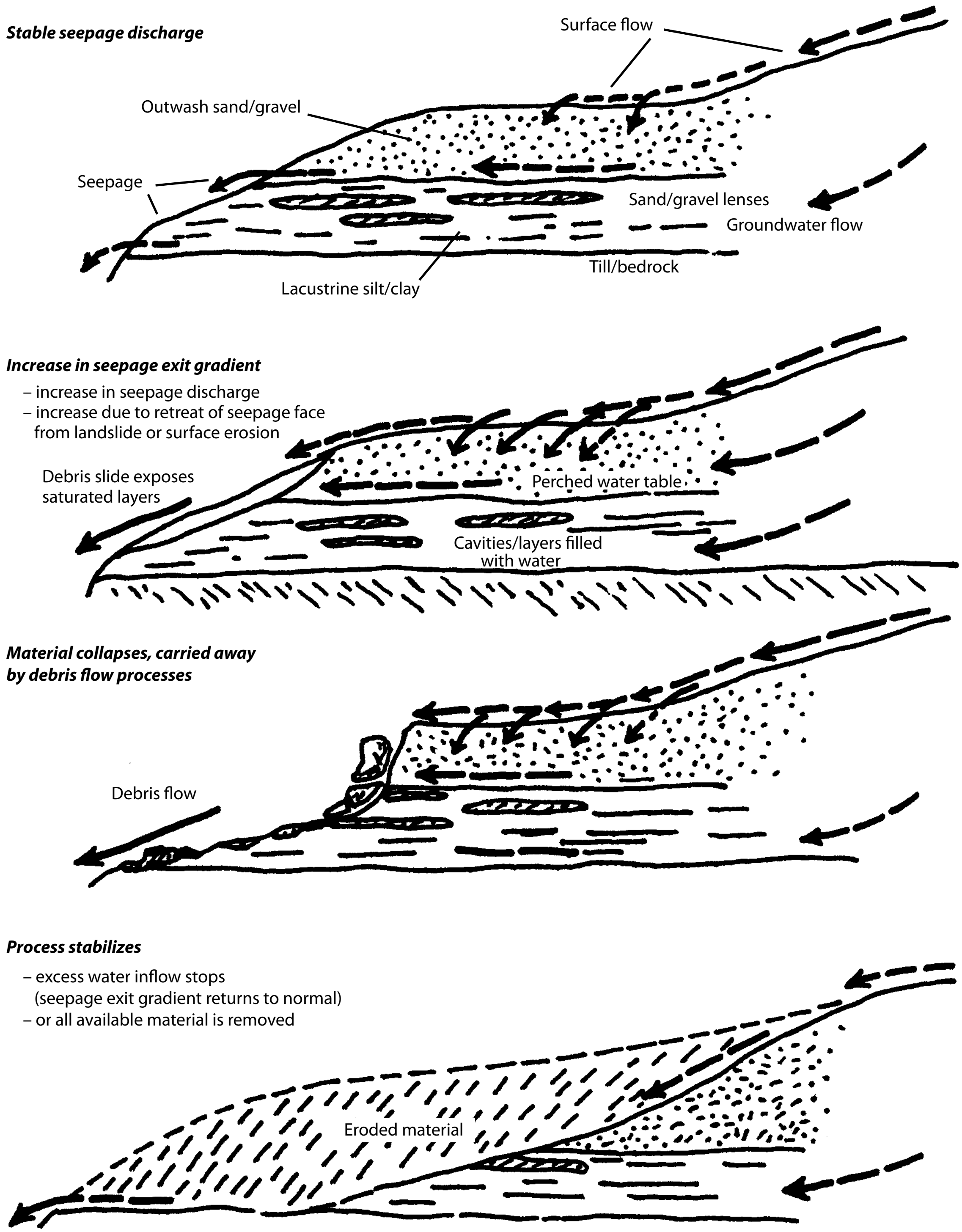

FIGURE 8.42 *Development of catastrophic seepage-face erosion concept. (Adapted from Siebert 1987)*

Knowledge of the processes occurring during such a catastrophic event enables the event to be classified. The outcome is the formation of a large gully, eroded largely by running water, hence a "washout." Landslide processes are also involved: the collapse, dilation of material, and extremely rapid debris flow of liquefied material, hence a "flow" or "debris flow." The term "washout-flow" thus describes the overall complex landslide process. Catastrophic seepage-face erosion describes the sudden change in the discharge conditions that result in an increase in seepage exit gradient, the subsequent convergence of surface and subsurface flows, the retreating seepage face, caving, collapse, and debris flow surges.

Fire-generated Erosion and Landslides

Effects of wildfire on slope stability

An increased incidence of landslides, erosion, and flooding is common following wildfire in many parts of the world (Shakesby and Doerr 2005). Most literature on the subject has focussed on surface erosion, but more recently the risks associated with large mass movement events, especially debris flows, have been appreciated (Cannon and Gartner 2005). In the western United States, considerable resources are expended on assessing the risk of post-wildfire erosion and mass movement, and on applying treatments to burned areas to reduce these risks (Robichaud et al. 2000).

Wildfire typically causes several changes in soil and vegetation, which can affect hydrology and slope stability (Scott and Pike 2003; Shakesby and Doerr 2005; Curran et al. 2006). These effects include:

- A water-repellent layer can be formed just below the surface of the mineral soil, caused by the deposition of hydrophobic compounds that are volatilized by burning of the forest floor. Water repellency can cause the generation of overland flow across large areas, especially during high-intensity rainfall under dry soil conditions. The effect typically persists for 3–5 years after the fire. Many post-wildfire debris flows have been caused by runoff and eroded sediment generated from areas of water-repellent soils during intense rainfall, which accumulate in gullies or steep stream channels in or below the burn.
- Combustion of the forest floor and understorey vegetation results in reduced water storage and interception capacity. Also, the forest floor protects the mineral soil from raindrop impacts. Loss of the forest floor can therefore increase the likelihood of overland flow and splash erosion of the mineral soil.
- The structure of the surface soil can be altered by fire, and by ash, which can clog soil pores or form a crust. This can further reduce the permeability of the surface soil.
- Death of trees results in increased snow accumulation, increased snowmelt rates, and reduced evapotranspiration. This can cause increased runoff from both snowmelt and long-duration rainfall, and higher groundwater levels, which can reduce slope stability. This change may be long-term until the forest regenerates, and may be equated to similar effects of clearcutting or mountain pine beetle infestation.

Debris flows are the most common mass movement process observed following fires, although the processes described above can affect many types of slope instability. In areas of stable terrain, debris floods caused by a combination of overland flow during rainstorms and an accumulation of eroded sediment in stream channels are common.

In northern regions underlain by permafrost, shallow debris slides caused by detachment of the active layer are common following wildfires (Lipovsky et al. 2005). The main contributing factor for these failures is probably loss of the insulating organic surface layer, as well as loss of shade and the decreased albedo of the burned area.

In 2003, many large wildfires occurred in the southern Interior of British Columbia, including several in population interface areas. In the following 2 years, flooding and mass movement events occurred in several of the burns, and included the following notable examples (Jordan et al. 2004; Jordan and Covert 2009).

- Okanagan Mountain Park fire near Kelowna: In October 2003, debris floods and flooding occurred in several creeks, caused by a short-duration rainstorm with a return period estimated at 10–100 years. Municipal roads, culverts, and private property were affected.
- Cedar Hills fire near Falkland: In June 2004, several small debris flows and flooding occurred in gullies below the burn, caused by a short-duration rainstorm (Figure 8.43). A highway and private property were affected.

FIGURE 8.43 *Evidence of flooding and mass movement events (June 2004) following the 2003 Cedar Hills fire, near Falkland: (a) overland flow, (b) concentration of flow and eroded sediment in gully, and (c) debris flow. (Photos: B. Grainger)*

- Kuskonook fire near Creston: In August 2004, debris flows destroyed two houses and blocked a highway for several days (see below).
- Lamb Creek fire near Moyie Lake: In August 2004, a large debris flood damaged a forest road and affected fish habitat.
- Ingersoll fire near Burton: In October 2005, during a 3-day, low-intensity rainstorm, about 15 debris flows and debris slides occurred in gullies in and below the burn (Figure 8.44). Several of these were very large and damaged forest roads and forested private property below.

Smaller flooding or debris flow events that affected highways or private property occurred in several other 2003 burns, including the McClure, Vaseaux Lake, McGillivray, and Strawberry Hill fires. A debris flow also occurred in the 2004 Botanie fire near Lytton.

The number and severity of events following the 2003 fires was unprecedented in British Columbia. A possible reason is that the summer of 2003 was exceptionally hot and dry, which may have resulted in unusually high soil burn severity due to a dry forest floor. An additional factor may be the unusual number of short, high-intensity rainstorms in fall 2003 and summer 2004.

Until the 2003 fire season, landslides, debris flows, and other major erosion events following fires were rarely reported in British Columbia. Some explanations may include:

- Rainfall intensities are lower than in more southern latitudes.
- Soil burn severity may be lower at most British Columbia locations compared with fires further south, because of soil and forest floor differences and a relatively cooler climate.

- The hydrology in the British Columbia interior is dominated by snowmelt, and even with a substantial increase in peak flow on small streams from summer rainstorms, these flows still may not exceed typical spring snowmelt peaks.
- Most wildfires in British Columbia are in relatively remote areas, and landslide or erosion events may have occurred but not been observed.

Despite the well-documented events that followed several of the 2003 fires, most of the burnt-over areas had no significant erosion or mass movement events. Several areas of large wildfires, which occurred earlier near populated areas (e.g., near Kimberley and Canal Flats in 1983), have been observed for many years by Forest Service staff and others, and no major events have been noted. However, debris flow events were reported anecdotally following several earlier fires, including the 1973 Eden Fire and the 1998 Silver Fire, both near Salmon Arm. No systematic documentation of mass movement events, rainfall events, or burn severity has taken place. It appears likely that, although localized erosion may be common, most large wildfires in British Columbia do not produce significant mass movement events, erosion, or flooding. Nevertheless, the probability of such events is increased over pre-fire conditions, especially where soil burn severity is high. The occurrence of major erosion events may depend on whether unusually intense rainstorms fall in the critical 3–5 years after a fire, when water repellent conditions may still persist (see "Hydrophobicity," below).

FIGURE 8.44 *Evidence of debris flows (October 2005) following the 2003 Mt. Ingersoll fire, near Burton. (Photo: P. Jordan)*

Case study: Kuskonook Creek debris flow (2004) near Creston The Kuskonook fire, east of the south arm of Kootenay Lake, burned 4800 ha of forest in the late summer of 2003 (Figure 8.45). This was one of several large wildfires in the southern Interior of British Columbia in the severe fire season of 2003. In the following 2 years, unusual landslide or erosion events occurred in or below at least six of these burns.

Around midnight on August 6, 2004, debris flows occurred in both Kuskonook and Jansen Creeks during a short, intense rainstorm. Both of these blocked Highway 3A, which was closed for several days. The larger flow caused severe damage in the small community of Kuskonook, destroying two houses and damaging several other buildings. Fortunately, no fatalities or injuries were inflicted, as the houses were unoccupied that evening.

The investigation of the Kuskonook event showed that the debris flow was caused by water-repellent soils in the burned area, which generated an unusual volume of overland flow in the headwaters of the drainage (Jordan et al. 2004; VanDine et al. 2005; Curran et al. 2006) (Figure 8.46). This runoff was concentrated in steep gullies below (Figure 8.47), and became a debris flow as the exceptionally high discharge entrained sediment from the channel bed and sidewalls. An estimated 20 000–30 000 m^3 of debris was deposited on the Kuskonook Creek fan (Figure 8.48).

This event is a good example of a debris flow caused by changes in soil hydrology following a severe wildfire. Although commonly reported in the United States (Kalendovsky and Cannon 1997), such events were believed to be unusual in Canada, until the events following the 2003 fires.

FIGURE 8.45 *View over Kuskonook and Jansen Creek watersheds, burned in 2003 fire. (Photo: D. Nichol)*

Hydrophobicity Soils that repel water are hydrophobic. Normally, soils have a high absorptive capacity for water such that water applied to coarse-textured soils will readily infiltrate the soil (DeBano et al. 1967); however, some soils repel water. These soils are referred to as "non-wettable," "water repellent," or "hydrophobic." In wettable soil, a strong attraction exists between dry soil particles and water. In hydrophobic soil, mineral particles are usually coated with non-polar organic materials that repel water. In water-repellent soils, drops of water form surface beads. Measuring the slope of the wetting angle between a droplet of water and the ground surface indicates the degree of hydrophobicity (Figure 8.49).

In British Columbia, forest fires sometimes create hydrophobic layers in soil. Thick acidic humus, coarse-textured soils, high-intensity fires, and prolonged periods of intense heat are important factors for the creation of water-repellent soil. Hydrophobic substances are released from humus forms and condense at depth. Extremely high surface temperatures may actually destroy surface hydrophobicity, but cause a subsurface water-repellent layer to form deeper in the soil (perhaps 1–10 cm in depth).

With a hydrophobic layer at depth restricting infiltration, surface soil may become saturated following precipitation, leading to overland flow, soil detachment, and erosion. The eroded material may coalesce into shallow landslides, or fill gullies and transform into debris flows.

FIGURE 8.46 *Soil erosion caused by overland flow on burned soils. (Photo: M. Curran)*

FIGURE 8.47 *Gully erosion in headwaters of Kuskonook Creek. (Photo: P. Jordan)*

FIGURE 8.48 *Debris flow deposit on Kuskonook Creek fan. (Photo: P. Jordan)*

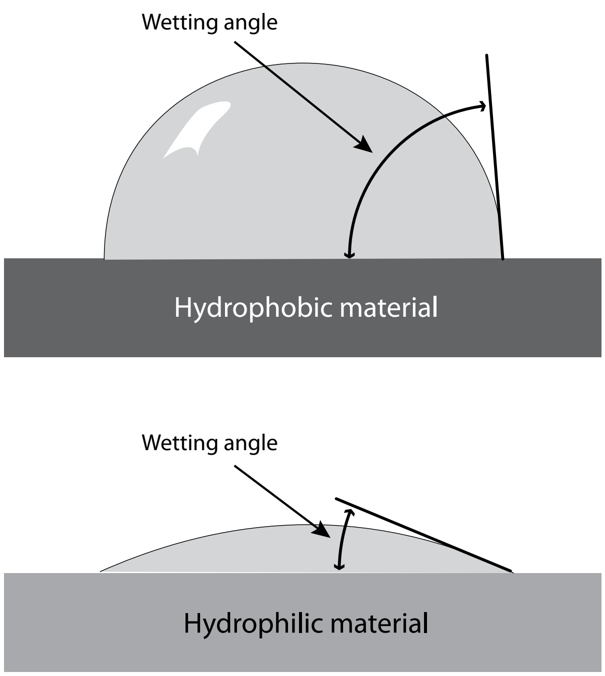

FIGURE 8.49 *The wetting angle between a droplet of water and a solid surface is an indication of the degree of hydrophobicity: the steeper the angle, the more hydrophobic the soil. (Modified from DeBano et al. 1967.*

INDICATORS OF INSTABILITY

Field terrain stability analysis requires an eye for detail to read subtle changes in the landscape and on aerial photographs. It not only requires an understanding of geomorphic processes, but also of local ecosystems.

Aerial Photograph Interpretation

Field investigations are usually preceded by aerial photograph interpretation. Sometimes unstable terrain is easy to recognize. Fresh, unvegetated scars often indicate recent landslide activity. Unstable terrain and old landslides that are overgrown with vegetation are more difficult to discern. Unvegetated and vegetated linear tracks are usually indicative of debris slides or flows. Debris flows usually have moderately steep fans, but the debris may also enter directly into powerful streams that remove the material, preventing fans from developing.

Flows tend to have ridges aligned in the direction of movement, whereas spreads and rotational slides (slumps) have ridges transverse to movement. Sometimes spreads and slumps transform into flows as the slope over which they move becomes steeper, or as material becomes progressively weaker.

Deep-seated landslides tend to have arcuate main scarps, often with wet depressions, referred to as “sag ponds.” Over time, the ponds may drain or become filled with peat. The negative surface expression of zones of depletion is to be contrasted with the positive surface expression of zones of accumulation. Lobate and hummocky surfaces are also characteristic of landslide deposits.

Field Interpretation

Many diagnostic features of past and potential landslides can be noted in the field. Indicators of instability are manifested in surface expression, vegetation, rock, and soil.

Tension fractures, spoon-shaped depressions, ridges, scarps, antislope scarps, seepage, and ponds on hillslopes may all be indicative of unstable terrain. Lower down the slope, one might find ridges, hummocks, lobes, or rubble. Rotational slides may daylight (emerge at the surface) in rivers or streams, pushing up the river bed (Figure 8.50).

Water-loving vegetation on slopes indicates seepage and is often associated with unstable terrain. The linear tracks of vegetation seen on aerial photographs may also be recognizable on the ground. Older trees adjacent to the younger vegetation may bear scars or have persistent mud lines on their trunks. In addition to scars, trees may display pronounced inclination or curvature. Trees tend to lean upslope at the head of landslides, and downslope on lobes, but this can be highly variable. In spreads and rotational landslides, trees tend to lean in the direction of rotation of the ridges and blocks formed beneath them. In some cases, trees may be split when they straddle a tension crack. Although trees typically display swelling at their bases, debris on fans may bury the lower parts of trees. Absence of buttswell should alert an observer to a burial event (Figure 8.51). Logs entrained in debris flows tend to be severely abraded (Figure 8.52).

Deposits formed by landslides may display a wide range of textural and morphological characteristics depending on the type of landslide. Deposits range from coarse angular blocks resulting from rock avalanches to fine-textured mud flow deposits. In general, landslide deposits display perturbation, but some translational landslides transport material with little to no internal disturbance. Typical colluvial soils are loose, with angular clasts. Debris flow clasts are often matrix-supported. Buried soils indicate repeated deposition events.

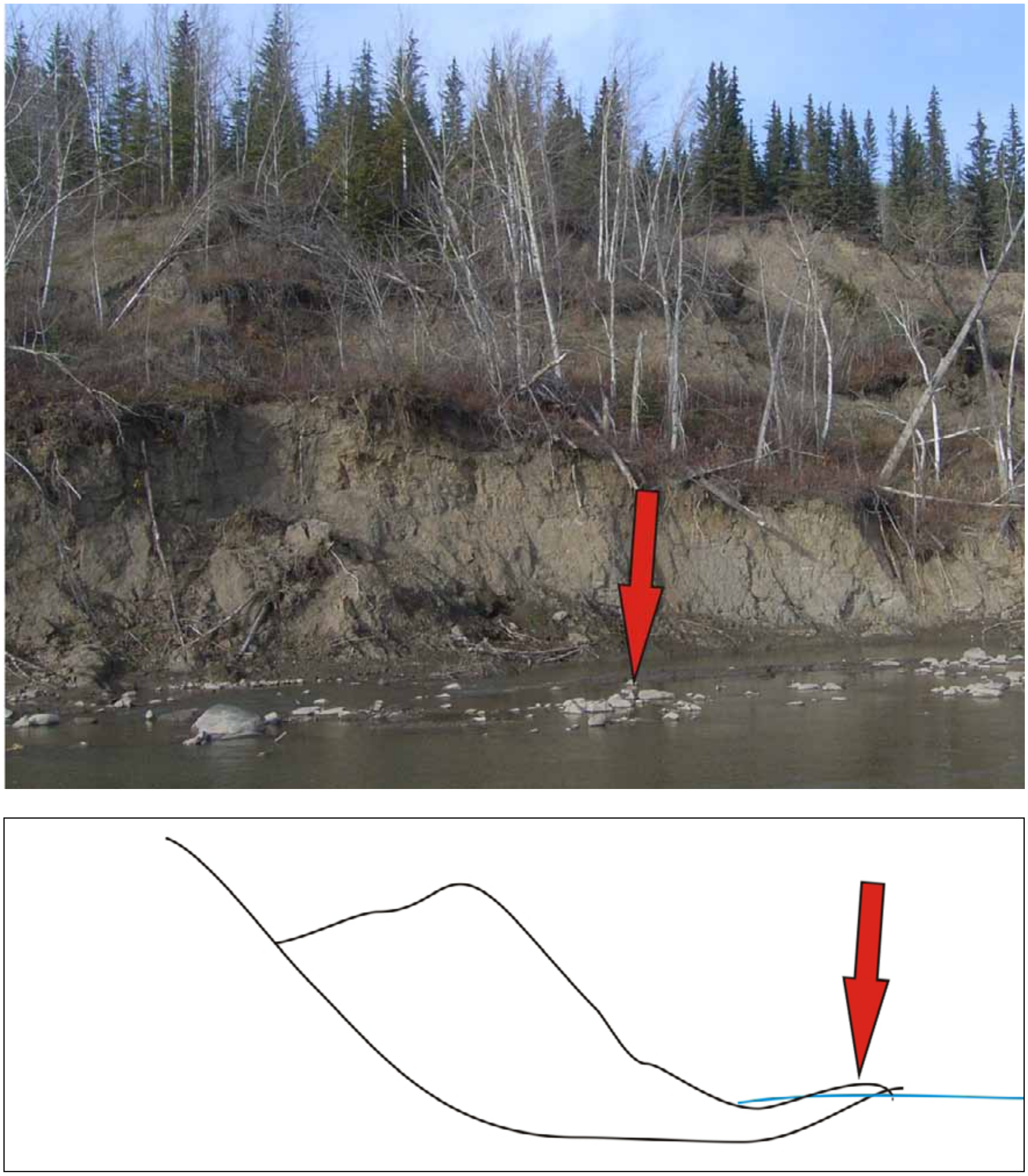

FIGURE 8.50 *Upturned rotational landslide toe exposed (arrow) in Kiskatinaw River, near Dawson Creek.; this landslide is sensitive to toe erosion by the river. (Photo: M. Geertsema)*

FIGURE 8.51 *Balsam poplars buried in debris-flow deposits near Chetwynd; note the absence of buttswell. (Photo: M. Geertsema)*

FIGURE 8.52 *Severely abraded log entrained in the 2002 Zymoetz rock slide–debris flow, near Terrace; the log is oriented in the direction of flow. (Photo: M. Geertsema)*

An important component of hazard and risk analysis is the determination of recurrence intervals and magnitude or frequency relationships. Both the spatial and temporal distributions of hazardous events are required for such analyses. This section discusses the determination of the distribution of hazards in time.

Landslides occur every year in British Columbia, and have probably been happening since deglaciation. They leave their marks as records on the landscape—as visible scars and lobes, and through burial of soil and plant materials. Over time, the visible records diminish in clarity and buried records may be eroded. Nonetheless, numerous techniques are available to date landslides.

Human Records

The most reliable sources of information are well-documented, eye-witness accounts. Newspaper accounts may provide an accurate date of some event, but spatial descriptions are not always accurate. Aerial photographs and satellite images help constrain the times of events, provided the events are sufficiently large to be recognizable at the image scale. Old maps, reports, Hudson's Bay Company records, and even landscape paintings (e.g., Nilson 2005) have been recorders of landslides. Some large rock slides generate their own seismic signatures, and landslides that dam rivers may be recorded indirectly by hydrometric stations as dips and spikes on hydrographs, corresponding to the formation and rupture of the landslide dams. Some catastrophic landslides are recorded in the oral traditions of First Nations people; however, the dates of the events are uncertain and are usually tested with radiocarbon dating methods. For example, stories of a large rock slide that buried a village near Hazelton some 3000 years ago (Gottesfeld et al. 1991) and of a rockslide–generated tsunami that wiped out a village in Knight Inlet about 500 years ago (Bornhold et al. 2007) are told to this day.

Natural Records

It is important for researchers to establish the occurrence of landslides over time. Landslides may have occurred more frequently under certain past climates (e.g., Geertsema and Schwab 1997) or be clustered around strong earthquake events. Knowledge from the past can be extrapolated to help with predictions for the future.

Even though landslides may be dated in numerous ways, the data are often scarce and difficult to obtain. One must choose an appropriate sampling location and interpret the data correctly. In British Columbia, the most common landslide-dating methods are dendrochronology (tree ring dating) and radiocarbon dating. Recent events, or events that recur on decadal time scales, require high-resolution dating techniques. Tree rings and varves[3] can provide excellent annual resolution. Radiocarbon dating is best left for century-scale resolution.

Tree ring dating

In forested landscapes, trees often provide the best records of past slope processes (Shroder 1980; Braam et al. 1987a, 1987b). Several properties of trees make them particularly useful for landslide dating: (1) they produce annual growth rings; (2) their rings are sensitive to environmental conditions and traumatic stresses; and (3) trees usually attempt to grow vertically—if something causes them to lean, they attempt to correct the lean by resuming vertical growth. Leaning trees will produce dark eccentric rings called "reaction wood."

Landslides affect forests in different ways. Shallow debris slides and avalanches tend to remove trees from their paths (nudation) and accumulate woody debris along their lower margins. Standing trees beyond the limit of the slides are often pushed off their vertical axes. Slow-moving landslides may result in the production of curved tree trunks. Debris flows and rock falls tend to scar trees in their paths. A single tree may bear scars from multiple events. Rotational slides, spreads, and low-gradient translational slides tilt trees. In some cases, blocks of ground are rafted horizontally, or dropped vertically without tilting trees. Flows may bury standing trees in zones of accumulation. Exposed surfaces on soil landslides in forested areas tend to be rapidly recruited by trees and shrubs; however, large rock slide rubble may remain unvegetated for many decades. A fan-delta collapse west of Houston caused trees weighed down

3 Annual lake sediments usually separated by winter clay and summer silt layers.

by sediment in root wads to float vertically in a lake for some weeks before sinking. Landslides that impound streams may drown trees upstream of the landslide dam. Over time, sediments accumulating in the temporary lakes may bury the trees.

Master chronologies can extend tree ring records back in time over a considerably longer span than living trees. The technique involves the statistical matching of ring sequences from living and fossil trees. Such cross-dating is far superior to radiocarbon dating of fossil trees, but established chronologies are region-specific and time consuming to obtain. Brian Luckman (University of Western Ontario) has developed a master chronology for the Rocky Mountains, and Dan Smith (University of Victoria) has developed one for the Coast Mountains.

Nudation

Landslides tend to remove vegetation from slopes, resetting the plant successional clock. New recruitment of trees and shrubs on landslides facilitates landslide dating by providing a minimum age for an event's deposit. Some species are rapid invaders, such as *Alnus, Betula, Salix*, and *Picea*, and these are useful for obtaining minimum dates of a landslide.

Dating landslides from new colonizing trees is not as simple as it might seem, as many potential sources of error exist. Trees may not appear within the first year following the landslide and may colonize at varying times. It is not always sufficient to date the oldest trees either, as many translational landslides transport trees in the vertical position (Figure 8.53). These trees simply continue to grow. Even if large trees are tilted during landslide transport, seedlings and small saplings may be transported without tilting, and these may be difficult to distinguish from post–landslide-colonizing plants. Another source of error lies in the difficulty of finding the first growth ring. Minor subsequent movements may result in staggered colonization.

If the colonizing vegetation is significantly different from that of the surrounding forests (e.g., *Salix* in a *Picea* matrix), one can be more confident in the trees providing a reasonable minimum date of the landslide. The technique is best used on debris slides, and rock and debris avalanches where little of the original surface is preserved. In landslides that carry rafts of surface material, large sample sizes are recommended. More precise and accurate dates can be obtained from the analysis of reaction wood in leaning trees and from tree scars.

FIGURE 8.53 *Rafted red alder in the Khyex landslide near Prince Rupert. Ring counts of these trees would overestimate the age of the landslide. (Photo: M. Geertsema)*

Inclination

The inclination (or leaning) of trees produces reaction wood, which in cross-section shows dark eccentric rings rather than normal light concentric rings (Figure 8.54d). Reaction wood can be used to directly date landslide events. In sudden movements, the change can be abrupt, often with the tree curving up to grow vertically again. Sometimes, a lateral branch takes over vertical dominance (Figure 8.54b).

Gradual or episodic movement can also be recorded in trees as can a change in the direction of tilting. Trees tend to be tilted upslope on rotational slide blocks and downslope on lobes. Trees can be tilted in multiple directions ("jackstrawed") in hummocky deposits or in areas with many small failures within a larger landslide body.

Trees can lean for many reasons other than from landslide movement (Figure 8.55). Trees often tilt in random directions in hummocky, permafrost terrain. They may tilt towards openings in the forest, exhibiting phototrophism, or growing towards the light. Trees may lean due to wind, particularly in shallow soils. Trees may be curved due to snowpack creep.

Corrasion (Scarring)

Corrasion of tree bark can result from the impact of rocks, debris, or other trees (Figure 8.56). Scars yield direct dates of landslide events. A single tree can record multiple events by displaying multiple scars of varying ages. However, scars can also result from numerous non-landslide events, including the fall of neighbouring trees, animal gnawing, and frost damage.

FIGURE 8.54 *Trees displaying variable responses to mass movement: (a), (c), and (d) show slow, episodic mass movement; (b) shows sudden, rapid movement that topples a tree and causes a lateral branch to assume apical dominance. (d) Note the various times and directions of reaction wood (arrows), which indicates leaning in different directions. (Photos: M. Geertsema)*

Burial
Flows in the lower, distal portions of landslides may partially bury trees without causing mortality. Evidence of burial may be lack of buttswell, and the sprouting of adventitious roots in the newly formed ground surface (Wilford et al. 2005a). Trees may also respond to the stress of burial by undergoing growth suppression. Response is often not immediate, and thus less reliable than scars and reaction wood for landslide dating.

Drowning
Many landslides form lakes when streams are dammed, or when closed depressions (e.g., sag ponds) are formed on the landslide surface. Trees inundated by standing water may die as a result of drowning (Figure 8.57). Careful cross-dating of drowned tree ring sequences to those of pre-slide living trees can yield a landslide date; however, trees may not die in the first season of inundation, and

FIGURE 8.55 *Trees can lean for reasons other than slope instability: (a) trembling aspen near Dawson Creek display growth towards the light; (b) black spruce leaning due to ice dynamics in hummocky permafrost terrain west of Fort Nelson. (Photos: M. Geertsema)*

FIGURE 8.56 *Tree scars: (a) scarred western hemlock on debris flow levee; (b) four separate debris flow scars shown on one tree cookie. (Photos: (a) M. Geertsema; (b) J.W. Schwab)*

FIGURE 8.57 *Landslides often dam rivers, inundating their floodplains: (a) in 2007 a rock slide–debris avalanche in northwestern British Columbia impounded the Todagin River, drowning spruce trees; (b) spruce trees drowned by a landslide that occurred in the mid-1600s at Halden Creek in northeastern British Columbia (Geertsema and Clague 2006). The lake filled with sediment, preserving the trees. After the dam broke, and the lake drained, bank erosion exhumed the tree. (Photos: [a] M. Geertsema; [b]. J. Clague)*

large sample sizes are needed. In cases where the drowned trees do not overlap in age with living trees, radiocarbon dating of organic layers in lake sediments behind the landslide dam can be used.

Varve chronologies

Varves, or annual rhythmic sedimentary couplets that occur in certain types of lakes, typically include a silty summer layer and a clayey winter layer. Clay-size particles take longer to settle out of suspension than larger grain sizes, and are thus the only sediment left to settle out of lake water in winter. Episodes of increased sedimentation point to periods of increased instability in watersheds (e.g., Menounos et al. 2006). Sometimes, the distal portions of large debris flows enter lakes. Varve counting above and below such event beds, often in combination with other marker beds such as tephras, can provide precise and accurate dates of the events, but potential sources of error exist. For example, varves can be difficult to count because of ambiguities in the sediment record. The spatial distribution of the sources of the sediment and the landslide deposits usually remain unknown. The data collection is also expensive and time consuming. The method has been applied to the dating of large submarine mud flows in Saanich Inlet (Blais-Stevens and Clague 2001).

Radiocarbon Dating

Radiocarbon dating is one of the most common and useful methods of dating prehistoric landslide deposits. It is useful for century-scale resolution, but can also be used for decadal timescales by using statistical techniques on more than one date.

Living plants take up CO_2 from the atmosphere. Animals also take up this carbon because they either eat plants or eat other animals that have eaten plants. Two stable isotopes of carbon occur: ^{12}C and ^{13}C. The approximate relative ratios of ^{12}C and ^{13}C are 99% and 1%, respectively. The unstable ^{14}C isotope undergoes radioactive decay and has a half-life of 5730 years. ^{14}C occurs in trace amounts (1.2×10^{-10}%) that vary slightly over time. Because the atmospheric production of ^{14}C is not uniform, tree rings have been used to produce absolute calibration of radiocarbon data. The calibration curve plots radiocarbon age against calendar age.

When an organism dies, it ceases to take up CO_2, and thus sets the radiocarbon clock. By measuring the declining proportion of ^{14}C relative to total C, the approximate age of the item can be determined. Conventional radiocarbon dating involves the counting of beta-rays given off as a decay product. The newer Accelerator Mass Spectrometry (AMS) dating method actually counts atoms, and requires much smaller sample sizes.

Radiocarbon dates can be improved by statistical methods such as wiggle matching (Figure 8.58). Using a technique described by Bronk Ramsey et al. (2001), Geertsema and Clague (2006) radiocarbon dated both the pith and outer rings of trees buried by landslide debris near Fort Nelson. Counting the rings between the inner and outer sampled wood allowed for wiggle matching of the radiocarbon dates to improve the statistical precision (Figure 8.58).

Buried wood, charcoal, and basal peat are commonly radiocarbon dated to determine landslide ages. Where datable material overlies a landslide, a minimum age is given, whereas a landslide overlying a datable organic layer yields a maximum age for the event.

Lichenometry

Lichenometry is the dating of lichens by measuring their nominal diameter (Figure 8.59). This technique has been applied to the dating of landslides that expose rock faces and produce boulders. It has been especially useful in Europe, where calibration of *Rhizocarpon* spp. growth on gravestones is possible. Although tried in British Columbia, the lack of growth curves and rigorous statistical sampling has thus far precluded its widespread use. The combination of lichenometry with dendrochronology, AMS radiocarbon dating, and surface exposure dating could help establish growth curves for localized areas of the province.

Simple Comparative and Observational Dating

In many cases, the age of landslides exceeds the age of the trees on their surfaces. Datable elements may exist, but in routine terrain stability field assessments and mapping it is impractical and prohibitively expensive to use expensive dating techniques. Fortunately, landscape indicators provide clues about the age of landslides. These include soil development, degree of acidification, thickness of peat in depressions, rounding and degradation of ridges and scarps, rock colour, and lichen cover. A few examples are given below.

When a landslide removes surface soil, it sets back the pedogenic clock (Geertsema and Pojar 2007).

a

b

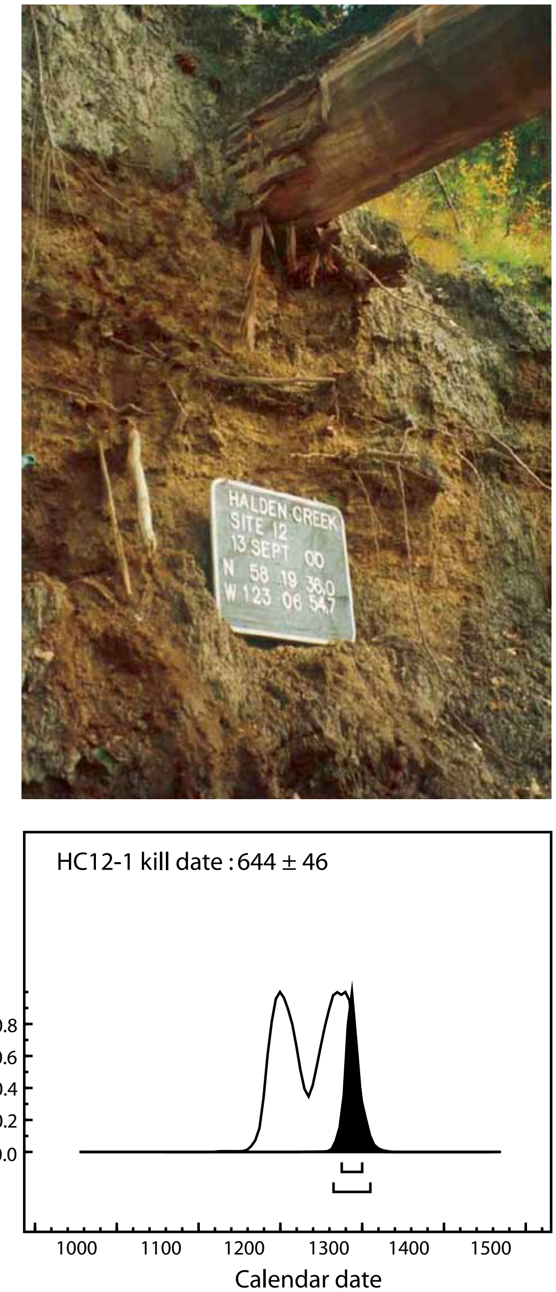

FIGURE 8.58 *Wiggle matching improves the precision of radiocarbon dating: (a) the rings of this cottonwood log buried by landslide debris near Fort Nelson were counted, and radiocarbon dates obtained from the pith and outer wood; (b) the dates were then wiggle matched with the ring count, improving the precision of the calibrated radiocarbon dates from 110 and 130 year errors to 25 and 45 year errors at 68 and 95% probabilities, respectively. This suggests that the kill date of the tree occurred sometime around 1388 AD ±12.5 years at the 1 sigma confidence level. (Modified from Geertsema and Clague 2006)*

Soil-forming processes begin anew in the unweathered soil or rock. For example, a Podzol (a well-developed soil; Soil Classification Working Group 1998) may be removed, restarting the cycle from Regosol to Brunisol to Podzol. Although soil development rates vary with climate and texture, a landslide with a Brunisol should be younger than that of a neighbouring landslide with a Podzol. It needs to be stressed that pre-slide soil horizons can be preserved in rafted blocks and, therefore, a careful evaluation of the context is essential.

Surface soils in forested areas are often more acidic than deeper soils. Testing pH, or simply testing for effervescence with hydrochloric acid, can distinguish between younger and older surfaces in areas where carbonate-rich parent material is evident. For example, exposed depth material at the Mink Creek landslide near Terrace had a pH of 8 in sharp contrast to acidic surface soil with a pH of less than 5. Over time, the high-pH soil would be expected to become acidic.

Many landslides contain water-filled depressions, both in zones of depletion (usually sag ponds) and in zones of accumulation. Organic sediments often accumulate in these ponds, eventually forming peat. Although peat compacts over time, variable thickness of peat in landslide depressions may give some indication of relative age (e.g., Geertsema and Schwab 1997).

Sharp edges of landslide scarps and ridges degrade, becoming more rounded with time (Keaton and Degraff 1996). This is especially true for soil landslides. In some landslides, ridges degrade rapidly, whereas in others ridges persist for millennia, especially in the drier interior of the province. The relative ages of fractures in bedrock can be distinguished, in part on the basis of physical weathering of the rock. In general, older landslides will have a more established drainage network and more subdued relief than fresh landslides.

Rock slide and rock fall scars expose fresh, unweathered rock faces. These often have a lighter tone than the greyish, weathered (often lichen-covered) areas surrounding unweathered rock. On aerial photographs, fresh rock slides have light tones. Without resorting to full lichenometric analysis, a comparison of lichen absence to extensive lichen cover can indicate the relative ages of two surfaces.

FIGURE 8.59 *(a) Lichen-covered rubble in a bedrock spread southeast of Tumbler Ridge; (b) close-up of* Rhizocarpon *species. In lichenometry, several measurements are made from the centre to the outside of the thallus to provide an average measurement of radial growth. (Photos: M. Geertsema)*

SUMMARY

British Columbia's diverse physiography, geology, soils, and climate zones provide a unique template that hosts a wide variety of landslide types and erosion processes. Our mountainous topography results in steep slopes, great local relief, and high precipitation, all of which predispose much of the province to landslide activity. Because landslide types can have drastically different consequences, it is important to distinguish between them. For instance, a slow-moving earth slide has different impacts than a rock avalanche, although both are classified as "landslides." Preconditioning factors, such as weathering, glacial erosion, river erosion, geologic structure, earthquakes, permafrost thaw, wildfire, deforestation, and road construction, can all lead to landsliding. The ultimate trigger may be heavy rainfall, snowmelt, undercutting, an increase in surface loading, or an earthquake.

Erosion is often a precursor of landsliding, and plays an important role in landscape evolution. For example, rain splash, and sheet, rill, and channel erosion contribute to the formation of gullies. Erosion can also occur beneath the ground surface in pipes. Wildfires can cause surface erosion, partly through the creation of water-repellent soils, as well as through the loss of the protective forest floor. Erosion can lead to the "charging" of gullies and stream channels with sediment, which can ultimately lead to damaging debris flows.

The landscape holds records and provides clues about past and potential hillslope processes. Vegetation, slope geometry, and microtopography can reveal important information. For example, wet-site indicators on slopes, tilted trees, spoon-shaped depressions, and hummocky lobes are all signs of slope instability. This evidence can be seen on air photos, and also in the field.

Determining when a landslide occurred is an important component of establishing return periods or landslide frequencies. Dating landslides is accomplished in many ways, including the examination of air photos and written records. A common method of dating landslides is through dendrochronology. Trees may bear scars, or reaction wood, from multiple landslide events. Interpreting the tree rings permits the establishment of landslide frequencies. Other dating methods used in British Columbia include lichenometry, varve counting, and radiocarbon dating.

Land use changes, such as road construction, timber harvesting, and urbanization, can significantly increase the likelihood of landslides by changing the course of surface and subsurface drainage or by loading or undercutting slopes. Chapter 9 ("Forest Management Effects on Hillslope Processes") discusses some of the effects of land use, in particular forest management, on landslide hazard.

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Forest Management Effects on Hillslope Processes

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INTRODUCTION

British Columbia's geologic history and climate contribute to a landscape that is geomorphically active (see Chapter 2, "Physiography of British Columbia," and Chapter 8, "Hillslope Processes"). In many areas, hillslopes are highly sensitive to impacts from changes in forest cover, hydrology, and soil conditions. Soils exposed from logging or road-building operations can erode through rain splash erosion, rilling, or ditch erosion. Vegetated soils can erode if subjected to concentrated flow such as a culvert discharge. Landslides occur when gravitational forces and hydrologic conditions exceed the strength of the soil.

In coastal British Columbia in the 1950s, forest harvesting began to shift from lowland valley-bottom areas onto hillslopes in search of new timber supplies. With some earlier exceptions, forest development in the Interior began to access steep-slope areas only in the 1970s. Harvesting and road construction were done with poor understanding or consideration of potentially unstable terrain, of erosion from roads and exposed soils, or of risks to downslope resources. As a result, forest practices caused increased erosion and increased landslide frequency, which affected water quality, fish habitat, public and worker safety, soil and forest resources, infrastructure, and the public perception of the forest industry. High-profile landslide events in coastal British Columbia in the late 1970s and early 1980s focussed the need for changes in forest management practices.

This chapter provides an overview of the effects of historic forest management practices and how newer practices address issues related to slope stability, sediment production, and alluvial fans. A brief discussion of remote sensing applications is also provided.

Information in this chapter comes from three general sources: (1) research and applied work published in the scientific literature or in government reports; (2) unpublished reports from British Columbia Forest Service files, typically related to operational work such as landslide investigations; and (3) unpublished data and observations collected by the authors in their operational or research work. The authors are practitioners in their professional fields, and collectively have conducted investigations into geomorphic and hydrologic processes and the relationship of these processes to forest management practices in all regions of the province. Where appropriate in this chapter, conclusions reflecting the authors' experience are presented. In this section, some research dealing with the general effects of forest development on watershed geomorphology is summarized. Following sections in the chapter give more detailed information on landslides, erosion, gullies, and alluvial fans in the context of forest practices.

Figure 9.1 shows the Norrish Creek watershed in the lower Fraser Valley, an area heavily affected by landslides caused by roads and logging. Although

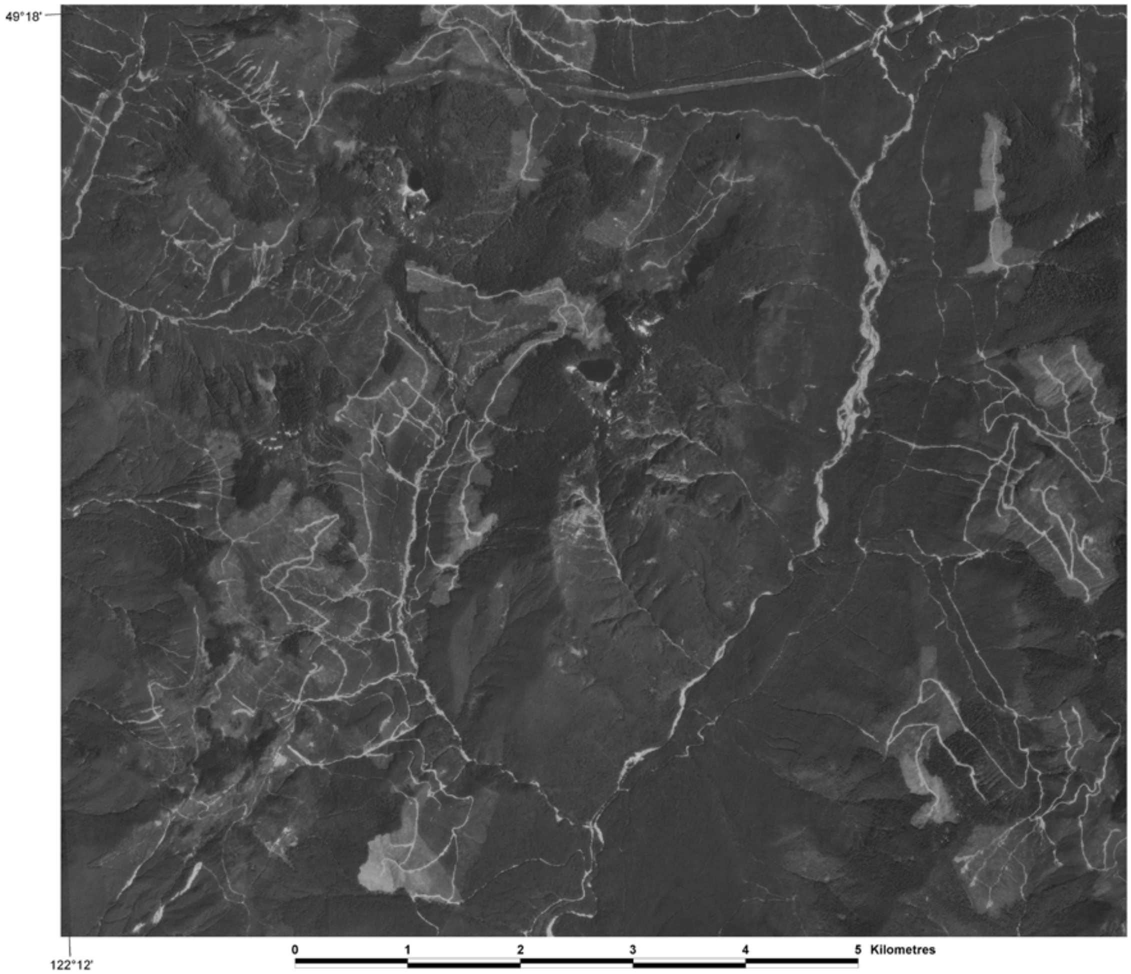

FIGURE 9.1 *Orthophoto prepared from 1995 air photos, showing part of the Norrish Creek watershed near Mission in the lower Fraser Valley.*

this example is not typical of the entire British Columbia coastal region, it is reasonably representative of many areas of the high-rainfall parts of the region (windward side of Vancouver Island, Haida Gwaii, and Coast Mountains), which were logged significantly from the 1960s to the 1980s (Figure 9.2).

Slaymaker (2000) reviewed the impacts of forest development on stream channels, slope stability, and sediment yield, and suggested that landslides and erosion have produced a 10-fold increase in sediment production from harvested areas in British Columbia.

Since the 1970s, several sediment budget studies have examined various aspects of sediment production from logging and roads, and its effects on fluvial systems in the U.S. Pacific Northwest (Beschta 1978; Kelsey 1980; Swanson et al. [editors] 1982; Reid and Dunne 1984; Megahan et al. 1986) and in British Columbia (Roberts and Church 1986; Jordan and Commandeur 1998; Hudson 2001; Toews and Henderson 2001). All of these studies identified the significance of forest roads as sediment sources due to both surface erosion and their role in causing landslides.

In coastal British Columbia streams, coarse sediment (bedload) from landslides has long been recognized as an important part of the sediment budget (Roberts and Church 1986). In watersheds that have been heavily affected by logging, increased landslide frequency is believed to have caused approximately an order of magnitude increase in sediment supply (Rood 1984; Schwab 1988). However, in one detailed study, Hudson (2001) examined coarse- and fine-sediment budgets during rainstorm events in Russell Creek, a 30-km^2 partially logged watershed on Vancouver Island. He found that landslides were the

FIGURE 9.2 *Numerous landslides caused by road-fill failure and harvesting effects in an area that was heavily logged in the 1970s. Many slides occurred 10 years or more after logging or construction. Mamquam River, Squamish Forest District. (Photo: P. Jordan)*

most important contributor of suspended sediment (followed by gully erosion), and that development-related sediment sources were minor compared to natural sources within the watershed.

In the Interior of British Columbia, long-term watershed studies with a sediment budget component have been conducted in the Stuart-Takla watersheds northwest of Prince George, the West Arm Demonstration Forest near Nelson, and the Penticton Creek watershed in the Okanagan.

In the Stuart-Takla project, Macdonald et al. (2003) measured streamflow and suspended sediment concentration in three small watersheds (0.4–1.0 km^2). A significant increase in suspended sediment occurred following logging and lasted for 2–3 years. Stream crossings were considered the biggest contributor of the sediment.

On Redfish Creek, a 26-km^2 partially logged watershed near Nelson, which had 10 years of data, Jordan (2006) found that the amount of suspended sediment from forest development (mainly from road erosion) was comparable to that produced by natural sources. In comparison, at Gold Creek near Cranbrook, a less mountainous and lower-precipitation area, development-related suspended sediment was insignificant. Based on limited measurements and observations in other watersheds in the region, Jordan (2001b) concluded that, on the average, sediment from development-related landslides was much less significant than erosion from roads, but that in isolated instances, landslides could dominate the sediment budget for several years in watersheds where they occurred.

Many mountainous areas of British Columbia have high levels of geomorphic activity, especially where there are recent (Neoglacial) glacial deposits, recent volcanism, or weak sedimentary rocks, and areas with high rates of precipitation (Guthrie 2005; see Chapter 2, "Physiography of British Columbia"). However, many forested mountain and plateau areas in the British Columbia Interior have little geomorphic activity and very low sediment yields when compared with other mountain regions of the world (Slaymaker 1987; Church et al. 1989; Jordan 2006). In these areas, where many streams naturally carry very little sediment, erosion and landslides from forest development can have a significant impact on watershed sediment budgets.

In the 1970s and 1980s, several reviews and inventories of landslides and erosion caused by forest development in the Pacific Northwest were conducted in response to increasing concern about the environmental effects of forest practices. Swanston and Swanson (1976), reviewing data from several study areas in Washington, Oregon, and British Columbia, found that the occurrence of debris avalanches and debris flows increased by 2–9 times in clearcuts, and by 25 to over 300 times due to roads, compared to undeveloped forest (based on area covered by roads, not total development area). Rood (1984), in an inventory of landslides on Haida Gwaii (formerly the Queen Charlotte Islands), reported that frequencies of debris slides increased by 30 times and debris flows increased by 41 times, and estimated that 80% of the resulting sediment yields came from clearcuts, whereas 20% came from roads. Sidle et al. (1985), reviewing data from Oregon, found that the rate of landslides increased up to 23 times in clearcuts, and that increases in mass movement rates were typically observed for 10–15 years.

Beginning in the 1980s, terrain attribute studies were introduced in British Columbia to more rigorously study the frequencies and causes of development-related landslides. A terrain attribute study is a statistical analysis relating landslide occurrence to variables that describe terrain features (attributes) and that can be measured or mapped (Howes 1987; Rollerson 1992). These variables can be quantitative (e.g., slope angle or gully depth) or qualitative (e.g., terrain category, bedrock lithology, or slope position). The purpose of terrain attribute studies is to provide quantitative predictions of landslide frequencies that are likely to occur from forest development in a particular region, and to determine which terrain attributes are most significant in predicting landslide occurrence. Most terrain attribute studies have been conducted in coastal British Columbia (Rollerson 1992; Rollerson et al. 1998, 2001, 2002; Millard et al. 2002). These typically consist of an inventory of landslides caused by forest development in a study area and detailed terrain mapping of the study area to collect the required terrain attribute data. Only a few terrain attribute studies have been conducted in the Interior (Jordan 2002). Since landslide frequencies are relatively low there compared to the Coast, and large study areas must be used, terrain attribute studies in the Interior have used operational terrain stability mapping done by the forest industry, or have used terrain attributes not based on terrain mapping.

Landslide frequencies in terrain attribute studies have usually been reported as the number of landslides per square kilometre per year (Ls/km^2 per year, where Ls = number of landslides) in areas where the approximate dates of logging and landslide occurrence are known. In some studies, including most early air photo–based inventories, if the dates of landslides are not well known, areal frequencies (or landslide densities) are reported as the number of landslides per square kilometre.

Landslide rates due to harvesting are variable in coastal British Columbia, and are generally higher in areas of higher precipitation. Steep slopes (> 20°) on Haida Gwaii are reported to have the highest post-clearcut landslide rate—about 1–1.7 Ls/km^2 per year—based on the assumption of a 10-year post-harvest period of landslides (Schwab 1988; Rollerson 1992). The west coast of Vancouver Island and the Cascade Mountains south of the Fraser Valley also have high post-clearcut landslide rates on steep slopes of about 0.7 Ls/km^2 per year (Rollerson et al. 1998, 2002; Millard et al. 2002). Horel (2006) reported landslide frequencies on steep slopes on northern Vancouver Island of 8.6 Ls/km^2 before the implementation of the *Forest Practices Code of British Columbia Act* (FPC) and 4.9 Ls/km^2 post-FPC. Again, assuming a 10-year period of post-harvest landslides, the pre-FPC landslide rate of 0.86 Ls/km^2 per year is similar to other Vancouver Island studies. When entire watershed areas are considered, rather than just steep slopes, landslide rates are much lower. Rates on Vancouver Island are 0.07–0.17 Ls/km^2 per year (Jakob 2000; Guthrie 2002), and in the Coast Mountains, they average 0.13 Ls/km^2 per year in exposed windward areas and 0.04 Ls/km^2 per year in drier leeward areas (Rollerson et al. 2001).

Studies in the Kootenay and Columbia regions of southeastern British Columbia reported landslide densities for development-related landslides of 0.38–1.07 Ls/km^2, with the highest densities occurring in the higher-relief and wetter terrain of the northern Selkirk and Monashee Mountains (Jordan 2001a, 2002) (Figure 9.3). Landslide frequencies in these areas increased 4–9 times because of forest development. The landslide densities for these regions are based on a time frame of about 20–30 years—that

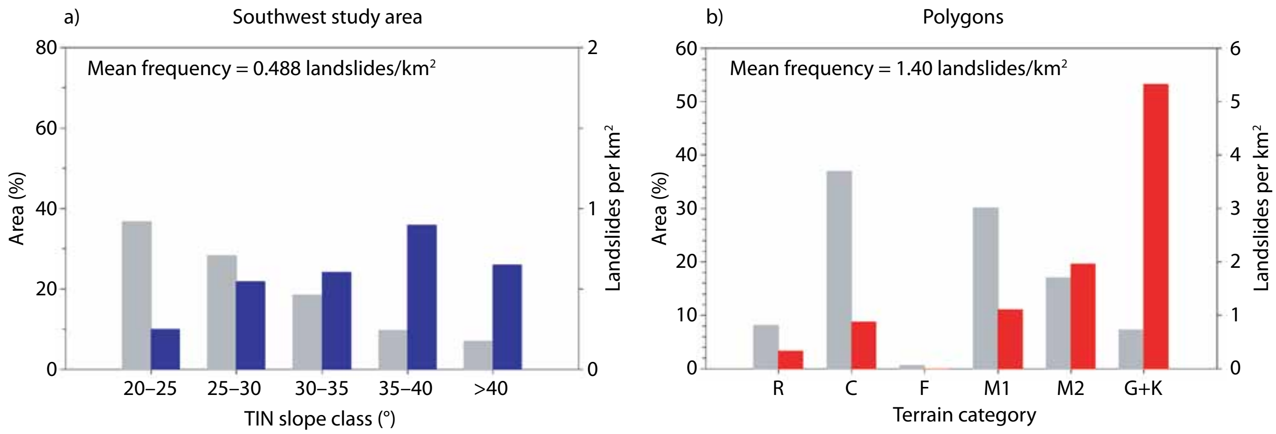

FIGURE 9.3 *Results from terrain attribute studies in the southeastern interior of British Columbia. The graphs show the relationship between densities of development-related landslides vs slope class and terrain category in two different study areas. Grey bars indicate abundance of each class or category in the study area; coloured bars indicate landslide density. (a) Southwest study area (southern Selkirk Mountains) landslide density by slope class. Areas with slope less than 20° are not included in the analysis. "TIN" is "triangular irregular network," a form of slope map that is more precise than grid-based slope maps. (b) Detailed terrain map study areas, Selkirk and Purcell Mountains, landslide density by terrain category. Categories are rock (R), colluvial (C), fluvial (F), shallow morainal (M1), deep morainal (M2), glaciofluvial/kame (G+K). (From Jordan, 2003, unpublished report)*

is, the period in which landslides can be reliably identified on air photos. Therefore, they translate to landslide frequencies of about 0.01–0.05 Ls/km^2 per year—an order of magnitude lower than most of the landslide frequencies measured in the coastal studies.

The time period used for estimating landslide frequencies on the Coast is shorter than that used in the Interior because revegetation is faster on the Coast, obscuring small landslides sooner. Also, most landslides on the Coast occur in clearcuts within 10 years of logging, whereas most landslides in the Interior occur in unlogged forest below roads, and can occur many years after construction.

Terrain attribute studies have typically shown that the most significant variables in explaining development-related landslide occurrence in coastal British Columbia are slope angle, presence of gullies, gully depth, presence of natural landslides, and slope configuration; however, these results vary somewhat between study areas. In the southeastern interior, the most significant variables were terrain category (with glaciofluvial and deep morainal deposits having the highest landslide densities), presence of gullies, and presence of natural landslides.

There is an important difference in study results for interior and coastal British Columbia. In the Interior, most development-related landslides are caused by roads (about 80%)[1] and most cutblock-related landslides are caused by skid trails. On the Coast, some studies indicate that most landslides are caused by clearcuts (Schwab 1988; Rollerson 1992; Rollerson et al. 1998, 2002; Millard et al. 2002), whereas others show that roads and harvesting are of about equal importance (Rollerson et al. 2001; Guthrie 2002). The proportion of coastal road-related slides may be related to the inherent instability of the area. Areas with high post-logging landslide rates, such as Haida Gwaii, tend to have a higher percentage of cutblock-related slides, whereas areas with lower post-logging landslide rates, such as the Coast Mountains, tend to have a greater proportion of road-related landslides.

1 Jordan, P. 2003. Landslide and terrain attribute study in the Nelson Forest Region. B.C. Min. For., Res. Br., Victoria, B.C. FRBC Proj. No. KB97202-ORE1. Unpubl. report.

Harvesting can result in changes to soil moisture and soil strength that can cause landslides. Harvested areas tend to receive more snow and undergo more rapid snowmelt than unharvested areas; consequently, higher inputs of moisture into the soil can occur in harvested sites in both coastal areas, which are subject to rain-on-snow events, and interior areas, which are subject to spring melt (Chapter 7, "The Effects of Forest Disturbance on Hydrologic Processes and Watershed Response"). Increased intensity of water input to the soil may result when the attenuating effect of a forest canopy is removed. Keim and Skaugset (2003) compared 1-minute-long rainfall intensities in sites beneath a forest canopy and adjacent sites without a forest canopy. In general, lower precipitation intensities were recorded in sites underneath a forest canopy. It is possible that short-term spikes in rainfall intensity may result in higher pore pressures in soils that are not protected by a forest canopy (Horel 2006). Although rainfall intensity is an important determinant in landslide initiation on the Coast, other factors such as antecedent conditions and snowmelt from rain-on-snow events are also important causes (Jakob and Weatherly 2003).

Harvesting and yarding of trees may affect the soil structure, which can disrupt preferential drainage pathways in the soil. For example, soil macropores can be crushed when cut trees land on the soil or when trees are dragged across the surface during yarding. This can result in increased pore pressures and subsequent landslides (Ziemer 1981).

The loss of root strength caused by roots rotting or yarding disturbance can also result in landslides. Root strength, which is generally treated as an added cohesive element in slope stability analysis, is lost when trees are cut and the roots rot. Ziemer (1981) found that soil strength is most strongly correlated with the total weight of live roots less than 5 mm in diameter; roots greater than 5 mm in diameter had less of an effect on soil strength. Small roots rot quickly. Hemlock and Sitka spruce roots less than 5 mm in diameter lose 40–90% of their strength within 2 years of harvesting (Ziemer and Swanston 1977). For an extensive discussion of root strength, see Sidle and Ochiai (2006).

Few methods are available to reduce landslide occurrences that result from harvesting. Some of the effects of harvesting, such as changes in water inputs and loss of root strength, are inevitable once a slope is harvested. Therefore, the primary method of reducing landslide occurrences is to avoid slopes that are likely to have post-harvest landslides, particularly if those slopes are above high-value resources or pose a health or safety concern. Some harvest methods can theoretically reduce landslide rates, but no studies have been conducted to quantify these effects. Partial harvesting, particularly single-stem harvesting, is less likely than clearcutting to cause significant loss of root strength due to root decomposition and increased delivery of water to the soil. "Snap-and-fly" single-stem harvesting, in which a helicopter is used to lift a tree directly off its stump, should not cause any damage to soil drainage pathways; therefore, it is less likely to cause post-harvest landslides, but has not been the subject of detailed research.

An important terrain stability factor to consider in the placement of cutblock boundaries is the likelihood of windthrow. A new study of landslides near Bamfield on Vancouver Island showed that windthrow was associated with half of the open-slope landslides that occurred during winter 2006–2007 (J. McDonald, MSc candidate, Simon Fraser University, pers. comm., Jan. 30, 2008). Wind-exposed edges of cutblocks should not be located on the edge of potentially unstable terrain such as gully edges. For a more detailed discussion of windthrow management, see Stathers et al. (1994). Figures 9.4–9.9 show examples of harvesting-related landslides.

FIGURE 9.4 *An unusual clearcut-caused debris slide/avalanche complex at Gorman Creek near Golden, in the Interior of British Columbia. The slide occurred the spring (about 1990) following harvesting and probably resulted from yarding disturbance combined with increased snowmelt rate following logging. (Photo: D. Toews)*

FIGURE 9.5 *Debris slides that initiated within a helicopter-yarded cutblock a few years after harvesting at Indian Arm in the Chilliwack Forest District. (Photo: T. Millard)*

FIGURE 9.6 *Headscarp area of a typical harvesting-related debris slide, Klanawa River, South Island Forest District. (Photo: T. Millard)*

FIGURE 9.7 *Transport zone of debris slide shown in Figure 9.6. Note the role of the tree roots in retaining sediment, and how the slide eroded to bedrock downslope of the rooting zone. (Photo: T. Millard)*

FIGURE 9.8 *Slopes in north Clayoquot Sound after a 4-day storm event in 1996 resulted in 273 landslides. (Photo: T. Millard)*

FIGURE 9.9 *Another view of Clayoquot Sound after the 1996 storm event, showing a mix of harvesting and road-related landslides. (Photo: T. Millard)*

Forest roads have been identified as the cause of many landslides in both coastal and interior British Columbia. Before about the 1980s, most forest roads, including those on steep slopes, were built using bulldozers and employed sidecast construction (Fannin et al. 2005; see Chapter 5, "Forest Practices"). This resulted in unstable fills on steep slopes, and combined with inadequate drainage control and inspection, caused many road-fill landslides.

In a study in the Kalum Forest District, VanBuskirk et al. (2005) found that the terrain factors that contributed significantly to road-fill landslides were presence of natural slope instability, gullied terrain, deep surficial materials, and moderate to imperfect drainage. Significant contributing engineering factors included perched fills (supported by logs, stumps, or trees) and poor drainage control.

Table 9.1 shows probable causes of development-related landslides based on a sample of air photos of 429 landslides in the Arrow and Kootenay Lake Forest Districts.[2] Field checking of about half the landslides confirmed the general distribution of these causes, as did numerous operational investigations of landslides. In this study area, road-fill failures were more common on older roads where weak fills on steep slopes were more likely.[3] On newer roads (especially those built after 1995), most failures were found to be caused by inadequate or poor road drainage. The authors' operational experience in this region supports these general conclusions, and provides additional information about the causes of typical landslide events. Landslides related to road drainage frequently occur below a culvert. Often, sufficient cross-drain culverts have been installed but minor errors have been made in culvert location, resulting in water being discharged at an unstable location. Sometimes more water is intercepted by the road cut than was anticipated by the design, but the appropriate drainage upgrades were not made. Landslides may also occur when a ditch or culvert is blocked and water is diverted down a road to a potentially unstable location. In many cases, more frequent inspections or the use of seasonal water bars could have prevented the landslide, which suggests that maintenance is an important factor in reducing road-related failures.

TABLE 9.1 *Landslide characteristics in the Arrow and Kootenay Lake Forest Districts (Jordan 2003, unpublished report)*

Landslide characteristic	**Observed cases (%; *n* = 429)**
Road-fill failure	41
Road-cut failure	5
Road drainage	31
Skid trail drainage	10
Skid trail fills or cuts	2
Clearcuts (unrelated to roads or skid trails)	2
Other causes unrelated to forestry	9

Figures 9.10–9.17 show several examples of landslides caused by road-fill failures and drainage diversions. In particular, Figure 9.14 illustrates an example of several very large debris avalanches that originated as road-fill failures in old roads containing wood in the fill. The incident, which occurred near Chilliwack in 1990, was instrumental in motivating the B.C. Ministry of Forests to initiate a program to deactivate old roads in landslide-prone terrain. This became the Watershed Restoration Program (Atkins et al. 2001).

2 Jordan, P., 2003. Unpublished.
3 Jordan, P., Unpublished data, Southern Interior Forest Region.

FIGURE 9.10 *Debris avalanches (1993) caused by road-fill failures and drainage diversions on slopes logged in the 1970s at Airy Creek in the Arrow Forest District. (Photo: P. Jordan)*

FIGURE 9.11 *Road-fill failure in the Kid Creek area of the Kootenay Lake Forest District, 2007; failure was caused by drainage diversion that saturated the road fill. (Photo: P. Jordan)*

FIGURE 9.12 *Debris slide that progressed to a large debris flow below a culvert in the Blueberry Creek area of the Arrow Forest District, 1993. The slide happened the spring following logging, occurring below a culvert that was misplaced by only a few metres from its optimum location. (Photo: P. Jordan)*

FIGURE 9.13 *Debris slide/avalanche that progressed to a large debris flow below a culvert in the Fortynine Creek area of the Kootenay Lake Forest District, 1996. Water flow to the culvert increased because of the cumulative effects of roads and cutblocks over a 20-year period, although culverts, ditches, and road maintenance generally followed accepted standards. (Photo: P. Jordan)*

FIGURE 9.14 *Landslides at Wahleach Lake, Chilliwack Forest District. These landslides occurred during a rainstorm in 1990, and originated in old road fills that were partially supported by wood. Poor drainage control on the roads, which were not deactivated, may also have been a contributing factor. (Photo: P. Jordan)*

FIGURE 9.15 *Debris slide/avalanche that involved a fill failure of a mainline Forest Service road near Giveout Creek in the Kootenay Lake Forest District, 2002. The debris slide was caused by drainage diversion on skid trails on private land above. Although ground-based logging of steep slopes is no longer practised on Crown land, it still sometimes occurs on private land. (Photo: P. Jordan)*

FIGURE 9.16 *Two road-fill slope slides associated with uncontrolled road drainage in the San Juan watershed, South Island Forest District. (Photo: T. Millard)*

FIGURE 9.17 *Road surface and slide headscarp from Figure 9.16. Note heavily eroded road surface. Several blocked culverts led to large volumes of water flowing onto the fillslope at the headscarp of both slides. (Photo: T. Millard)*

Drainage diversions from forest roads and trails have been recognized as an important cause of landslides in the British Columbia interior since the late 1980s. In many cases, forest development has occurred on relatively gently sloping terrain, plateaus, or benchlands above steeper slopes. Landslides occurring in such topographic situations are commonly referred to as "gentle-over-steep" landslides (e.g., Figures 9.18–9.20).

Forest development in such areas has been widespread in the British Columbia interior for several reasons. The main reason is that large areas of commercially valuable forests on plateaus and uplands lie above incised valleys; however, most of the potentially unstable terrain occurs on the steep valley sides. Most elements at risk, such as populated areas, infrastructure, and fish habitat are at the base of these slopes. The gently sloping terrain was developed first, since logging costs are lower on gentler slopes, and most logging was ground-based, which creates opportunities for drainage diversion by skid trails and roads. In areas where forest development spanned steeper and gentler slopes and used cable-harvesting systems, roads were typically located on the gentler slopes to minimize costs and avoid unstable terrain.

Historically, many gentle-over-steep landslides were caused either by drainage diversions on roads that had insufficient or substandard culverts or other drainage structures, or by skid trails that had not been deactivated. With the introduction of the *Forest Practices Code of British Columbia Act* in 1995, or through changes implemented some years earlier by many forest companies, it became standard practice to deactivate roads and skid trails; however, many landslides continued to occur because of drainage diversions from older roads and logged areas. An important benefit of improved road engineering practices was a reduction in landslides caused by road-fill failure. Nevertheless, landslides that

FIGURE 9.18 *"Gentle-over-steep" debris avalanche below road and ground-based harvesting near Shaw Creek in the Kootenay Lake Forest District, 1996. In this case, the road and skid trails were properly deactivated, and there were no drainage diversions observed; subtle changes to slope hydrology may have been responsible. (Photo: P. Jordan)*

FIGURE 9.19 *"Gentle-over-steep" debris avalanche caused by drainage diversions on road and skid trails near Ferguson Creek in the Arrow Forest District, about 1990. (Photo: P. Jordan)*

FIGURE 9.20 *"Gentle-over-steep" debris flow caused by road drainage diversion on Lower Arrow Lake in the Arrow Forest District, 1999. On a newly constructed road, two culverts were not installed on a small ephemeral stream, and water was discharged off a switchback. This caused the debris flow, which began as a small debris slide below the switchback. (Photo: P. Jordan)*

occurred below roads continued to be a significant problem (Grainger 2002). Investigation of many such incidents by B.C. Ministry of Forests engineering and geoscience specialists showed that FPC regulatory requirements had a significant weakness; that is, the requirement for a terrain stability assessment by a qualified professional was based on the terrain stability at the site, not on terrain conditions below the proposed development. Thus, in many cases, drainage control on roads or in proposed logging developments was never reviewed or prescribed by a qualified specialist in gentle-over-steep situations (although in some high-risk areas, assessments were done voluntarily by forest companies, or were required by the forest districts).

In the British Columbia interior, drainage diversions by roads in gentle-over-steep situations were first recognized as a serious problem in an engineering report to the Arrow Forest District.[4] In 1990, a group of gentle-over-steep debris flows near Slocan resulted in a major investigation,[5] and recommendations for more regulation and improved practices for road and skid trail deactivation were made. The hydrologic and engineering causes of such landslides were reviewed by Jordan (2001a), Grainger (2002), and Jordan and Nicol (2002).

Gentle-over-steep landslides (and other drainage-related landslides downslope from forest development) are caused by the hydrologic changes brought about by roads or logging. These hydrologic changes can include:

- dispersed subsurface flow brought to the surface and concentrated into discrete pathways by ditches, culverts, water bars, or water discharged off switchbacks;
- overland flow generated from road or trail surfaces, which is discharged onto a slope below at discrete locations;
- accidental discharge of water at an unplanned location because of the failure of a ditch or culvert; and
- increased generation of runoff from snowmelt or rainfall that is due to the effects of forest harvesting.

The physical processes that lead to increased landslide hazard are the same whether or not the contributing drainage area upslope is of lower slope steepness; landslides are frequently caused by drainage concentration or diversion below roads that are on steep terrain. However, the gentle-over-steep configuration is common in areas of plateau topography because of the prevalence of forest development on gentle slopes.

Figures 9.21 and 9.22 are conceptual sketches showing the typical effects of roads on slope hydrology and landslide initiation. Shallow subsurface flow is intercepted and concentrated by road cuts and ditches. A landslide may occur downslope if the increased flow of water encounters a location where the slope is barely stable (factor of safety close to 1.0) and requires only slightly increased pore water pressure to fail. Typically, these failures occur where the slope steepens, and also where the soil thins downslope, bringing the water table close to the surface.

Case Study: Belgo Creek Debris Avalanche

Three case studies are presented below (Belgo Creek, Donna Creek, and Hummingbird Creek) as examples of gentle-over-steep failures with significant downslope or downstream impacts.

Six slope failures occurred in the vicinity of Belgo Creek, 30 km east of Kelowna, on June 11–13, 1990. The most destructive event (Figure 9.23) was a large debris avalanche on June 12, which destroyed a house on Philpott Road and resulted in three fatalities (Schwab et al. 1990; Fannin et al. 2005).

The landslides occurred during a 3-day rain-on-snow event, which was heavy but not exceptional (Schwab et al. 1990). This followed 1.5 months of very wet weather; May and June of 1990 had the highest 2-month rainfall on record at Kelowna.

The landslide started as a group of four small debris flows in an area of old logging (pre-1963). The lowest of these rapidly increased in width and volume downslope, flowing a total of 1.5 km and eroding an area of 5 ha. The deposits of saturated debris spread out over an area of 12 ha in the valley bottom. The magnitude of the event was estimated at 23 000 m^3 (Cass et al.1992). The event was originally classified as a debris flow; however, according to the more recent classification of Hungr et al. (2001), it would more properly be called a "debris avalanche."

The slope in the landslide start zone averages 33°

4 Chow, B. 1988. Review of logging related slope failures of April 1988 in Arrow Forest District, Nelson Forest Region. B.C. Min. For. Unpubl. report.

5 Curran, M., B. Chow, D. Toews, and D. Boyer. 1990. Landslide study: Cape Horn bluffs area. B.C. Min. For. and B.C. Min. Environ. Unpubl. report.

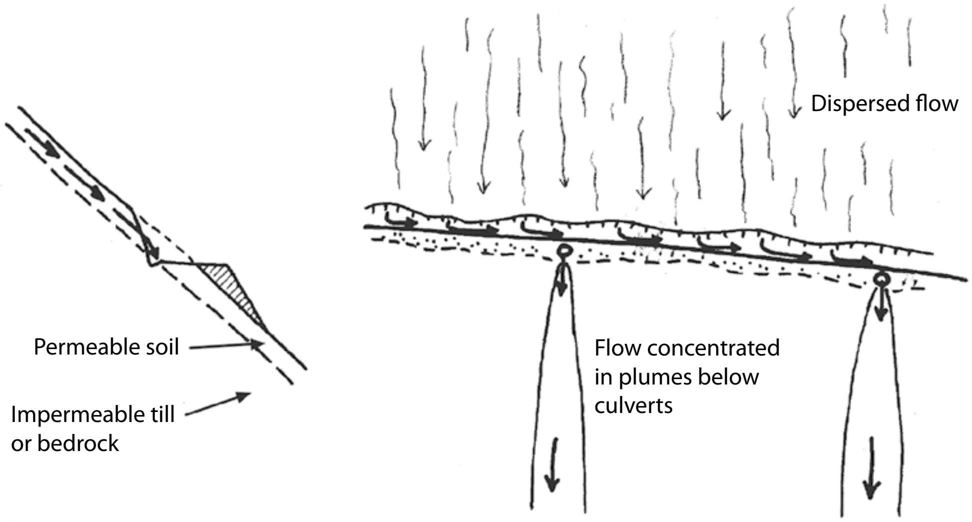

FIGURE 9.21 *Sketch showing the typical effect of roads and culverts on slope hydrology for uniform slopes with shallow soils and low drainage density (Jordan 2001a).*

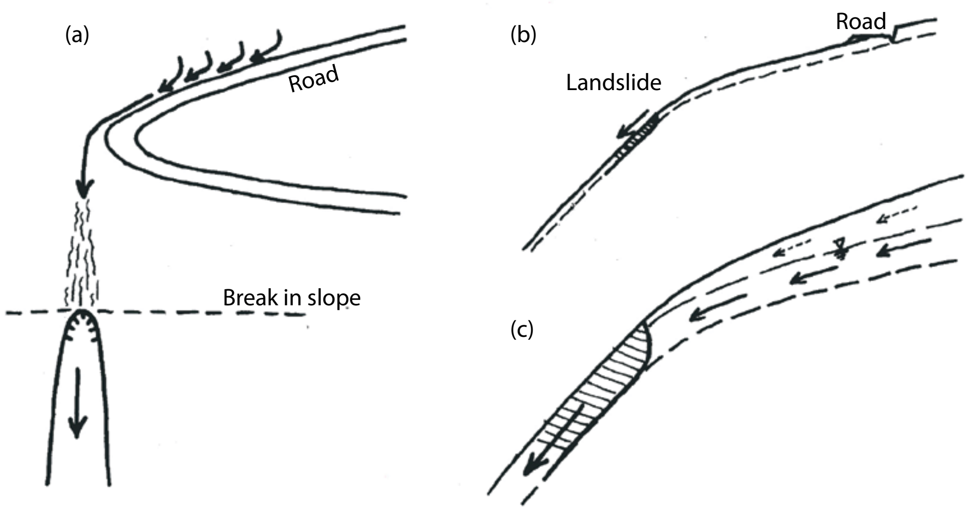

FIGURE 9.22 *Sketch showing the typical role of a road drainage diversion in triggering a "gentle-over-steep" landslide: (a) Plan view, landslide below a switchback;(b) cross-section showing typical point of landslide initiation; and (c) enlargement, showing convergence of groundwater caused by the thinning soil depth at break in slope. Often, landslides initiate in a concavity on the contour where horizontal convergence of flow also occurs (Jordan 2001a).*

(65%), and is covered with veneers and blankets of sandy till and colluvium. The investigation following the event showed evidence of extensive small-scale slope instability in the area and a lack of defined surface watercourses, but no evidence of previous large landslides. On the plateau above, large clearcuts had been created from 1983 to 1988. The roads accessing these clearcuts had been deactivated with water bars. Selective logging had occurred before 1963 in and above the landslide start zone and in the upper part of the slide path, on both Crown and private land. Most of the logging roads in that area had not been deactivated. Roads in the clearcut above the landslide increased the contributing drainage area by about 20%, and it is estimated that the effects of the clearcut increased the rate of runoff during the storm by about 15%.

This event was important in the history of forest geomorphology in British Columbia because it led to awareness, for the first time, of the risks presented by landslides following forest development in the Interior. This contributed to:

1. the development of regulations in the FPC that were aimed at minimizing landslide hazards;
2. the development of the Watershed Restoration Program, which aimed to reduce risks by deactivating abandoned logging roads throughout the province; and
3. the use of terrain stability mapping as a planning tool.

FIGURE 9.23 *Aerial view of the Belgo Creek debris avalanche. (Photo: J. Schwab)*

Case Study: Donna Creek Washout Debris Flow

The catastrophic washout / debris flow at Donna Creek, near Mackenzie, in June 1992 emphasizes a very basic concept in our knowledge of water source areas (drainage basins) and how roads modify water movement. Roads constructed on the hillslope intersected poorly defined microtopographic drainage basins (Figure 9.24). Water was captured and routed along road ditches. Sufficient cross-drains or culverts were not provided to allow water to cross the roads and flow into natural drainage basins (Figure 9.25). The capture and redirection of surface and subsurface water expanded the drainage area to an existing basin outlet. The expanded drainage represented a catchment increase of 9.8 times (5.8 ha to 56.9 ha). An excessive volume of water, in the order of 193 000 m^3 over 9 days during a period of rapid snowmelt, was delivered onto a glaciofluvial-glaciolacustrine terrace composed of highly erodible sands, silts, and gravels. Even though the site appeared to be well drained, it could not handle a rapid increase in groundwater discharge at the terrace scarp. A rapid increase in seepage exit gradient resulted from the sudden increase in groundwater discharge combined with a possible sudden removal of material at the terrace scarp by a landslide and (or) running water. Catastrophic seepage face erosion describes the sudden change in the discharge conditions that resulted in a rapidly retreating seepage face, caving, collapse, and debris flow surges (Figure 9.26).

FIGURE 9.24 *Natural flow pathways downslope (right to left in photo) are intersected by roads on the hillslope above Donna Creek (far left). (Photo: J. Schwab)*

FIGURE 9.25 *Insufficient cross-drainage resulted in the capture and routing of water along road ditches. (Photo: J. Schwab)*

FIGURE 9.26 *Water discharged onto a highly erodible terrace resulted in the catastrophic seepage face erosion. (Photo: J. Schwab)*

Case Study: Hummingbird Creek Debris Avalanche and Debris Flow

On July 11, 1997, a large debris flow descended Hummingbird Creek and caused considerable damage to property and infrastructure on its alluvial fan. The fan is located on Mara Lake, 8 km south of Sicamous. About 150 houses were located on the fan, and Highway 97A crosses the upper part of the fan. Several unoccupied summer homes and other buildings were damaged by the debris flow, the highway was washed out, and further minor damage occurred below the direct impact zone due to inundation by water and fine debris. There were no direct fatalities or injuries. The incident was investigated by a team of government staff and consultants (Anderson et al. 1997; Jakob et al. 2000). Wilford et al. (2009) undertook a retrospective analysis of the event; Figure 9.27 provides an overview of the area.

The event started as a debris avalanche 35 m below a culvert on a logging road. At this location, roads and logging are on relatively gentle terrain, which drops off steeply below the road to the creek. The terrain consists of shallow till and colluvium over coarse-textured metamorphic bedrock that dips downslope. The slope at the landslide scarp was about 35° (70%). Pre-landslide terrain stability mapping rated this location as "potentially unstable." [6] The landslide occurred after a week of heavy rain; the 7-day rainfall at three nearby climate stations averaged 112 mm, an event that is estimated to have a return period of about 60 years. This followed the melting of an above-average snowpack, so antecedent soil moisture was likely high. During the May–July period, numerous other landslide and erosion events in the area were caused by wet weather and a rapid melting of a deep snowpack.

The culvert directly above the landslide drains an area within a cutblock that was logged in 1994. A spur road in the cutblock caused the drainage area above the culvert to more than triple in size from 1.6 to 5.6 ha. A second culvert 65 m up the road drains a larger area, and contributed water to the slope in the lower part of the debris avalanche. Concentration

6 Terratech Consulting Ltd. 1997. Reconnaissance terrain stability study (TSIL D), Riverside Forest Products Ltd. Hunters Range Operating Area, Salmon Arm Forest District. Draft submitted to Riverside Forest Products Ltd., Lumby, B.C. Unpubl. report.

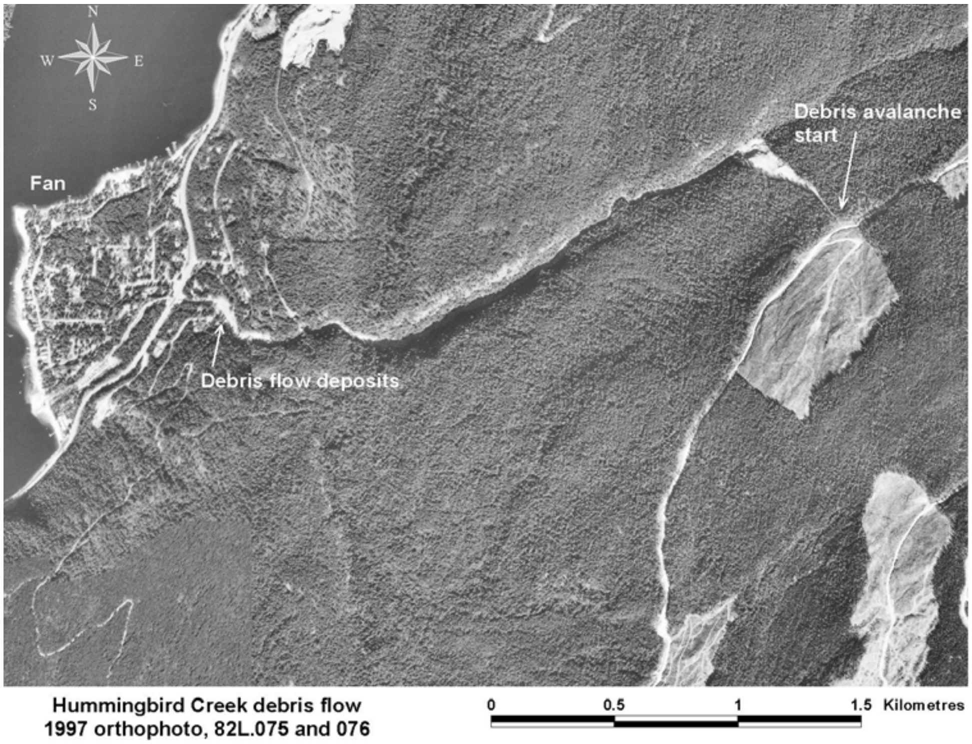

FIGURE 9.27 *Orthophoto prepared from air photos taken soon after the debris flow event.*

of runoff by the cutblock road and the culverts was identified as the main development-related factor that caused the landslide.

The debris avalanche was exceptionally large, with a volume estimated at 25 000 m^3. On entering the creek, the additional water transformed the mode of transport to a debris flow (Figure 9.28). The debris flow scoured gravel, soil, and trees from the creek bed and banks, and continued for about 2.5 km to the fan. Once on the fan, the debris flow lost confinement and began depositing sediment. Most of the coarse debris was deposited in the first 400 m, above the highway on the upper part of the fan, where the slope is about 6–10° (10–18%). Finer debris and flood deposits were spread over a large area of the lower fan, which has a gentler slope and is occupied mainly by roads and houses. The total volume of debris flow deposition was about 92 000 m^3 (Jakob et al. 2000).

Except for its unusually large size, this event is fairly typical of "gentle-over-steep" landslides, which are common in the Columbia Mountains and Shuswap Highlands (Grainger 2002). These landslides are caused when runoff is concentrated by logging roads and trails and is discharged onto steeper slopes below. In this region, such events can present risks to life and property in the inhabited valley bottoms (Wilford et al. 2009).

A Forest Practices Board investigation concluded that the forest licensee adequately maintained the road and complied with FPC requirements. The report recommended revisions to FPC regulations to include assessment of risks to life and property, and to reduce hazards from road drainage being discharged onto potentially unstable slopes (Forest Practices Board 2001).

Photograph 1

Looking downstram along the scoured Hummingbird Creek channel toward Mara Lake, with the toe of the debris scarp in the foreground.

Photograph 2

Debris flow scarp in relation to the Ashton - Mara East FSR, CP 733 - Block 3, and Hummingbird Creek.

FIGURE 9.28 *Photographs of the initial landslide (right) and the scoured channel upslope of the alluvial fan (left) (Anderson et al. 1997).*

SURFACE EROSION

In the Pacific Northwest, numerous studies have indicated that surface erosion from forest roads is an important source of sediment (Reid and Dunne 1984; Megahan et al. 1986; Hudson 2001). These studies have generally shown that sediment production is greatest during and soon after construction, and that the level of industrial traffic on roads greatly affects sediment production. An important distinction between landslides and surface erosion is that landslides are episodic, whereas surface erosion from roads is chronically occurring.

Factors affecting the importance of roads as sediment sources include bedrock and surficial geology, soil texture, presence of seepage, and the relative contributions of surface and subsurface flow during spring runoff [7] or storm events. Engineering factors are also important, and include road location, road gradient, adequacy of drainage control, the use of road surfacing material, and inspection and maintenance practices (B.C. Ministry of Forests 2002). Additional factors include connectivity to water bodies, amount of vehicle traffic, frequency of maintenance or inspection, and precipitation amounts. Figures 9.29 and 9.30 show examples of severe road erosion caused by some of the engineering factors; in particular, inadequate drainage control and failure of road drainage structures.

In sediment budget studies conducted in the southern Interior of British Columbia, Jordan (2001b) found that on a watershed scale, erosion from forest roads was a highly significant source of suspended sediment in Redfish Creek, a mountainous watershed underlain by granitic rocks that has had a moderate amount of forest development (10% logged). However, in Gold Creek, a more heavily developed watershed underlain by fractured sedimen-

7 Thompson, S.R. 2000. A system for assessing surface erosion hazard on forest roads. BC. Min. For., For. Sci. Sect., Nelson Forest Region, Nelson, B.C. Unpubl. report.

FIGURE 9.29 *Surface erosion on a road near Redfish Creek in the Kootenay Lake Forest District in 1993, caused by a plugged culvert and inadequate water bars. (Photo: P. Jordan)*

tary rocks, road erosion was a relatively insignificant source of sediment. Surface erosion from clearcuts was not a measurable source of sediment in either watershed.

In the British Columbia Interior, where landslides are relatively rare, surface erosion is probably the most significant source of sediment resulting from forest development in most areas. This is especially the case where development has occurred on erodible lacustrine soils, which are widespread in the central Interior. The low relative runoff (streamflow as a proportion of precipitation) in many areas of the dry central Interior may contribute to high potential impacts on stream channels and water quality from surface erosion, if connectivity exists between roads and stream channels.

Until the late 1980s, most logging in the Interior was done by ground-based methods, which often resulted in a high density of skid trails and excessive soil disturbance from ground skidding. This probably resulted in considerable erosion and sediment contributions to streams; however, quantitative studies are lacking. In the 1980s, soil disturbance guidelines were developed by the B.C. Ministry of Forests (Lewis 1991), and were included in the FPC in 1995. Since then, improved practices, including the deactivation of skid trails following logging and increased use of cable logging on steep slopes, have greatly reduced the incidence of erosion and sediment pro-

FIGURE 9.30 *An extreme example of road erosion in the Lemon Creek drainage, Arrow Forest District, caused by diversion of a small creek down an old road during spring runoff (2002). (Photo: C. Pettitt, Valhalla Wilderness Society)*

duction from logged areas. Operational experience of Ministry of Forests staff, and the few sediment budget studies that have been done, indicate that erosion from logged areas is usually a minor source of sediment compared to erosion from roads.

Toews and Henderson (2001) presented methods to estimate sediment production from roads for use in the Interior Watershed Assessment Procedure (B.C. Ministry of Forests 1995b). They presented preliminary sediment budget data for four watersheds, which showed that sediment from road erosion was, on average, relatively minor compared with that from natural sources, and that most erosion came from a small part of the road system. Most of the road segments they surveyed had erosion rates of 0.3–30 m^3/km per year, and an average of 32% of eroded sediment was delivered to streams.

In a related, unpublished study, Thompson measured road surface erosion on newly constructed roads throughout the Kootenay region.[8] Median erosion rates ranged from 0.4 m^3/km per year for dry-climate sites with no seepage to 10 m^3/km per year for moist-climate sites with seepage. The major factors affecting erosion were failures in the road drainage system, climate, and the presence of seepage in the cutbank. Most erosion came from the cutbank-ditch system; the road surface was second in importance.

GULLIES

Steep gully wall slopes, concentration of groundwater, and steep stream channels create difficult terrain conditions for forestry operations. Figures 9.31–9.37 show examples of gullies and forest management.

Yarding of timber often results in damage to gully sidewalls, and deposition of logging slash into the gully channel can alter flow pathways. In gullies with streams, the slash may be transported, resulting in debris jams and sediment wedges. In some cases, these jams fail catastrophically and a debris flow or debris flood can result. In other cases, the jams are stable, but if a debris flow initiates farther up the gully, the increased sediment and woody debris will add to the volume of the subsequent debris flow. Sometimes, debris jams and sediment wedges may force channel-water flows against a sidewall and erode into the toe of the slope, which can result in a sidewall slope failure. Harvesting of timber within coastal gullies results in some of the highest post-harvest landslide rates of any terrain type (Rollerson 1992; Rollerson et al. 1998, 2001; Millard et al. 2002).

Road crossings of gullies can create several problems. If culverts are blocked by woody debris, the road fill material may saturate and fail. Road cuts into steep gully sidewalls may fail, and if a steep road fill is constructed, that may fail as well. All these failures may result in debris flows that travel down the gully. Road ditches that deliver water from outside the gully onto the gully walls can initiate slope failures, or if the water is delivered directly to the channel, it can initiate an in-channel debris flow. Conversely, roads that climb into a gully may direct gully water along the ditchline to open slopes beyond the gully, which can cause subsequent open-slope failures.

Recent cutblock layouts tend to leave potentially unstable gullies partially or completely unharvested. Commonly, the cutblock edge is near gully boundaries, or is buffered for gullies inside cutblock boundaries, with the buffer designed to be windfirm at the gully edge. A recent increase has been noted in the number of windthrow-related landslides, particularly where windthrow along a buffered gully penetrates into the buffer farther than expected and causes a gully wall failure. Where a gully appears stable, harvesting occurs, but if the stream is large enough to transport woody debris, the logging slash is removed from the gully. Individual trees within a gully may be selected for harvest, with the whole tree removed by a helicopter ("snap and fly"), which results in very little disturbance to the gully.

Designing roads to descend into and climb out of a gully reduces the likelihood that the channel will become blocked and cause an avulsion (Wilford et al. 2005b; Millard et al. 2006). Road crossings should be large enough to transport both water and debris. Where debris flows are likely to occur, the road crossing should be designed to either pass the debris flow or minimize the effect of the road on the passage of the debris flow (Jakob and Jordan 2001).

8 Ibid.

FIGURE 9.31 *Gullies in the Gordon River watershed, South Island Forest District. (Photo: T. Millard)*

FIGURE 9.32 *Closely spaced gullies ("dissected terrain") at MacMillan Creek, Haida Gwaii Forest District. (Photo: T. Millard)*

FIGURE 9.33 *Steep headwall area of gully in Rennell Sound, Haida Gwaii Forest District. (Photo: T. Millard)*

FIGURE 9.34 *Unstable gully sidewall area in Rennell Sound, Haida Gwaii Forest District. (Photo: T. Millard)*

FIGURE 9.35 *Small helicopter-yarded patch cuts, with a buffered gully (middle of the photo), at Foley Creek watershed, Chilliwack Forest District. This approach to harvesting near unstable terrain is becoming more common. (Photo: T. Millard)*

FIGURE 9.36 *Debris slides in a gully near the Klanawa River in the South Island Forest District. The gully was left unlogged but was heavily damaged by windthrow. (Photo: T. Millard)*

FIGURE 9.37 *Debris flows that originate in gullies can present a significant hazard to public safety and to property in populated areas. (Photo: P. Jordan)*

ALLUVIAL FANS

Alluvial fans are deposits of sediment formed where streams or debris flow transport zones lose confinement. Generally, these fans are found in valley-bottom locations where tributary streams or gullies enter a main valley. Fans also occur on mid-slope benches or other locations where confinement is lost, lowering the power to transport material. Sediment originates in a watershed, which can be of any scale, ranging from a single hillslope gully to a valley system hundreds of square kilometres in area. Sediment in small (less than about 10 km^2), steep watersheds can be transported to fans by debris flows (VanDine 1985; Wilford et al. 2005a; Millard et al. 2006). In British Columbia, the terrain classification system identifies these as *colluvial fans* (Howes and Kenk [editors] 1997). As watersheds become larger, it is common for fluvial processes to dominate sediment movement from watersheds to fans (e.g., floods and debris floods) (Wilford et al. 2005a; Millard et al. 2006). In British Columbia, these are classified as *fluvial fans* (Howes and Kenk [editors] 1997). In the literature, it is common to use the term "alluvial fan" to refer to fans formed by fluvial or colluvial processes, or a combination of the two. Slope gradients on fluvial fans are characteristically less than 4°, with gradients of up to 18° on colluvial fans (at steeper gradients the landform is referred to as a "cone") (Jackson et al. 1987; Howes and Kenk [editors] 1997; Wilford et al. 2005a). In graded (i.e., mountainous as opposed to plateau) watersheds, the dominant hydrogeomorphic process influencing fans can be correlated with the relative relief number, which is watershed relief divided by the square root of watershed area—sometimes known as the "Melton Ratio" (Melton 1965). Watersheds that produce floods typically have a relative relief number of less than 0.3 (Jackson et al. 1987; Bovis and Jakob 1999). Based on regional studies, it is possible to differentiate debris flow from debris flood watersheds (Wilford et al. 2004; Millard et al. 2006). Debris flood events are similar to flood events, but they have a higher sediment load and commonly result in greater disturbance on the fan surface (Hungr et al. 2001). Regardless of the hydrogeomorphic process, landslides are important processes for delivering sediment to fans—either directly as in a debris flow or delivered by fluvial transport of bedload.

Many alluvial fans in British Columbia are considered to be paraglacial—formed with sedminents delivered by the melting of Pleistoscene glaciers in denuded watersheds (Ryder 1971a, 1971b). High volumes of water and sediment resulted in large fans—larger than contemporary hydrogeomorphic conditions would suggest. Under current conditions, hydrogeomorphic processes may result in comparatively minor additions or removals of sediment from paraglacial fans; however, these additions can have a profound influence on the surface of fans. In addition, the spatial extent and complexity of the channel network on a fluvial fan surface reflects boundary conditions, such as sediment and water supply from the watershed, fan vegetation, the space into which the fan is building, and whether hydrogeomorphic processes are removing sediment from the toe of the fan (Millard et al. 2010). Fluvial fans that receive a high sediment supply from the watersheds have a greater spatial extent of the channel network than fans that receive a lesser supply of sediment.

One change on alluvial fans in British Columbia compared to paraglacial time is the presence of forest cover. It is clear that forests on fans, particularly along stream channels, provide a range of hydrogeomorphic roles: storage of sediment, maintenance of stream channel location, and soil reinforcement by roots. Site features can be used to identify these "hydrogeomorphic riparian" zones (Wilford et al. 2005b).

PLANNING AND PRACTICES TO REDUCE FORESTRY EFFECTS ON HILLSLOPE PROCESSES

Before the establishment of the FPC in 1995, terrain stability management was either voluntary or a contractual obligation (Fannin et al. 2007). However, environmental concerns and recognition of the effects of forest management on landslides and sediment production have resulted in many changes to forestry practices over the past two decades. During this time, terrain stability mapping and terrain stability assessments began to be widely used in coastal British Columbia. The FPC established province-wide regulations regarding terrain stability and sediment production.

An investigation by the Forest Practices Board (2005) found that, based on a limited sample from both the Coast and Interior, the frequency of landslides resulting from cutblocks and roads had decreased following the introduction of the FPC. Probable factors contributing to this reduction included establishment of reserves and management zones around gullies and streams, better road location, and improved road construction practices.

The FPC required a terrain stability field assessment if one of the following criteria was met: unstable or potentially unstable terrain as shown on terrain stability mapping, slopes greater than 60%, field indicators of instability, or a District Manager's request. Although the trigger for an assessment was based on hazard, the FPC required an evaluation of risk, and in certain situations, restricted forest management activities if the risk was too high (see section on "Landslide Risk Management," below). The FPC also required watershed assessments that evaluated sediment sources and their effects on the stream channel or drinking-water quality. A risk-based approach to assessments was emphasized both in the FPC's guidebooks and in standards developed by the Joint Practices Board (representing members of the Association of Engineers and Geoscientists of British Columbia and the Association of British Columbia Forest Professionals to provide guidance to its members on issues of professional practice). Professionals who make the assessments and prescriptions develop practices and methods that are appropriate to the site. On-the-ground practices were generally not prescribed by the FPC, although information on recommended practices and methods was given in the guidebooks.

The transition to the *Forest and Range Practices Act* started in 2004. This Act represents a "results-based" approach and does not require any assessments. Licensees are expected to choose management strategies that will not cause "material adverse effects." Since a greater reliance is placed on professionals under the Act, professional associations are providing greater direction for guiding professional practice. The Association of Professional Engineers and Geoscientists of British Columbia (APEGBC) and the Association of British Columbia Forest Professionals (ABCFP) have developed two documents that address professional practice for terrain stability. The first document, *Guidelines for Management of Terrain Stability in the Forest Sector*

(Association of Professional Engineers and Geoscientists of British Columbia and Association of British Columbia Forest Professionals 2008), describes a comprehensive process or model to be developed by professionals managing forest operations in steep terrain, including identifying situations in which a terrain stability assessment is needed. The second document, *Guidelines for Terrain Stability Assessments in the Forest Sector* (in preparation), is based on a 2003 document (Association of Professional Engineers and Geoscientists of British Columbia 2003) and provides guidance to professionals conducting terrain stability assessments. Recently, under the Forest and Range Evaluation Program, a pilot effectiveness evaluation survey to assess sediment from roads and cutblocks has been developed (see Carson et al. 2009) to assess performance relating to water quality.

Strategies for managing the risk of forestry-caused landslides and erosion events can be grouped into four general categories, in increasing order of cost and difficulty (Chatwin et al. 1994):

1. avoidance: for example, road location and cutblock layout
2. prevention: for example, road construction and drainage design
3. stabilization: for example, construction of reinforced slopes or retaining walls in landslide areas
4. protection: for example, the building of berms or catch basins to protect elements at risk from hazards upslope

The discussion below focusses on the first two categories: avoidance and prevention.

Terrain Stability Mapping

Terrain stability mapping in British Columbia was initially developed by the coastal forest industry in the 1970s, and was modified in the 1980s by Ministry of Forests and forest industry mappers. For a review of this mapping system and its limitations, see Schwab and Geertsema (2006).

Terrain mapping as it is now practised in British Columbia evolved from a classification system introduced by R.J. Fulton of the Geological Survey of Canada (Fulton et al. 1979). The British Columbia Terrain Classification System was developed from Fulton's system (Environment and Land Use Committee Secretariat 1976) and has evolved into an elaborate and comprehensive system (Howes and Kenk [editors] 1997; B.C. Ministry of Forests 1999). When terrain maps are created by experienced mappers, they form a sound basis for many derivative maps—among them, terrain stability maps.

Terrain stability mapping is now an integral part of forest development planning in British Columbia (Association of Professional Engineers and Geoscientists of British Columbia 2003). Its focus is landslide hazard—specifically, the likelihood of development activity causing a slope failure. Polygons (map units) are rated based on the likelihood that a landslide might initiate within the polygon but not on whether a landslide may impact it. Therefore, fans would be in a low terrain stability class despite the vulnerability of this feature to impact by landslides. A separate analysis is sometimes undertaken to evaluate the travel distance of a landslide, and hence the likelihood of sediment entering a stream (Hogan and Wilford 1989). Because the stability classes were created for application to forest development activities, this method of landslide hazard mapping is not suitable for general land use zonation[9] associated with infrastructure or residential development. Furthermore, since the terrain stability classification system was developed for steep terrain, shallow soils, and relatively simple landslides (debris slides and flows on open slopes or in gullies), large, complex landslides are not adequately addressed (Schwab and Geertsema 2006).

Although the standards for terrain stability mapping are set by government (B.C. Ministry of Forests 1999), operational mapping in areas of proposed development is conducted by the forest industry. The version of mapping adopted for the FPC in 1995 included reconnaissance terrain stability mapping (RTSM) and detailed terrain stability mapping (DTSM) with three and five hazard classes, respectively (B.C. Ministry of Forests 1999). The three RTSM hazard classes are: stable (S), potentially unstable (P), and unstable (U). On the maps, classes P and U include a terrain symbol, geomorphic process if applicable, and a slope range. The five DTSM hazard classes range from stable (I) to the most unstable (V). The two types of mapping differ mainly in extent of landscape coverage: DTSM covers the entire landscape, whereas for RTSM, only those polygons judged

9 Hungr, O., B. Gerath, and D. VanDine. 1994. Landslide hazard mapping guidelines for British Columbia. Earth Sciences Task Force, B.C. Government Resources Inventory Committee. Unpubl. report.

by the mapper as unstable or potentially unstable are mapped. This results in considerable cost savings, especially in areas such as the interior plateaus where most of the landscape is stable. In complex, mountainous terrain, or in areas with high downstream consequences, such as community watersheds, DTSM is generally preferred.

All terrain maps, including terrain stability maps, are based on air photo interpretation followed by field checking. Polygons of approximately homogeneous surficial material and landform are identified on a stereo pair of photos, and preliminary terrain labels are assigned. The accuracy and reliability of the maps is determined mainly by the amount of field checking that is done. The level of field checking (terrain survey intensity level, or TSIL) for various types and scales of mapping is specified in the standards (B.C. Ministry of Forests 1999). Typically, TSIL B or C (50–75% and 20–50%, respectively, of polygons ground checked) and a scale of 1:10 000 or 1:20 000 is used for DTSM. For RTSM, TSIL D (1–20% of polygons ground checked) and a scale of 1:20 000 or 1:50 000 is used. Following field checking, final polygon boundaries, terrain labels, and interpretations are completed. Various photogrammetric methods (such as monorestitution), or sometimes transfer to base maps by eye for RTSM, are used to transfer the polygon linework from the air photos onto topographic maps. With the increased use of computers for mapping purposes, GIS (geographic information systems) is replacing drafting on mylar, and Roman numerals are falling out of use.

A complete description of the terrain classification system used for the terrain labels is given in Howes and Kenk (editors, 1997). In its simplest form, the label consists of symbols for soil texture, genetic material (morainal, colluvial, fluvial, etc.), surface expression or landform, and modifying geomorphic processes (in that order). For example, the label "zsMb-V" means "silty sandy morainal blanket, modified by gullying." Figure 9.38 provides an example of a terrain stability map, prepared at TSIL C and a scale of 1:20 000, showing both the air photo and the final map.

The accuracy of terrain stability mapping is highly dependent on the skill and experience of the mapper, as well as on the extent of field checking. No two mappers will produce exactly the same polygon configuration, labels, or stability ratings, although usually the products of experienced mappers will generally be in agreement. Individual mappers typically develop terrain stability rating criteria tables based on slope, surficial material, drainage conditions, and evidence of past slope movement, and the mapper's professional subjective opinion often is a factor in assigning a stability rating. To provide objective and quantitative ratings, data from terrain attribute studies can be used in those areas where available. It is generally accepted (although no longer required by government regulation) that mappers must be licensed professionals with adequate training and experience.

A number of remote sensing and digital elevation model (DEM) approaches have possible applica-

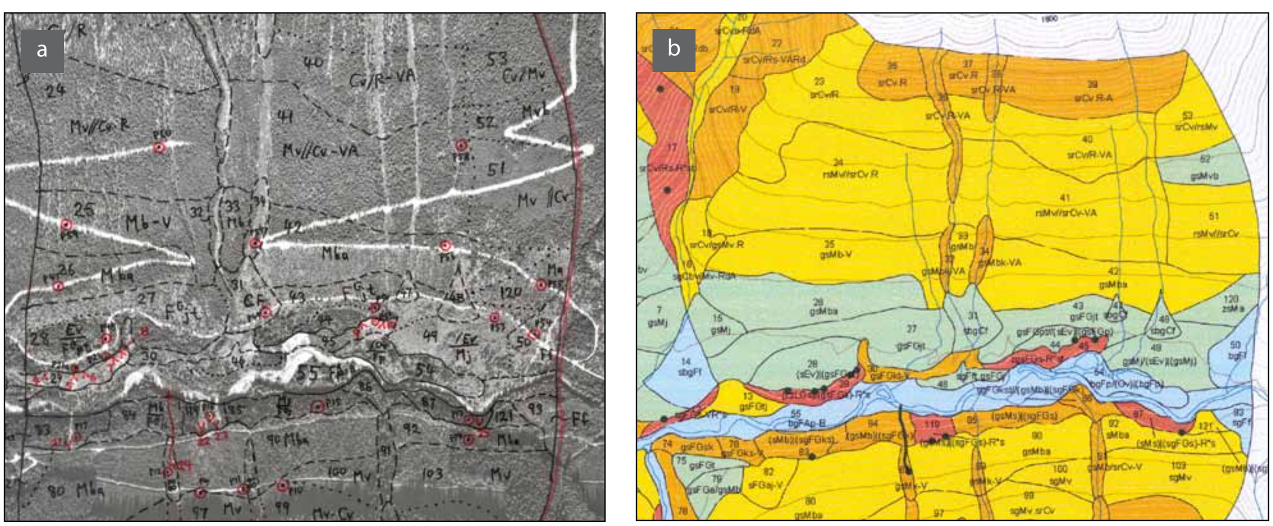

FIGURE 9.38 *Example of a terrain map: (a) air photo interpretation and (b) completed terrain map. On this map, terrain stability classes are indicated by colours: classes I, II, III, IV, and V are coloured blue, green, yellow, orange, and red, respectively.*

tion to mapping terrain stability (Slaymaker 2001). Although attempts have been made to apply some of these to forestry-related landslide hazard (Niemann and Howes 1992; Pack et al. 1998), digital mapping methods have not gained wide acceptance in British Columbia.

Terrain Stability Assessments

Terrain stability assessments evaluate terrain stability hazards and risks that may be affected by forest development or that may affect forest development. Terrain stability assessments originated in coastal British Columbia in the 1980s as forestry-related landslides became a management issue (Fannin et al. 2007). These assessments were required under the FPC, and although not required under the *Forest and Range Practices Act*, they are an important part of due diligence in managing steep terrain (Horel and Higman 2006). The following summary is based on the *Guidelines for Terrain Stability Assessments in the Forest Sector* (Association of Professional Engineers and Geoscientists of British Columbia 2003), and on more recent developments in terrain stability assessment practice.

The objectives of terrain stability assessments include, but are not limited to:

- characterizing the existing landslide hazards (terrain and terrain stability conditions) in areas within, adjacent to, or connected to the forest development area;
- evaluating potential or existing effects of the forest development on terrain stability;
- determining the landslide hazards and potential or existing effects of the forest development on identified elements at risk, including worker and public safety;
- comparing the landslide hazards or risks with available guidelines or standards of acceptable hazard or acceptable risk, including those determined by government and those contained within a licensee's terrain stability management model; and
- recommending site-specific actions to reduce and/or manage the landslide hazards and risks resulting from the forest development.

In most cases, the terrain specialist does extensive office and field work to evaluate terrain conditions and the planned forestry development. Terrain conditions evaluated include:

- slope configuration and geometry;
- surficial material type, strength, and drainage;
- evidence of past or historic landslides, or signs of instability (Figure 9.39);
- hillslope hydrology;
- potential landslide triggering events, such as rainstorms, rain-on-snow events, snowmelt, or windthrow; and
- characteristics of any nearby landslides, including initiation-zone characteristics and travel distance.

The terrain specialist also considers the potential effects of forest development, including:

- the method and effects of harvesting on hillslope stability;
- the stability of any proposed roads, including cutslopes and fillslopes;
- the effects of road drainage on hillslope hydrology, particularly the concentration of drainage that may lead to fillslope or hillslope failure; and
- the effects of existing development.

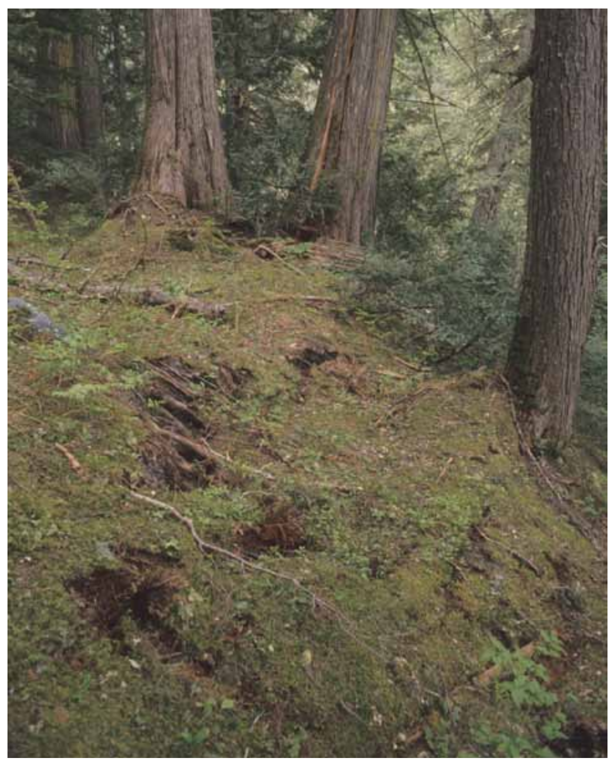

FIGURE 9.39 *Tension cracks, Tangiers River in the Columbia Forest District. An important part of terrain stability assessments, as well as layout work by forest technicians, is to observe indicators of slope instability such as this. (Photo: P. Jordan)*

The terrain stability assessment report often includes recommendations for managing terrain stability risks. Management options can include avoiding potentially unstable areas, modifying harvesting or road building methods, or in rare cases, constructing stabilization or protective works such as debris flow basins.

Terrain specialists generally use a combination of theoretical understanding of terrain stability, professional experience, and observations of nearby slopes to assess landslide likelihood. On the Coast, terrain attribute studies are important sources of information for quantifying landslide hazard in some areas. In other areas of the province, landslide densities are lower than on the Coast, and there have not been a sufficient number of terrain attribute studies to practically quantify landslide likelihood.

Landslide Risk Management

Landslide risk management is the systematic application of management policies, procedures, and practices to the tasks of identifying, analyzing, assessing, mitigating, and monitoring risk resulting from landslides (Fell et al. 2005). It is often used in forest development planning and operations to weigh the benefits of a particular forest development against the potential of that development to cause landslides, which may result in subsequent damage. It is also used to assist with the management, mitigation, and monitoring of landslide risks. The benefits of landslide risk management include (Fell et al. 2005):

- uses a rational, systematic approach to the assessment of risk;
- can be applied to non-deterministic situations;
- can be applied to land use planning;
- allows comparisons of risks across a landscape;
- focusses attention on what happens if a landslide does occur;
- provides an open and transparent process on the nature and key contributors to landslide risk; and
- allows for the systematic consideration of risk mitigation options and cost/benefit ratios.

The systematic determination of the degree of risk is referred to as *risk analysis* and is one step in the risk management process (Figure 9.40). Risk analysis includes the systematic use of information to identify hazards and to estimate the chance for, and severity of, injury or loss to individuals or populations, property, the environment, or other things of value (Canadian Standards Association 1997). *Risk assessment* combines risk analysis and the step of *risk evaluation* to determine whether the risk is acceptable or tolerable. Risk assessment includes neither the consideration of options for *risk control*, nor the actions to control risk or monitor performance of site works over time. *Risk management* is a complete process involving all the steps in the decision-making framework and communication about risk issues (Wise et al. [editors] 2004). Risk analyses are often completed by terrain stability professionals, whereas

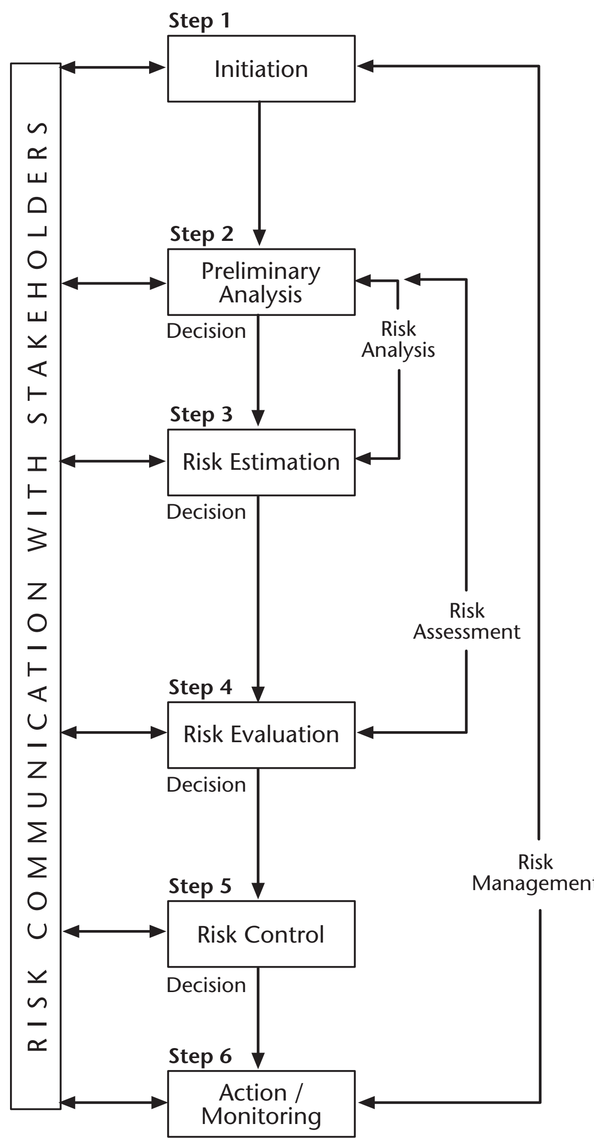

FIGURE 9.40 *Six steps in the decision-making framework for risk management. (Source: Wise et al. [editors] 2004, as adapted from Canadian Standards Association 1997)*

risk evaluation and risk control requirements are often the responsibility of the forest resource managers.

Risk is the chance of injury or loss as defined as a measure of the probability and the consequence of an adverse effect to health, property, the environment, or other things of value (Canadian Standards Association 1997). When considering risk in the context of forest management and hillslope processes, risk can be evaluated with respect to the adverse effects of landslides, sediment delivery, and snow avalanches to elements at risk such as roads, highways, fish and fish habitat, water systems, water quality, public safety, timber values, visual values, buildings, and other infrastructure. The level of risk depends on the value placed on the item at risk and the likelihood of an adverse process affecting the element at risk.

In the forestry sector, *landslide risk assessments* are commonly conducted when roads and cutblocks are proposed on or above steep slopes, when important downslope values could be affected by forestry developments, when interface wildfire occurrence could trigger landslides, or when unstable areas are identified for inspection or monitoring.

In its simplest form, risk can be mathematically expressed as:

$$R = P \times C \qquad (1)$$

where: R = risk, P = probability of an adverse process occurring, and C = consequence. Because its determination is usually based on judgement, the probability of an adverse process occurring can be referred to as judgemental or subjective probability (Vick 2002).

Risk can be estimated quantitatively or qualitatively. A simple example of a quantitative risk analysis follows.

> *The subjective probability of a road-related landslide occurring over a steep road segment is estimated at 1 in 10 (over a 20-year-design road life), and the consequence of the landslide is the value of the road segment potentially damaged (say $50 000). The 20-year risk can then be expressed monetarily as P × C = 0.1 × $50 000 or $5000. If no other values are at risk and the cost to eliminate the hazard is greater than $5000, then, based only on the present-day monetary road value, it would not be economical to eliminate the hazard.*

The degree of risk for non-monetary elements at risk is not easily quantified, and can be expressed in qualitative terms such as low, moderate, or high. For more complex situations involving social, environmental, and economic elements at risk, a single risk value or description cannot describe the risk for all elements. For example, a hazardous sediment source may pose a high risk to fish habitat and only a moderate risk to a community water system; likewise, a community water system may be exposed to multiple independent concurrent hazards of varying probabilities of occurrence. Additional terms can be added to the risk equation in situations where a possible adverse process could occur, but where it is uncertain whether, or to what degree, the adverse process will result in a negative consequence. Terms such as the probability of spatial impact to the element at risk, the probability of temporal impact to the element at risk, and the degree of damage to an element (should impact occur) can all be added as follows:

$$R = P \times P(S{:}H) \times P(T{:}S) \times V(L{:}T) \times E \qquad (2)$$

where: $P(S{:}H)$ = probability of spatial impact of the element, $P(T{:}S)$ = the probability of temporal impact, $V(L{:}T)$ = vulnerability or an estimate of the proportion of loss or damage to the element, and E = the value of the element. For example:

> *A proposed forestry development results in three hazards that include a landslide hazard, a sediment production hazard, and a surface water diversion hazard. Elements at risk include domestic water use, fish habitat, a public highway, and the travelling public. A risk analysis would consider each element at risk and each possible hazard.*
> *Domestic water use may have both an infrastructure value and a water quality value. All three of the noted hazards could affect these values and therefore would be considered separately.*
> *The fish habitat component of the risk analysis may use the concept of partial risk where R = P(HA), and P(HA) is the probability of a landslide occurring and affecting the element at risk. If the likelihood of a landslide occurring is low and the probability that the landslide will reach and affect the fish habitat is high (if a landslide occurs), then the resulting partial risk to fish habitat as a result of a landslide would be low × high = moderate (Table 9.2). The advantage of using partial risk is that the degree of damage to the resource value does not need to*

TABLE 9.2 *Example of a simple qualitative risk matrix for partial risk (Wise et al. [editors] 2004)*

P(HA), annual probability (likelihood) of occurrence of a specific hazardous landslide and it reaching or otherwise affecting the site occupied by a specific element		**P(S:H) x P(T:S) Probability (likelihood) that the landslide will reach or otherwise affect the site occupied by a specific element, given that the landslide occurs**		
P(HA) = P(H) x P(S:H) x P(T:S)		**High**	**Moderate**	**Low**
P(H), annual probability (likelihood) of occurrence of a specific hazardous landslide	Very High	Very High	Very High	High
	High	Very High	High	Moderate
	Moderate	High	Moderate	Low
	Low	Moderate	Low	Very Low
	Very Low	Low	Very Low	Very Low

be estimated. A more refined fish habitat analysis may consider the species of fish, spawning areas, spawning times, stream channel characteristics, and recoverability.

The highway has an infrastructure value, a safety value (travelling public), and a nuisance value (delays to those using the highway). The hazard to the highway is likely limited to possible landslides. When considering the risk to the travelling public using the highway, more terms are required that consider the likelihood of the slide reaching the highway, the likelihood that the slide will impact a vehicle (or the vehicle driving into slide debris), and the likelihood of injury or even death. Landslide characteristics such as the estimated landslide size, travel speed, runout distance, and trigger mechanisms may be used to refine the estimated hazard and risk.

This example discusses six values with up to three hazards for each value, resulting in up to 12 separately described risk levels that require assessment and evaluation. For more detailed examples of landslide risk analyses in the forest sector, see the eight case studies in Wise et al. (editors, 2004).

When a risk analysis is completed, an evaluation is made to determine whether the risk is worth taking. This evaluation may compare the risk to established risk criteria or to locally determined levels of tolerable risk. Acceptable or tolerable risk levels depend on many factors (Morgan 1997) including societal attitudes and response, legal attitudes, quantified tolerable societal and individual risks, and efficient expenditure of funds. The evaluation may also utilize a cost–benefit approach in which the risk cost can be directly compared to the benefits of proceeding with the development. This can become very complex and caution must be exercised when the benefits are financial and the risk costs are related to the environment or public safety.

After the risk evaluation is completed, consideration must be given to reducing the risk (risk control or risk mitigation) through whatever means are available that are consistent with the intent of the development and with financial constraints.

Practices to Avoid Harvesting-related Landslides and Sediment Production

Harvesting-related landslides are caused by changes in rooting strength, water inputs to the soil, or changes in soil structure and drainage. In many cases, the most effective method of avoiding harvesting-related landslides is to avoid slopes that may be subject to landslides. These slopes are identified by terrain mapping or terrain stability assessments and are excluded from cutblocks. Cutblock layout is much more selective now than in the past, and typically avoids locations such as gullies. Where risks to resources are deemed not critical, approaches can be used to achieve protection from landslides and still make some timber supply available.

Partial harvesting methods may result in fewer landslides than clearcutting on steep hillslopes. Patch cuts are unlikely to reduce landslide incidence unless the patches are located on small areas of relatively stable ground and do not leave edge trees on unstable ground that may be subject to windthrow. Dispersed harvesting may result in fewer landslides as there will be less effect on rooting strength and water inputs. Although theory suggests that dispersed harvesting should have less effect on hillslope stability than clearcutting, there are no field studies to show the effect of reduced levels of harvesting on post-harvest landslide rates.

Helicopter or full-suspension yarding of timber may cause less damage to soils than conventional cable yarding. Krag (1996) found that cable-yarded slopes in Rennell Sound, Haida Gwaii, had 8% mineral soil exposure but adjacent heli-logged areas had 0.1–2.4% mineral exposure. The associated effect on landslide rates is uncertain since the trials were conducted over a range of terrain stability types (Millard and Chatwin 2001). Roberts et al. (2004) compared landslide rates in helicopter-yarded areas with those in conventionally cable-yarded areas. In gullied areas, no significant difference was evident in landslide rates between the two methods. In open-slope areas, a significant difference was evident: cable-yarded areas had a rate of 0.02 landslides per hectare, whereas no landslides occurred in the helicopter-yarded areas; however, the relatively small sample size and the contrary results for gullied terrain versus open-slope terrain make these study results somewhat tentative.

Sediment production from harvested areas is related primarily to soil disturbance that occurs during timber yarding. Most forest soils are covered by an organic duff layer that prevents rain splash erosion, but when mineral soils below the duff layer are exposed, erosion can occur. Cable-yarding systems may drag logs along the ground and create extensive areas of soil disturbance. Ground-based yarding systems expose soils by dragging logs along the ground and by the actions of skidder wheels or tracks. The FPC regulations contained maximum soil disturbance limits, as do the *Forest and Range Practices Act* regulations. Soil disturbances can be minimized during cable-yarding operations by designing cutblock layouts and choosing cable systems that ensure that most of the log is suspended. Ground-based systems commonly use low ground pressure equipment or designated trails that are rehabilitated following harvest.

Practices to Avoid Landslides and Sediment Production from Forest Roads

In the last 20 years, road design and construction have progressed substantially, reducing the likelihood of landslides and sediment production from new road construction (Fannin et al. 2005). Major factors contributing to this progress include:

- improved technology, such as the use of excavators instead of bulldozers for most construction, and the use of geotextiles for engineered fillslopes;
- increased involvement of engineers and other qualified professionals in designing roads for difficult terrain, and in conducting terrain stability assessments; and
- increased regulation, and industry response to increased public awareness of forest practices.

In 1991, the B.C. Ministry of Forests published *A Guide for Management of Landslide-Prone Terrain in the Pacific Northwest* (Chatwin et al. 1994). This guide was aimed at forest industry practitioners and provided basic information on landslide processes, terrain stability assessments, recommended road locations, design and construction practices to minimize landslide hazard, road deactivation techniques, and general management of landslide risks. The Ministry of Forests, along with forest industry co-operators, conducted an extension program, which included a series of workshops based on the guide, for operational industry and government staff throughout the province.

With the introduction of the FPC in 1995, the *Forest Road Engineering Guidebook* was published (the most recent revision is B.C. Ministry of Forests 2002). This guidebook provides recommendations for measures to maintain slope stability when designing roads, including prescriptions made by a qualified professional. It also provides guidelines and information on stream crossings, erosion control, construction methods, drainage structures, and deactivation techniques, and includes methods for a qualitative risk assessment procedure.

The *Forest and Range Practices Act* brought increased reliance on professionals, and a more risk-based approach to management of landslides and other hazards. In support of this, the Ministry of Forests published *Landslide Risk Case Studies in Forest Development Planning and Operations* (Wise et al. [editors] 2004). This handbook includes an explanation of qualitative and quantitative risk assessment procedures, and provides a number of examples of the application of risk assessment to topics such as road location, reconstruction of landslide-damaged roads, alluvial fan risk management, and cutblock layout.

The publications listed above provide specific techniques for minimizing landslide risk. Some basic principles to prevent landslides and erosion from forest roads include:

- using qualified professionals to conduct terrain stability assessments, design roads, prescribe road construction methods, and provide advice

on related subjects such as sediment control and drainage plans;
- using risk assessment principles to choose optimum road locations to minimize landslide risk;
- using engineered road designs in unstable and potentially unstable areas;
- using techniques such as full benching, end hauling, and engineered fills to avoid unstable road fills;
- recognizing that drainage control on roads is the key to avoiding landslides and erosion; and
- inspecting, maintaining, and properly deactivating roads.

Many road-caused landslides are related to drainage control problems: poor drainage design, insufficient culverting, inadequate or poorly implemented deactivation, failure of road drainage structures, and lack of adequate maintenance. Landslides often happen where multiple levels of roads are built on a hillside but culverts and cross-ditches on each level are not properly aligned with those on the levels below. In cases where a road on an ascending grade crosses a small stream but the culvert is prone to blockage, failing to back up the culvert with a cross-ditch is another common cause of landslides. One key measure to prevent such landslides is having a professionally prepared drainage plan for landslide-prone areas (e.g., Green and Halleran 2002). Other important measures (not requiring professional input) include frequently inspecting stream crossings, ditches, and cross-drain culverts, and installing temporary water bars or other seasonal deactivation works in areas with potential landslide or erosion problems.

Surface erosion is inevitable on all forest roads, especially ones that are actively used. Nevertheless, measures can be taken to reduce sediment inputs to streams (Carson and Younie 2003). For example, culverts, water bars, or grade dips are commonly used to divert local runoff into a stable, vegetated area instead of directing it into a stream channel. Numerous construction and maintenance methods are available to minimize road surface and ditch erosion. There is extensive literature on the subject, for example, Gillies (2007).

FOREST MANAGEMENT AND ALLUVIAL FANS

Forest management can have direct and indirect effects on alluvial fans. Direct effects involve practices on fans such as road construction, forest harvesting, windthrow, and drainage structures. Indirect effects to fans involve practices in the watershed that influence the delivery of sediment and water to the fan.

The effects of forest management on fans were not generally recognized until the 1990s, when the *Gully Assessment Procedure Guidebook* (B.C. Ministry of Forests 1995a) included a brief fan assessment procedure for coastal fans downslope of gullies. Although problems with roads, drainage structures, and stream avulsions frequently occurred on fans, these had not been placed in a landform context. Previous landform mapping undertaken for forest management was principally focussed on the identification of unstable hillslopes, and therefore hillslope processes were mapped in detail but valley-bottom processes and landforms were often lumped into large polygons. Thus, valley-bottom units could include alluvial fans as well as morainal blankets and glaciofluvial deposits. When identifying landforms associated with road and drainage structure problems, it is possible to develop prescriptions appropriate for the hydrogeomorphic processes occurring on fans (Wilford et al. 2003; Wilford et al. 2005b). Such fan-specific prescriptions are appropriate to avoid destabilizing fan surfaces on the Coast (an offence under the *Forest and Range Practices Act*, Section 54), or to have an impact on forest soils or fish habitat on the Coast and in the Interior (Sections 35 and 57).

An appropriate step in planning forest development in an area is to identify alluvial fans. Office and field investigations identify the nature and extent of hydrogeomorphic activity on the fan. Design of drainage structures takes into account not just peak flow runoff but also the sediment and debris loads associated with the hydrogeomorphic process. Debris floods can have peak discharges up to two times greater than associated flood peak flows, and debris flows can have discharges one to two orders of magnitude higher than flood peaks (Hungr et al. 2001; Jakob and Jordan 2001). Attention must also be paid to the direction of the road gradient relative to the stream channel because a road could become a new channel if, for example, a drainage structure is not able to accommodate the water and sediment load or water is broadcast across a fan surface. The portion of the fan surface influenced by hydrogeomorphic processes is identified in the field through a series of features, including buried tree butts, log steps, splays of sediment, and cohorts of trees. This zone is referred to as the "hydrogeomorphic riparian zone." Removal of trees from this zone can lead to an increase in the extent of disturbance on fans from hydrogeomorphic events (Sato 1991; Wilford et al. 2003).

Indirect effects of forest management activities on fans centre on changes in the supply of sediment or water from a watershed. These changes can result in geomorphic adjustments and impacts to elements at risk on the associated alluvial fan. For example, an increase in sediment delivery to a fan can result in channel infilling and avulsions, and an increase in active channel area (Millard et al. 2010). An increase in water flows can also lead to erosion of the channel on the fan (channel entrenchment). The increase in frequency of landslides related to forestry activities, as discussed in the previous sections, has led to elevated disturbances on alluvial fans throughout British Columbia. For example, sediment from logging-related landslides in the Shale Creek watershed on Haida Gwaii was transported out of the watershed by periodic floods and deposited in the stream channel on the fan. This resulted in a channel avulsion (see Wilford et al. 2009, Appendix 3). Concentration of runoff by a logging road in the Hunter Range near Sicamous in the southern Interior led to a significant debris flow that resulted in the destruction of two houses and damage to several other houses, as well as to a major highway and local roads on the fan of Hummingbird Creek (see Wilford et al. 2009, Appendix 4; see also, "Case Study: Hummingbird Creek Debris Avalanche and Debris Flow" above). Debris flows from logged areas above Port Alice, a coastal town built on a fan, damaged homes and led to the construction of debris flow diversion structures.

Forest management in British Columbia has focussed prescription development at the site level (e.g., terrain stability assessments, silvicultural prescriptions based on ecosystem units). However, given the potential changes in sediment and water supply that result from forest management activities in the watersheds of fans, it is necessary to take a watershed perspective. In *Managing Forested Watersheds for Hydrogeomorphic Risks on Fans*, Wilford et al. (2009) presented a framework to gain this perspective when developing forestry plans in a watershed. This framework includes the following steps.

1. Identify fans and delineate watersheds.
2. Identify elements at risk on fans.
3. Investigate fan processes.
4. Investigate watershed processes.
5. Analyze risks and develop plans.

This framework may be of value when considering forestry operations in either first- or second-growth stands on a fan if previous forest development had occurred in the associated watershed. In this case, planning activity would focus on the determination of potential changes in sediment budget or peak flows and on forecasting the net impacts to the fan (see Wilford et al. 2009, Appendix 3).

FOREST MANAGEMENT AND SNOW AVALANCHES

Snow avalanches are a common occurrence throughout much of British Columbia. These avalanches are a concern for forest management because of the potential damage to productive forest lands and infrastructure, the potential for increased avalanche hazard following harvesting, and the risk avalanches pose to forest workers, winter recreationists, and other members of the public.

Avalanche activity is determined by the combination of slope conditions, snow properties (e.g., crystal structure), and weather. (For a more comprehensive discussion of these factors and how they contribute to avalanches, see McClung and Schaerer 1993.) Seasonal snowpacks are inherently dynamic. During the colder winter period, precipitation falls as snow. Each weather event adds a layer of snow on top of the layers laid down previously in the season. Each layer of snow within the snowpack has unique snow crystal structure, which ultimately determines the strength within and between snow layers. Factors such as temperature, moisture, energy exchange, wind, and snow loading all contribute to changes in snow crystal structure over time.

Avalanches occur when the force of gravity acting on a layer or layers of snow exceeds the strength of the layer(s). Loose snow avalanches result from a local loss of cohesion near the surface, and typically spread out from a central trigger point. Slab avalanches occur when a weak layer of snow at depth within the snowpack fails, releasing the slab of overlying snow. Avalanche paths are typically separated into three zones: (1) the start zone, (2) the track, and (3) the runout zone (Weir 2002). The start zone is where an avalanche is released; it then accelerates through the track and deposits in the runout zone.

Forests influence microclimate conditions, such as wind, precipitation, temperature, and radiation, which in turn affect snow stability (Table 9.3). As a result, microclimate conditions can vary from forested to open areas (e.g., Frey and Salm 1990). Forest harvesting, therefore, can influence avalanche activity through its effect on these microclimatic conditions.

Harvested areas typically have increased wind, precipitation, incoming shortwave radiation, and outgoing longwave radiation compared to closed canopy sites. Tree removal also results in reduced snowpack strength on harvested slopes. The net effect of these factors can be an increase in snow avalanche activity on steep harvested slopes. (For a comprehensive discussion on snow avalanches and forest management, see Weir 2002.)

McClung (2001) suggested that approximately 10 000 clearcuts in British Columbia are significantly affected by snow avalanches. Most of these have been affected by avalanches that initiated upslope from the clearcut. Approximately 10% of these cutblocks had avalanches that initiated within the cutblock and impacted downslope resources.

The primary concerns about avalanches in a forest management context are related to safety, physical impacts, and silvicultural problems (Jordan 1998). Safety concerns focus on the safety of workers, recreational users of forest lands (such as snowmobilers and backcountry skiers), and highways and other areas downslope of harvested areas. Physical impacts from avalanches include loss of productive forest land, soil degradation, damage to timber, subsequent impacts on stream channels and water quality, and impacts to infrastructure. Silvicultural problems include poor regeneration in cutblocks affected by avalanches.

Mitigation of avalanche risks in a forest management context includes implementing measures that ensure worker and public safety, minimize damage to plantations from avalanches originating either inside or outside the cutblock, and preventing large avalanches from originating in clearcuts. Safety measures have been addressed by Worksafe BC and by the Canadian Avalanche Association, which has developed safety guidelines, avalanche control standards and procedures, and safety education programs. The Joint Practices Board has also been involved with setting standards for assessing snow avalanche risks related to forest management.

At present, little information exists on strategies to reduce risks from avalanches that initiate in cutblocks. Weir (2002) summarized the state of current knowledge. Unlike in Europe, where clearcutting in avalanche terrain rarely occurs, in British Columbia clearcutting is the most widely practised silvicultural system in steep mountain environments. Since about 2000, snow avalanche hazard assessments have been done for several cutblocks, mainly in the Columbia

TABLE 9.3 *Forest cover and climate factors affecting snow stability (Weir 2002, after Frey and Salm 1990)*

Parameter	In forest openings	Under closed canopy forest
Wind	Modified by terrain and forest margins. Cornice formation is common at ridges. Winds may scour the snow surface.	Wind speed is markedly reduced within and below the canopy. Wind causes snow to fall, which disturbs the surface of the snowpack below.
Precipitation	Accumulation rate is equivalent to precipitation rate unless it is modified by wind.	Canopy interception and sublimation losses reduce accumulation on ground typically by 30% (snow water equivalent).
Air temperature	Strong air temperature gradients develop immediately above the snow surface.	Weak air temperature gradients develop immediately above the snow surface.
Radiation	Receipt of shortwave radiation is affected only by topographic shadowing and forest margins. Losses of longwave radiation promote surface cooling, favouring surface hoar formation and upper-level faceting.	Snow surface is shaded by canopy. Long-wave radiation balance is damped and energy losses from snow surface are reduced. Surface hoar formation is reduced, and generally any surface hoar is destroyed by falling snow from the canopy.

Mountains, and various risk reduction strategies were tried; however, there has been little documentation or research on the results of those treatments.

Most of the prescribed treatments involved reducing the size of clearcuts to decrease wind loading and limit the size of potential avalanches. This includes leaving reserves at obvious starting zones, breaking cutblocks up into a series of smaller patches, feathering the edges of cutblocks or leaving irregular edges to reduce wind effects, pulling back in-block road fills (which can form starting zones), and leaving high stumps or slash piles to increase surface roughness.

Figures 9.41–9.45 show examples of snow avalanche and forest management issues.

FIGURE 9.41 *Damage to a plantation caused by an avalanche that originated in a clearcut near Shannon Creek in the Arrow Forest District. The avalanche in the more recent upper cutblock occurred about 2 years after logging and created what may become a permanent avalanche path. The smaller feature to the left is a small landslide that resulted from drainage diversion on an old skid trail. (Photo: P. Jordan)*

FIGURE 9.42 *New avalanche track created by a large avalanche that originated in an 8-year-old clearcut and caused damage to timber and a plantation near Nagle Creek in the Columbia Forest District, 1996. (Photo: P. Jordan)*

FIGURE 9.43 *Avalanches in these two cutblocks in 1988 created avalanche tracks that threaten the highway below on Slocan Lake in the Arrow Forest District. (Photo: P. Jordan)*

FIGURE 9.44 *Bridge on Bull River in the Cranbrook Forest District destroyed in 1997 by a large avalanche in an established track. (Photo: D. Toews)*

FIGURE 9.45 *Close-up of the destroyed bridge on Bull River (see Figure 9.44 above). (Photo: D. Toews)*

REMOTE SENSING APPLICATIONS

Although terrain mapping and landslide inventories have traditionally relied on aerial photographs, commercially available satellite imagery is now a suitable alternative as the imagery approaches aerial photograph resolution at a reasonable cost (see case study below). Satellite imagery can be easily acquired soon after events occur, and sequential imagery can be used effectively to detect changes. Analysis of stereo satellite imagery may also be used for change detection purposes. Table 9.4 shows satellites that currently provide commercially available imagery. In addition, Google Earth provides an efficient platform for viewing imagery. Much of British Columbia is shown in high-resolution satellite or orthophoto imagery in Google Earth; however, Google does not provide the actual date of the image.

Optimal atmospheric conditions for collecting remotely sensed data include no clouds or extreme humidity; therefore, the collection of imagery is largely restricted to the summer months. Images with cloud cover greater than 20% are considered unacceptable.

In 2005, the Forest Practices Board of British Columbia produced the report *Managing Landslide Risk from Forest Practices in British Columbia*, and used high-resolution satellite imagery as a component of its review (Forest Practices Board 2005). The following are some of the conclusions and recommendations from this report.

TABLE 9.4 *Satellites that provide commercially available imagery*

Satellite	Temporal resolution (days)	Ground swath (km)	Spatial resolution (m)
SPOT-5	26	117	10 × 10 (Multispectral Linear Array) 5 × 5 (Panchromatic Linear Array) 2.5 × 2.5 (Panchromatic Linear Array)
IKONOS	3–11	13	4 × 4 multispectral 1 × 1 panchromatic
QUICKBIRD	1–3	16.5	2.44 × 2.44 multispectral 0.61 × 0.61 panchromatic
GeoEye-1	2–8	15.2	1.65 × 1.65 multispectral 0.41 × 0.41 panchromatic but resampled to 0.5 m

- High spatial resolution, multispectral data collection and stereoscopic viewing greatly improve the ability of a sensor for landslide detection.
- IKONOS and QUICKBIRD imagery outperformed SPOT imagery for landslide detection.
- Imagery used for landslide detection should combine three properties: high spatial resolution, multispectral data collection, and if possible, stereoscopic viewing.

The use of high-resolution satellite imagery for mapping and monitoring the forests of British Columbia is an environmentally responsible choice. Once the satellite has been launched, imagery is available at regular intervals, with no further use of fossil fuels, thereby reducing the carbon footprint of environmental monitoring.

Case study: Using satellite imagery for terrain and landslide evaluation

A remote sensing project has been conducted in the Coast Forest Region since 2000 as a means of studying the effectiveness of using commercially available high-resolution (61 cm to 1 m resolution) satellite imagery for resource feature mapping and compliance and enforcement surveillance (Kliparchuk and Collins 2003, 2008). The results have shown that the following features can be identified: types of disturbance, origins or initiation points of disturbance, area of disturbance and runout zones, connectivity to streams, and degree of revegetation. The spatial distribution of the sediment sources is also readily apparent.

A comparative analysis of IKONOS (81 cm resolution) imagery acquired in 2000 and QUICKBIRD (61 cm resolution) imagery from 2003 was also conducted using Principal Component Analysis (Figure 9.46). The analysis enabled the mapping of the spatial extent and magnitude of change over time across the landscape. Changes in harvesting patterns, revegetation, windthrow, road deactivation, and landslides, and subtle changes on the margins of cutblocks or within standing timber were detectable and provided input to compliance and enforcement risk rating and inspection prioritization. Other types of images were also used in the project. True colour composites allowed for more intuitive interpretation; however, colour infrared images provided better separation between vegetation and bare ground. In addition, the colour infrared highlighted areas of recent revegetation that was due to an increase in vegetation vigour (Figure 9.46b).

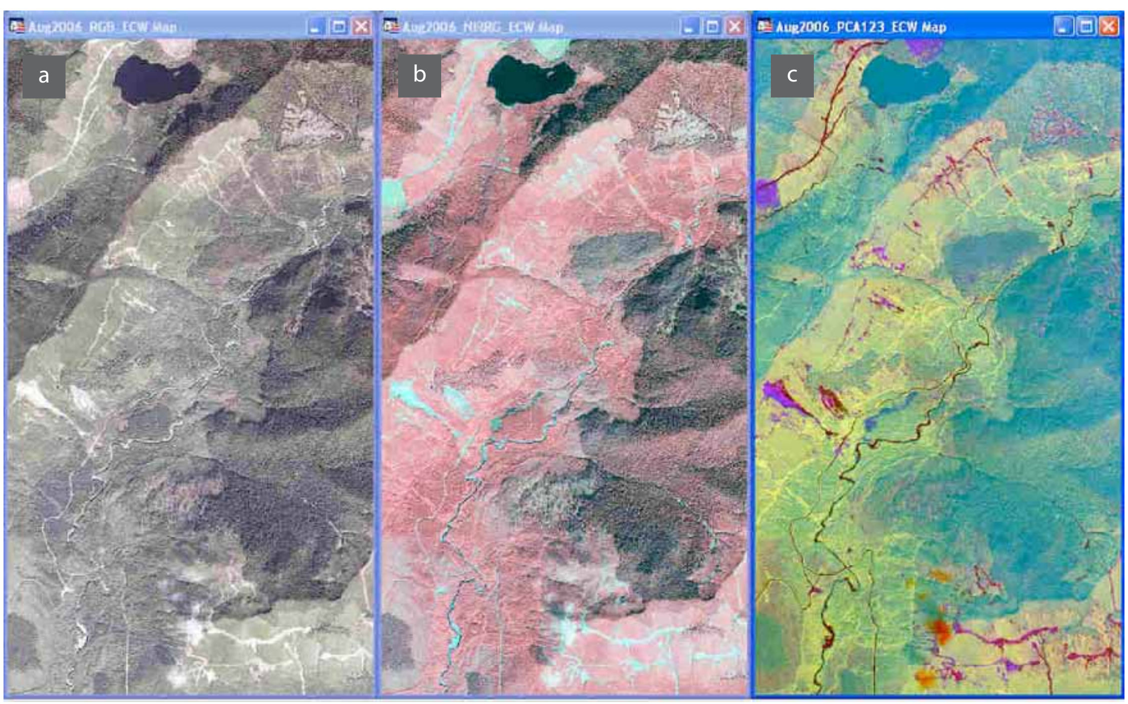

FIGURE 9.46 *Different colour composite images for August 2006 imagery: (a) Red/Green/Blue; (b) Near IR/Red/Green; and (c) Principal Component Analysis 1/2/3. © GeoEye. (Kliparchuk et al. 2007)*

In addition, stereo IKONOS imagery was acquired in 2006 for an area on the west coast of Vancouver Island (Kliparchuk et al. 2007). The stereoscopic capability of IKONOS imagery aided in the interpretation of slope morphology and provided resolutions comparable to air photos and offered the added advantage of multispectral digital data and large area coverage, which eased viewing (due to the larger field of view compared to a photo stereopair) and image enhancement (Figure 9.47). In addition, the imagery, when loaded into a softcopy photogrammetric workstation, can be directly viewed in stereo on a computer screen using polarized goggles, analyzed, and interpreted in three-dimensional perspective (3D), or it can be draped over a digital elevation model (DEM) to provide a 3D view of a landscape, which can be viewed in various perspectives. A DEM is produced by digitizing spot heights and breaklines from the stereo imagery, then turning the elevation points and lines into a regular elevation grid. The DEMs created from IKONOS stereo imagery appear to be more accurate and sensitive to microscale terrain features than those created from digital contour data (Nichol et al. 2006).

Linework digitized using the imagery enables interpretation of slope morphology and allows detailed contour lines to be generated (Figures 9.48a and 9.48b). The derived contour map can then be compared with the TRIM mapsheets, and statistics can be calculated on the minimum, maximum, mean, variance, and standard deviation values for the grid difference values. Kliparchuk et al. (2007) asserted that this process is more efficient and less time consuming than stereo air photo interpretation.

Kliparchuk et al. (2007) sampled spot heights and breaklines between 4 and 9 m within a slide area and in a 15-m area around the margin of the slide area. The spot heights and breaklines were saved as 3D AUTOCAD files (Figure 9.49). The breaklines were saved as points, and then both point files were imported into MapInfo Professional format. A hypothetical pre-landslide DEM was created by removing the spot heights and breakline points from the interior of the landslide, and then was regridded based on the sample points at the edges of the slide. Kliparchuk et al. (2007) did not want to make any assumptions about the convexity of the across-slope curvature, so did not use a gridding algorithm to add convexity. Figures 9.50a and 9.50b show the theoretical pre-slide grid, the post-slide grid, and the differenced grid. The authors summed up the differenced values from the differenced grid to determine the volume of material removed by the landslide: the value calculated was 368 000 m^3. If the pre-slide area had a drainage channel, then the actual volume would be less than the calculated volume.

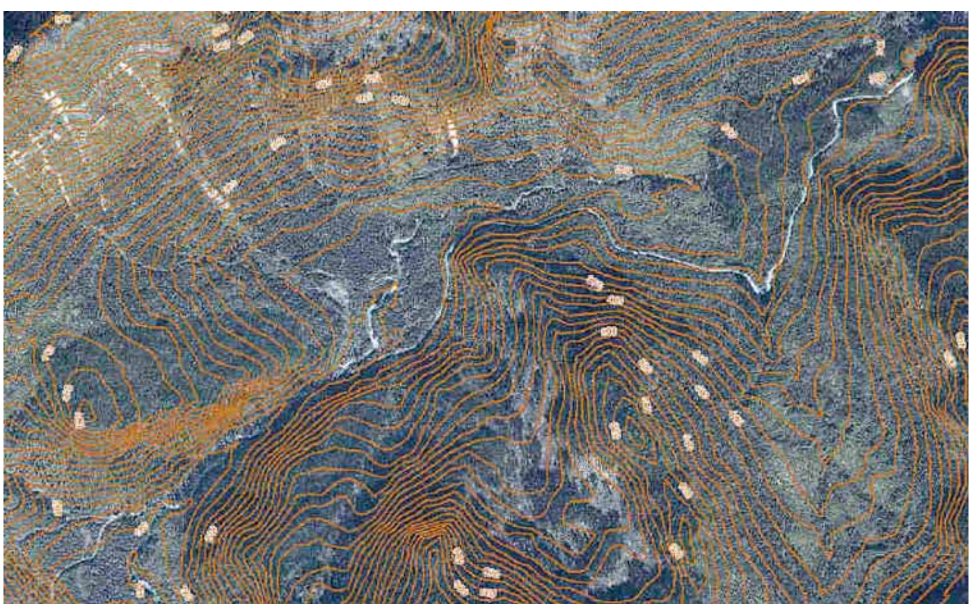

FIGURE 9.47 *IKONOS stereo image-derived contours draped over IKONOS image. © GeoEye. (Kliparchuk et al. 2007)*

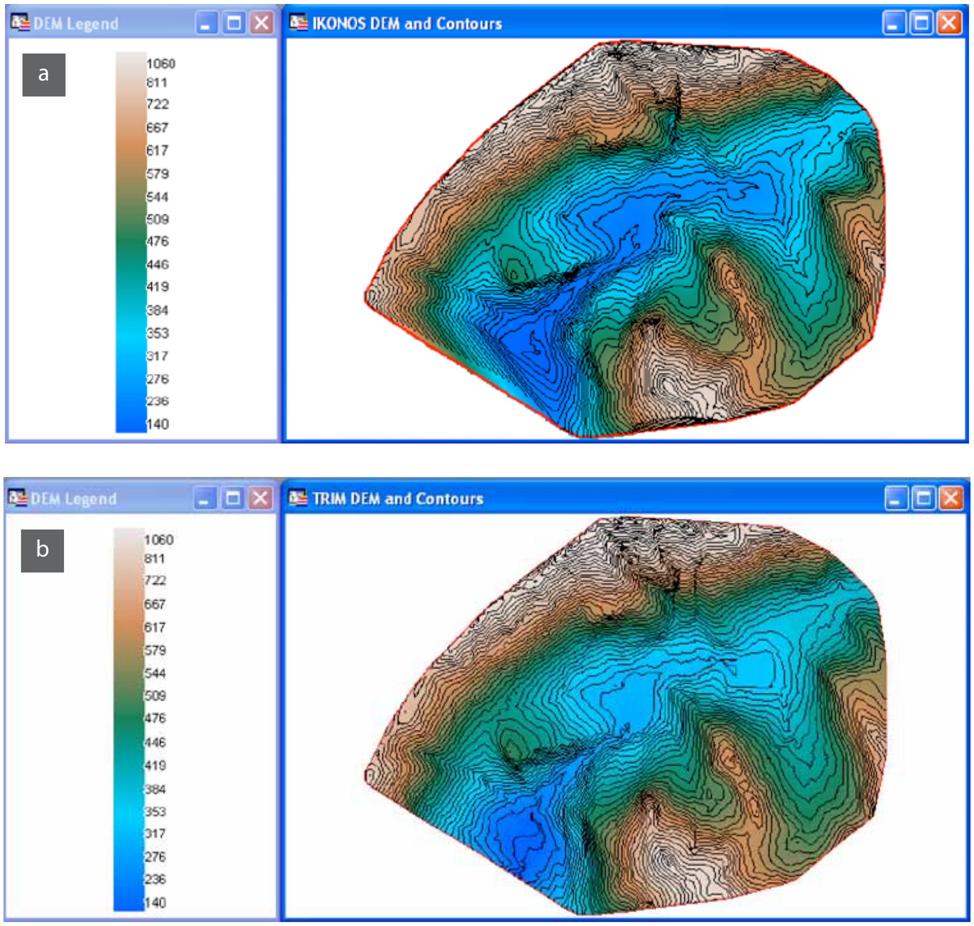

FIGURE 9.48 *(a) Stereo IKONOS-derived contours, and (b) corresponding TRIM contours (Kliparchuk et al. 2007).*

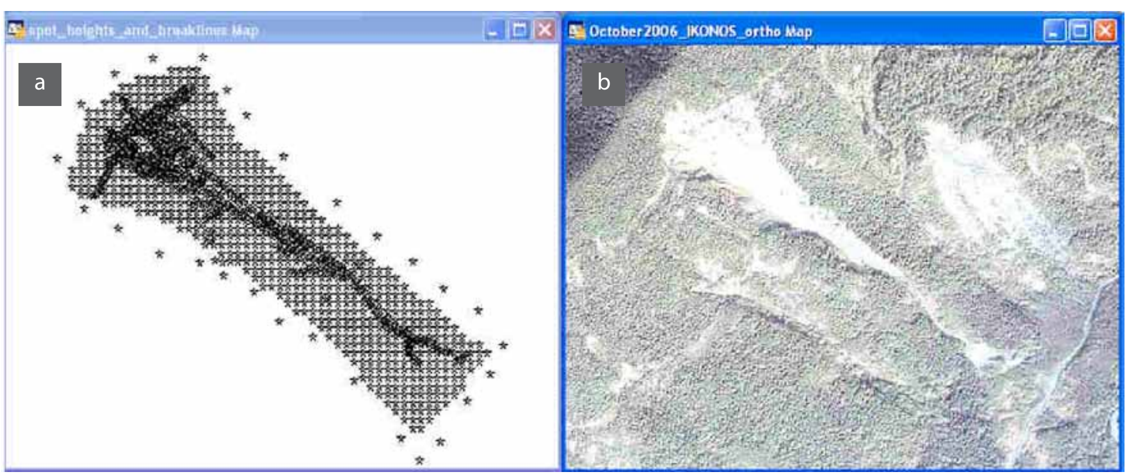

FIGURE 9.49 *(a) Spot heights and breakline points, and (b) image. © GeoEye (Kliparchuk et al. 2007).*

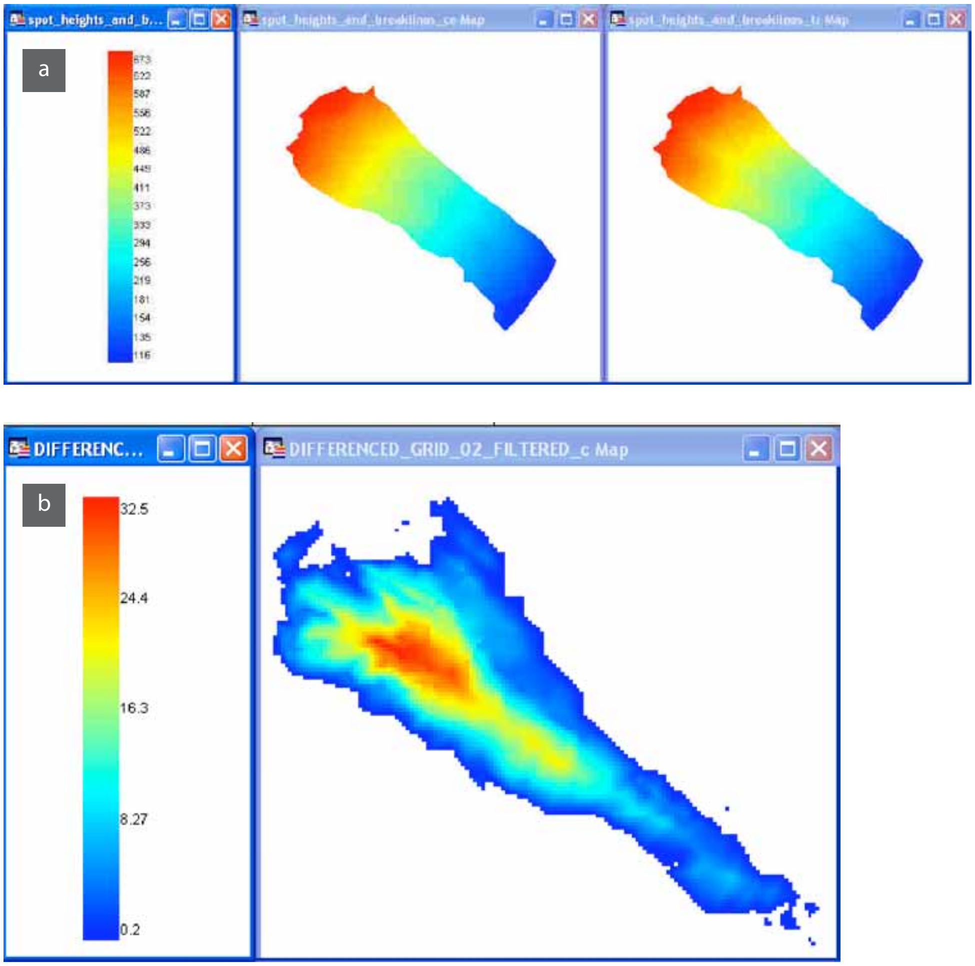

FIGURE 9.50 *(a) Pre-slide grid in the middle and post-slide grid on the right; (b) differenced grid: largest difference is in the red area (elevation legend on the left; values are in metres) (Kliparchuk et al. 2007).*

Kliparchuk et al. (2007) also generated a slope magnitude digital landform model (DLM) from the post-slide grid, which is shown in Figure 9.51. Descriptive statistics for the slope magnitude DLM were also produced (Tables 9.5 and 9.6).

TABLE 9.5 *Descriptive statistics for a slope magnitude digital landform model differenced grid (in metres; data from Kliparchuk et al. 2007)*

Minimum	0.2
Maximum	32.5
Mean	9.1
Variance	61.9
Standard Deviation	7.9

TABLE 9.6 *Descriptive statistics for a slope magnitude digital landform model slope magnitude grid (in metres; data from Kliparchuk et al. 2007)*

Minimum	0.8
Maximum	94.1
Mean	26.6
Variance	246.5
Standard Deviation	15.7

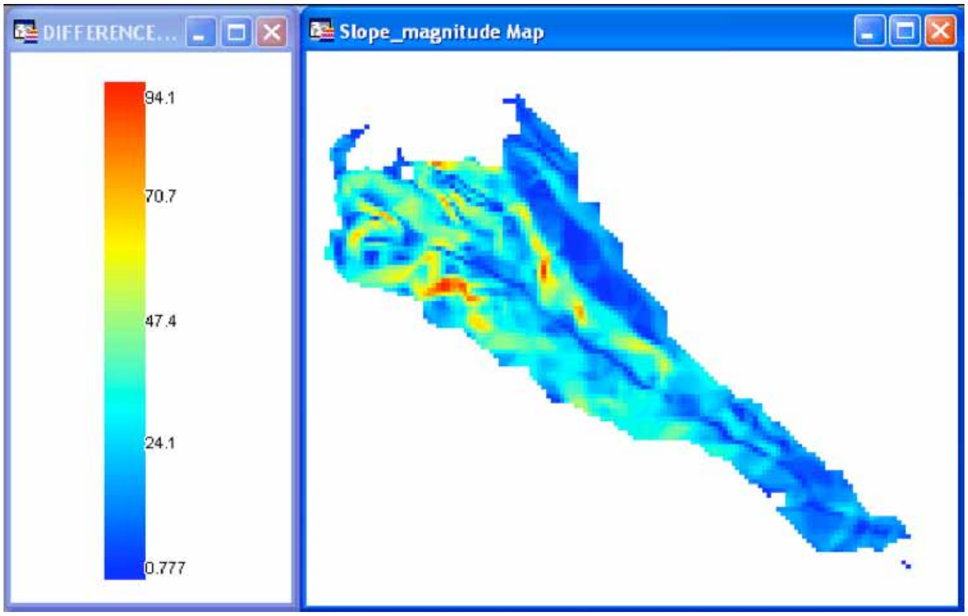

FIGURE 9.51 *Slope magnitude digital landform model (slope in degrees) (Kliparchuk et al. 2007).*

SUMMARY

Historical forest practices have increased hillslope erosion and sediment delivery into streams, which has resulted in significant environmental damage as well as costs to society, such as worker and public deaths, extensive damage to infrastructure, and impacts on human health. Landslides and surface erosion are the primary processes responsible for these effects, but significant differences exist between coastal and interior areas of British Columbia: landslides are generally a greater concern on the Coast, whereas surface erosion from roads is more of an issue in the Interior. Coastal landslide rates increased about 10-fold as a result of historical forest practices, with landslides resulting from both road construction and harvesting. As the costs associated with these impacts became recognized, forest practices began to change.

Practices to prevent landslides or soil erosion typically start with mapping of hillslope stability and surface erosion potential so that sensitive locations are identified before cutblock or road layout begins. Avoidance of unstable or sensitive areas is often the best approach. Detailed terrain stability assessments are often used when operations in potentially unstable areas are being considered and generally include measures to reduce the likelihood of landslides occurring. As understanding of terrain issues increases, applications of new technology, such as remote sensing, or use of more sophisticated analysis, such as formal risk assessments, are utilized. As with any endeavour that entails risk, a better understanding of the processes involved and the use of more careful practices results in fewer problems.

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CHAPTER 10

Channel Geomorphology: Fluvial Forms, Processes, and Forest Management Effects

DAN L. HOGAN AND DAVID S. LUZI

INTRODUCTION

This chapter deals with fluvial geomorphology—that is, with the landforms developed by flowing water. This seemingly simple subject encompasses a rather complex set of interrelated processes that produce a diverse array of fluvial forms. We restrict the content of this compendium chapter to those aspects most relevant to the professional hydrologist and geomorphologist working in British Columbia, a province that is physiographically diverse (Chapter 2, "Physiography of British Columbia") and, therefore, includes many types of streams and rivers. For simplicity, we use "streams" as an all-encompassing term that includes all channels with flowing water (creeks, brooks, tributaries, rivers, etc.), regardless of the absolute size or the timing and frequency of flows carried by the channels. This chapter provides an overview of the key factors determining the morphology and dynamics of forested watershed streams in British Columbia and shows how forest management influences these factors. It is not intended to provide a complete and comprehensive treatise of the subject; the reader is referred to the many textbooks and technical journals for further, detailed coverage of fluvial geomorphology (e.g., Leopold et al. 1964; Schumm 1977; Calow and Petts [editors] 1992; Knighton 1998; Naiman and Bilby [editors] 1998; Wohl 2000; Bridge 2003; Bennett and Simon [editors] 2004). We do, however, show the links to other chapters in this volume and then consider the effects on stream channel conditions and dynamics that result from human influence on the landscape.

Streams of all sizes have the same basic function: they move water, sediment, and other matter (organic, chemical, biological materials, etc.) over the land surface and, ultimately, empty all of these materials into the ocean or a lake. The details regarding time and space scales are numerous; to enable us to consider these factors, we approach this chapter within a watershed context. A watershed includes the entire stream network upstream of a point on the mainstem, as well as the hillslopes contributing water, sediment, and woody debris to the network. The fluvial dynamics at a given point along a stream are influenced by the various processes occurring in the watershed upstream, which is why it is necessary to consider processes that function at a watershed scale. Although most management decisions are made to protect a particular stream reach, it is important to understand that this protection can only be achieved by maintaining properly functioning conditions within the contributing watershed. A stream reach is defined here as a length of channel with homogeneous morphology, discharge, and hillslope-channel coupling.

For this chapter, we do not intend to prescribe best management practices for forestry operations, as these will vary regionally, but rather provide a foundation on which decisions can be made to minimize channel impacts. To do this, we consider the following points, which underlie our objective.

- The watershed is the fundamental landscape unit that must be understood when considering stream channel processes.
- Important differences exist among watersheds in British Columbia that are related to variations in climate, physiography, and land use, as well as ecologic and geomorphic legacies.
- Specific factors within a watershed determine channel morphology; some of these factors are determined by the conditions of the surrounding terrain and others are associated with the properties of the fluvial network.
- Depending on the type of watershed, the production, delivery, and calibre of sediment supplied to the channel are often the most significant of the potential factors influencing channel morphology.
- Forest management activities can interfere with some factors that determine channel morphology and can, if not conducted in a prudent manner, have long-term and widespread harmful impacts on stream environments.

These points are expanded in the following sections.

FACTORS CONTROLLING CHANNEL MORPHOLOGY

Factors controlling channel morphology can be divided into those that are imposed on the watershed (i.e., independent) and those that adjust to the imposed conditions (i.e., dependent). Only the dependent factors can be influenced by forest management activities. The independent landscape factors controlling channel morphology are geology, climate, and human (Figure 10.1). The geology of a watershed is determined by processes acting at the landscape scale, and can include volcanism, tectonics, and, to a lesser extent, surficial processes such as glacial erosion and deposition (see Chapter 2, "Physiography of British Columbia"). Within a watershed, these processes control the distribution, structure and type of bedrock, surficial materials, and topography (Montgomery 1999). Climate is considered an independent factor at the landscape scale, as it is driven by synoptic conditions related to global atmospheric circulation patterns (see Chapter 3, "Weather and Climate"). Human alteration of the landscape can also significantly change watershed conditions.

The geologic, climatic, and human conditions to which a watershed is subjected determine the dependent landscape variables of sediment supply, stream discharge, and vegetation (Montgomery and Buffington 1993; Buffington et al. 2003). Channel morphology is the result of the combined influence of the dependent landscape variables, and the channel responds to changes in these variables by adjustments in one or many of the dependent channel variables (Figure 10.1). An additional important independent variable is time since disturbance.

Sediment supply is determined by the frequency, volume, and calibre of material delivered to the channel. Stream discharge includes the frequency, magnitude, and duration of streamflows. Both temporal and spatial variability in discharge can have a large influence on channel morphology (for a discussion of flood-generating mechanisms in British Columbia, see Chapter 4, "Regional Hydrology"). Riparian vegetation has an important influence on bank erodibility and near-bank hydraulic conditions, and is also a source of in-channel large woody debris (LWD). Classic conceptual models depicted channel morphology as primarily a function of streamflow and sediment transport rate, where transport rate equals sediment supply for equilibrium conditions (e.g., Lane 1955; Blench 1957; Schumm 1971). However, these models did not explicitly address the role of vegetation or other boundary conditions, which often play a critical role in determining channel morphology.

In addition to riparian vegetation, important boundary conditions include elements found within the stream channel, as well as those that may influence the channel's ability to migrate laterally and (or) build vertically. The most important boundary conditions include:

- bank composition and structure, which influence bank erodibility as determined by the sedimentology and geotechnical properties of the material bounding the channel;
- bedrock and other non-erodible units (such as colluvial material, compact tills, and lag glaciofluvial deposits), which may limit lateral and vertical channel migration and determine stream channel alignment;
- erodible sediment stored in valley bottoms in floodplains, fans, or terraces (including alluvial

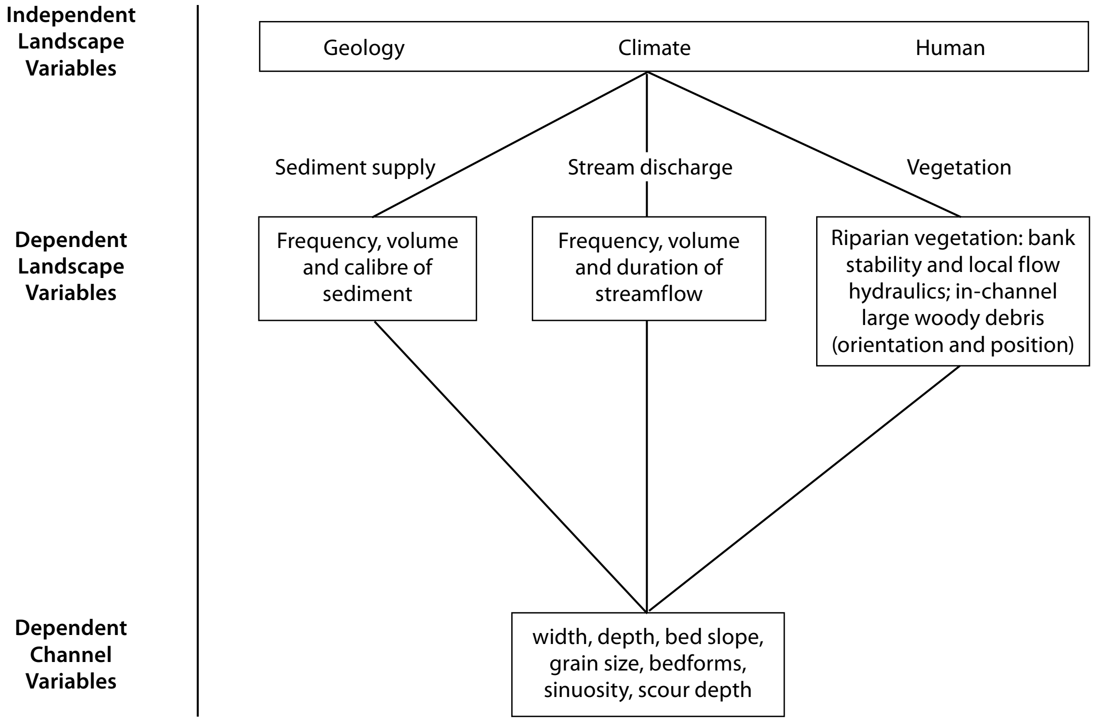

FIGURE 10.1 *Governing conditions as independent landscape and watershed variables and the dependent channel variables (modified from Montgomery and Buffington 1993 and Buffington et al. 2003).*

sediments; lacustrine, marine, and glacial outwash deposits; and fine-textured colluvium);
- valley slope, which, although related to the volume of stored sediment in the valley, represents the short- to medium-term maximum possible gradient that a stream channel can attain; and
- human channel alternations, such as culverts, riprap bank protection, bridge crossings, and flood protection works.

These boundary conditions are primarily influenced by the geomorphic history of a landscape, as well as the history of human intervention. Church and Slaymaker (1989) suggested that, because many streams in British Columbia are still constrained by boundary sediments deposited during glaciation, the streams have not completely adjusted to post-glacial conditions. The current morphology of a stream is, therefore, a product of both present-day and historic watershed processes. Thus, an awareness of both present and historical influences on current stream morphology is important (see the Yakoun case study below for an example).

CHANNEL TYPES, MORPHOLOGY, AND INDICATORS OF DISTURBANCE

Channel type classifications can be based on the type of material through which streams flow and in which channels form. Schumm (1985) proposed a channel classification that included three categories: (1) bedrock, (2) semi-controlled, and (3) alluvial; however, this classification does not address the variable geotechnical properties associated with the glaciated landforms found across British Columbia. Categories should be based on the materials that determine channel bed and bank strength and the channel's threshold of erodibility (Kellerhals et al. 1976). Three categories of materials are used in this chapter: (1) non-erodible, (2) semi-erodible, and (3) erodible. These terms (as opposed to the conventional "non-alluvial" and "alluvial") are more useful from a forest-operations perspective. Although, by definition, all contemporary alluvial material is erodible, many non-alluvial materials are also highly erodible (e.g., marine, lacustrine, and glaciofluvial deposits). Similarly, some alluvial materials are far less erodible than others; for instance, armoured channel beds developed by fluvial processes are far more resistant to movement than other alluvium such as gravel-bar deposits, which are rearranged on an annual basis. Here, we are interested in a classification scheme that distinguishes features on the basis of their susceptibility to changes from forest management activities.

For channels that flow through non-erodible materials (e.g., bedrock, coarse colluvium, and non-erodible glacial deposits), boundary conditions tend to dominate the channel morphology. This type of channel usually has a limited sediment supply and a morphology that is largely determined by the structure and composition of the material through which it flows. Bedrock channels, for example, frequently run along faults or other geologic planes of weakness within the rock. Overall, these channels are relatively insensitive to disturbances, including disturbances from changes occurring upstream (i.e., the channel is relatively stable), but bedrock channels are very effective at transferring disturbances from upstream to downstream reaches. Although bedrock channels resist erosion, Montgomery et al. (1996) observed that LWD could promote sedimentation that potentially causes bedrock reaches to change from non-erodible to "forced" erodible (alluvial) zones. The opposite is also evident where long expanses of former erodible (alluvial) channels are degraded to non-erodible bedrock zones downstream of logjams, which impede the downstream transfer of sediment (Hogan and Bird 1998). These situations support watershed and riparian protection initiatives, particularly in environments where sediment transfer changes can occur, even though the particular zone is non-erodible.

Channels flowing through semi-erodible material may have reaches that alternate between zones flowing through non-erodible, partially erodible, or fully erodible materials. This becomes a scale issue and leads to classification problems; for instance, how extensive should a non-erodible channel section be before it is classified as "non-erodible"? The degree of erodibility can also vary along either the channel's banks or bed, depending on local boundary conditions (i.e., the degree of erodibility of the boundary). Although the transitional nature of this type of channel can make classification problematic, its identification is important because these channels are relatively sensitive to changes in the governing factors.

Alluvial channels comprise the major type of channel within the erodible material category. Alluvial channels often develop within larger alluvial landscapes, such as along main valleys with fan complexes and floodplain features. This type of channel frequently flows through erodible material that has been previously eroded, transported, and deposited by flowing water. Streams bounded by alluvial sediment are active, and relatively major pattern changes may occur as the channel migrates laterally across the alluvial deposit. Major pattern changes may also occur because of changes in the governing factors, such as those produced by upstream land use. Stream channels may form in other erodible, but non-alluvial, materials and these warrant careful consideration when planning forestry activities.

Channels can also be classified according to planform pattern, which in turn is a function of watershed properties. To classify fluvial landforms during air photo interpretation, Mollard (1973) identified 17 planform channel types that were related to both the physiographic environment in which channels flowed, and the materials that made up the channel bed and banks. In general, Mollard (1973) based this channel pattern classification on the factors controlling morphology, specifically streamflow, sediment supply, the relative dominance of fluvial transport processes, and the materials within which the channel is formed.

Church (1992) classified channel patterns on the basis of the calibre and volume of sediment supply, separating the patterns into phases related to how the supplied sediment was then transported (Figure 10.2). These patterns include phases dominated by bed material supply and wash material supply, and a transitional phase where neither bed nor wash material dominates. Bed material generally forms the coarser part of the sediment load a channel is able to transport and constitutes the bed and lower banks of the stream. The wash material is the finer part of the sediment load and is generally transported long distances, being deposited in the upper banks and on the floodplain. Generally, bed material consists of coarser sands, gravel, cobbles, and boulders; wash material consists of finer sands, silts, and clays. The flow energy of the particular channel determines whether sand-sized sediment is assigned to bed or wash material (Church 2006).

For applied or operational purposes, much of the work of Mollard (1973), Schumm (1977), and Church (1992) can be used to assess channel form

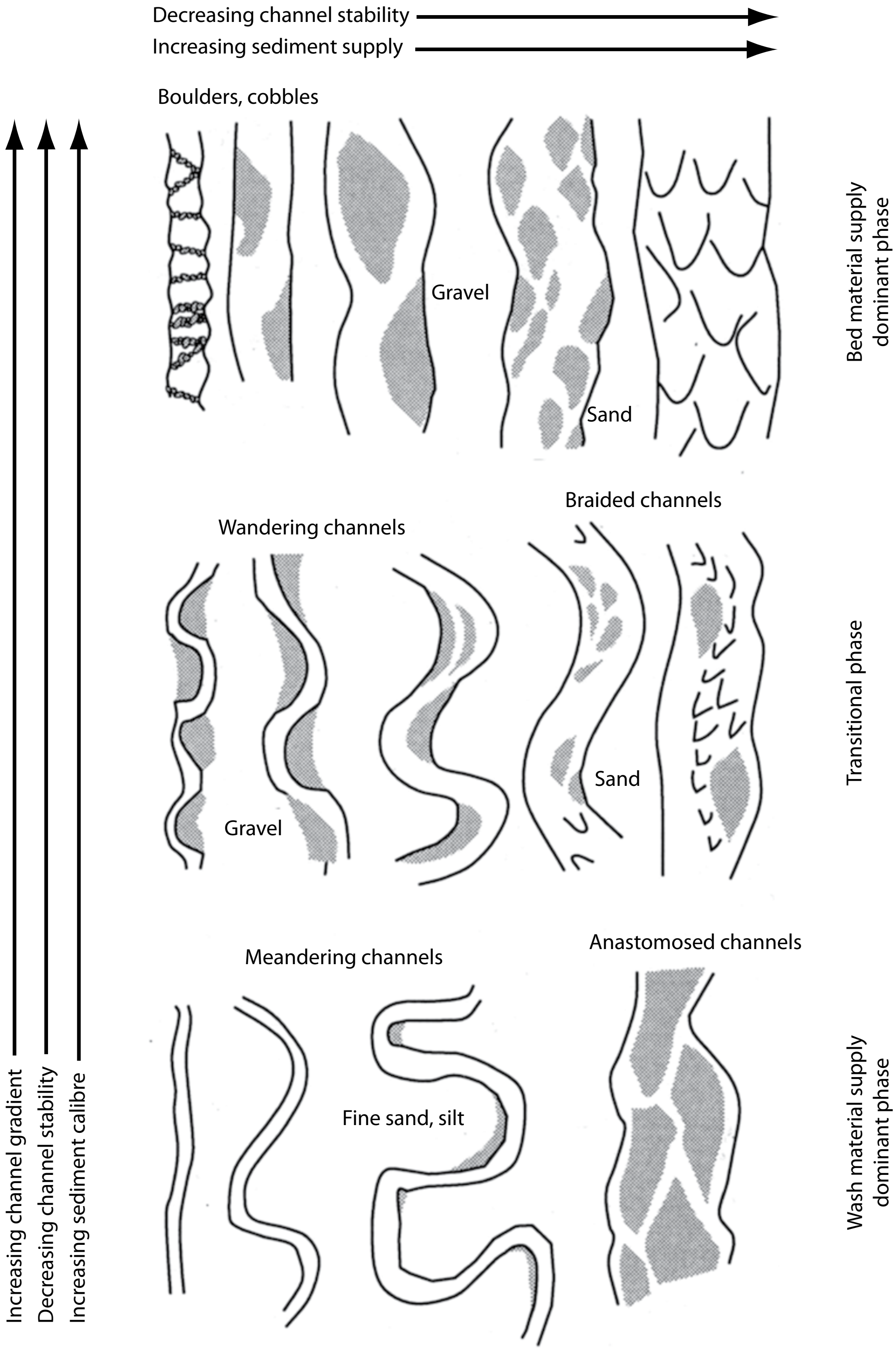

FIGURE 10.2 *Channel form (B.C. Ministry of Forests and B.C. Ministry of Environment, Lands and Parks 1996a, after Church 1992).*

and function. Figure 10.2 shows general trends in channel form as these relate to the governing factors, and the direction of channel stability as it relates to channel form. In this context, "stability" refers to a channel's propensity for vertical or lateral movement (Church 2006). This particular diagram was modified for the province's Channel Assessment Procedure (B.C. Ministry of Forests and B.C. Ministry of Environment, Lands and Parks 1996a) and is used as a preliminary assessment tool to establish channel attributes and to document channel pattern changes over time. As sediment supply increases above the transport capacity of the channel, sediment is deposited (aggradation), which increases the channel width-to-depth ratio, and the level of channel stability decreases. Channel aggradation is evident on aerial photographs as an increase in the size, number, and extent of sediment accumulations within the channel, when compared to earlier photographs. For channels with moderate-sized bed material (such as gravel-bed streams), channels with moderate sediment supply usually have a straight or sinuous planform. As the supply rate approaches or exceeds the channel's capacity to transport the additional sediment, the channel may break into two or more individual channels. When the channel is not too active it can divide and recombine around stable, vegetated islands; these are called "wandering channels" (Figure 10.2). In other situations, the channel becomes too active for stable vegetated islands to develop, and the system divides into numerous individual channels that divide and recombine around unstable gravel bars; these are called "braided channels." Characterized by rapid lateral migration rates, and often undergoing net vertical aggradation, braided channels are amongst the most active of all the stream channels in British Columbia.

Detectable changes in channel pattern indicate important changes in both the watershed and the factors controlling morphology. Managers can use the evidence of channel changes, as prepared by a hydrologist and (or) geomorphologist, as an indicator of the environmental health of the watershed. In general, as sediment supply or streamflow increases, channel pattern becomes straighter (Figure 10.3). Changes in the patterns of in-channel sediment storage in bars and islands are also an early indicator of future channel problems. Bars are non-vegetated accumulations of sediment typically exposed above the low-water level that often develop on the sides

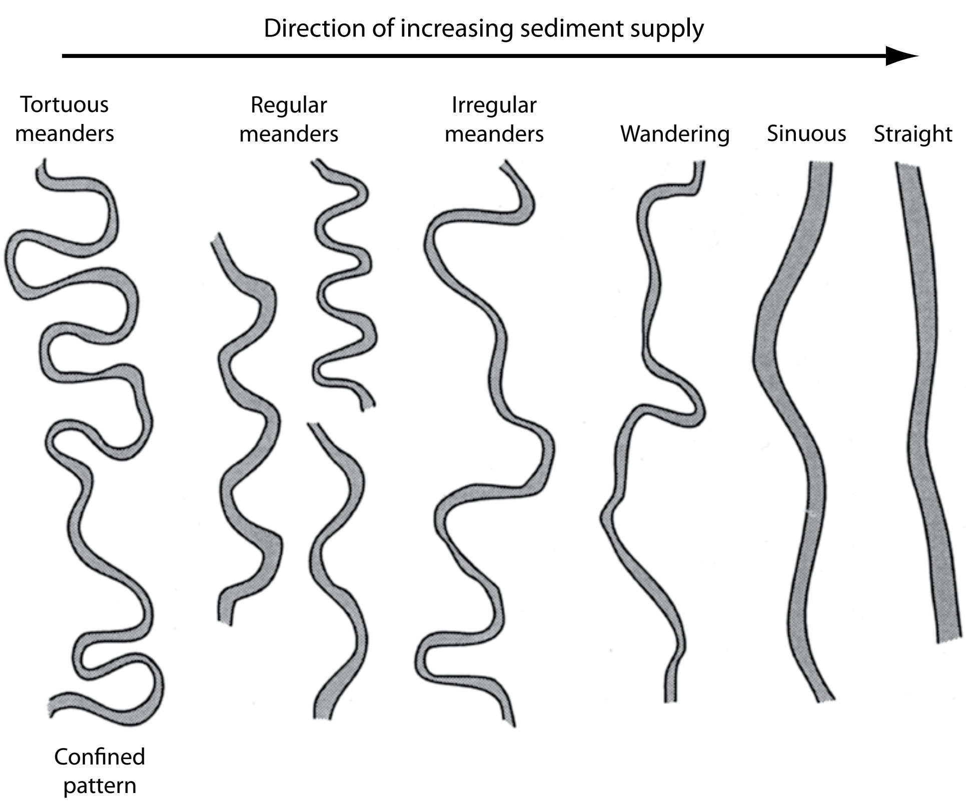

FIGURE 10.3 *Channel pattern classification (modified from Kellerhals et al. 1976).*

of the channel, although these accumulations may also form in the middle of the channel (Figure 10.4). Bars are aggregate features, the stability of which is a function of the interlocking nature of many smaller particles into a larger feature. Changes in bar morphology over time usually indicate variations in upstream sediment supply. For example, if the bars of a stream reach are predominantly medial bars when historically they had been point bars, this may indicate a general increase in sediment supplied to the reach. In contrast, channel islands are vegetated with the top surfaces occurring at or above bankfull channel height (Figure 10.5). Islands are relatively stable over time, but expand and contract in response to long-term sediment supply rates; an increase in the number of islands generally indicates increased sediment supply.

Changes in the lateral activity of the channel (i.e., displacement of the channel laterally across a valley flat surface) may also indicate variations in the conditions upstream (Figure 10.6). Lateral movement is often caused by progressive bank erosion or channel avulsion. Progressive bank erosion can be the result of sediment aggradation within the channel or can occur simply from natural meandering processes. In contrast to bank erosion, channel avulsion is usually a relatively sudden and major shift in the position of the channel to a new part of the floodplain (first-order avulsion), a sudden re-occupation of an old channel on the floodplain (second-order avulsion),

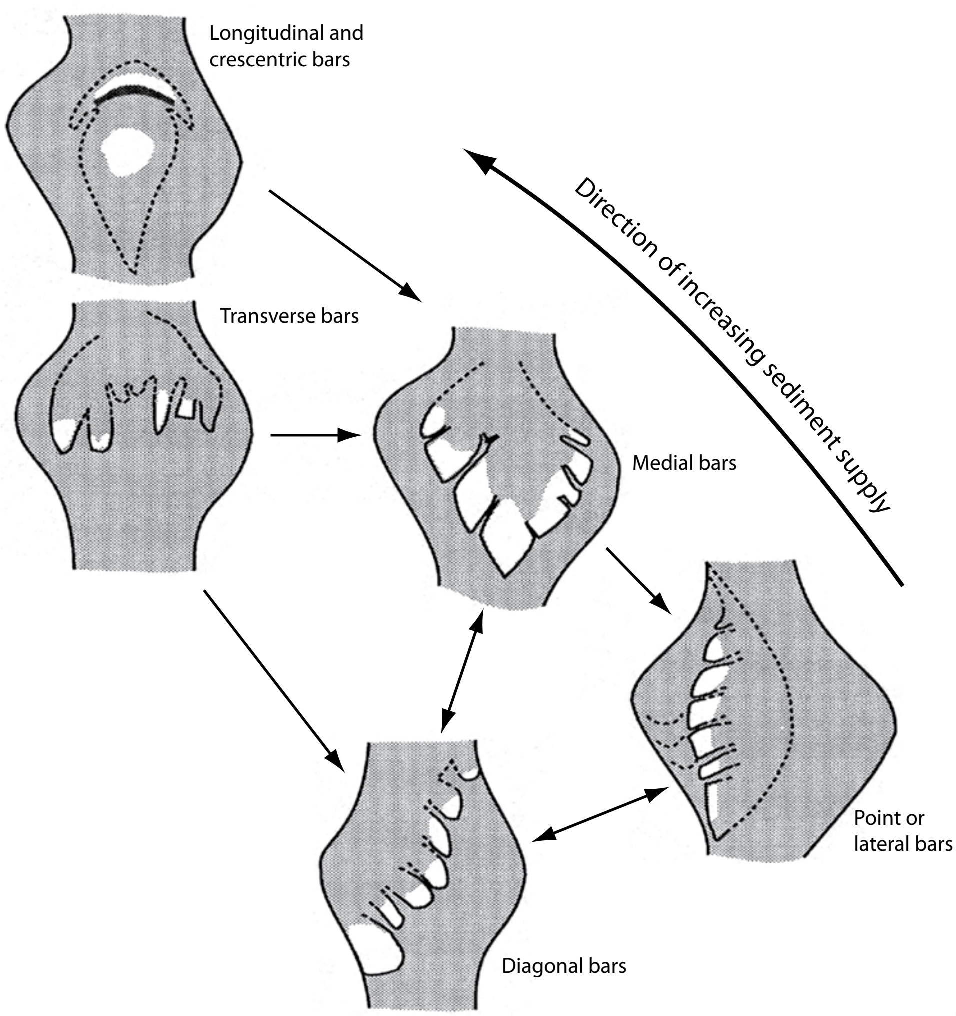

FIGURE 10.4 *Channel bars (B.C. Ministry of Forests and B.C. Ministry of Environment, Lands and Parks 1996a, after Church and Jones 1982).*

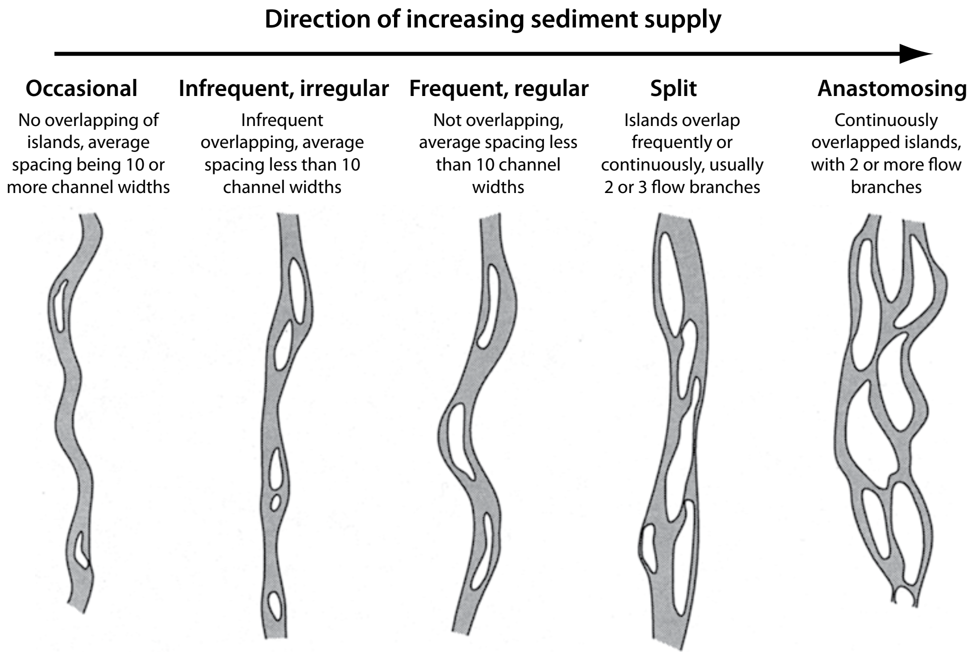

FIGURE 10.5 *Channel islands (after Kellerhals et al. 1976).*

or a relatively minor switching of channels within a braided channel or other similarly active channels (third-order avulsion) (Nanson and Knighton 1996). Logjams accumulating along certain streams are an interesting aspect in British Columbia's forest lands. These features, discussed later in the chapter, can have a dramatic influence on both bank erosion and avulsion processes over long time periods and large areas.

Lateral channel movement influences the riparian zone, eroding some areas and building up others. The channel's boundary conditions and the relationship between the stream and the valley through which it flows will determine the limit of lateral channel movement. If no imposed constraints are present, such as valley confinement,[1] bridges, or dykes, and the valley flat is filled with erodible material, then the channel is usually capable of eroding across the entire extent of its floodplain. Wherever valley width exceeds channel width, a potential for lateral channel movement exists, although in confined systems in which the valley is only marginally wider, the extent of lateral movement is limited. In forested valleys, the additional bank strength provided by riparian vegetation can limit lateral channel movement and enable a stable channel morphology to exist in an environment in which it otherwise may not occur. Millar (2000) illustrated how Slesse Creek evolved from a stable sinuous gravel-bed river to an unstable braided morphology after the removal of riparian vegetation. For additional information on lateral channel movement, refer to Rapp and Abbe (2003) who present a detailed discussion of channel migration and the methods and tools to delineate boundaries for historic, current, and potential lateral channel movement.

Several systems are used to differentiate the various channel types found in British Columbia. The Channel Assessment Procedure (B.C. Ministry of Forests and B.C. Ministry of Environment, Lands and Parks 1996a, 1996b) uses aerial photographs followed by field verification, or just field studies for streams not reliably visible on photographs. Both approaches rely on obtaining data on basic channel dimensions (gradient, width, depth, and sediment size) to provide a systematic, repeatable, and objective method of channel type determination. For intermediate- and smaller-sized streams (bankfull width < 20 m), the procedure identifies three morphologies at low flow conditions: (1) riffle-pool, (2) cascade-

1 Valley confinement refers to the degree to which a channel is deflected by the valley walls or by resistant terraces (Kellerhals et al. 1976).

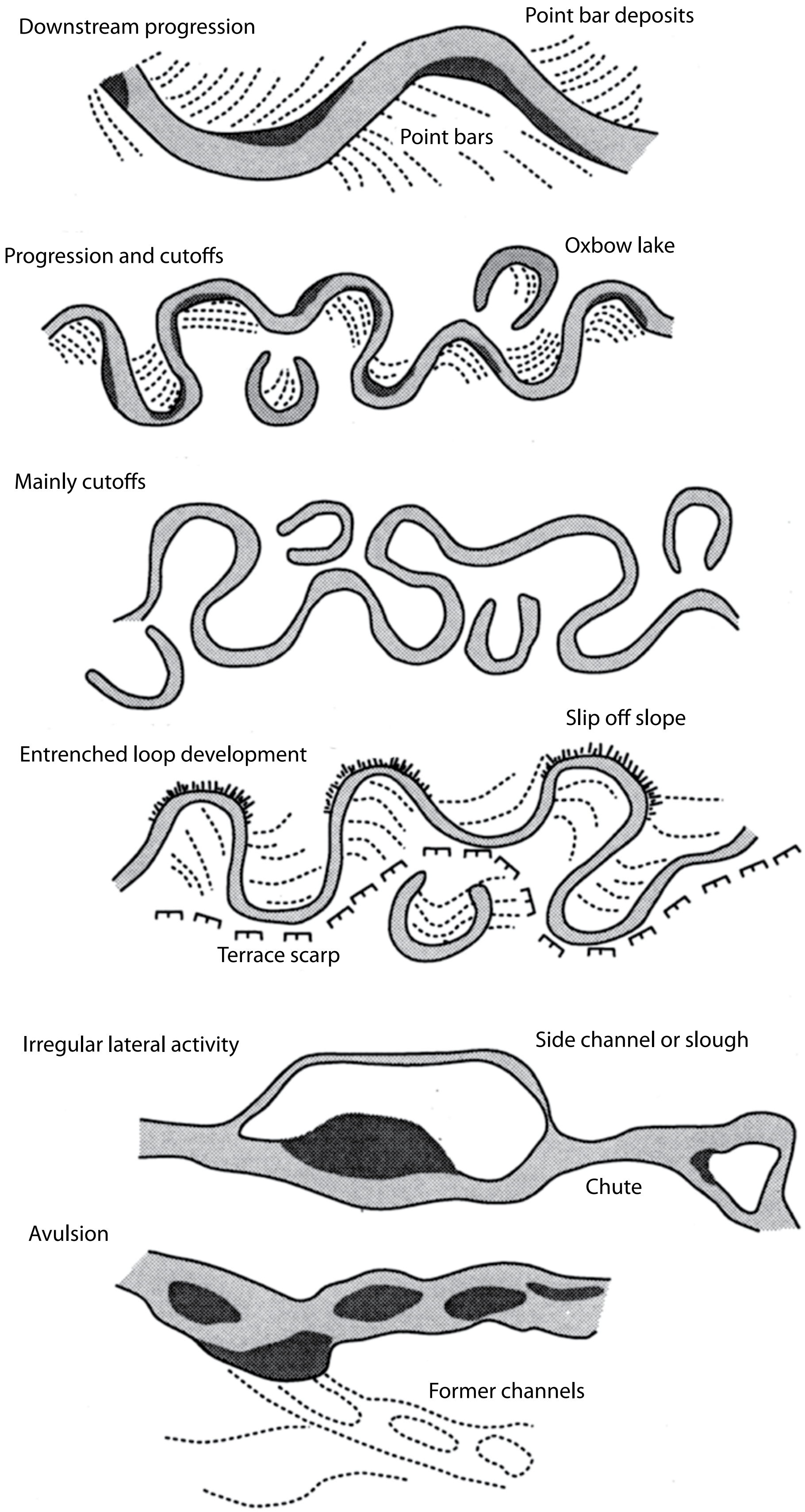

FIGURE 10.6 *Lateral activity associated with a large channel (after B.C. Ministry of Forests and B.C. Ministry of Environment, Lands and Parks 1996a, and Kellerhals et al. 1976).*

pool, and (3) step-pool (Table 10.1). In addition, the three morphologies are further subdivided by dominant bed material clast size and whether LWD is functioning (i.e., influencing morphology), present, or absent in the channel. The riffle-pool morphology consists of riffle, bar, and pool units, with the bar representing the major storage site for sediment storage. In general, pools are topographically low areas with relatively slow-moving water, and riffles are topographically high areas with locally steeper gradients and faster-flowing water (Figure 10.7a).

Riffles are accumulations of sediment that extend diagonally across the channel to the head of a bar, which extends downstream on the opposite side of the channel. As bars are typically deposited on alternate sides of the stream, riffles will cross the channel in alternate directions, shifting from one side to the other as water flows downstream. On average, the distance between riffles is about two to seven times the channel width (Leopold et al. 1964; Hogan 1986; Montgomery et al. 1995). Pools occur upstream of each riffle, and are both narrower and deeper than the riffles at low flow. Pools are often modified by the scouring action of water flowing around obstructions such as bedrock outcrops, large boulders, channel bends, and often wood in forested watersheds (Lisle 1986; Montgomery et al. 1995).

Although riffle-pool morphologies are stable configurations, they are not static. As sediment supply is increased, channel bars expand into the centre of the channel, become less stable, and move more frequently (Figure 10.4). As the bar expands, the riffle attached to the bar expands, and the pool extent is reduced, creating a simplified morphology with minimal depth variability. Bar expansion can also lead to bank erosion, which further increases sediment supply to the reach. The increased sediment supply also changes the composition of the bed material, frequently resulting in a bed surface of finer texture. A reduction in sediment supply causes extensive riffles and bars, reduced pool volumes and depths, and coarser bed surface.

TABLE 10.1 *Channel types and associated characteristics (modified from B.C. Ministry of Forests and B.C. Ministry of Environment, Lands and Parks 1996b)*

Morphology	Sub-code	Bed material	LWD
riffle-pool	$RP_{g\text{-}w}$	gravel	functioning
riffle-pool	$RP_{c\text{-}w}$	cobble	functioning
cascade-pool	$CP_{c\text{-}w}$	cobble	present
cascade-pool	CP_{b}	boulder	absent
step-pool	$SP_{b\text{-}w}$	boulder	present
step-pool	SP_{b}	boulder	absent
step-pool	SP_{r}	boulder-block	absent

At the majority of streamflows occurring in riffle-pool channels in a normal year, bedload does not move; as streamflow stage rises, sediment eventually becomes entrained, usually at or near the stream's bankfull discharge. Bed material is initially entrained from the riffle surface, then from the pool, and as discharge approaches bankfull, sediment is transported over, or deposited on, the next riffle downstream (Pyrce and Ashmore 2003). Therefore, sediment does not move at most flow rates, but does move during infrequent, annually occurring high flows. On the Coast, bankfull discharges commonly occur during the late fall and winter as a result of heavy rainfalls and rain-on-snow events; in the Interior, these discharges occur in spring and early summer as a result of snowmelt (see Chapter 4, "Regional Hydrology").

Similar to riffle-pool morphology, cascades are aggregate structures (generally a series of repeating stone lines), but have cobble- and boulder-sized particles, with water flowing over and around each clumped feature (Figure 10.7b). Pools located between the cascades are usually as long as the channel is wide and tend to be of lower-gradient. The cascade-pool morphologies, which are considered partially erodible features (fully alluvial to semi-alluvial), represent a transitional phase between conditions found in lower gradient riffle-pool channels and the higher-gradient step-pool channels. Increased sediment supply can result in fewer distinct pools and lead to localized bank erosion. Decreased sediment supply can lead to the erosional displacement of the stone lines, leaving no recognizable pattern.

In step-pool morphology, steps are created through the interlocking of a few large particles (usually < 10 stones) aligned across the channel (Figure 10.7c). The steps consist of diagonally arranged stone lines in diamond- or oval-shaped cells and represent abrupt breaks in the longitudinal profile. Pools with finer-textured sediment are positioned between steps. Steps form by the progressive movement of large stones over short distances. These stones eventually jam together, producing very stable features. Step formation also depends on the relative Shields number (i.e., the ratio between the applied shear stress and the stress needed to mobilize the bed) and the ratio between bed material supply and

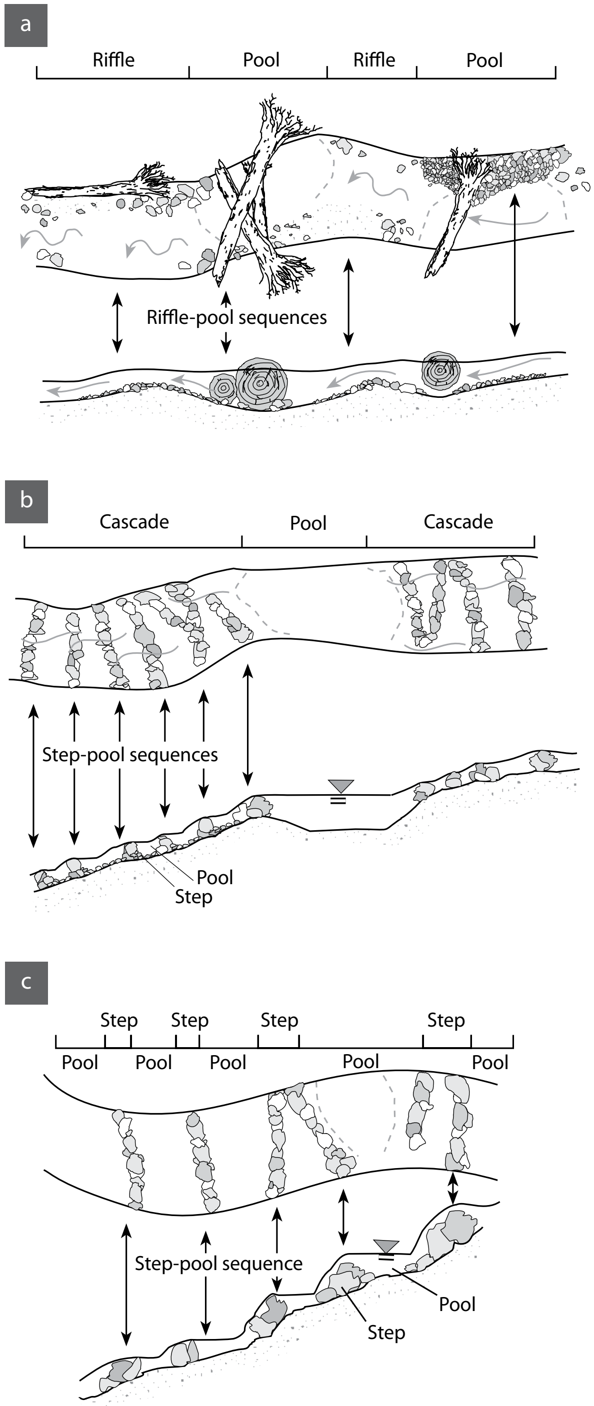

FIGURE 10.7 *Channel morphological units (B.C. Ministry of Forests and B.C. Ministry of Environment, Lands and Parks 1996b): (a) riffle-pool morphology; (b) cascade-pool morphology (after Grant et al. 1990); and (c) step-pool morphology (after Church 1992).*

discharge (Church and Zimmermann 2007). Large woody debris also contributes to step formation; if it is incorporated into the step riser, then step heights increase along with channel resistance (Curran and Wohl 2003). Once steps are established, large storm floods with recurrence intervals of 30–50 years (Grant et al. 1990) and debris flows are required to disturb them, although the actual time frame of step-pool disturbance can vary greatly (Church and Zimmermann 2007). Channels exhibiting step-pool morphology are partially erodible to non-erodible features (semi- or non-alluvial). Channel banks are composed of similar materials (large interlocking clasts) and bank strength is less dependent on riparian vegetation than in strictly alluvial zones.

Each channel type responds differently to changes in sediment supply or discharge. General responses include either vertical shifts (aggradation or degradation, evident by the upward or downward position of the channel bed) and (or) lateral shifts (sideways movement of the bed and banks, evident by old or abandoned channels on a floodplain). The riffle- and cascade-pool types are free to move both vertically and horizontally in the erodible deposits, but the step-pool type is usually restricted to vertical shifts within its non-erodible boundaries (except in steep fans). The Channel Assessment Procedure considers the expected response of each channel type (Figure 10.8). See the *Channel Assessment Procedure Field Guidebook* (B.C. Ministry of Forests and B.C. Ministry of Environment, Lands and Parks 1996b) for a detailed description, including photographs, field examples, and indicators of disturbance.

The Channel Assessment Procedure is intended to evaluate a channel's response to changes in the forces that shape its morphology and does not explicitly assign causes to these changes; the responses will be the same whether produced by natural or human-related influences on sediment supply, riparian vegetation, or streamflow. However, understanding both the cause and result of the response is critical to all aspects of forest management, ranging from initial planning and operational practices to restoration activities (see Chapter 18, "Stream, Riparian, and Watershed Restoration").

Now that the driving factors determining channel morphology have been placed in a watershed context, we next discuss how these factors, watersheds, and stream channels vary spatially and temporally across the province.

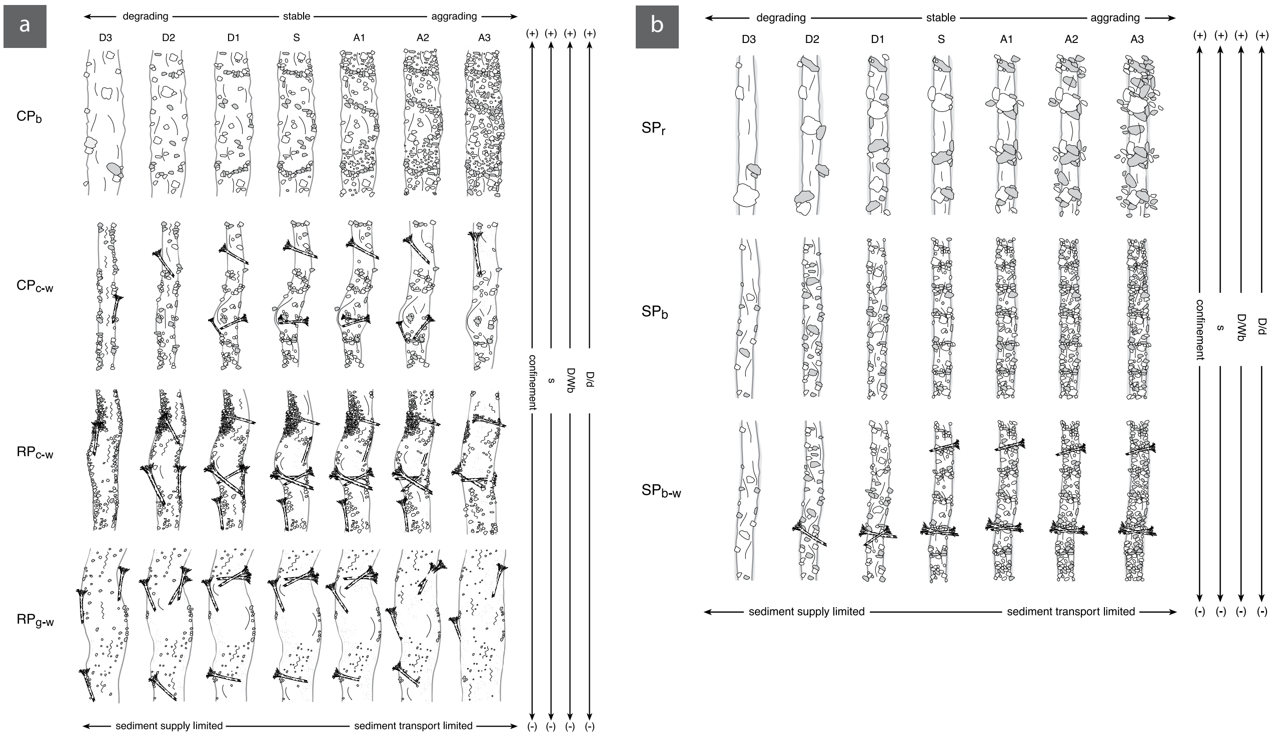

FIGURE 10.8 *Channel morphology matrix showing levels of disturbance (aggradation and degradation): (a) cascade-pool (CP_b, $CP_{c\text{-}w}$) and riffle-pool ($RP_{c\text{-}w}$, $RP_{g\text{-}w}$) morphologies; and (b) step-pool (SP_r, SP_b, $SP_{b\text{-}w}$) morphology. See Table 10.1 for morphology definitions; S = channel slope, D = largest stone moved by flowing water, Wb = bankfull channel width, d = bankfull channel depth (after B.C. Ministry of Forests and B.C. Ministry of Environment, Lands and Parks 1996b).*

STREAMS OF BRITISH COLUMBIA

The factors governing channel morphology do not differ geographically, and are thus considered universal. Nevertheless, each factor's relative importance to a specific channel does vary, as does the factor's internal attributes, which are determined by local watershed characteristics. Although the importance of sediment supply is universal, its actual attributes—that is, whether coarse-textured sediment is delivered directly (but episodically) by landslides, or whether fine-textured sediment is delivered continually from streambank erosion—have a fundamental effect on channel morphology. Therefore, it is the relative differences in the attributes of each factor that lead to the diverse nature of streams in British Columbia. The key to understanding these different functions lies in grasping the nature of watershed characteristics; that is, the type of watershed will determine factor characteristics.

British Columbia's diverse biophysical environments leave a distinct imprint on its streams. The physiography, climate, hydrology, soils, forests, and other components of provincial geography are discussed elsewhere in this compendium (see Chapters 1–4); all provide valuable background for understanding stream environments. Also important to stream development are other landscape features, particularly those strongly associated with the local hydrology. Cheong (1996) explored specific geomorphic and hydrologic variables that influence channel morphology in British Columbia and identified 11 distinctly different watershed types. Table 10.2 reclassifies these into four types and Figure 10.9 shows examples. The watershed types are differentiated on the basis of the percent area covered with:

- perennial snow or ice, which influences the stream discharge regime, especially summer flows;

TABLE 10.2 *Watershed types reclassified and summarized according to connectivity of hillslope and channel sediments (based on 1:50 000 maps)*

Watershed type (%)[a]	Terrain attributes[b]	Physiographic zones[c]
I(25)	Steep, coupled (gullies, fans)	Coast Mountains, Northern and Rocky Mountains, Kootenay and Columbia
II(16)	Steep, decoupled (gullies, floodplains)	Exposed Coast, Southern Rockies, Northeast Mountains
III(33)	Flat, coupled (incised plateau)	Northern Interior, Okanagan, Cariboo and Monashee
IV(26)	Flat, decoupled (floodplains)	Northern Plains, Northern Interior, Exposed Coast

a The percent of the total number of watersheds (*N* = 87).
b For details of the dominant morphological setting, see Cheong (1996).
c Zones taken from Cheong (1996); the first zone has the greatest proportion of the particular watershed type.

- steep lands (greater than 60% gradient), which influence timing of discharge as well as erosion potential;
- lakes or open water, which modulate stream discharge and are sediment sinks;
- valley flats (gradients < 7% and connected to the channel network), which can store both water and sediment; and
- other landforms (extensive gully networks, fan complexes, terraces, etc.), which are sources or sinks for sediment.

The different watershed types are more prevalent in certain physiographic zones; for example, type I occurs most frequently in the Coast Mountains and type III in the Northern Interior (Table 10.2). The watershed attributes determine the stream channel boundary conditions. British Columbia's topography is often thought of as primarily steep and coastal. In fact, large areas of the province—almost 60% of the landscape—are covered with low-gradient topography.

The shape of a watershed controls the overall longitudinal profile of stream channels. Type I watersheds have a concave-up longitudinal profile (Table 10.2; Figure 10.9a). In this setting, streams are expected to exhibit riffle-pool morphology in the wider, lower-gradient, and finer-textured reaches near the drainage outlet, and step-pool morphology in the narrow, higher-gradient, and coarser-textured reaches near the headwaters.

Conversely, type IV watersheds have streams with convex-up longitudinal profiles (Table 10.2; Figure 10.9d). The same channel morphologies exist in this watershed type, but at different positions along the profile. For example, the steepest channels with step-pool morphology are located near the drainage outlet (Figure 10.10a). The smaller headwater channels are characterized by riffle-pool morphology, and LWD is fundamental in providing structure and physical strength and influencing form (Figure 10.10b).

Type II and III watersheds exist in other areas of the province with channel longitudinal profiles reflecting the imposed conditions (Table 10.2; Figures 10.9b and 10.9c). Longitudinal profiles are a blend of alternating convex-up and concave-up patterns—profiles common in the province's low-relief interior—with frequent hanging valleys produced by glacial erosion. Channel morphology is a function of slope (Buffington et al. 2004; Brardinoni and Hassan 2007) and can be predicted on the basis of watershed type. In addition, watershed types provide information on other terrain attributes, such as the degree of hillslope-channel coupling, which is another important factor in determining channel morphology.

As the absolute size of a particular watershed increases, the overall shape of the composite drainage will depend on the arrangement of internal sub-basins. Each sub-basin can be a different watershed type and, since each watershed type has specific topographic characteristics (steep headwaters, flat valleys, lakes, etc.) with associated hillslope-channel coupling properties, channel morphology may vary greatly along the watercourse. Depending on scale, the possible combinations are virtually limitless, as is reflected in British Columbia's extremely diverse range of stream types.

Within a watershed, stream sediment supply has a critical influence over channel morphology; however, sediment delivery to, and movement within, a stream has only been implied thus far. A sediment budget addresses this issue and is commonly defined as an accounting of the sources, storage, transfer, and fate of sediment within a watershed. (For more information on constructing a formal sediment

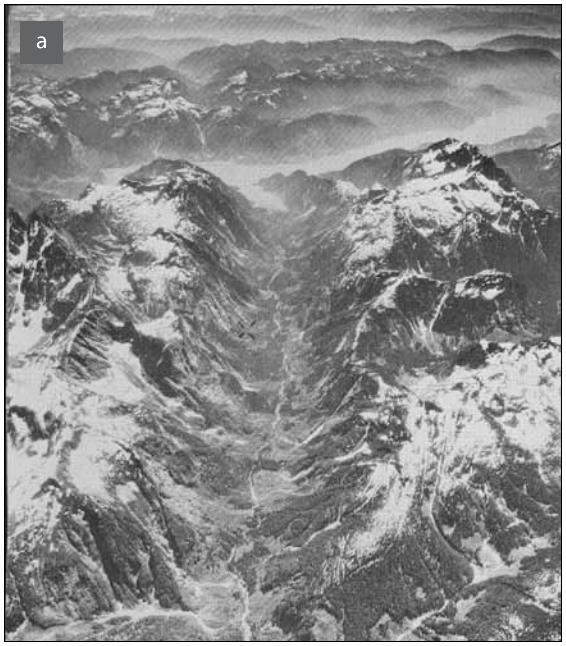

Plate XIIb. (22) Coast Mountains, Pacific Ranges. Looking southeast down the glaciated valley of Tingle Creek to Stave Lake near the southern edge of the Pacific Ranges. Mountain Baker (10,778 feet), a volcanic cone in the Cascade Mountains of Washington, is in the right distance. Photo B.C. 499:82.

Plate XIb. (20) Coast Mountains. Chilcotin Ranges. Looking northwestward across Taseko River toward the abrupt front of the Chilcotin Ranges against the Fraser Plateau. Elevation of Taseko River is just below 4,500 feet. Mount Tatlow (10,058 feet) is in the left distance and Mount Waddington is the high peak on the skyline. Photo B.C. 654:35.

Plate XXXVIII. (71) Alberta Plateau. Looking southwest across a remnant of the upland surface of the Alberta Plateau at an elevation of 2,500 to 3,000 feet between the Fort Nelson and Muskwa Rivers. Notice the scarp, which is the outcrop of a flat-lying sandstone member. Photo R.C.A.F. T27R-196.

Plate XLa. (74) Fort Nelson Lowland. Looking east across the Fort Nelson Lowland, elevation 1,300 to 1,400 feet, from the junction of the Kahntah and Fontas Rivers. The large meltwater channel on the left runs southwestward from Ekwan Lake. The relief on the surface is not more than 300 feet. Photo B.C. 1198:71.

FIGURE 10.9 *Examples of watershed types from Holland (1976): (a) type I watershed; (b) type II watershed; (c) type III watershed; and (d) type IV watershed.*

budget, see Dietrich and Dunne 1978 and Reid and Dunne 1996.) A simplified sediment budget includes both terrestrial and aquatic sources, and storage and transfer (hillslopes and channels) components (Figure 10.11). Sediment is delivered to the channel through three main hillslope processes (landslides, soil creep, sheetwash) or it can be stored as colluvium along valley floors and floodplains until it is then transferred to the channel by colluvial and fluvial processes (Figure 10.11; see also Table 2.1 in Chapter 2, "Physiography of British Columbia").

Although quantitative sediment budgets are rarely constructed for management purposes, conceptual sediment budgets can be developed that identify the watershed processes most important to channel morphology in a particular basin. The simplified budget (Figure 10.11) illustrates three important points.

1. Several process types exist, with each type producing different sediment amounts and textures; sediment delivery to the channel depends on the channel's location within the watershed. In the hillslope zones, landslides produce the greatest amounts of sediment (relative to other processes) and the textures vary by several orders of magnitude (from boulders to sand), although the material is primarily coarse textured. Lesser amounts

FIGURE 10.10 *Examples of channels in type IV watersheds: (a) step-pool morphology near drainage outlet; and (b) pool-riffle channel near drainage divide. (Photos: D. Hogan)*

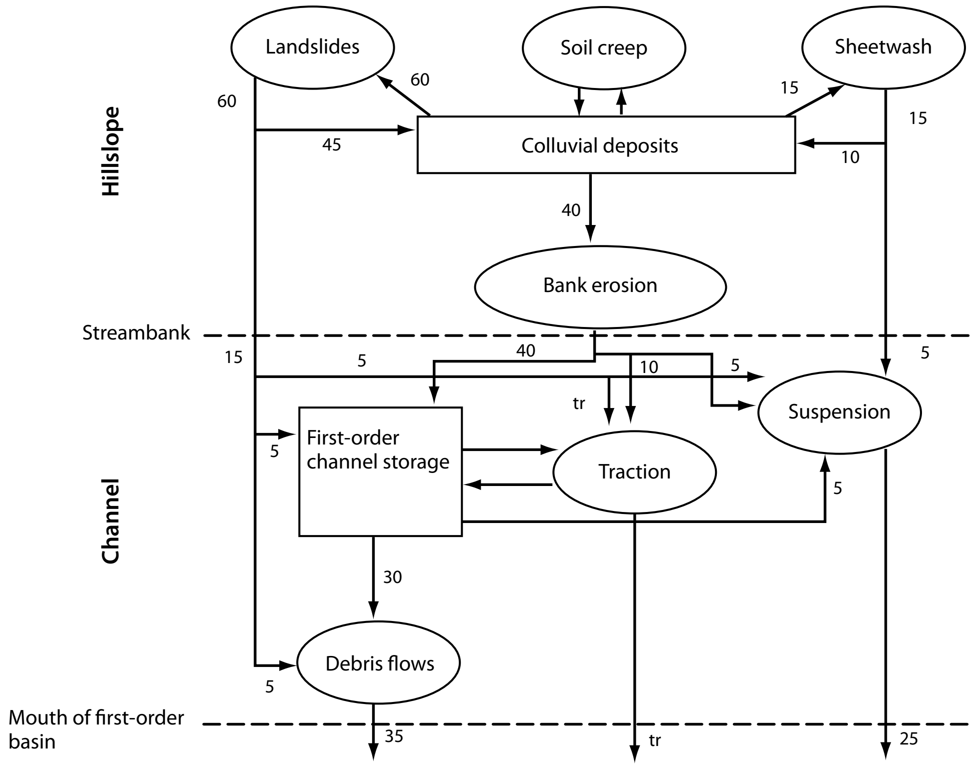

FIGURE 10.11 *Hypothetical sediment budget for a first-order basin. Processes are noted as ovals, storage elements as rectangles, and transfers as arrows; streambank and basin mouth are noted as dashed lines. Sediment transfers values are given in t/km² per year (after Reid and Dunne 1996).*

(i.e., less than a quarter of landslide production) of finer-textured, in-channel sediment are produced by soil creep and sheetwash and through road-related erosion. Much of this material is stored as colluvium in fans or valley fill along the hillside footslopes. In the channel zone, sediment produced through streambank erosion is stored for differing durations within the channel margin and LWD-related storage areas.

2. As the watershed becomes larger and the distance from the headwater zone becomes greater, the amounts of sediment produced and delivered to the channel by soil creep and sheetwash increases relative to that delivered by landslides; the coupling of the hillslope and stream channel becomes less direct (the channel becomes increasingly isolated from the hillslope due to the presence of a valley flat).
3. Nearby sediment sources, such as floodplains, channel banks, and in-channel sediment storage, increase downstream.

Consequently, all aspects of a sediment budget, as constrained by the outlet of the drainage basin, will depend on watershed type. The delivery of landslide, soil creep, and sheetwash material from upslope to the stream network is conditioned by watershed properties. In type I watersheds, where steep slopes are directly coupled to the stream network, landslides will clearly be the primary sediment source, if the materials are susceptible to mass wasting (see Chapter 8, "Hillslope Processes," and Chapter 9, "Forest Management Effects on Hillslope Processes"). The nature of sediments derived from landslides depends on the parent material, but the textures will generally be coarser than those derived from other input mechanisms such as from upstream reaches. In these cases, source mechanisms other than landslides are secondary. However, if watersheds are less steep and (or) hillslopes are not coupled to the channel (types IV and II), then soil creep and sheetwash are relatively more prevalent. These mechanisms will deliver finer-textured material to the stream system than those derived from mass wasting. In other watershed types (types III and II), the relative rates of sediment production and delivery to the channel will vary. For larger watersheds, the configuration and type of individual sub-basins can also influence sediment budget dynamics. For example, a steep, coupled sub-basin (type I) flowing into a channel that originates in a flat, uncoupled sub-basin (type IV) can strongly influence sediment dynamics at and downstream of the confluence.

Time is another aspect of sediment supply implicitly included in a sediment budget. For example, landslides occur episodically and thus deliver large volumes of material infrequently over a given time period. In watersheds prone to episodic landslide inputs, the channel must constantly adjust to the natural rate of landslide disturbance. Soil creep and sheetwash occur chronically and thus deliver relatively smaller volumes to the channel. This sediment is much finer and will likely not alter channel morphology (Figure 10.2), but it can have adverse effects on aquatic biota. If channel gradient is constant but sediment supply is increased, then the stream will change from a single-thread channel (with meanders in low gradients and step-pools in steeper gradients) to a multiple-thread channel that is either anastomosing at lower gradients or with braids and chutes or cascade morphologies at steeper gradients.

In addition to sediment, LWD is an important component influencing stream morphology and is common in many forested streams in British Columbia. Although characterized in various ways (see review by Hassan et al. 2005), it is most frequently defined as wood material 1 m or longer with a mean diameter of greater than 0.1 m. Large woody debris enters a stream section by several mechanisms, including as inputs from landslides (Figure 10.12a), windthrow or blow-down (Figure 10.12b), bank erosion (Figure 10.12c), tree mortality and fall (Figure 10.12d), and flotation from upstream (Figure 10.12e). The type of watershed largely controls the dominant LWD input mechanism. Woody debris from landslides will predominate in steep, coupled watersheds, whereas windthrow and bank erosion are important in lower-gradient, decoupled basins. In the Interior, tree mortality is important, especially in areas where vast expanses of insect-infested forest and (or) forest fires occur. Downstream flotation of LWD depends on stream size and related scale factors, but this mechanism is generally more important in larger

FIGURE 10.12 *Large woody material input mechanisms: (a) landslides, (b) windthrow, (c) streambank erosion, (d) tree mortality, and (e) flotation from upstream. (Photos: D. Hogan)*

channels where the typical tree height is less than the channel width (Montgomery et al. 2003).

After its delivery to the channel, LWD has a range of effects that will depend on the relative size of wood compared to channel dimensions and the arrangement of wood within and along the channel. The input mechanism and dominant tree species often determine the size of wood entering a stream channel. For example, woody debris introduced by landslides has a range of sizes, including both intact and broken stems, and many smaller pieces; tree mortality, blowdown, and streambank erosion deliver mostly intact stems resulting in larger pieces. The size of these trees varies by tree species and forest type.

Figure 10.13 shows the range of tree heights and bole diameters for the dominant species in all the biogeoclimatic ecosystem classification (BEC) zones in British Columbia. Larger trees affect a greater range of stream sizes than smaller trees and this influence depends on biogeoclimatic factors. Channel complexity and diversity are attributed to LWD characteristics. When the wood is large compared to channel width and is predominantly oriented across the channel, it directly affects water flow as well as sediment movement and storage. Where LWD is oriented across or perpendicular to the channel, channels are wide, sediment textures are highly variable spatially, and banks are undercut. Where LWD is lying parallel to the channel, channels are narrow, beds are scoured, and banks are vertical or sloping away from the channel. This LWD architecture is typical in all forested watershed streams across all forest types (Bird et al. 2004).

Landslides and windthrow are commonly responsible for the entry of large numbers of trees to the channel at a single point, although bank erosion and mortality will also deliver many trees to a single location. As channel size increases and becomes less connected to the hillslope, LWD is more commonly floated in from upstream. This episodic delivery of substantial amounts of LWD has particular significance to the spatial and temporal variability of stream channel conditions. Hogan et al. (1998a) linked landslide frequency to the presence of logjams, showing that logjams invariably occur where landslides in the forest enter stream channels. Accumulations of LWD that interfere with water flow and the transfer and storage of sediment within the channel are also referred to as "jams." Jams are

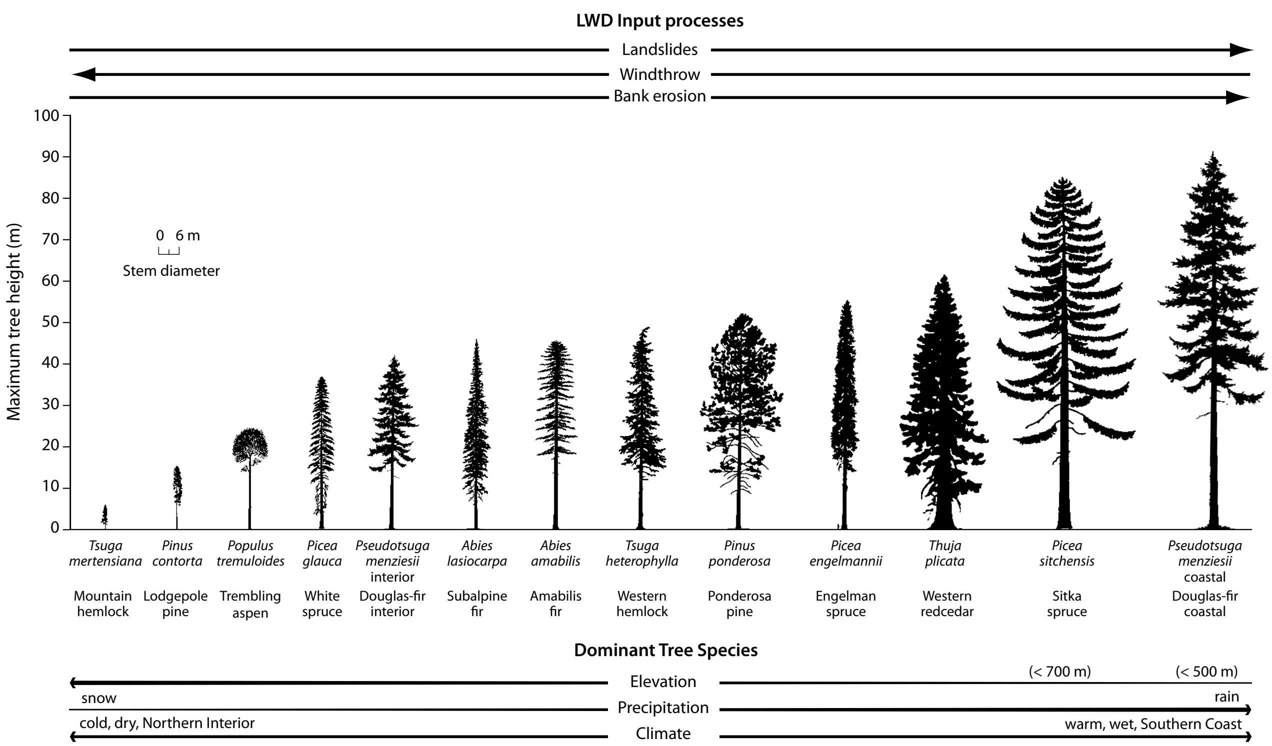

FIGURE 10.13 *Large woody debris input processes and maximum tree height for dominant tree species in all the biogeoclimatic zones in British Columbia. General patterns of LWD input processes based on preliminary data analysis and regional physiography. All sizes are approximate and for illustrative purposes only.*

either wide and low structures in zones with free lateral movement (as in streams with a channel migration zone or floodplain), or narrow and high structures in zones confined by erosion-resistant materials, thereby creating areas of vertical sediment deposition in an otherwise erosional or bedrock reach, the "forced" alluvial reaches of Montgomery and Buffington (1997).

Formation of in-channel logjams at the terminus of landslide run-out paths leads to many channel modifications. Soon after the jam forms, the channel both upstream and downstream of the structure undergoes major changes (Hogan 1989). For example, the channel tends to fill with sediment upstream as the intact barrier of wood interrupts the downstream transfer of sediment. In addition, surface gradients decrease, channel banks erode as the channel expands to accommodate increased storage of sediment, surface sediment textures become finer, and pools in-fill and decrease in overall extent with a corresponding increase in riffles and braided zones. Overall, the channel comes to resemble a simple run or glide. Downstream of the logjam, the channel is deprived of sediment from upstream and adjusts by downcutting, which causes the loss of pools, locally steeper gradients, coarse surface sediment textures, and fewer pieces of functional LWD. The combined zone of influence (both upstream and downstream) may exceed distances equivalent to 100 channel widths in length (Hogan et al. 1998b).

In addition to the spatial influence of logjams, temporal adjustments occur (Hogan 1989). In the first decade after jam formation, wood begins to deteriorate and the jam structure gradually becomes more open, allowing sediment and woody material to pass around or through the jam. After a decade, the processes occurring immediately after jam formation reverse—the upstream channel begins to downcut, pools develop, surface sediment textures become coarser, previously buried LWD becomes exhumed, and the extent of riffles decreases. Downstream processes are also reversed with increased bar development, pool formation around LWD, and textural fining. This trend continues and channels gradually return to the complex, diverse environments that existed before the landside inputs. The temporal adjustments last for about 50 years. Similar trends are evident in non-coastal areas, although documentation is not yet completed (D. Hogan, B.C. Ministry of Forests and Range, unpublished data).

In British Columbia, debris budgets (wood input, storage, and output) have been developed for several decades; however, many difficulties are associated with the conclusive determination of input mechanisms. Preliminary results suggest that input mechanisms vary by BEC zone, with landslide inputs more prevalent in the steeper coastal zones, windthrow-related inputs being more common in northern interior zones and streambank erosion more important in interior and northern areas (Figure 10.13). Early results also indicate that the pattern of in-stream storage, that is LWD predominantly stored in jams, is similar across the province. Little is known about the output of woody material from the stream system; this budget term is commonly deduced from the other two terms and includes the inherent uncertainties of both.

This section considered the factors controlling channel morphology from the context of British Columbia's different watershed types and the processes occurring within each. This background is necessary for the following discussion of the influence of forestry activities on stream conditions.

FOREST MANAGEMENT INFLUENCES ON CHANNEL MORPHOLOGY

Resource managers have been interested in the influence of forestry activities on watershed conditions, and streams in particular, for decades. Of prime importance is the need to protect public and worker safety and the environment. Safety issues are usually related to landslides, but also involve stream crossings and roadways on floodplains. Terrain specialists deal with many of these issues, but considerations regarding structure location and flood mitigation remain important. The federal *Fisheries Act* regulates fish habitat protection, and several government and industry agencies have implemented forest practices regulations and guidelines to both safeguard aquatic environments and fish habitat, and ensure channel integrity.

A great deal of research has dealt with management and channel morphology (see Fish–Forestry Interaction Program [FFIP] references at end of chapter). Here, we restrict our discussion to the influences of management practices on the factors controlling channel morphology (Figure 10.1). Case studies illustrate many of the points included in Table 10.3, which lists a ranking of these factors according to the potential of forest management activities to alter a channel. The factor ranking is based on our assessment of forest management impact causing stream channel change, although its validity is the subject of an ongoing debate amongst geomorphologists and hydrologists. This debate centres around the relative importance of management activities on the dependent factors (sediment supply, riparian vegetation, and streamflow changes), and not the independent, or geologically imposed, factors.

Case Studies

The five case studies reviewed here are drawn from diverse physiographic and biogeoclimatic settings. The focus is on watersheds, with an emphasis on relatively low-gradient, riffle-pool streams. These are generally the most sensitive to the effects of forest management and have been the subject of B.C. Forest Service research efforts over the past three decades.

The first case study concentrates on a suite of examples from Haida Gwaii (Queen Charlotte Islands). The first example provides an overview of the range of different channel types found within a single old-growth forested watershed. The next three examples illustrate the importance of LWD in modifying channel and riparian zone character over both time and space. The role of forests, in terms of their natural disturbance patterns and susceptibility to management modifications, is a fundamental key to understanding channel morphology in British Columbia.

Another coastal case study, from Carnation Creek on Vancouver Island, provides finer temporal and spatial resolution on the nature of channel alterations documented on Haida Gwaii. Carnation Creek is part of a comprehensive fish–forestry interaction program, and channel monitoring began in 1971 (see www.for.gov.bc.ca/hre/ffip/CarnationCrk.htm). Extensive upslope logging and landslides have occurred within this watershed, and the channel experienced riparian logging that has affected streambank stability and LWD supply.

The Donna Creek case study considers channel response to a single, large landslide event in north-central British Columbia, and addresses several geomorphic and anthropogenic factors. The watershed has undergone specific land use changes (for details, see Chapter 9, "Forest Management Effects on Hillslope Processes"), and logging activities have influenced runoff patterns, sediment delivery to the channel, and riparian zone disturbance. The Fubar Creek case study summarizes a monitoring program that was designed to document channel adjustment and recovery of small, interior streams following riparian logging and disturbance to the channel banks. The Yakoun River case study illustrates the

TABLE 10.3 *Factors governing channel morphology according to the potential of forest management activities to influence channel conditions (in descending order of influence)*

Factor	
1. Sediment supply - source (size; timing)	a. Landslides (coarse sediment and LWD; episodic) b. Roads – function of road use/maintenance (fine sediment; chronic) c. Gullies (mixed sediment and LWD; episodic and chronic) d. Fans (fine sediment; episodic)
2. Riparian	a. Streamside vegetation removal b. Streambank disturbance
3. Land use	a. Road drainage b. Road crossings (culverts and bridges) c. Road protection (e.g., riprap) d. Timber harvesting
4. Streamflow	a. Peak flows b. Low flows c. Direct channel capture
5. Climate	a. Streamflows b. Indirect land use effects

importance of watershed type on sediment production and delivery to a larger mainstem channel and the spatial response of the larger channel. It also highlights the importance of historical management strategies and practices, a factor that is often misidentified or simply overlooked when considering forest management practices and stream channel integrity.

Government Creek Government Creek is an intermediate-sized (17 km^2), old-growth, coastal watershed on the northern tip of Moresby Island in the Haida Gwaii archipelago (seen from aloft in Figure 10.14a). The watershed has steep hillslopes in the headwater zones and lower-gradient terrain closer to the stream's mouth (type I watershed). The channel network has a single gravel-bed stream at its outlet; a series of smaller tributaries join the mainstem upstream and away from the stream mouth. The main sediment source for Government Creek is landslides that deliver sediment directly to the channel network. Streamflow, channel width and depth, and sediment storage all increase as the watershed area becomes larger, and channel gradient and bed sediment textures decrease.

Figure 10.14b shows the channel morphology near the stream's outlet. Both the channel and near-bank sediments are erodible (alluvial), and the morphology is characterized by riffle-pools and associated bars. Continuing upstream along the mainstem channel and entering the first main tributary (Figure 10.14c), the channel gradient and bed material size increases and channel width decreases along the tributary. The morphology changes from riffle-pool to cascade-pool (Figure 10.14d). In zones near the headwaters, the channel is characterized by step-pool morphology (Figure 10.14e).

The Government Creek watershed includes all three of the commonly found stream morphological types. This is typical and expected if the entire length of channel (from the stream mouth to the basin's headwater) is considered, although the stream order along the channel length may vary according to watershed conditions.

Channel structure and large woody debris To consider the role of LWD in controlling channel structure and to provide background from which to consider forestry activities later in this compendium, this Haida Gwaii example compares an unlogged, old-growth forest stream with a stream in a logged watershed. Figure 10.15 illustrates the morphological characteristics of a typical steep-coupled watershed (Government Creek) in an old-growth forest in the Coastal Western Hemlock (CWH) biogeoclimatic zone. The channel is diverse, with complex longitudinal and planimetric forms (Figure 10.15a). The longitudinal profile has riffles and distinct, well-defined pools, which account for 65% of the overall channel area. The channel width is variable, alternating between narrow sections with stable banks and wide sections where the channel becomes locally unstable. The channel banks are commonly undercut and bars consist of cobble, gravel, and sand-textured sediment. The complexity and diversity of the channel is attributed to LWD characteristics (Figure 10.15a). The wood is large compared to channel width and is predominantly diagonally oriented across the channel, directly influencing water flow and sediment movement and storage (Figure 10.15b).

Another Haida Gwaii example (Mosquito Creek) illustrates the influence of riparian logging practices on channels not affected by landslides (see example in Chapter 9, "Forest Management Effects on Hillslope Processes"). Besides logging—57% of the watershed was logged during the 1940s and 1960s by skidder and high-lead methods without riparian leave strips—Mosquito Creek (Figure 10.16) is similar in most biogeophysical aspects to Government Creek. The two creeks are similar in watershed area, drainage density and shape, average channel and hillslope gradient (including uncoupled sections), geology, forest type (CWH), and climate. The post-harvest channel of Mosquito Creek is relatively morphologically simple, with minimal variability in longitudinal, planimetric, and sedimentologic characteristics. Its longitudinal profile shows long pools with relatively uniform depths. Although riffles and glides are more prevalent in the logged Mosquito Creek channel (Figure 10.16a) compared to the forested Government Creek channel (Figure 10.15a), their shapes are commonly long and shallow. Channel width is not only wider than expected for the drainage size and hydrological conditions, it is constantly wide with minimal variability. Channel banks are rarely undercut and most channel bars consist of uniformly textured gravels.

The LWD characteristics associated with each channel explain the underlying differences in the channel morphologies of the logged and forested watershed streams. The most important difference is a shift in LWD orientation—significantly more

FIGURE 10.14 *A typical coastal basin, Government Creek, Haida Gwaii: (a) view looking upstream to watershed at the mouth of Government Creek; (b) view looking downstream at the channel bed near the stream outlet (note riffle-pool morphology); (c) view looking upstream at the confluence of two tributaries; (d) view looking upstream at one of the tributary streams (note cascade-pool morphology); and (e) view looking upstream near headwaters (note step-pool morphology). (Photos: D. Hogan)*

LWD is oriented parallel to the channel banks of the logged channel, whereas the forested stream shows a predominantly diagonal arrangement of LWD. The shift in orientation reduces the interaction among LWD, streamflow, and sediment transport, and therefore the same amount of woody material has less influence on scouring and trapping of sediment in the logged stream (Hogan 1986). Although the total volume of LWD is similar in each stream, the size distribution also shows a shift, with more

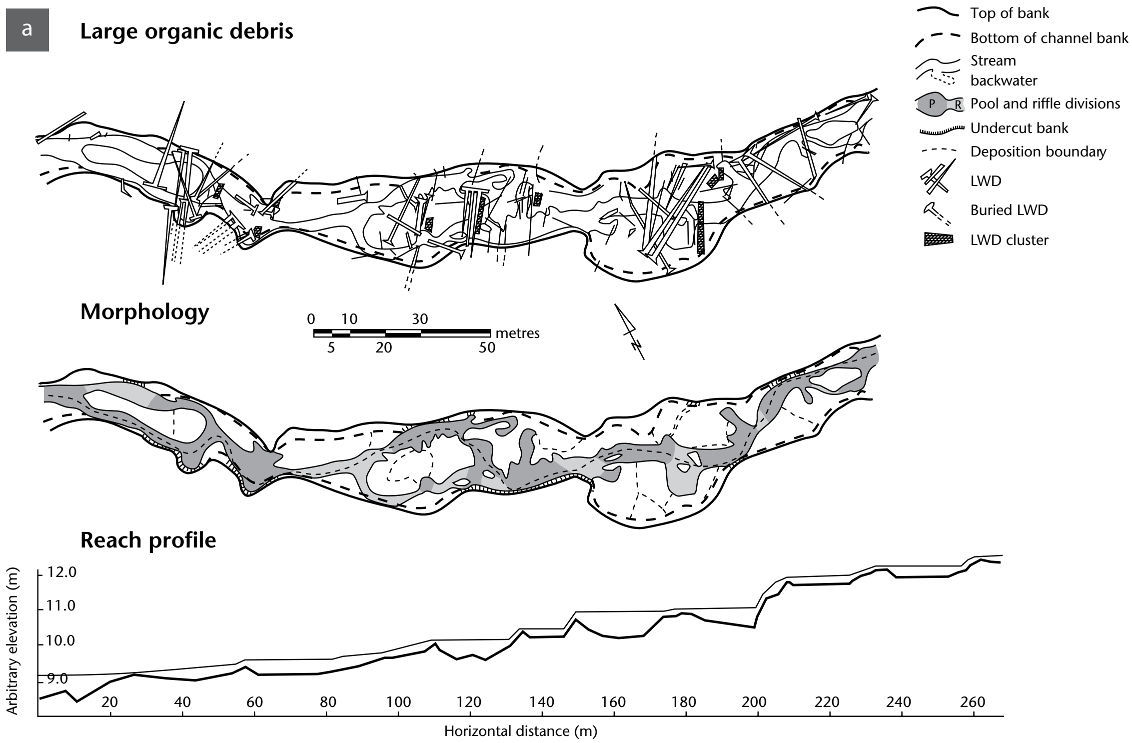

FIGURE 10.15 *Morphological characteristics of an old-growth coastal stream, Government Creek, Haida Gwaii: (a) large woody debris location map, planimetric map, and longitudinal profile (Hogan 1986); and (b) photograph from site. (Photo: D. Hogan)*

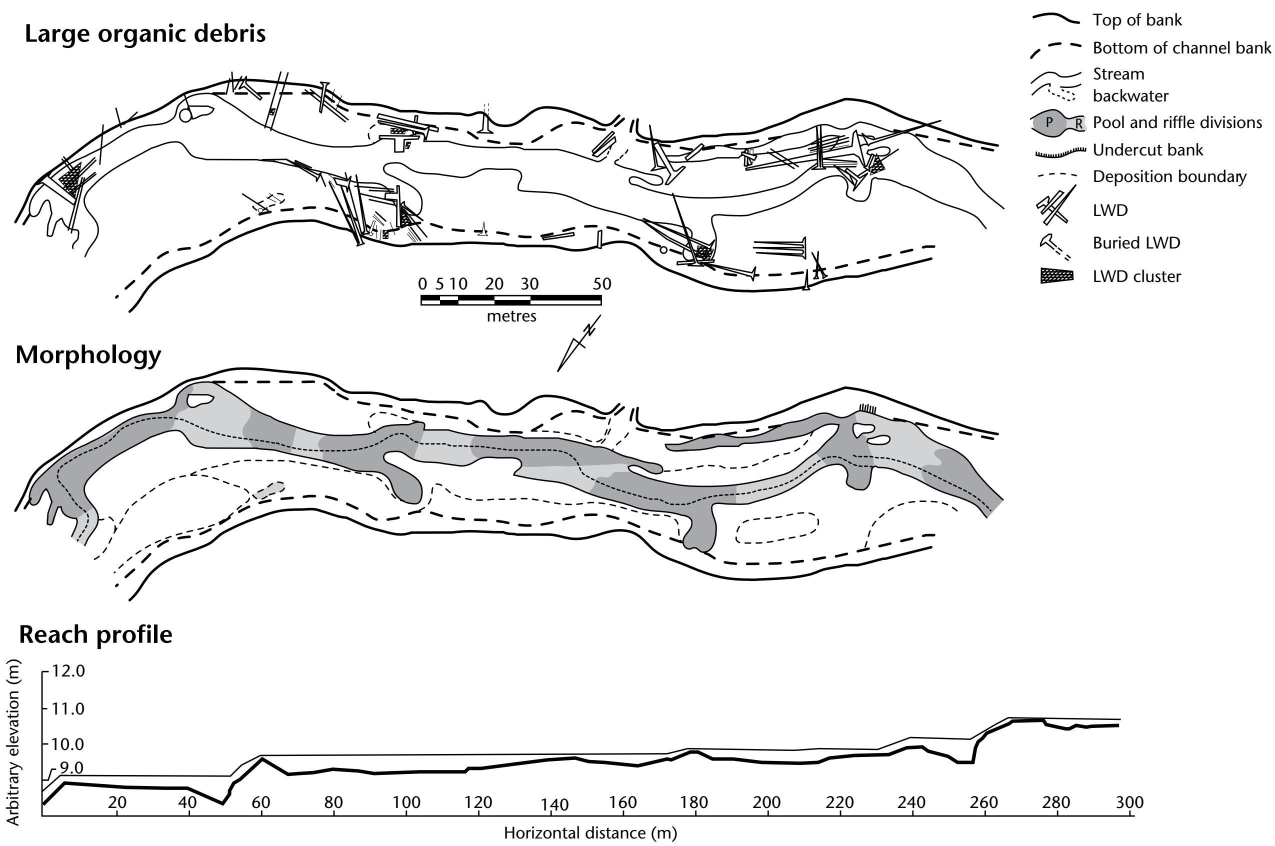

FIGURE 10.16 *Morphological characteristics of a logged coastal stream, Mosquito Creek, Haida Gwaii; large woody debris location map, planimetric map, and longitudinal profile (adapted from Hogan 1986).*

small material evident in the logged stream. The smaller absolute size of the pieces, and the relative size reduction due to the wider channel, makes this material more mobile at similar flow stages than that in the forested stream. The increased mobility leads to a reduction in overall channel stability.

The preceding example serves to illustrate the differences between two essentially identical streams (with the only exception that one was logged). The two streams flow through watersheds of similar type and both have steep headwaters with the downstream zones uncoupled from hillslope processes. The main difference between the channels is related to the removal of riparian vegetation and the direct physical disturbance of streambanks, and not landslides that may have occurred in distal, uncoupled areas of the watershed. Removal of riparian vegetation and direct disturbance of streambanks resulted in very different channel conditions in each stream. The logged watershed stream is relatively simple geomorphically, with long, shallow, and uniformly shaped pools and riffles. The forested watershed stream is geomorphically complex with diverse features. The differences are attributed to the loss of bank strength, which causes channel widening, which can dramatically increase sediment supply and lead to a decrease in channel complexity if transport capacity is exceeded. Additionally, the removal of the riparian vegetation as a source of LWD input to the stream reduces the channel's ability to store additional sediment and lengthens the channel's recovery time.

Channel structure and natural disturbance (large woody debris jams) Much of Haida Gwaii has steep, unstable terrain and a wet climate with high-intensity rainstorms. The most prevalent watershed type is steep and coupled to the channel. Landslides are common in this coastal setting, occurring both episodically (as infrequently occurring, large-magnitude events) and on a more frequent basis (as annually occurring, small-magnitude events) (see

Chapter 9, "Forest Management Effects on Hillslope Processes"). Furthermore, Schwab (1983, 1998) and many others have documented an increase in landslide occurrence as a result of certain forestry activities (see additional FFIP references). Hogan et al. (1998b) summarized the influence of landslides on channel conditions. For channels flowing through old-growth watersheds it was found that:

- landslides occur episodically (Figure 10.17), with the largest generally attributed to a combination of geological and meteorological factors;
- LWD jams form along the stream channel at or near where landslide materials are deposited into the channel;
- LWD jams and channel conditions evolve over time as the jams' influence on sediment transport and storage patterns change (Figure 10.18). Major disturbances in the channel (e.g., bed aggradation, multi-branched flows, channel widening with streambank erosion, infilled pools upstream, and severely scoured channels eliminating morphological features downstream of the newly formed jam) coincide with jam formation, but normal channel processes, such as rhythmic channel scour and fill, preferential flow paths (single rather than multi-branches), pool formation, bar development, and riparian bank revegetation may recover over timespans approaching 50 years.

This sequence produces a complex and diverse channel due to the mosaic of channel states, with an approximately equal frequency of recently disturbed (major alterations) to old, recovered, and essentially non-disturbed conditions developing over time, all within the same stream. However, along with the increase in landslide rates after logging, there is a corresponding increase in the number of recently formed in-channel LWD jams (Figure 10.19). This leads to a second peak of the bimodal logjam age distribution. This shift in LWD jam age distribution causes the channel in logged watersheds to have relatively greater channel lengths in a disturbed state with simplified, less complex channel morphologies than the unlogged streams. The reasons for acceler-

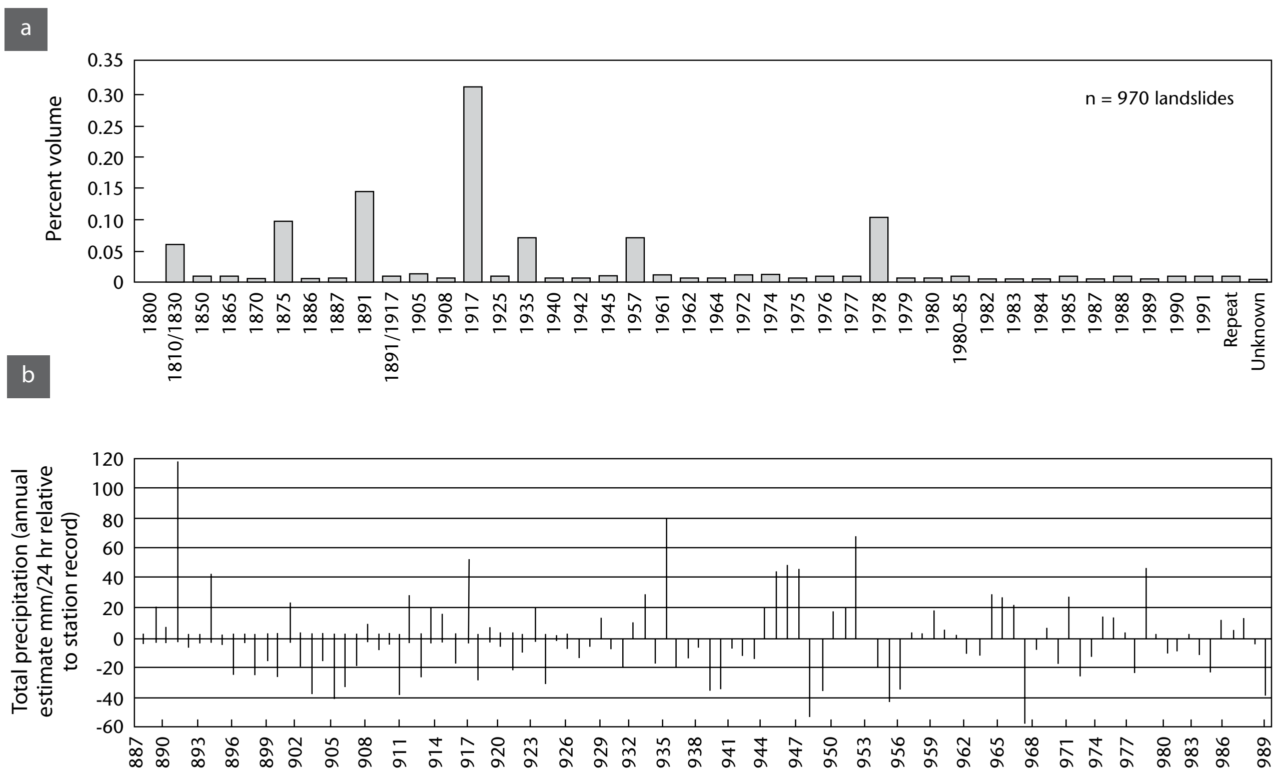

FIGURE 10.17 *Historical landslide and precipitation records: (a) landslide events occurring on Haida Gwaii (Queen Charlotte Islands) 1810–1991 (from Schwab 1998); (b) annual maximum 24-hr precipitation records for selected stations (aggregate record for: Port Simpson, 1887–1909; Masset, 1910–1914; Queen Charlotte City, 1915–1948; Sandspit, 1949–1962; Tasu, 1963–1972; Sewell Inlet, 1973–1989) (after Hogan et al. 1998b).*

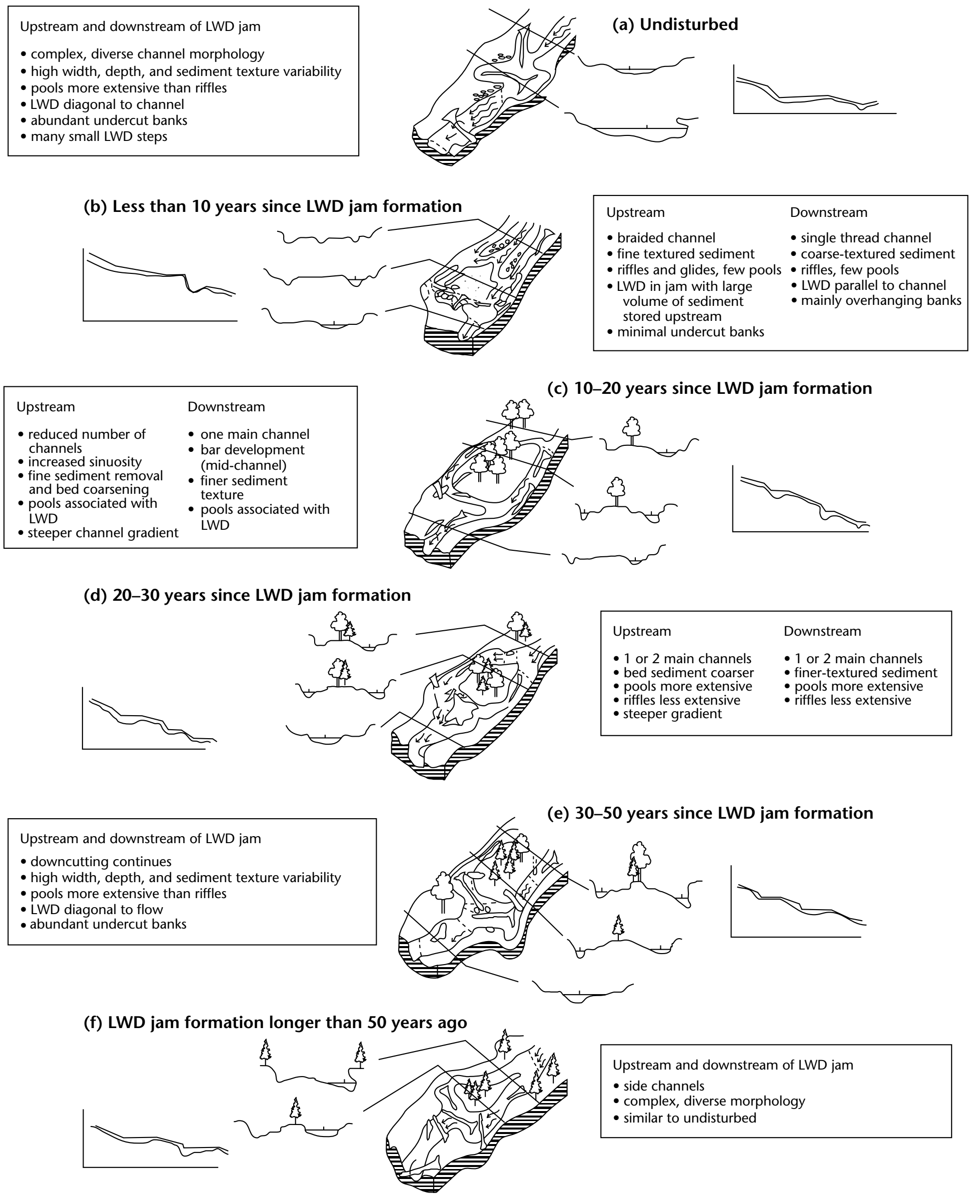

FIGURE 10.18 *Adjustment of channel morphology in response to large woody debris (LWD) jam formation and deterioration (modified from Hogan et al. 1998b).*

ated landslide rates in logged areas are discussed elsewhere (see Chapter 9, "Forest Management Effects on Hillslope Processes," and the additional FFIP references).

Although the connection between landslides, material input, LWD jam formation, and channel evolution has been documented mainly in coastal watersheds, evidence suggests that similar processes and conditions exist in the province's non-coastal, steep, coupled watersheds. The in-channel adjustments to LWD jam formation are similar across the province, with the following important differences:

- material input remains episodic, but may be from other mechanisms besides landslides, such as windthrow, particularly in areas affected by fire, insect infestations, and flood-induced bank erosion (large snowpack or ice melt);
- some areas produce smaller trees that are less able to form jam complexes, which are highly immovable and impermeable barriers to sediment transfer (Figure 10.14); and
- LWD jam deterioration processes (decomposition,

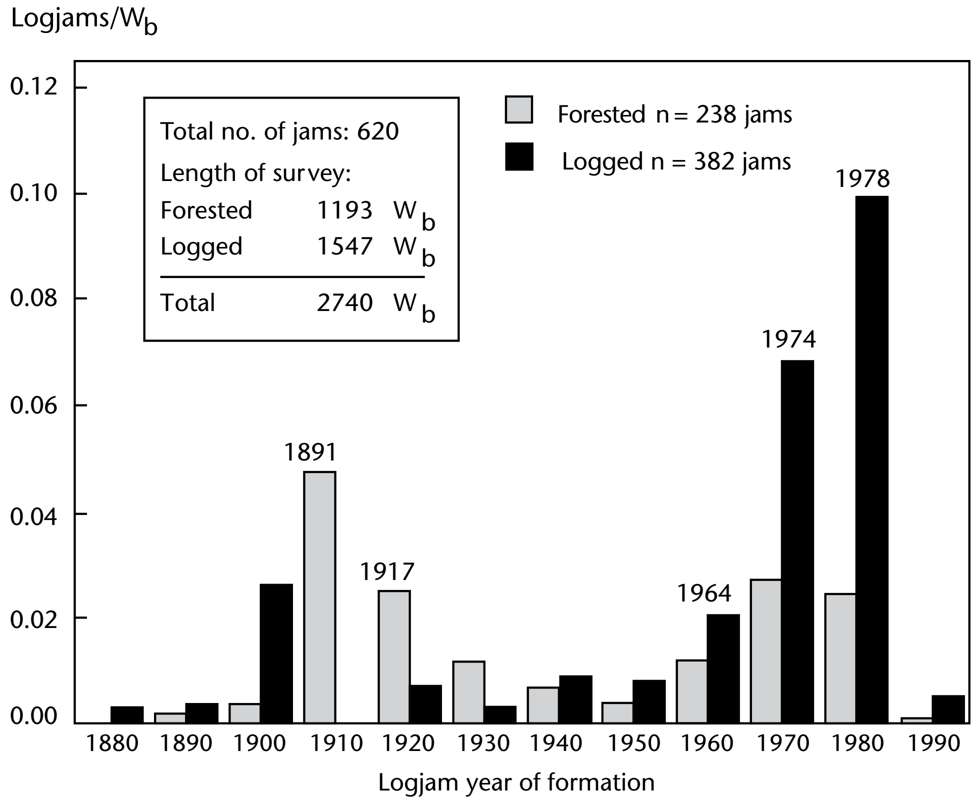

FIGURE 10.19 *Large woody debris jam age distributions for forested and logged watershed streams on Haida Gwaii (W_b = bankfull width; after Hogan et al. 1998a).*

physical resistance to abrasion) vary by tree species and climate (Harmon et al. 1986).

Natural disturbance and riparian development

Natural disturbances are important determinants of channel morphology, but these disturbances also influence riparian stand composition. Figure 10.20 illustrates the interaction between the stream channel and the riparian zone for an intermediate-sized coastal stream (Gregory Creek; from Bird 1993). The major storm events of 1891, 1917, and 1978 (see Figure 10.17) induced landslides that introduced large quantities of wood into the channel and created logjams. Subsequent flooding, following these events, forced streamflows around the logjams and into the floodplain, leading to channel avulsions into the riparian zone (Figure 10.20). With the jams no longer within the active channel, sediment wedges that had accumulated upstream were rapidly colonized by riparian species. The temporal dynamics (storms→landslides→logjams→avulsions) eventually led to a mosaic of diverse and complex riparian stand ages, ranging from 12 years to over 300 years, which is typical of old-growth forests (Bird 1993). The channel migrated across the valley flat, cutting channels and building bars, islands, and the floodplain; this sequence of erosion–deposition cycles over centuries produces the old-growth riparian mosaic. The older, abandoned channels can convey waters and provide refugia for aquatic biota during flood events.

Bird's (1993) work highlights important operational implications. It shows a link between hillslopes and riparian zones that is as important as the link between stream channels and riparian zones. Traditionally, riparian zone management (Chapter 15, "Riparian Management and Effects on Function") has been based on the maintenance of existing streamside vegetation to protect riparian zones as sources of LWD to streams (e.g., buffer strips); however, riparian zones are also controlled by, and depend on, hillslope processes. Landslides initiate the formation of instream logjams that then create a local base level in the channel, which disrupts sediment transport and initiates the aggradation of a sediment wedge. Opportunistic, pioneering riparian plant species colonize the infrequently flooded por-

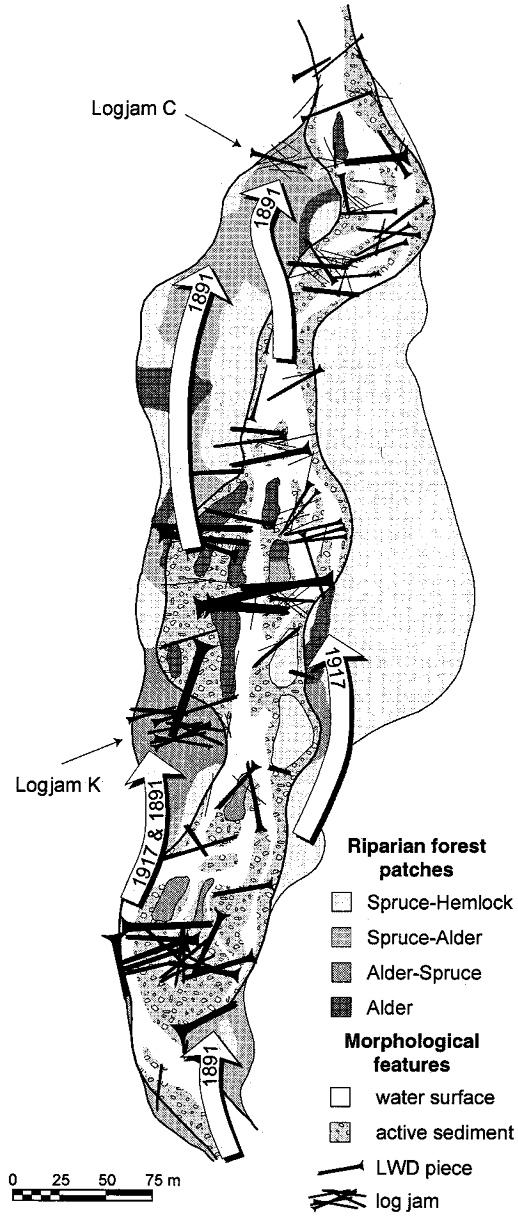

FIGURE 10.20 *Pathways of fluvial disturbance in a riparian area (after Bird 1993). The arrows identify two events occurring in 1891 and 1917 (indicated by Spruce-Alder and the Alder-Spruce patches, respectively), when the channel avulsed into the riparian area. Logjams C and K formed during these events and are now abandoned by the channel. A third event in 1978, indicated by Alder patches, was responsible for the creation of several islands. The riparian area occupied by Spruce-Hemlock patches has been undisturbed for at least 300 years (after Hogan et al. 1998b).*

tions of the sediment wedge, increasing the stability of the stored sediment. Consequently, the formation of a sediment wedge influences the distribution of riparian vegetation that, in turn, affects the stability of the stream channel. Large floods force water and sediment loads around logjams and into the riparian zone, creating patches of riparian vegetation. Historically, only the influence of the riparian zone on the channel has been of concern from a management perspective; however, Bird's (1993) work indicates that the channel can influence the composition of the riparian zone, and therefore riparian zone management should consider hillslope processes as well as stream channels.

Riparian zone management has traditionally considered the physical and biological attributes of channels that are not expected to move laterally (i.e., lateral movement is not considered); however, as Bird (1993) has shown, laterally unconfined channels are capable of moving across entire floodplains within a single forest stand rotation. The ability of channels to move laterally depends on both the erodibility of valley-flat material and the relative width of the valley to the channel. Designated riparian management zones that do not account for potential channel movement can be rendered ineffective by subsequent channel migration. Prudent riparian zone management should consider the potential for lateral channel movement when delineating riparian buffers.

Carnation Creek Carnation Creek is a relatively small watershed, draining 11.2 km^2 on the west coast of Vancouver Island. It has a warm, wet climate; annual precipitation ranges from 2100 to 4800 mm, with 95% falling as rain between November and April. Stream discharges range from 0.025 m^3/s in the summer to over 60 m^3/s during winter freshets. The watershed consists of two steep-coupled basins that are linked longitudinally. This creates an upper basin, with steep headwaters and a flat, bowl-shaped section downstream, that leads into a steep-sided canyon and is then connected to a lower basin, with a broad, valley flat, which extends for approximately 1 km before the stream enters the ocean (Figure 10.21).

Two phases of logging were undertaken. The first phase occurred from 1976 to 1981, and involved riparian logging; the second phase occurred from 1987 to 1994 in the creek's headwaters. Three riparian harvesting treatments were applied at Carnation

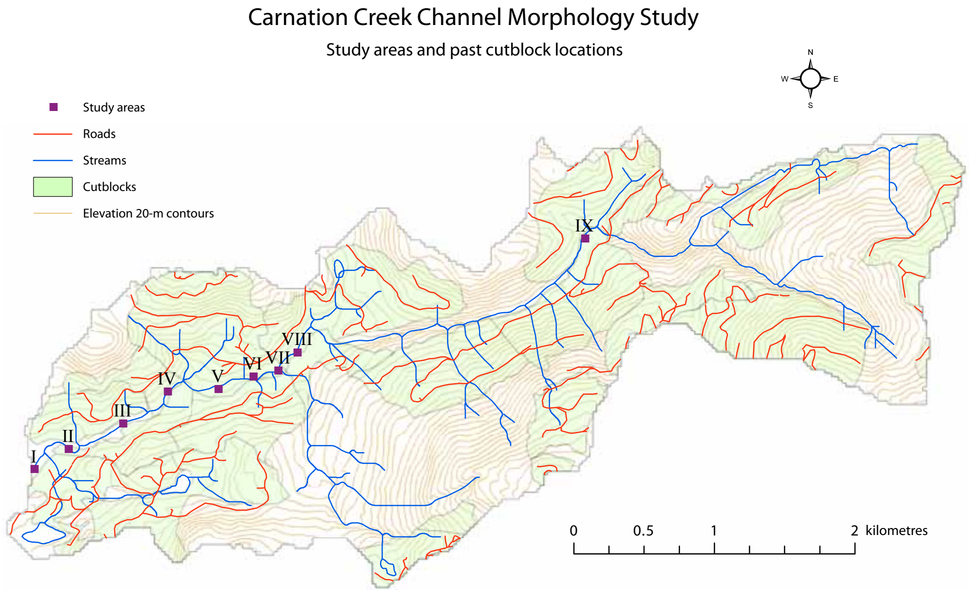

FIGURE 10.21 *Carnation Creek watershed, showing study areas and cutblock locations (A. Zimmermann).*

Creek, but only one "careful" riparian treatment (logged to the streambank, but with no direct in-stream activities) is discussed here.

Figure 10.22a illustrates the pre-logging (1977) morphology of the study area. At this time, the channel exhibited a complex morphology as a result of individual LWD pieces and an intact riparian zone. Figure 10.23 shows the bankfull widths of four study area cross-sections.

This confirms that little change in cross-sectional width occurred between 1971 and 1980. During 1978 and 1979, the study area was logged, adding more in-channel LWD from slash left along the channel banks upstream and within the study area. A key piece spanning the channel (see Figure 10.22a) trapped this additional material and resulted in the formation of a logjam. A majority of the key pieces in the channel were a legacy from the old-growth stand. It is likely that as these pieces decay, the stability of the jams in Carnation Creek will be reduced. As the logjam grew in size and began to impede downstream sediment transport, the channel widened (Figure 10.22b); additional reduction in bank strength caused by riparian tree removal was also a factor. Most of the cross-sections upstream of the logjam (3, 6, and 9) experienced widening; cross-sections 12 and 18 were within and downstream of the jam, respectively (Figure 10.23).

In January 1984, a large storm event resulted in multiple gully failures that delivered sediment and wood into a steep, confined canyon segment of the creek. This material was transferred and deposited downstream of the canyon into the study area, where further downstream transport was prevented by the logjam (Figure 10.22c). This resulted in significant aggradation upstream of the jam and started the cycle of jam–channel interaction. All of the cross-sections experienced significant channel widening due to a sediment wedge, which was deposited upstream of the jam (Figure 10.23). After 10 years, another peak flow event broke through the jam and allowed the sediment to be evacuated, returning the width in this area to almost pre-disturbance values (Figure 10.23).

The Carnation Creek example highlights the importance of the connectivity of hillslopes to channels, and the transfer of disturbances to downstream reaches. Riparian logging added to the scale of the disturbance by contributing additional LWD to the reach and reducing bank strength. The full temporal extent of the logging impacts in Carnation Creek are uncertain, since a number of legacy logs still span the channel and act as key pieces for logjams; however, as these pieces decay, the extent and durability of current logjams will be compromised and the extent and durability of future jams will be limited.

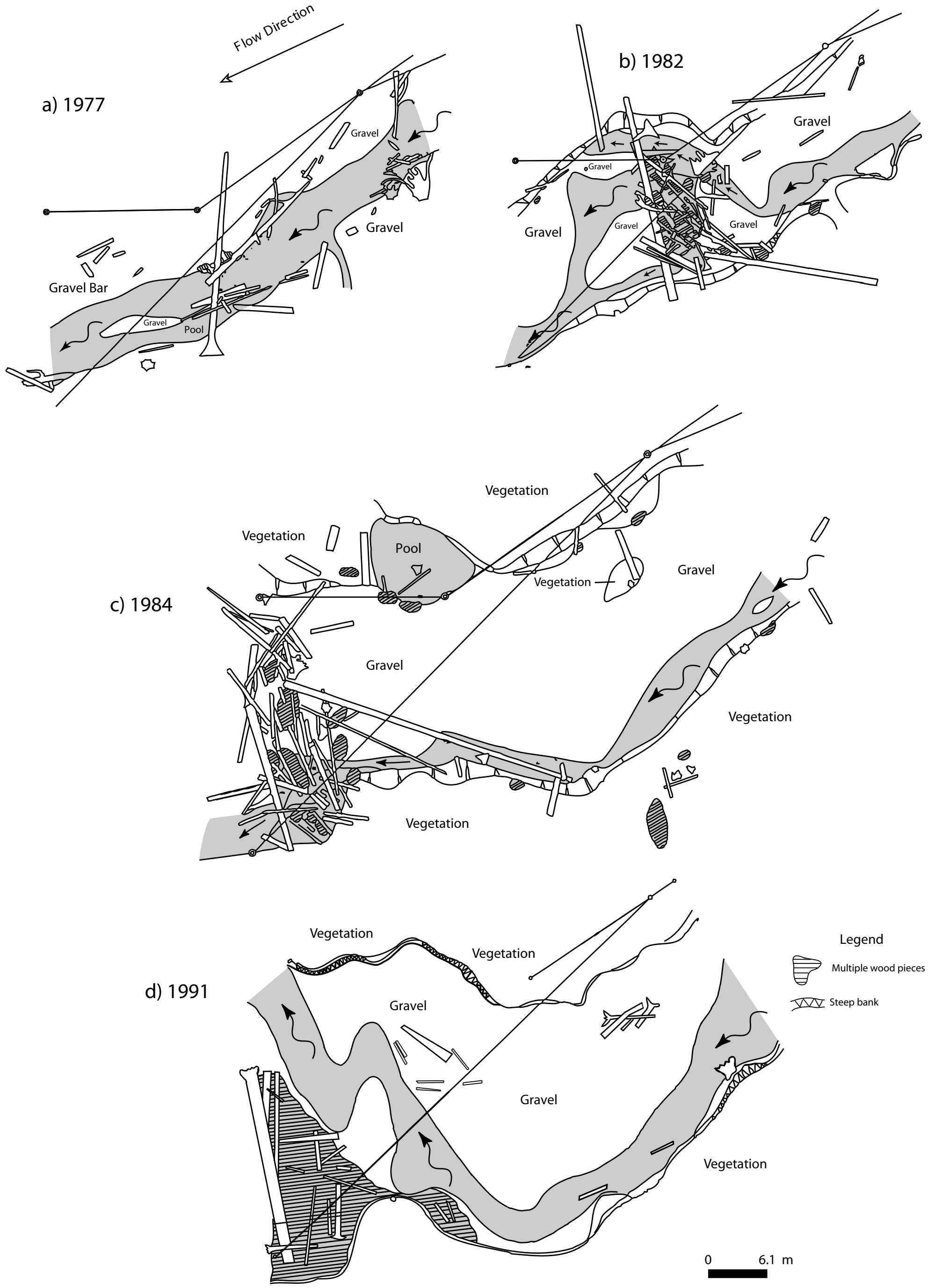

FIGURE 10.22 *Carnation Creek morphometric maps of study area during summer low flows 1977–1991. Shaded areas indicate wetted surface at time of survey (Luzi 2000).*

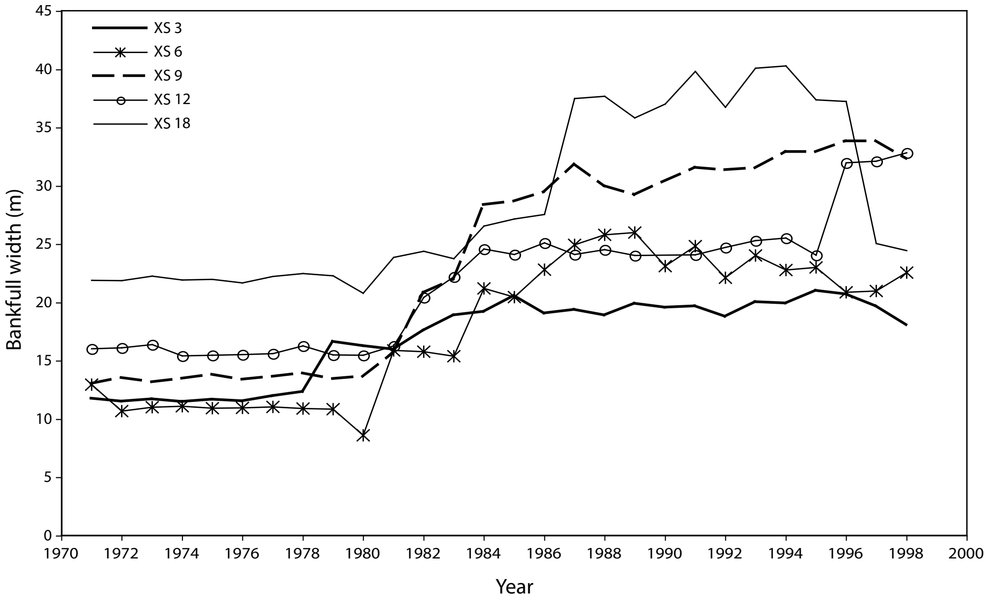

FIGURE 10.23 *Bankfull widths of selected cross-sections from Carnation Creek study area; logging in study areas occurred during 1978–1979 (Luzi 2000).*

Donna Creek Donna Creek, a relatively large tributary of the Manson River, is located in the Omineca Mountains 75 km northwest of Mackenzie. Drainage basin area is 126 km^2 and channel width near the outlet approaches 30 m. Fisheries values in Manson River and its tributaries are significant and, in particular, provide habitat for kokanee salmon. The upper reaches of Donna Creek are generally coupled to the hillslopes with only a narrow valley bottom to buffer sediment transfers from hillslopes to the channel. Road-building activities along the upper slopes of the watershed altered the natural hillslope runoff pattern by water capture and routing along road ditches, which resulted in a mass wasting event. A full analysis of the upslope processes, forest practices, and resulting mass wasting event is included in Chapter 9 ("Forest Management Effects on Hillslope Processes") and should be consulted when considering the channel impacts discussed here.

During the spring snowmelt of 1992, the altered runoff patterns resulted in the delivery of an excessive amount of water (approximately seven times greater than natural conditions) to a glaciofluvial/glaciolacustrine terrace. This initiated a series of debris slides and flows that transferred 420 000 m^3 of sediment to Donna Creek (see Schwab 2001 for details). Sediment delivered to the channel was either stored along the channel margin or transferred to the Manson River by debris flood (about 7 km downstream from the erosion scar).

Upstream of the erosion scar (stream input location), the banks were stable and undercut, and the channel bars were primarily of the diagonal and point bar type; these occupied less than half of the channel width at low flow (Figure 10.24a). Woody debris was abundant, with approximately 12–14 m^3 of debris in any 100-m length of channel, and was important in controlling sediment storage and movement. Woody debris steps and jams were critical structural channel elements. Pool-riffle sequences were typical of small channels with abundant woody debris; the average pool-riffle spacing was 4.1–4.6 bankfull widths.

Immediately below the erosion scar, most of the in-channel debris was either parallel to the channel or elevated above the bed. The channel experienced extensive widening, and deep, fine sediment deposits extended 80 m across the valley flat (Figure 10.24b). Sediment depth was usually 1 m or less but depths exceeded 2 m in some areas. Riffles and riffle-glides were the most common morphological features, and pools were infrequent.

The physical features of Donna Creek were radically altered as a result of the sediment and debris introduced to the stream following the hillslope failure (Table 10.4). Impacts included severe channel

FIGURE 10.24 *The impact of a large landslide on Donna Creek, an intermediate-sized interior stream: (a) view of Donna Creek upstream of 1992 landslide (or washout flow, see Chapter 9) entry point; and (b) view of Donna Creek at 1992 landslide (or washout flow) entry point. (Photos: D. Hogan)*

TABLE 10.4 *Morphological channel conditions observed in the field during 1992 and 2007 (Donna Creek). Note that mean pool length and pool-riffle spacing are given in bankfull width (W_b) units. Subdominant bar types are shown in parentheses (Schwab et al., unpublished data).*

Reach number	Bankfull width (m)		Mean pool length (W_b)		Pool-riffle spacing (m)		Bar type[a]		D_{95} (mm)		Woody debris (m^3/100 m)	
	1992	2007	1992	2007	1992	2007	1992	2007	1992	2007	1992	2007
1	8	10	2.2	1.7	4.6	4.1	sc	sc (mc)	196	126	14	12
2	17	19	3.6	-	11.8	-	mc	sc	163	45	71	120
3	19	29	0.88	0.53	3.0	5.0	mc (sc)	sc	203	216	15	64
4	12	19	0.39	0.76	7.3	4.2	sc (mc)	sc (mc)	169	179	30	38
5	11	25	-	2.1	-	3.0	sc	sc	212	157	5.7	4.9
8	20	18	1.5	1.2	3.1	3.0	pb (sc)	sc (pb)	201	164	2.2	8.1

a sc = side channel bar; mc = mid-channel bar; pb = point bar

erosion and extensive sedimentation in upstream, near-source areas, and massive sedimentation in downstream channel reaches. These impacts have persisted in the channel for over 10 years (Figure 10.25); the channel has experienced alternating cycles of net sediment scour and deposition as sediment is reworked from temporary storage and transported downstream. In addition, woody debris that was re-mobilized from previous accumulations stored in the channel and (or) along the channel margin has been reorganized and augmented with wood recruited from the riparian area to form several new logjams.

The Donna Creek landslide and subsequent channel response is an extreme example of environmental damage that can result from poor forestry practices (specifically, the design and maintenance of forest roads). The example illustrates the importance of localized alterations to natural hydrological processes and the potential to trigger massive geomorphic change to a river channel (both temporally and spatially) by overwhelming natural patterns of sediment delivery to stream systems.

Fubar Creek The branching network of stream channels in a watershed ensures that more small streams are encountered as timber harvesting activities expand from valley-bottom areas into headwater areas. Concerns about the effects of forest management activities near small streams became a serious issue in the Prince Rupert Forest Region (now the Northern Interior Forest Region) in the early to mid-1980s. In an attempt to address these concerns, regional researchers established a program to document the disturbance and recovery to pre-logging conditions in a series of small streams, each with a specific logging history. One stream, Fubar Creek,

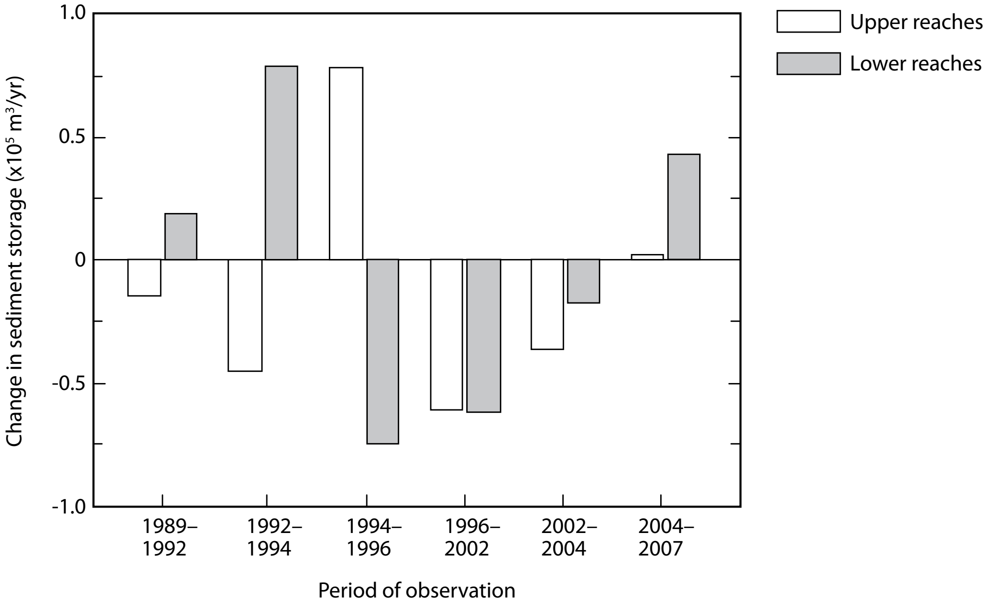

FIGURE 10.25 *Erosion and deposition patterns in Donna Creek (Schwab et al., unpublished data).*

was selected to evaluate the effect of clearcut logging an entire cutblock, including the complete removal of all riparian vegetation on both banks of the stream. An upstream section of channel (Fubar Upper), biophysically similar to the downstream logged section (Fubar Lower), remained unlogged and constituted a control. No upslope issues (e.g., landslides) existed to confound the effect of riparian logging.

Fubar Creek is a small drainage located in the upper Zymoetz River watershed, approximately 20 km west of Smithers. Its riffle-pool channel is generally less than 3 m in bankfull width. The reach was logged to the streambanks in the mid-1980s using conventional ground-based skidding techniques with equipment operating over and across the channel. These operations directly altered in-stream channel features.

During logging operations, the streambanks were disturbed and sediment was transferred to the channel (Figure 10.26). If channel conditions in the upstream reach were representative of those in the downstream reach before logging, then channel width doubled and sediment storage increased in the channel (note the growth of mid-channel bars). Channel recovery has been relatively slow, with evidence of channel disturbance persisting to 2005 (e.g., sediment storage remained relatively high compared to Fubar Upper).

The channel bed in Fubar Lower underwent extensive erosion since the initial disturbance, and the banks built into the active channel, resulting in a decrease in channel width (by about 1 m between 1992 and 2000). Generally, this indicates that sediment mobilized in the channel during and shortly after logging has been transferred downstream and (or) stored overbank as the channel recovers from the initial disturbance. The recovery process may have been prolonged, however, because logging to the streambanks has affected the supply of woody debris to the channel. Although more pieces were evident in the channel, these were generally smaller in length and radius and likely less functional as impediments to sediment transport. For example, between 1992 and 2000, the number of pieces increased from 41 to 51 and the mean length and diameter both decreased from 1.42 to 1.12 m and from 0.13 to 0.10 m, respectively. No statistical tests of difference were conducted on these data (this operational work does not include a complete sample of the stream so the data are from a single point), but the patterns are similar to those found in other streams (see above and additional FFIP references at end of this chapter). Because the treatments practised at this site were considered normal for the place and time, the channel differences presented here are attributed to overall logging activity; no attempt has been made to attribute channel differences to the effects of individual activities such as roads, stream crossings, yarding methods, and site degradation.

This example illustrates the importance of intact

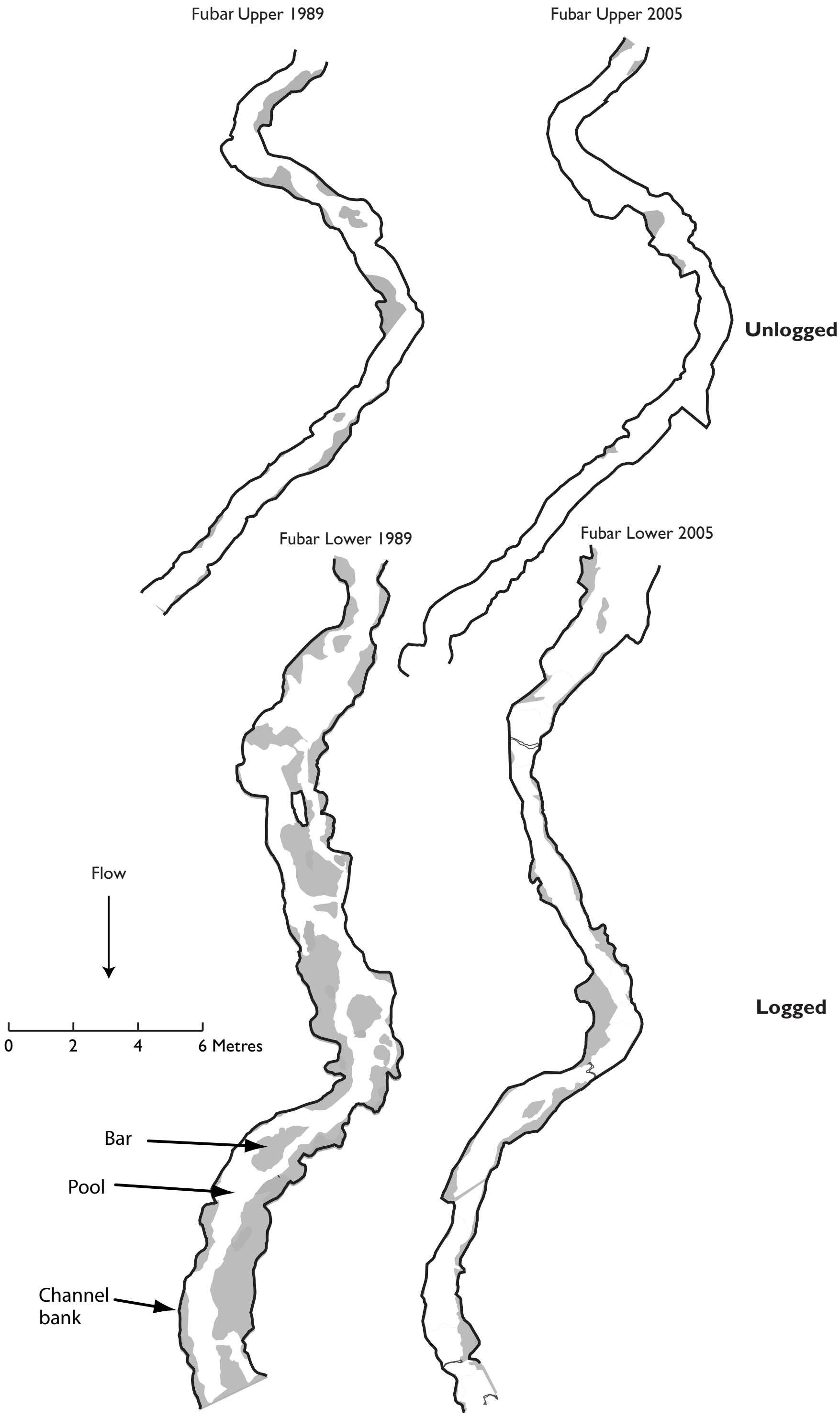

FIGURE 10.26 *Planimetric maps for unlogged (Fubar Upper) and logged (Fubar Lower) channels in Fubar Creek 1989 and 2005 (after Bird et al. 2010).*

streamside vegetation. Activities associated with the removal of riparian vegetation resulted in an immediate and dramatic change in channel conditions in this small interior stream. The channel disturbances have not been reversed in over a decade. Recent regulations require the maintenance of channel condition if harvesting is planned in the riparian management area of streams (for current British Columbia practices, see Chapter 15, "Riparian Management and Effects on Function").

Yakoun River The hydrological implications of extensive logging began to receive critical consideration in the 1960s and serious concerns were raised in coastal British Columbia in the 1970s. One approach to addressing the hydrological issues was to undertake watershed assessments. Several watershed assessment procedures were developed, each with its own individual strengths and weaknesses. An advantage of these procedures was that hydrological processes operating at a watershed level, as opposed to site-specific or plot level, were central to each (Wilford 1987; B.C. Ministry of Forests and B.C. Ministry of Environment, Lands and Parks 1995). A common goal of forest management plans was to avoid changes to stream channel morphology, thereby ensuring the protection of the physical habitat of fish and other aquatic organisms. An integral component of watershed assessments was the need to balance different aspects that may lead to channel change and thus direct the focus toward the hydrologic and geomorphic factors operating in a watershed. The Yakoun River case study provides an example of this balancing act.

The Yakoun River has a large watershed that drains 477 km^2 at its mouth near Port Clements on Haida Gwaii. This case study addresses several valuable aspects of stream channel morphology assessments; it shows the importance of partitioning the watershed into multiple internal sub-basins, the relative weight of hydrologic and geomorphic factors, and the consequence of disregarding historical management strategies and practices when evaluating channel conditions. This last aspect is often neglected because it is easy to assume that if forest management started in remote areas in the 1940s and 1950s, then human-influenced events had surely not occurred before that time. The Yakoun River area of Haida Gwaii was isolated until the 1960s and no significant industrial-scale logging had taken place. A detailed discussion of this case, which illustrates the importance of historical anthropogenic events in a coastal setting, is included in Hogan (1992). Similar examples could be cited for the interior far north, the Okanagan, and the Kootenays.

A common concern when developing forest plans centres around the amount or rate of logging within the watershed. Hydrological considerations could be assessed in the Yakoun River system because streamflow gauging records were available from the Canadian government (WSC Station 08Q002). The station, located close to the river's mouth, has daily streamflow records dating back to 1962. An evaluation of these discharge records in conjunction with local climate data showed that the main change in the hydrologic response of the Yakoun River during the period of accelerated timber harvest (1962–1989) was a decrease in the ratio of runoff to rainfall; there was also a modest increase in local precipitation and a slight decrease in annual discharge over the period.[2] This is not the expected response based on observations in most other research watersheds (Chapter 7, "The Effects of Forest Disturbance on Hydrologic Processes and Watershed Response"), and the decrease in runoff was attributed to greater evapotranspiration by the regenerating alder than the previous old-growth conifer forest.[3] Hydrograph analysis further supported the conclusion that the basic hydrologic response characteristics of the system appeared unchanged.

A river system's sediment supply is also a central issue when considering forestry plans. The Yakoun River watershed has steep headwaters, lower-gradient mid-slope sections, and a major, wide and long valley flat that includes a large lake. It has steep, connected tributaries and a flat-uncoupled lower mainstem. The overall watershed was portioned into 19 sub-basins to distinguish these different types from the downstream mainstem section. A series of historical aerial photographs helped identify channel changes along the mainstem. These included photographs from 1937 (before logging), 1961 (during an early logging phase), and 1988 (late and post-logging); additional photographs from 1954 and 1979 were added for selected sections. Photogrammetric analyses of 23 homogeneous reaches included map-

2 Hardy BBT Limited. 1991. Analysis of changes in the discharge of Yakoun River. Consulting Engineering and Environmental Services Report (G.E. Barrett, VG-05673), March 20, 1991. Unpubl.
3 Ibid.

ping of channel position and the type and location of bars, islands, and logjams.

Evaluation of the early (1937) aerial photographs showed evidence of extensive stream channel disturbance. This included reaches with indications of channel widening, reduced width variability, unstable bar configuration with many mid-channel bars (longitudinal, crescentric, and medial), and island development. Because no logging of any importance had occurred, the cause of this disturbance was questioned. A review of early government documents found a logjam inventory for the Yakoun River compiled in 1904 by Ells (1906; Table 10.5). This Geological Survey of Canada inventory was compiled to find a route up the Yakoun River from the ocean for a self-propelled drilling platform (18-m barge) to engage in mineral exploration at tributary confluences. The jams were characterized as prevalent, huge, old, and solid, and represented a serious constraint to river navigation. Starting in 1912, logjams were systematically removed from the lower Yakoun River to overcome this impediment (Dalzell 1968). A second river-clearing program commenced in 1913–1914 when all jams, including three that reportedly extended for 3 km each, were demolished by dynamite, enabling access to Wilson Creek (in reach 15 of the recent study). Further clearing occurred in 1958–1959, when roughly 5000 m^3 of logs and trees were yarded from the channel, largely to protect a bridge. The 31 jams listed by Ells compare with 51, 69, and 55 jams evident in the 1937, 1961, and 1988 air photographs, respectively.

Earlier in this chapter, we discussed the importance of logjam influence on streamflow and sediment transport, particularly in the early stages of jam development. Figure 10.17 showed that the largest storms on record (last 200 years) occurred in 1875, 1891, 1917, 1935 and 1978. These storms undoubtedly caused extensive bank erosion, and introduced large volumes of wood from riparian zones into the stream system. Note that both the large lake and extensive valley flats of the Yakoun effectively decouple the hillslopes, reducing the direct input from landslides, although the headwater tributaries were most likely severely affected by landslides during the storms. Removal of the logjams seriously affected stream channel morphology and stability by changing in-channel sediment storage patterns and reach-scale flow properties. These changes, in turn, resulted in altered bar patterns with more frequent mid-channel bars and reduced bank integrity and wider channels.

From the available information, the following conclusions were drawn (Hogan 1992).

- The mainstem channel had undergone significant changes over the last 50 years, including changes in channel width, bar structure, and logjam function. These changes were not evident in all areas; some reaches had undergone modifications and others remained relatively unaltered.
- Overall, the channel experienced the largest changes during the time when only a small proportion of the Yakoun watershed was logged. It seems unlikely that the amount of watershed logged had a locally significant impact on channel changes. More probably, site-specific impacts have been more important—the channel was probably stressed far more by historical removal of in-stream LWD through mining-related activities (1912–1914) than by sediment derived through spatially diffuse sources as a result of contemporary logging activities.
- Several factors, besides logging activity, have been identified as playing important contributing roles in the documented river behaviour. These include the direct impact of early mining activities on channel form and processes, and natural sequences of large, low-frequency storms.
- Localized instances appear to contribute to channel erosion (e.g., poor road layout and disturbed streambanks).

The Yakoun River case study highlights the potential range of factors to be considered when developing forest plans for an area, including watershed and sub-basin characteristics, location of development within the watershed, natural disturbance regimes,

TABLE 10.5 *Summary of logjam characteristics on the Yakoun River (after Ells 1906)*

Distance upstream of mouth (mi)	Logjam inventory
2 – 4	10 jams, logs with diameters > 1 yard
4 – 7	Few jams
7 – 10	10 jams, log diameters up to 60 inches
10 – 16	Rare jams
16 – 19	10 jams, very poor navigational potential
19 – 23	Except 1 large jam, clear upstream to lake

and historic anthropogenic disturbances. These examples show that each watershed responds differently to sediment inputs, whether from rapid (acute, episodic) events, or gradual and continuous (chronic) processes. The time required for fluvial systems to adjust to excess sediment following disturbances can range from decades in this and other studies (Madej and Ozaki 1996), to centuries (Knighton 1989). As Schumm (1977) emphasized, streams are physical systems with a history.

SUMMARY

This chapter emphasizes the overarching principles that control stream channel geomorphology, in light of resource management practices and planning. We have not attempted to provide a complete and comprehensive discussion of fluvial geomorphology or to prescribe best management practices for forest operations. We have addressed the fluvial geomorphology problems of British Columbia's forest lands through case studies from various geographic areas of the province, with the intention of dealing with applied problems encountered during management activities over the last two decades. We hope this offers a foundation upon which decisions can be made to minimize channel impacts.

We have presented the watershed as the fundamental landscape unit because its composition determines many factors that govern channel conditions and how watersheds respond to active management. Watershed shape is a function of geological processes including both independent and dependent factors. These, in turn, strongly influence the nature of the channel's sediment supply. The reworking of this sediment along the various sections of a watershed produces different channel types. In covering these topics, we have detailed the importance of the watershed as it influences the temporal and spatial nature of sediment and other materials entering the channel, and placed the watershed in relative context compared to other processes.

Our case studies illustrate the extent to which land use management affects a watershed and its channel system. The five case studies dealt with impacts that are both spatially and temporally extensive; these impacts can be severe or more subdued, but all contain important knowledge to consider with respect to forest management planning. By considering the factors governing channel morphology, it is possible to manage land use practices and simultaneously minimize deleterious impacts to stream channels.

Consideration of the processes that operate at a watershed level in British Columbia led to the development of the Channel Assessment Procedure (B.C. Ministry of Forests and B.C. Ministry of Environment, Lands and Parks 1996a and 1996b). This procedure embodies many of the concepts considered in this chapter and deals with a channel's expected adjustments to changes in materials supply, delivery mechanisms, bank integrity, and stream flow. The procedure endeavours to summarize much of what we have presented in this chapter, attempting to resolve past watershed impacts and to avoid future problems.

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CHAPTER 11

Karst Geomorphology, Hydrology, and Management

TIM STOKES, PAUL GRIFFITHS, AND CAROL RAMSEY

INTRODUCTION

The term “karst” applies to a distinctive type of landscape that develops from the dissolving action of water on soluble bedrock (Figure 11.1)—primarily limestone and marble but also dolostone, gypsum and halite.

Karst landscapes are characterized by fluted and pitted rock surfaces, shafts, sinkholes, sinking streams, springs, subsurface drainage systems, and caves. The unique features and three-dimensional nature of karst landscapes are the result of a complex

FIGURE 11.1 *Limestone: a soluble rock. (Photo: P. Griffiths)*

interplay between geology, climate, topography, hydrology, and biological factors over long time scales. Globally, examples of karst topography can be found at all latitudes and at all elevations, with rock types potentially containing karst covering approximately 20% of the Earth's land surface (Ford and Williams 2007). British Columbia's karst landscapes are of particular interest because they support coastal temperate rainforests (Figures 11.2 and 11.3), which are found only in a few other regions of the world such as southeast Alaska, Tasmania, New Zealand, and Chile (Ford and Williams 2007).

Limestone, marble, and dolostone are all examples of carbonate rocks. Carbonate rocks are primarily made up of carbonate minerals, such as calcite ($CaCO_3$) in the case of limestone and marble, and dolomite ($CaMg[CO_3]_2$) in the case of dolostone. The formation of karst landscapes in carbonate bedrock involves the "carbon dioxide (CO_2) cascade" (Figure 11.4).

In this process, rain falls through the atmosphere and picks up CO_2, which then dissolves into rain droplets. Once the rain hits and infiltrates the ground, it percolates through the soil and picks up more CO_2 and forms a weak solution of carbonic acid ($H_2O + CO_2 = H_2CO_3$). This slightly acidic water then exploits any existing joints or fractures in the bedrock, gradually dissolving the bedrock and creating larger openings or conduits for the water to flow through. Over many thousands of years, this process eventually creates underground drainage systems and caves. Mechanical processes such as stream corrasion (abrasion) also come into play once subsurface conduits are of a significant size.

Carbonate bedrock underlies approximately 10% of British Columbia (Figure 11.5), but this presence of carbonate bedrock alone does not necessarily indicate the presence of karst. The formation of karst depends on attributes such as bedrock type and purity,[1] physiographic location, and biogeoclimatic set-

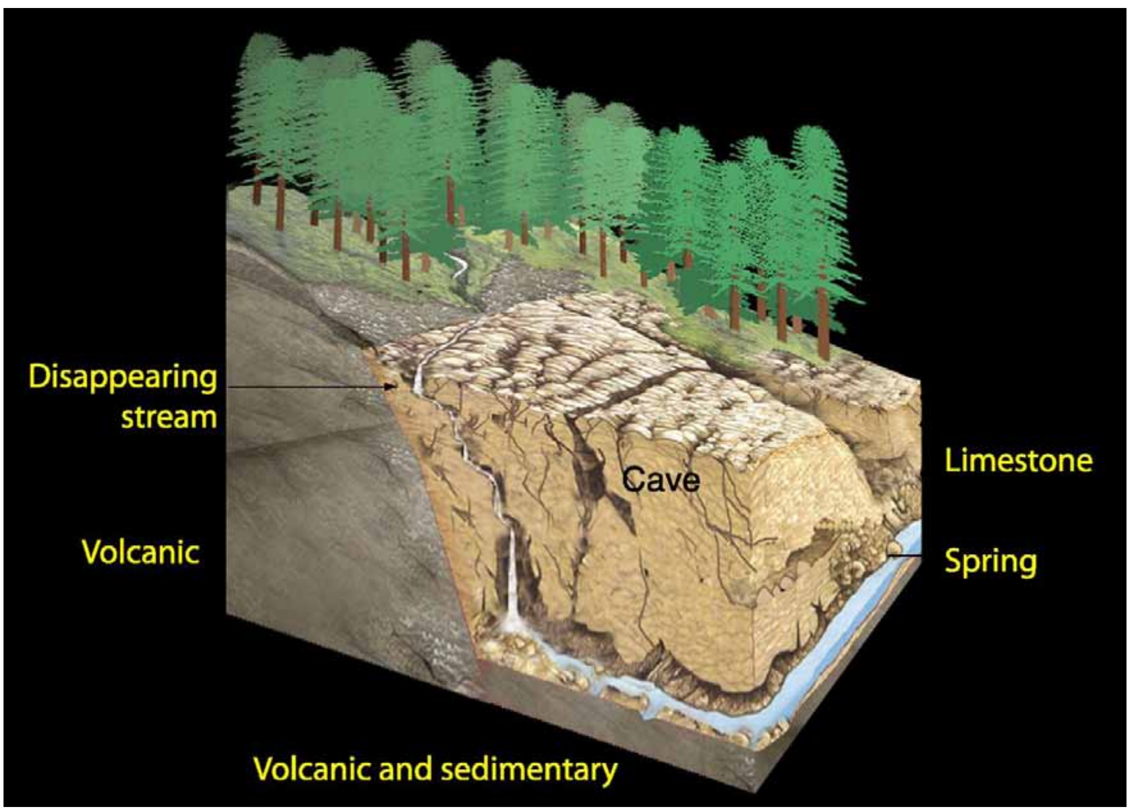


FIGURE 11.2 *A forested karst landscape system (P. Griffiths).*

1 Limestone purity (%$CaCO_2$) is one of the most important controls on karst development, the purer the limestone, the higher its potential for karst development. Karst development in carbonate rocks requires a calcium carbonate ($CaCO_3$) content of 70% or greater (Ford and Williams 2007).

FIGURE 11.3 *Forest-covered karst with small sinkhole (centre). (Photo: P. Griffiths)*

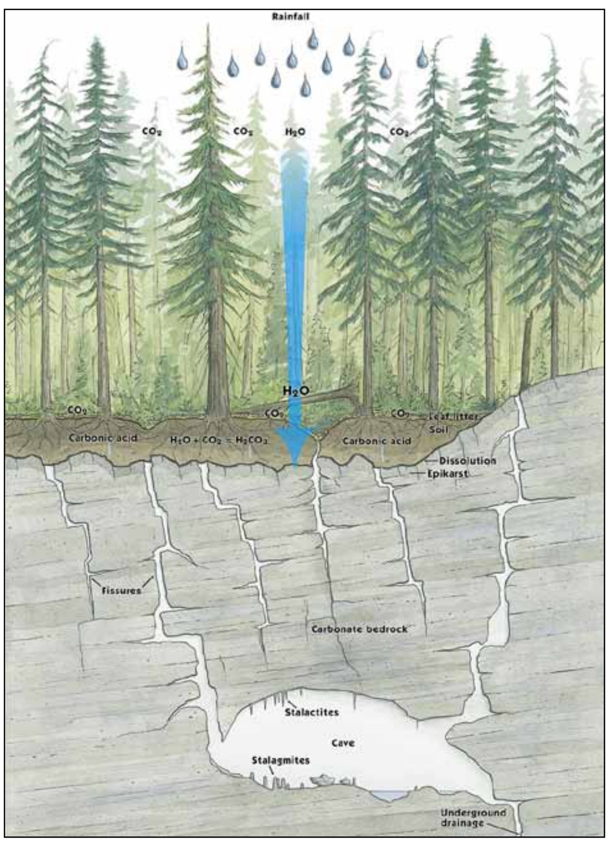


FIGURE 11.4 *The "carbon dioxide (CO_2) cascade" in the forested karst environment (B.C. Ministry of Forests 1997).*

ting and hence its development across the province is highly variable (Stokes and Griffiths 2000).

Extensive areas of alpine karst occur in the northern and southern Rocky Mountains. Well-developed karst areas are associated with carbonate units on Vancouver Island and Haida Gwaii (Queen Charlotte Islands), and smaller, isolated pockets of karst are known along the north and mid-Coast, Texada and Quadra Islands, the Sechelt Peninsula, and near Chilliwack. Less well-known karst areas occur in northwest British Columbia (e.g., Atlin, Stuart, and Babine Lakes, as well as along the Stikine, Nakina, and Taku Rivers), and in the Interior (such as the Purcell and Pavilion Mountains). Approximately 4% of Vancouver Island is underlain by limestone and much of it is forested. Vancouver Island's karst mostly occurs in the north within long and continuous belts of limestone 1–10 km wide and 10–100 km long (Figure 11.6).

A number of cave and karst parks have been established on Vancouver Island, including Horne Lake Caves Park, Clayoquot Plateau Park, Weymer Creek Park, White Ridge Park, and Artlish Caves Park (Figure 11.6).

A wide range of international material is available on the science of caves and karst. Particularly relevant to this province is the *Karst in British Columbia* brochure (B.C. Ministry of Forests 1997). The B.C. Ministry of Forests and Range's karst webpage

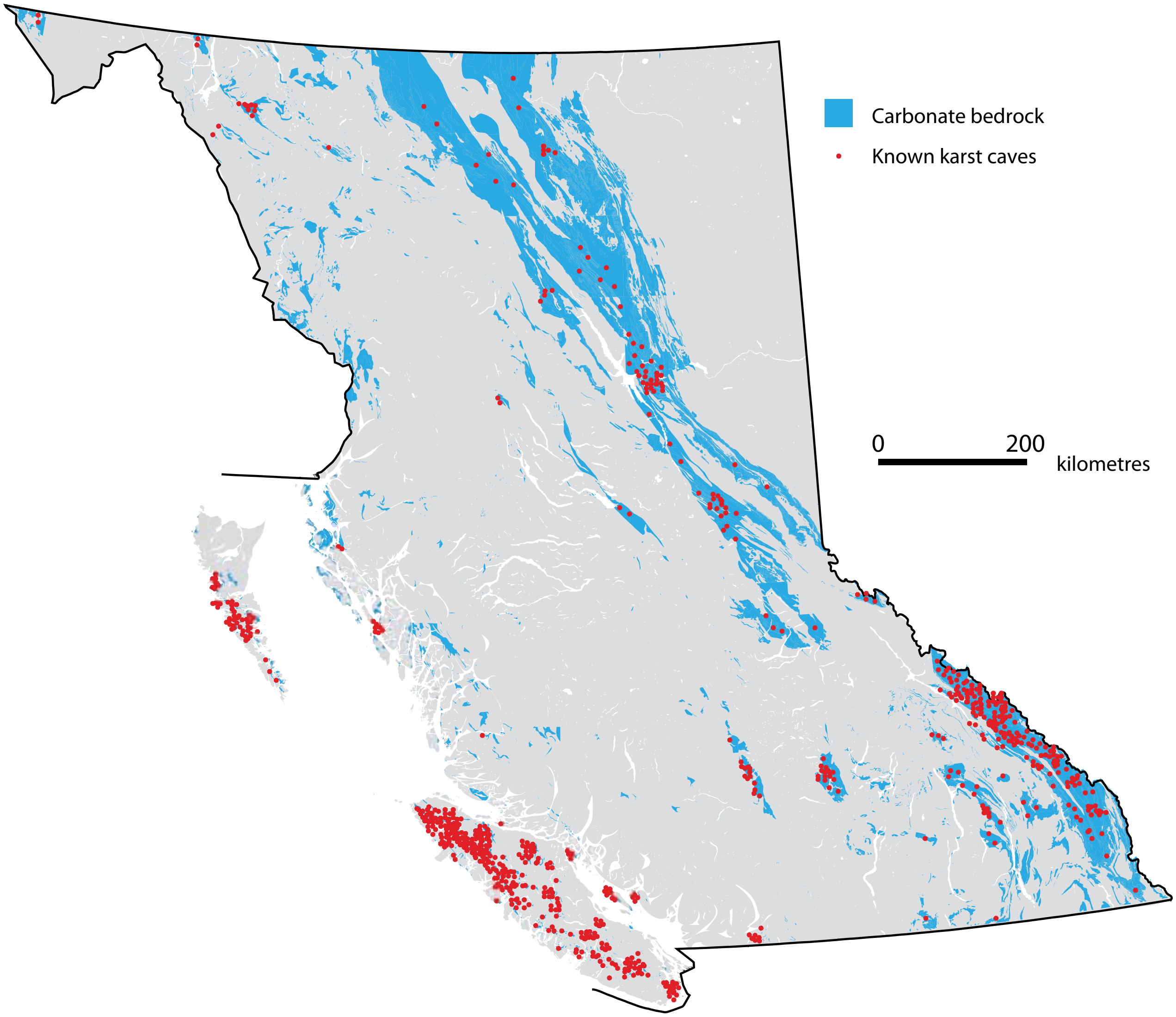


FIGURE 11.5 *Distribution of carbonate bedrock, potential karst lands, and known karst caves within British Columbia (P. Griffiths and B.C. Ministry of Forests 1997).*

is another source of useful information (see www.for.gov.bc.ca/hfp/values/features/karst/index.htm). Papers that provide good summaries of cave/karst landscape issues and processes include those of Baichtal and Swanston (1996) and White et al. (1995). Textbooks on caves and karst science include those of Jennings (1985), White (1988), Ford and Williams (2007), Gillieson (1998), Finlayson and Hamilton-Smith (2003), and Palmer (2007). Gunn (2004) provides an extensive and well-illustrated encyclopedia that covers all aspects of cave and karst science.

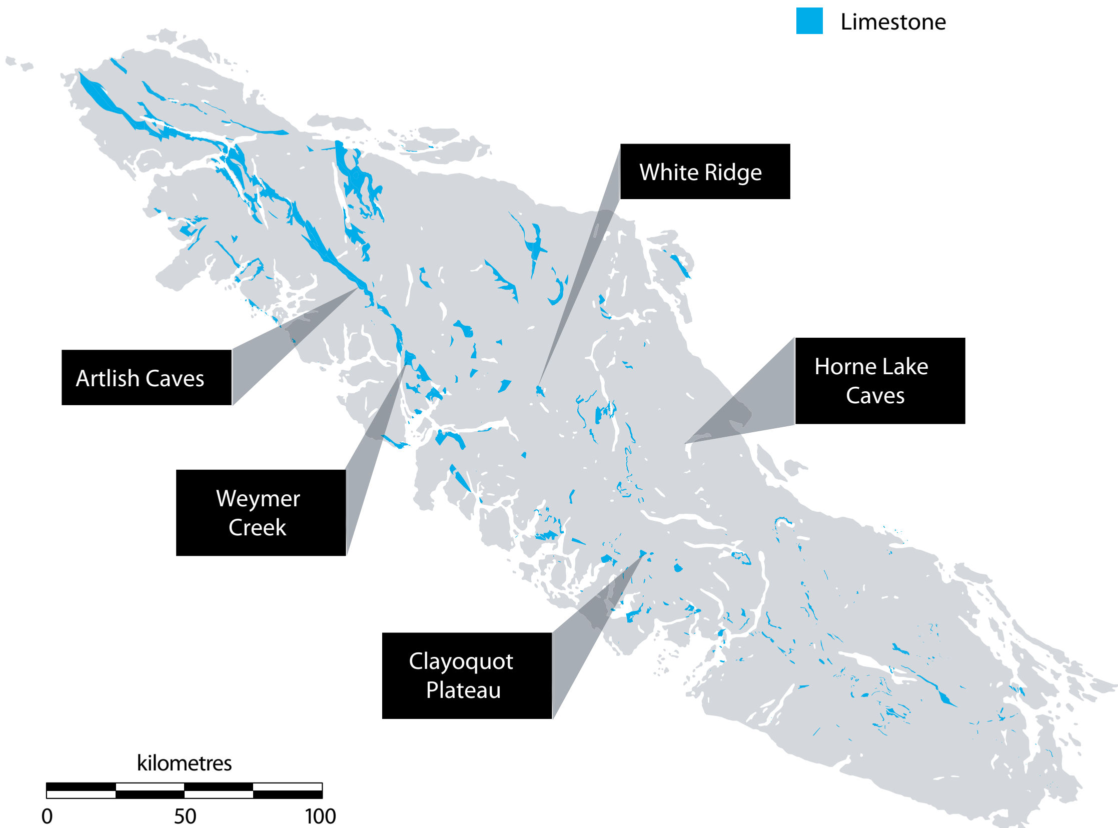


FIGURE 11.6 *Cave and karst parks of Vancouver Island.*

KARST LANDSCAPES AS FUNCTIONING SYSTEMS

All landscapes (e.g., desert, glacial, mountain, and karst) work as functioning systems, exhibiting continual movements of materials, energy, and biota, which in turn constrain the natural processes and balances of these environments. However, the complexity of systems associated with karst landscapes is enhanced because these landscapes include distinctive subsurface environments consisting of solutionally enlarged fractures and cavities that directly link the surface to the subsurface and vice-versa. These micro- to mesoscale fractures and cavities provide pathways for easy transfer of water, air, soil, rock, organic matter, and biota. The processes by which these materials are moved are integral to the character, and functioning of a karst system. Interruptions to these processes can result in adverse impacts to both the surface and subsurface environments (Baichtal 1995; Baichtal and Swanston 1996). Thus, karst can pose additional management considerations, challenges, and constraints relative to other types of landscape.

Karst systems are distinct from non-karst systems because of the processes of karst dissolution, the permeability of the solutionally developed landscape surface, the presence of a well-developed and open subsurface, fewer surface streams, and an overall calcium-rich environment (White et al. 1995; Gunn 2004). Given these differences, it is perhaps not surprising that specialized surface and subsurface biota can inhabit karst landscapes (Figure 11.7).

These can range from calcium-dependent flora on the exposed bedrock surface to cave-adapted fauna in the subsurface (e.g., *Stygobromus quatsinensis,* a rare freshwater crustacean found in underground pools on Vancouver Island; Holsinger and Shaw 1987). Some subsurface fauna in karst ecosystems

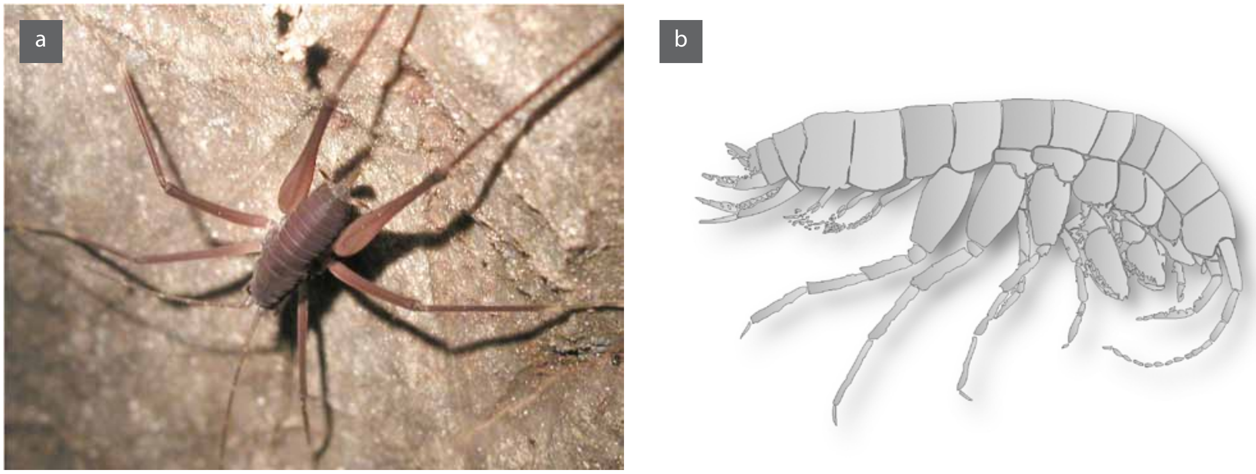


FIGURE 11.7 *Examples of karst fauna: (a) a cricket commonly found in caves; and (b) a cave-adapted crustacean. ((a) photo: P. Griffiths; (b) modified from Holsinger and Shaw 1987)*

have evolved over long time periods in this very stable and dark environment, resulting in adaptations such as reduced pigment and eyesight (Finlayson and Hamilton-Smith 2003). Such organisms, termed "troglobites," are obliged to live their entire lives in the subterranean world.

IDENTIFYING KARST LANDSCAPES FEATURES

A karst landscape unit, or more simply a "karst unit," is defined for the purposes of this chapter as "a three-dimensional belt or block of soluble bedrock area surrounded by other less soluble rock types." The first step in investigating a karst unit is to confirm the extent or intensity of karst development and to delineate its boundaries. Some understanding of bedrock geology is critical to determine the likely extent and boundaries of a karst unit. Bedrock geology maps[2] can be consulted to identify limestone or other soluble bedrock units. In many cases, the scale of bedrock mapping is not sufficient to outline the soluable bedrock unit in detail, and therefore field investigation is required to confirm the unit boundaries.

In the field, carbonate bedrock exposures such as limestone or marble are often visible in rock cuts, along creeks, under windthrown trees, or along topographic highs or bluffs. Carbonate bedrock can be readily identified by characteristics such as its white to grey colour, solutionally weathered surfaces, bedding layers, presence of fossils, and relative softness compared to other rocks. Carbonate rock types can be confirmed by dropping a small amount of dilute hydrochloric (HCl) acid onto its surface.[3] If the rock sample is limestone or marble, the HCl solution will effervesce (or bubble) with a visible and audible chemical reaction that gives off CO_2. Dolostone may require powdering of the rock and a more concentrated acid solution to produce effervescence. A useful tool for delineating the presence of carbonate bedrock at the regional level is to look for float material (e.g., cobbles and boulders of limestone) in larger creeks that drain these areas. In addition, the presence of anomalously high electrical conductivity values in streams as measured using a handheld electrical conductivity meter can indicate the occurrence of water that has been in contact with carbonate bedrock for extended periods; such water might emerge from a karst spring.[4]

2 In British Columbia, websites such as MapPlace BC (www.mapplace.ca) can be a good place to start.
3 Dilute HCl or muriatic acid is readily available from pharmacies and is commonly used by geologists for this purpose.
4 Water in contact with carbonate bedrock generally contains more free ions, owing to the chemical reactions that have dissolved the bedrock; therefore, when tested with a handheld conductivity probe, this water provides a high reading compared to water from non-carbonate rocks. As a general rule, karst waters have 5–10 times higher electrical conductivity values than non-karst waters.

Bedrock geology, in combination with surface contour maps, can assist in delineating karst units. Aerial photographs, high-resolution satellite imagery, and (in some cases) lidar can also be useful to identify diagnostic surface karst features (e.g., large sinkholes and disappearing streams or springs), distinct differences in bedrock lithologies (e.g., white marbles), or disrupted surface drainage patterns. In some aerial photographs and satellite images, forest cover may hinder the identification of these features; however, many characteristics of a karst landscape can be identified in recently harvested cutblocks or in areas above the tree line (B.C. Ministry of Forests 2003a).

Karst landscapes are usually recognized in the field by the presence of diagnostic surface karst features. Karst features that can readily be observed on the surface include: solutionally rounded or sculpted bedrock exposures; sinkholes; cave entrances; streams that disappear at discrete openings or sink points; and springs that emerge from bedrock openings or conduits. These types of surface karst features are common in karst areas of coastal British Columbia where relatively pure carbonate bedrock is present and precipitation levels are high (e.g., Vancouver Island, Haida Gwaii, and the mid-Coast). However, these diagnostic features can be more difficult to identify and confirm in the Interior where precipitation levels are generally lower and (or) the landscape is overlain by thick mantles of glacial material (Stokes and Griffiths 2000).

Surface karst features can vary dimensionally from small-scale features (millimetres to centimetres) to large-scale features that measure in the hundreds of metres (Ford and Williams 2007). Small-scale features on soluble rock outcrops can include distinctive linear channels or grooves known as "karren" (Figure 11.8) that are classified by dimensions and morphology (Gunn 2004).

Examples commonly found in coastal British Columbia include *rundkarren* (rounded channels separated by rounded ridges) and *rillenkarren* (shallow channels with sharp ridges 2–3 cm apart). Larger-scale surface karst features are commonly

FIGURE 11.8 *Karst solutional grooves or karren on steeply sloping bedrock surfaces. (Photo: P. Griffiths)*

encountered when traversing a karst landscape (Figure 11.9). In most cases, these larger features are classified by morphology, shape, and dimensions rather than genetic origin; however, in some cases, the feature's function (e.g., input/output of water and air) is used as part of the classification. Table 11.1 presents examples of some of the most common surface karst features.

Details on the types of surface karst features typically encountered within the forests of British Columbia are available in the appendices of the *Karst Inventory Standards and Vulnerability Assessment Procedures for British Columbia* (B.C. Ministry of Forests 2003a) and in the *Karst Management Handbook for British Columbia* (B.C. Ministry of Forests 2003b). Additional details about the origins and functions of various types of surface karst features are available in Jennings (1985), White (1988), and Ford and Williams (2007).

In the broadest sense, the three-dimensional nature of a karst landscape can be broken down into three parts: (1) exokarst, (2) epikarst, and (3) endokarst (Figure 11.10).[5]

FIGURE 11.9 *Examples of karst features found in forested regions of coastal British Columbia: (a) a sinking stream at a vertical sink point (or swallet); (b) a series of small sinkholes; (c) a karst spring; and (d) a dry karst canyon. (Photos: P. Griffiths)*

5 Although the terms "exokarst" and "endokarst" are rarely used, these terms are useful to illustrate the upper surface and subsurface components of the karst landscape.

TABLE 11.1 *Common surface karst features*

Dry valley	A valley that generally lacks a stream, although one may occasionally form during peak rainfall events.
Grike	A narrow and deep slot formed by dissolution along a pre-existing fracture in bedrock.
Karst canyon	A steep-sided canyon in karst sometimes exhibiting distinctive surface solutional rocky relief features (e.g., scalloping).
Karst spring	A site where an underground stream emerges from a karst conduit or cave.
Polje	A large, flat-bottomed karst depression with water periodically present across its floor.
Sinkhole	A topographically closed depression that is circular or elliptical in plan view, with enclosing sidewalls that can range from shallow and gradually sloping toward a central drainage focus to steep or almost vertical.
Solution tube	A circular or elliptical steeply inclined tube formed by dissolution, which is sometimes found on karst bedrock exposures.
Swallet	A point where a stream of any size sinks underground. In some cases, a swallet can also be a cave entrance.

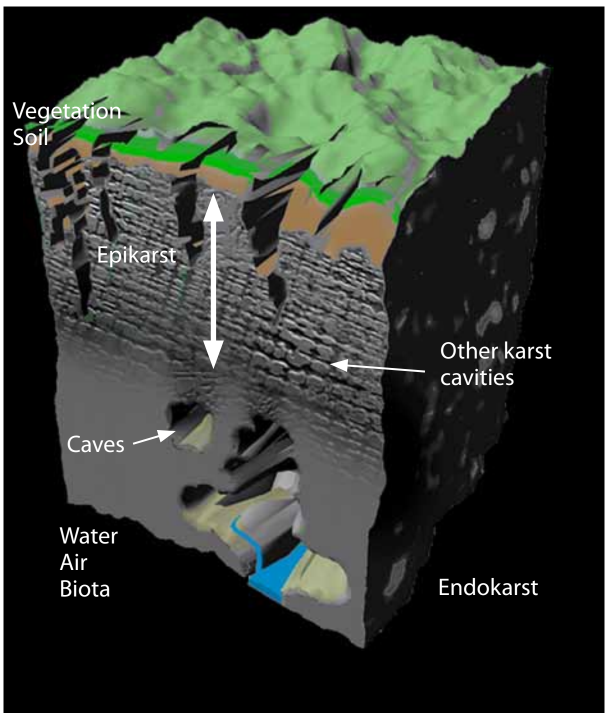


FIGURE 11.10 *The linkage between epikarst and endokarst; note that exokarst is the surface of the karst landscape (P. Griffiths).*

Exokarst describes all features found on the surface of the karst landscape, ranging from small-scale to large-scale features (e.g., from karren to sinkholes to poljes). Epikarst is the zone of solutionally enlarged openings or fractures that extends from the surface (the exokarst) down as much as 10–30 m below the surface to the underlying endokarst. The endokarst describes all deeper components of the underground karst landscape, including the smallest cavities, cave speleothems,[6] cave sediments, and cave passages. The epikarst zone therefore plays a critical role in the karst system, allowing water, air, and other materials (sediment, organic debris, and nutrients) to be readily transferred from the surface to the subsurface.

The term "cave" is often defined as "a natural cavity within the earth's crust that is connected to the surface, is penetrable by a human, and includes a zone of permanent and total darkness" (B.C. Ministry of Forests 2003a:72). Most people correctly associate caves with karst, although in the context of karst systems, caves tend to acquire a disproportionate amount of public attention. Caves are undeniably very important features and can contain a range of significant values and resources, including the geomorphological, paleontological, archaeological, and biological. A number of references are available that provide a more extensive discussion of these values and resources, plus the related cave management issues (e.g., Kiernan 1988; International Union for the Conservation of Nature 1997; New Zealand Department of Conservation 1999; Ramsey 2004). However, caves as karst features should also be placed into the

[6] A speleothem is any form of secondary deposit in a cave that forms by mineral precipitation (usually of calcite) and includes such features as stalagmites, stalactites, and draperies.

perspective of other subsurface openings in the karst system, as the vast majority of these openings or voids are not large enough for humans to enter but are, nevertheless, important biospaces with their associated eco-hydrological functions (Figure 11.11). As such, caves typically make up only a small portion of the cavities within a karst system (e.g., less than 0.01%; Ford and Williams 2007).

Caves may appear as complex or random patterns when displayed in maps or as cross-sections, but these features typically exhibit three basic components: (1) passages, (2) chambers, and (3) one or more entrances. In most cases, geological or hydrological factors dictate the location of a cave by defining its shape, extent, and dimensions. A cave's significance is not necessarily related to its size; even a very small cave can contain significant resource contents or values.

Regionally, Vancouver Island has the highest density of recorded caves of any karst region in Canada, as well as 5 of the 10 longest and deepest caves in Canada (e.g., Weymer Cave System is more than 13 km long and Thanksgiving Cave is over 400 m deep). Caves on Vancouver Island are predominantly found in carbonate bedrock units that are steeply dipping, and occur from sea level, on forested lower and middle slopes, and up to the highest peaks 2000 m above sea level. Many of these caves consist of multiple chambers and passages with associated vertical drops of up to 50 m, whereas other caves occur along river drainages with discrete sink points and emergences (e.g., Artlish River Cave). Calcium carbonate deposits, or speleothems, in the form of stalactites, stalagmites, soda straws, draperies, and helictites are present in many British Columbia caves, as are cave fills or sediments (e.g., layered clay, sand, gravel, and rubble deposits). Both speleothems and cave sediments contain important information for understanding scientific issues such as ancient flora/fauna, past glaciation events and climates, and, past human activities (e.g., migration patterns).

Because caves can occur in geological environments other than karst, caves are not necessarily diagnostic karst features. Non karstic examples include lava caves found in basalt flows, glacier caves, crevice/fracture caves along faults, sea caves caused by wave action and erosion, and talus caves under rock debris (Palmer 2007).

FIGURE 11.11 *Examples of the cave environment: (a) an active underground stream; (b) a sinking river and large swallet; and (c) viewing from within a cave passage towards a cave entrance. (Photos: P. Griffiths)*

Water is the key to understanding the formation and functions of a karst landscape. As precipitation falls on a karst landscape, it generally infiltrates downward through the soil towards the soil–bedrock contact (Figure 11.12). The water then percolates through the epikarst zone along small fractures or solutionally enlarged openings in the bedrock, gradually moving downward until it reaches larger conduits and (or) caves below. In general, the upper unsaturated (or vadose) part of a karst aquifer is where water partially fills openings or voids, and the lower saturated (or phreatic) part of the aquifer is where all voids are water filled. Water may be stored in these voids within conduits and fractures and, depending on flow stage, will eventually leave the karst landscape system at outflow sites such as springs. Some obvious hydrological features that distinguish karst from other types of landscapes include a general lack of surface drainage or streams, the presence of discrete sink points where streams disappear (swallets), and the occurrence of springs where water emerges.

An important characteristic of many karst landscape systems is the presence of an aquifer suspended above the phreatic zone and located within the epikarst zone (Williams 2008). The porosity and permeability of epikarst is typically greatest at the surface and decreases with depth (i.e., from 1 m down to 10 m below the surface). The net result is that rainfall can be temporarily detained and stored in the rock matrix and fractures of the epikarst before infiltrating downwards into the lower parts of the karst aquifer (Williams 2008). Water on reaching the lower part of the karst aquifer moves into larger subterranean conduits that not only provide the main flow paths for water within the aquifer but also allow flow into and out of fractures (Gillieson 1998; Gunn 2004; Williams 2008). In general, the storage of water in the rock matrix and fracture porosity is considered a longer-term phenomenon, and water storage in conduits is a shorter-term phenomenon.

Streams will run over the surface of karst when and where water flow exceeds what can infiltrate into the channel bed or into sink points within the karst landscape. Karst streams are often inactive during low flow periods and active only during high flow events. Year-round surface flow on well-developed

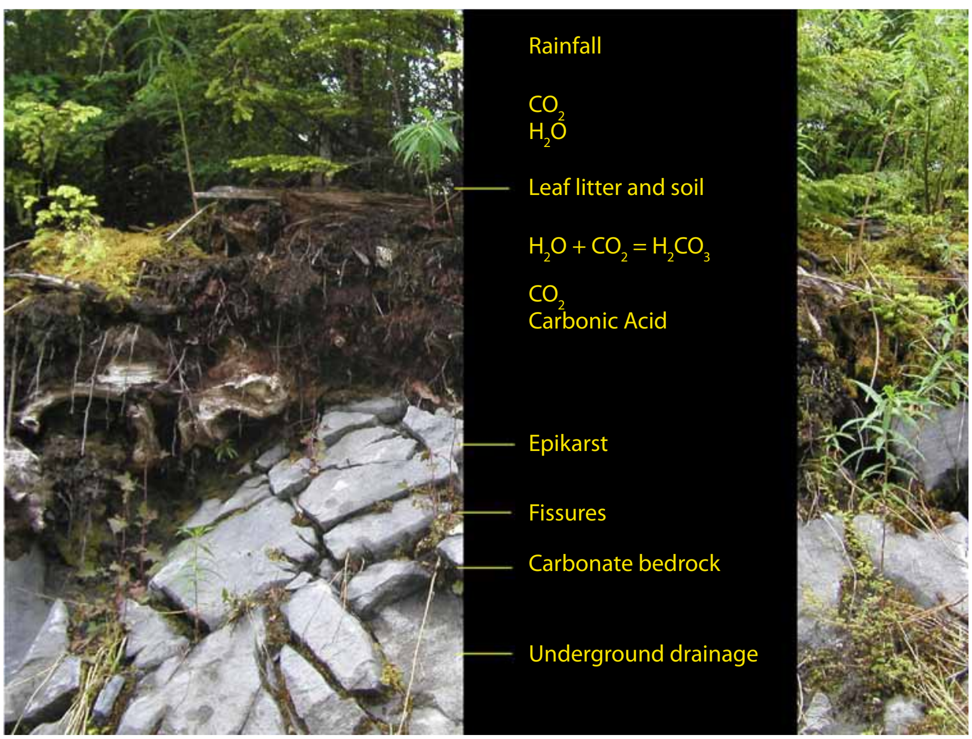


FIGURE 11.12 *Infiltration of water through soil and the epikarst. (Source: Griffiths and Ramsey 2009)*

karst landscapes is rare but can occur where the karst is covered by thick and (or) impermeable sediment cover (e.g., till). Some of the most spectacular features associated with karst streams are karst canyons, where water has dissolved the soluble bedrock, creating steep and sometimes overhanging sidewalls.

The recharge of karst aquifers can be either autogenic, allogenic, or a combination of the two (Figure 11.13; White et al. 1985; Ford and Williams 2007). In autogenic recharge, water falls directly onto the karst and infiltrates the soil and epikarst, and then enters into the underlying aquifer. In some cases, this autogenic recharge flow can be concentrated at point-input features such as sinkholes. Allogenic recharge occurs when water falls on adjacent or nearby non-karst landscapes and is transported onto a karst unit via surface streams.

This water may eventually disappear underground if it reaches discrete sink points in the karst unit such as swallets or sinkholes. The characteristics of water from allogenic recharge sources can vary depending on conditions upstream, but this water generally has lower electrical conductivity and lower pH values compared to water that has flowed through a karst system. In some cases, allogenic water can be very aggressive (acidic) when derived from non-karst wetlands or bogs. When such water encounters carbonate bedrock, it can result in more intensive karst development. On Vancouver Island, allogenic streams draining from non-karst slopes to a karst unit can form a line of swallets or sink points along the upper karst unit boundary.

An important concept in karst hydrology is the notion of the "karst catchment" (i.e., the drainage area that contributes water to a particular karst landscape unit).[7] Karst catchments can cross beneath topographic divides because the water flowing in underground conduit systems is not necessarily constrained by surface topography, and hence the catchment for any particular karst unit may bear little or no relation to the surrounding topographic divides. Water from adjacent or adjoining non-karst landscapes can also contribute significantly to the catchments of karst units (Figure 11.14; B.C. Ministry of Forests 2003a).

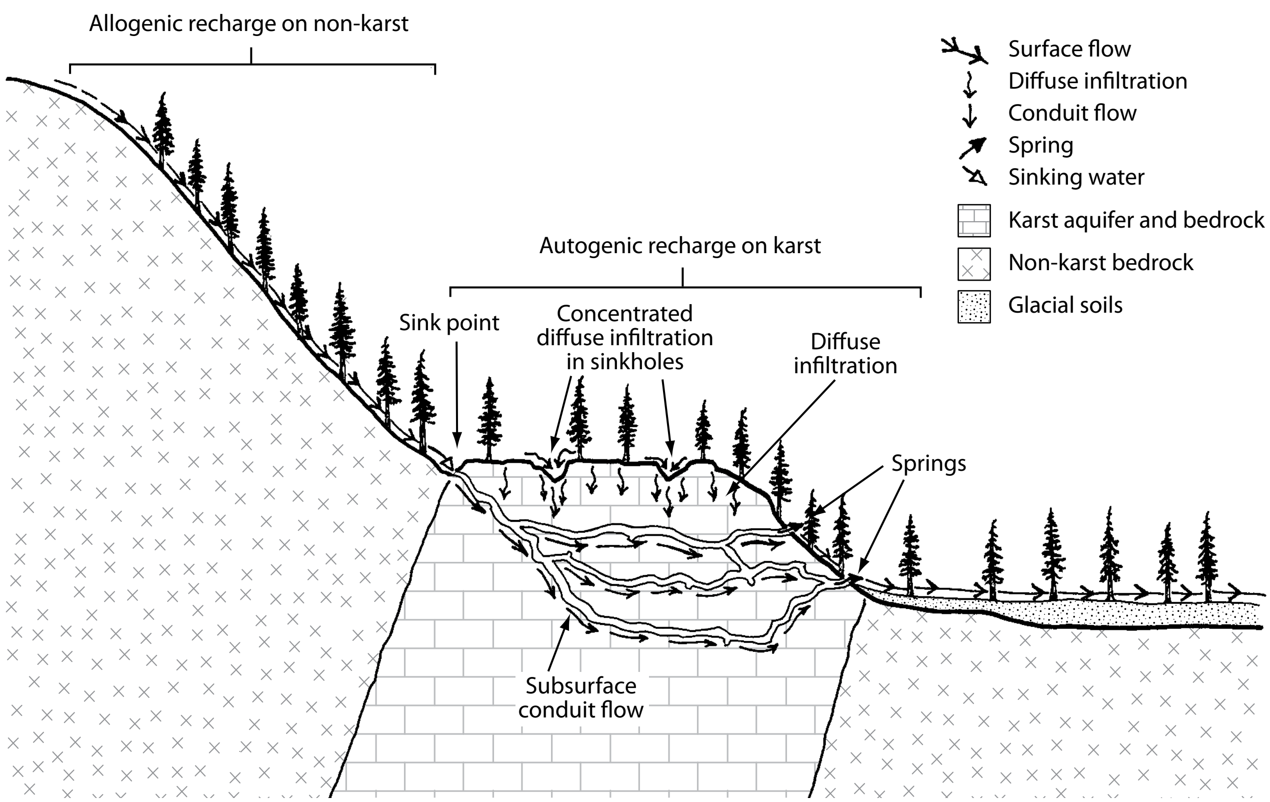
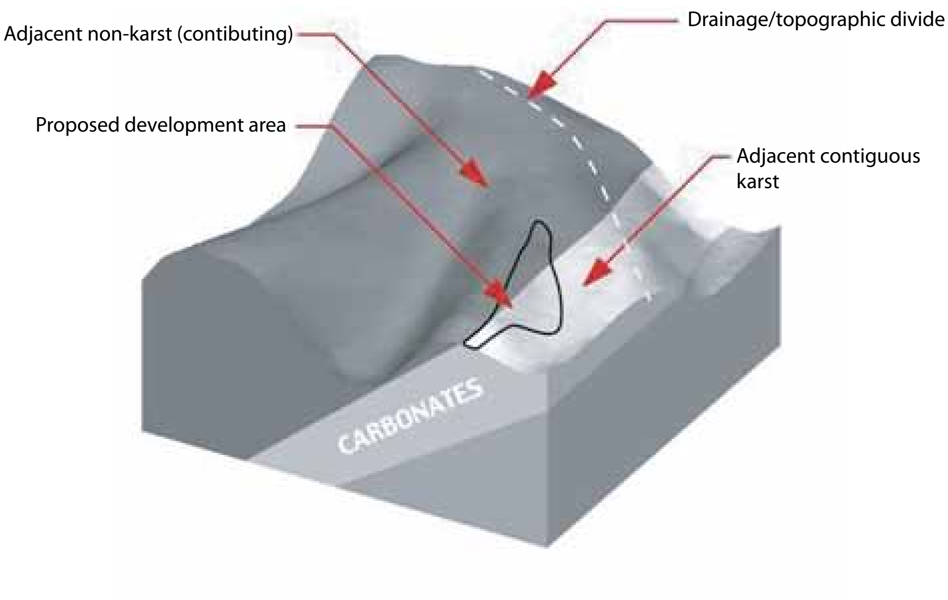


FIGURE 11.13 *Autogenic and allogenic recharge of karst aquifers (T. Stokes).*

7 The term "karst catchment" is used instead of "karst watershed," primarily because watershed implies a strong topographic control (i.e., watershed boundaries that occur along topographic divides). This is commonly not the case for karst landscapes.

A. Adjacent karst and non-karst catchments



B. Adjoining karst and non-karst catchments

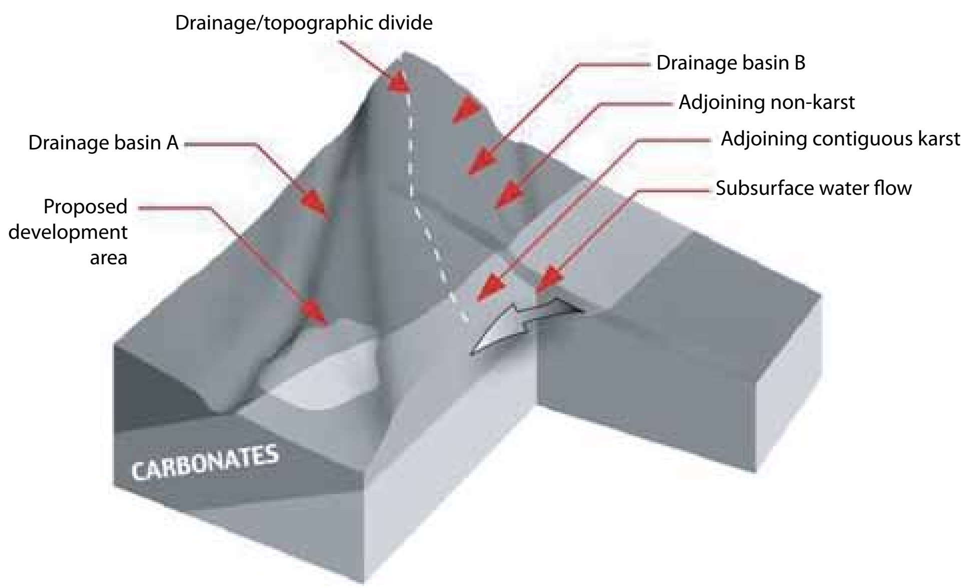


FIGURE 11.14 *Karst catchment hydrology; note the differences between the contributing catchment areas of the adjacent and adjoining karst, plus the adjacent and adjoining non-karst; also note the subsurface flow under topographic divide (B.C. Ministry of Forests 2003a).*

Techniques such as water tracing using fluorescent dyes are often required to fully determine a karst catchment's full extent (Stokes et al. 1998; Prussian and Baichtal 2007).

Water typically leaves karst aquifers by way of karst springs, which represent water that has flowed through a carbonate bedrock from a source area at a higher elevation. Karst spring discharges can range from small trickles of water to raging rivers tens of metres in width. Typically, karst springs are located at lower elevations—along valley floors, sides of lakes, or coastal shorelines. In some cases, these springs can occur beneath water bodies. Karst springs differ from those occurring in other rock types in that these springs are mostly fed via conduits. Discharge waters from karst springs are also used to infer some of the physical and chemical characteristics of a karst aquifer (Gunn 2004). Springs with continuous year-round flow suggest that the aquifer has some potential for storage relative to the amount of water flowing through the system. These springs typically occur at low elevations and are termed "outflow springs." Springs that are more active during high flows, or that have seasonal or intermittent flows, are termed "overflow springs." Overflow springs are typically at sites of slightly higher elevation than the corresponding outflow springs.

Although often overlooked in Canada, karst aquifers are recognized globally as important natural resources. An estimated 25% of the world's population depends on water from karst aquifers for daily use (Ford and Williams 2007). Subterranean karst aquifers have been included in the RAMSAR Wetland Classification System since 1971 (New Zealand Department of Conservation 1999), and provide habitat for underground-adapted aquatic fauna known as "stygobites" (Pipan 2005; Ford and Williams 2007; Pipan and Culver 2007). Research in southeast Alaska suggests that aquatic ecosystems associated with streams fed by karst waters can be more productive than those that are not. Streams flowing through or from karst landscapes have distinct water chemistry and appear to support more fish than non-karst streams (Baichtal et al. 1995; Bryant et al. 1998). This research likely has important implications for fisheries and karst landscapes of coastal British Columbia.

KARST LANDSCAPE UNITS: TWO CASE STUDIES FROM VANCOUVER ISLAND

To illustrate the typical conditions and characteristics of forested karst landscape units in British Columbia, we describe two case studies for Vancouver Island—one at a low elevation and one at a high elevation. In general, most of the karst landscapes on Vancouver Island are within the limestone of either the Quatsino or the Mount Mark formations. Generally, these limestone formations occur as moderately to steeply dipping linear belts less than one kilometre to tens of kilometres in length and hundreds of metres to kilometres wide. The limestone of both the Quatsino and Mount Mark formations is of a relatively high purity—typically greater than 90% $CaCO_3$. Karst development is also controlled by elevation and slope gradient. In general, high-elevation areas are more likely than low-elevation areas to develop a steep hydraulic gradient, and hence have a greater potential for karst development. Gentle slopes (e.g., benches) are also preferable to steep slopes for karst development, possibly as the former allows more time for water infiltration (Stokes 1999; B.C. Ministry of Forests 2003a). Figure 11.15 shows an example of a lower-elevation karst landscape unit on Quadra Island, which is one of the Northern Gulf Islands just to the east of Vancouver Island.

In this location, a belt of Triassic Quatsino Formation limestone extends north to south through the centre of the island. The karst unit is approximately 15 km long and 1–2 km wide, varying in elevation from sea level to approximately 100 m above sea level. The limestone unit is bounded by basaltic volcanic rocks to the west and by granitic rocks to the east, and is steeply to moderately dipping. Most of the limestone is located in or near a topographic low that is mantled by glacial materials. Based on reconnaissance mapping, the limestone in this region is considered to have high potential for karst development (Stokes 1999). Small (< 10 m diameter) sinkholes are common but variable in concentration. Large sinkholes (up to 40 m in diameter) do occur but are rare. Solution holes and grikes also occur on exposed epikarst sites, which are typically found on karst bedrock highs (e.g., hums) and (or) occasional ridges. A number of karst springs occur on the mid- and lower-elevation slopes, with at least one used for domestic water supply. Caves occur in several

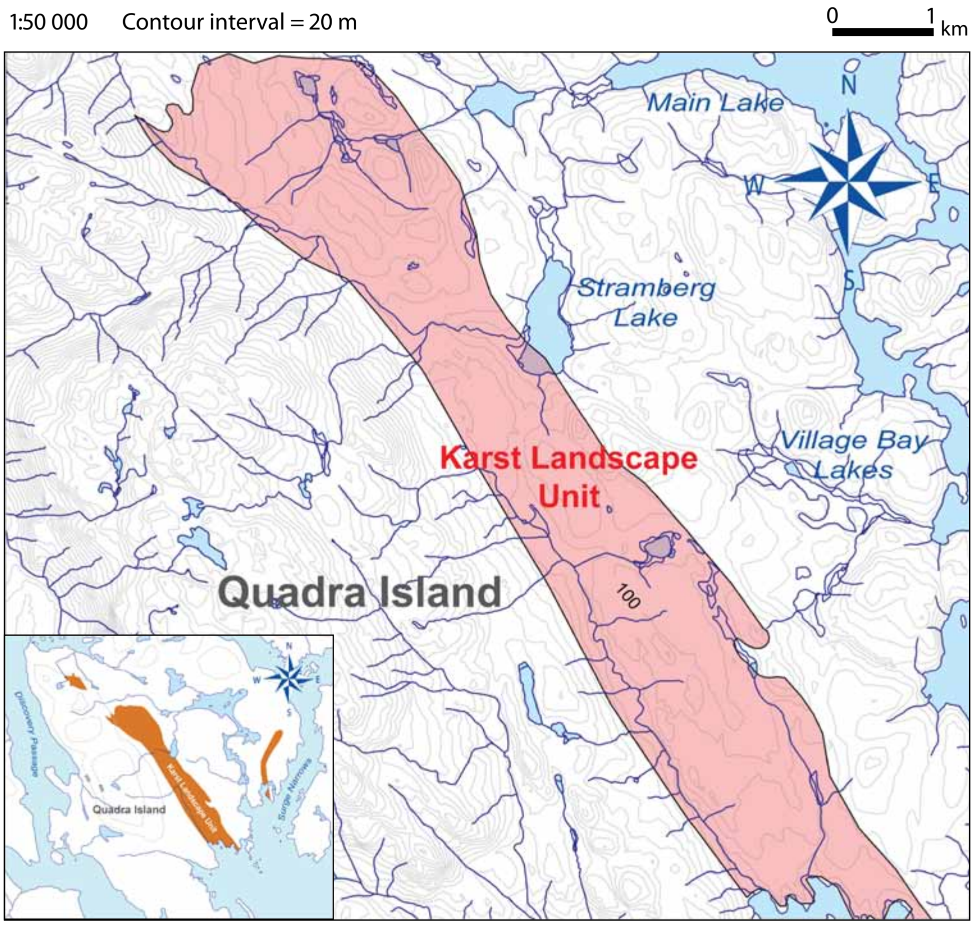


FIGURE 11.15 *Quadra Island karst unit, northern Gulf Islands (P. Griffiths).*

locations but are not of great length or depth. Most of Quadra Island was logged and forest stands consist of second-growth trees of various ages. Active logging continues both on and off the karst unit. In our experience, no systematic, planning-level karst inventory has been completed in this region, but a number of operational-scale karst field assessments linked to cutblock planning and development were carried out.

Figure 11.16 shows a higher-elevation example of a karst landscape unit to the east of Nimpkish Lake in the Noomas Creek and Kinman Creek areas of northern Vancouver Island. At this location, the karst occurs in a 10–15 km long, 4–5 km wide, northwest–southeast trending limestone unit of the Quatsino Formation, bounded by volcanic rocks to the east and west and by intrusive rocks to the south. This unit varies in elevation from approximately 700 to 1300 m above sea level and has variable soil cover. Significant portions of the unit were logged but some old-growth stands remain. The area has a high potential for karst development based on the reconnaissance mapping of British Columbia performed by Stokes (1999). Regional evaluation of the unit was carried out as part of a planning-level inventory of Tree Farm Licence 37 for Canadian Forest Products Ltd.[8] This project defined the limits of the karst unit, identified some of the major karst features, and stratified the karst unit into areas of different karst vulnerability potential. The unit contains a number of extensive cave systems such as Arch Cave and Glory'ole—both of which have provincial and national significance. In addition to these significant caves, the unit also includes several large and

8 Stokes, T.R. and P.A. Griffiths. 2003. Planning-level karst inventory for TFL 37 and FL A19233, northern Vancouver Island. Prepared for Canadian Forest Products Ltd. Unpubl. report.

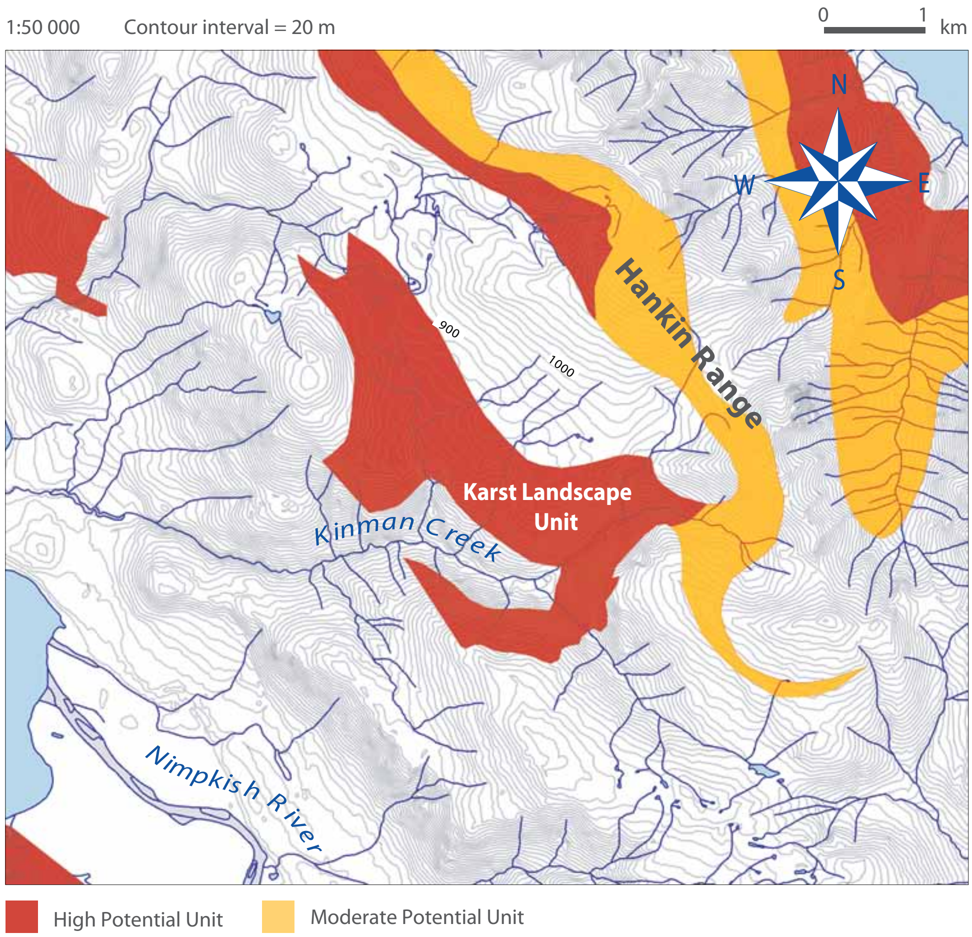


FIGURE 11.16 *Noomas Creek and Kinman Creek karst units, northern Vancouver Island (P. Griffiths).*

significant springs. Epikarst, visible on some exposed bedrock surfaces, is slightly to moderately well developed. A wide range of different karst features such as sinkholes, karst canyons, grikes, and swallets are also present. Dye tracing carried out at the northern end of the unit linked sink points associated with cave entrances and important cave systems to downslope springs. As well as defining the various subsurface flow paths, the dye tracing also identified potentially sensitive non-karst catchments upslope (Stokes et al. 1998).

IMPACTS OF FORESTRY ACTIVITIES ON KARST LANDSCAPES

Overall, forestry activities have a range of impacts on the karst landscape, including disturbance to surface karst features, subsurface environments, karst waters, and karst biota (Figure 11.17). These disturbances can also affect water quality and flow regime, scientific values (e.g., archaeological, paleontological, geological), recreation (both surface and subsurface), visual quality, and fisheries. Because of the inherent low-energy transfers present at these sites, impacts to subsurface environments and caves take hundreds to thousands of years to recover and restore to their previous states (Gillieson 1998). On the surface, as with non-karst landscapes, recovery of forest conditions after disturbances will also occur over time; however, soil loss into vertical solution openings or epikarst can significantly slow this process (Harding and Ford 1993). Removal of the forest canopy on well-developed karst during logging or other forestry activities can:

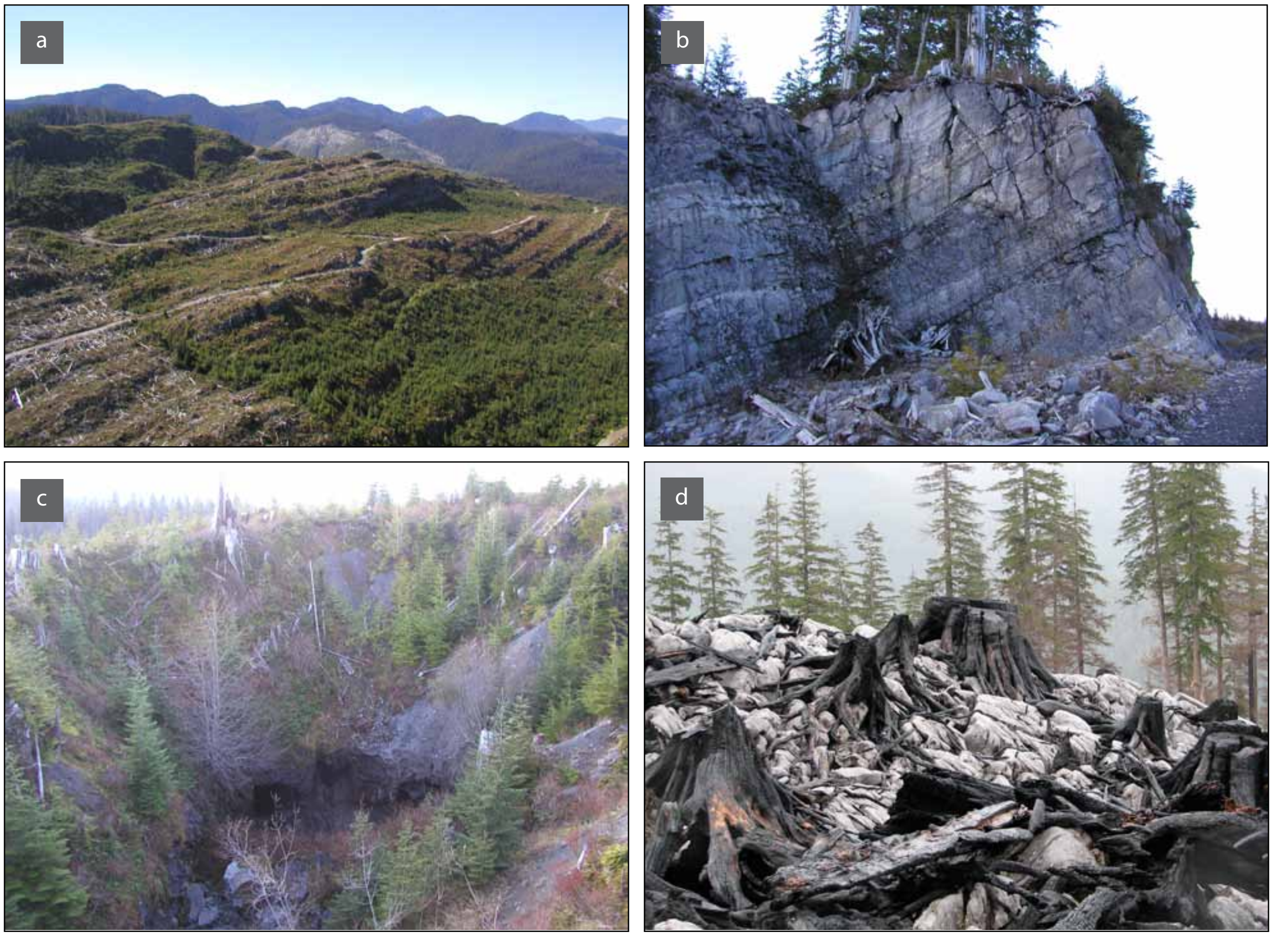


FIGURE 11.17 *Examples of past and potential disturbances to karst by forestry activities: (a) a contiguous clearcutting of karst lands; (b) a quarry excavated into karst; (c) a large sinkhole with destabilized sidewalls following logging activities; and (d) soil loss and burning on well-developed epikarst. (Photos: P. Griffiths)*

- change hydrology by redirecting surface flows that can dewater or flood subsurface conduits;
- increase input of organic debris and sediment into subsurface cavities;
- alter microclimates of larger surface features (e.g., a > 20 m diameter sinkhole);
- alter microclimates of shallow subsurface cavities; and
- increase surface desiccation and loss of thin surface soils into well-developed karst.

One of the key differences between karst landscapes and other types of landscapes is the presence of subsurface biota, of which little is really known, particularly in British Columbia. Detailed inventory information on the biodiversity of forested karst lands on Vancouver Island and in the rest of British Columbia is limited. Studies in other karst areas outside of British Columbia show that the better developed a karst landscape (i.e., the more openness between the surface and subsurface environments), the greater the likelihood of finding life forms that have adapted to it. In effect, the greater the variability (or topographical roughness) of the karst landscape, the more likely it is to possess a greater diversity of life forms that have developed in isolated niches. Caves and other karst cavities can host and support a wide range of cave-adapted life forms (e.g., blind and de-pigmented crustaceans; Gillieson 1998).

Karst Inventories in British Columbia

A framework for carrying out karst inventories in British Columbia is outlined in the *Karst Inventory Standards and Vulnerability Assessment Procedures for British Columbia* (B.C. Ministry of Forests

2003a). Under this framework, inventory activity can take place at three levels: (1) reconnaissance, (2) planning, and (3) operational.

The intent of these three different inventory levels is to provide a filtered approach to the karst inventory process, going from broader information to progressively greater detail. In 1999, a reconnaissance-level karst inventory project was carried out for all of British Columbia (Stokes 1999). During this project, all potential soluble bedrock units (i.e., limestone, dolomite, marble, and gypsum) were identified at a 1:250 000 scale using a series of digital bedrock geology maps. These soluble bedrock units were then rated for the potential to develop karst. Knowledge of specific cave and (or) karst features was also incorporated into the mapping.

Planning-level karst inventories can be carried out at the 1:20 000 or 1:50 000 scales, and are intended for the strategic management of forestry activities in karst landscapes. The primary aims of planning-level karst inventories are to:

- stratify the sensitivity or vulnerability of the karst landscape;
- identify major surface karst features; and
- provide a preliminary delineation of the karst catchments.

To date, only two planning-level inventory projects have been completed in British Columbia, both on northern Vancouver Island.[9] In both cases, considerable bedrock mapping was required in the field to verify the extent and boundaries of the karst units. Geographic Information System (GIS) analysis was also used to rapidly identify areas of differing karst vulnerability potential by using the attributes of elevation and slope gradient.

Operational-level inventories primarily involve the use of Karst Field Assessments (KFAs) carried out at 1:5000 or 1:10 000 scales. A KFA is a detailed evaluation of the karst attributes and features in the cutblock area. It covers not only the area of potential or suspected karst units within a proposed cutblock, but also other areas outside the cutblock. The process for completing operational-scale inventories is discussed below and is outlined in the *Karst Inventory Standards and Vulnerability Assessment Procedures for British Columbia* (B.C. Ministry of Forests 2003a).

Cave inspections can be done as part of a KFA, and in some cases more detailed inventories may be required to determine a cave's significance and vulnerability to disturbance. These should be undertaken only by qualified personnel (Kiernan 1988; Ramsey 2004).

At the cutblock level, KFAs may include karst areas 100 m beyond the cutblock boundaries, as well as reaches of sinking streams or watercourses outside of the cutblock, depending on various circumstances (see B.C. Ministry of Forests 2003a). Typical KFA field activities (Figure 11.18) include:

- identifying geological contacts and inferring or delineating the extent of potential karst units;
- locating, classifying, and evaluating surface karst features for relative significance;
- evaluating attributes such as the level of epikarst development, soil thickness and texture, density of surface karst features, roughness of the karst surface, and subsurface karst potential;
- assessing streams to see whether they sink or lose water into the subsurface;
- inspecting and mapping caves;
- identifying unique or unusual flora and fauna and (or) habitats; and
- identifying potential geomorphic hazards that could affect the karst unit (landslides, windthrow, etc.).

The data collected during a KFA can be used to broadly stratify the karst unit of interest into polygons of similar karst characteristics, which are then rated for vulnerability as low, moderate, high, or very high. Karst vulnerability is defined as the susceptibility of a karst ecosystem to change. Karst vulnerability ratings are determined by a four-step procedure (Figure 11.19) and are used to guide the selection of appropriate best management practices as outlined in the *Karst Management Handbook for British Columbia* (B.C. Ministry of Forests 2003b).

9 These projects are: (1) Stokes, T.R. and P.A. Griffiths. 2002. Planning-level karst inventory of TFL 19, Nootka Region, B.C. Prepared for Western Forest Products. Unpubl. report; and (2) Stokes, T.R. and P.G. Griffiths. 2003. Planning-level karst inventory for TFL 37 and FL A19233, northern Vancouver Island. Prepared for Canadian Forest Products. Unpubl. report.

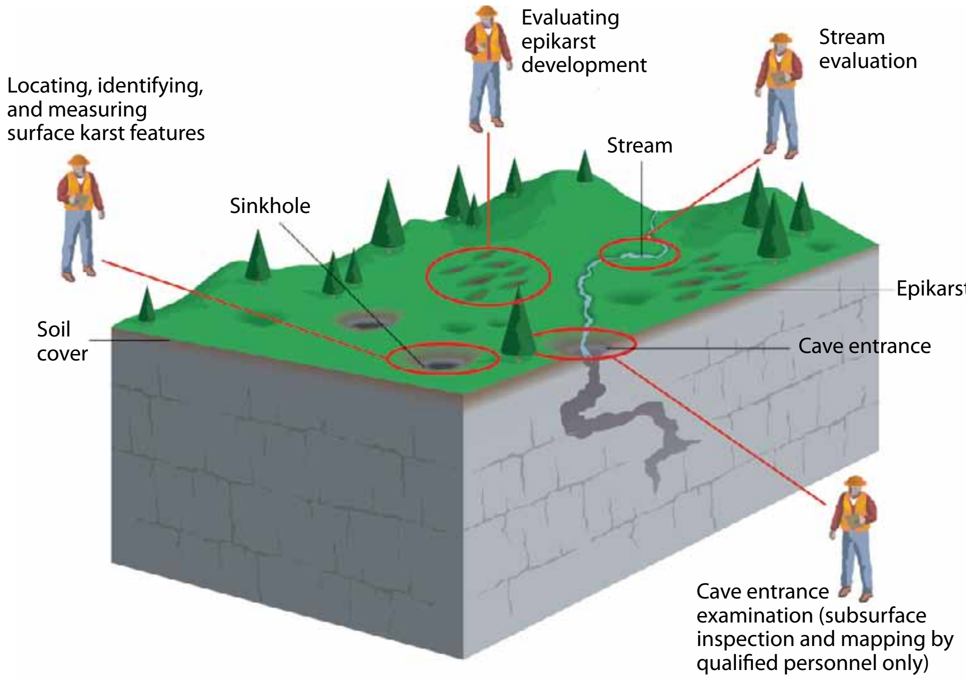


FIGURE 11.18 *Karst field assessment activities (B.C. Ministry of Forests 2003a).*

Karst Management and Best Management Practices for Forestry Activities on Karst Landscapes

The B.C. Ministry of Forests first acknowledged karst landscapes as complex ecosystems in 1997 (B.C. Ministry of Forests 1997; Beedle 1997), recognizing that management efforts should focus on protecting the integrity of the whole karst system rather than individual karst features and caves. This new approach to managing karst resources was embodied within a series of significant government initiatives and associated publications (i.e., Stokes and Griffiths 2000; B.C. Ministry of Forests 2003a, 2003b). These documents are available on the B.C. Ministry of Forests and Range website at www.for.gov.bc.ca/hfp/values/features/karst/index.htm.

A Preliminary Discussion of Karst Inventory Systems and Principles for British Columbia (Stokes and Griffiths 2000) proposes a scientific framework for a standardized inventory system for karst landscapes in British Columbia. The *Karst Inventory Standards and Vulnerability Assessment Procedures for British Columbia* (B.C. Ministry of Forests 2003a) outlines standards and procedures for evaluating and inventorying karst landscapes at various scales. The first version of this document was published in 2001.

The *Karst Management Handbook for British Columbia* (B.C. Ministry of Forests 2003b) provides best management practices for harvesting and forest road construction where these forestry operations impinge upon:

- surface karst features (e.g., sinkholes, karst springs, epikarst exposures, cave entrances);
- cave systems;
- sinking streams and sinking watercourses; and
- the broader karst landscape.

Some specific examples of best management practices include:

- developing reserve and management areas for more significant surface karst features and above shallow caves or caves with exceptional features;
- realigning roads and carefully designing road drainage systems to avoid sinkholes;
- using overlanding[10] road construction techniques with coarse rock fill to bridge small-scale karst

10 Overlanding is a construction technique whereby fill is imported to build the road up to a level grade rather than using conventional cut and fill.

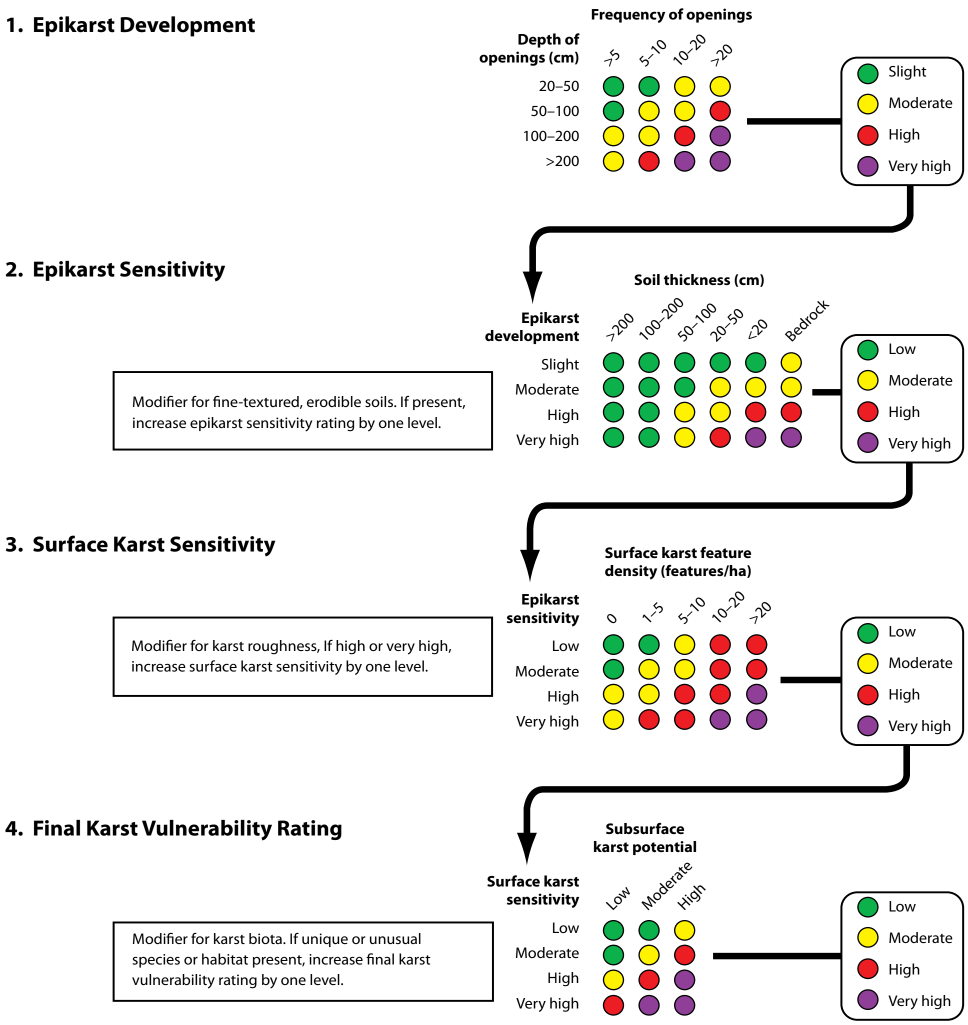


FIGURE 11.19 *The four-step karst vulnerability rating system (B.C. Ministry of Forests 2003a).*

features such as grikes or epikarst zones; and

- developing buffers along the edges of sinking streams that contribute to significant recipient features.

The best management practices for the broader karst landscape are linked incrementally to the four karst vulnerability ratings, such that the higher the level of karst vulnerability the more numerous and comprehensive the management practices (see Figure 11.20).

Case Studies of Karst Management Practices on Forested Cutblocks

To illustrate how inventories, best management practices, and harvesting or road construction activities are completed on cutblocks underlain by karst in the province, we provide three case studies from Vancouver Island.[11]

Case study 1: Large sinkholes and small caves

Site conditions This rectangular 17-ha cutblock is located near a ridge top in an old-growth forest stand. The northern half of the cutblock is underlain by Quatsino Formation limestone. The cutblock is at an elevation of approximately 200–300 m above sea level. A road was built to access the cutblock from the north. The landscape is characterized by gentle to moderate rolling topography with occasional limestone outcrops and interrupted drainage linears. Glacial sediments cover much of the cutblock area.

11 For expediency and to allow comments and opinions on the results of harvesting and road construction activities, we have excluded the location of the cutblocks and names of licensees.

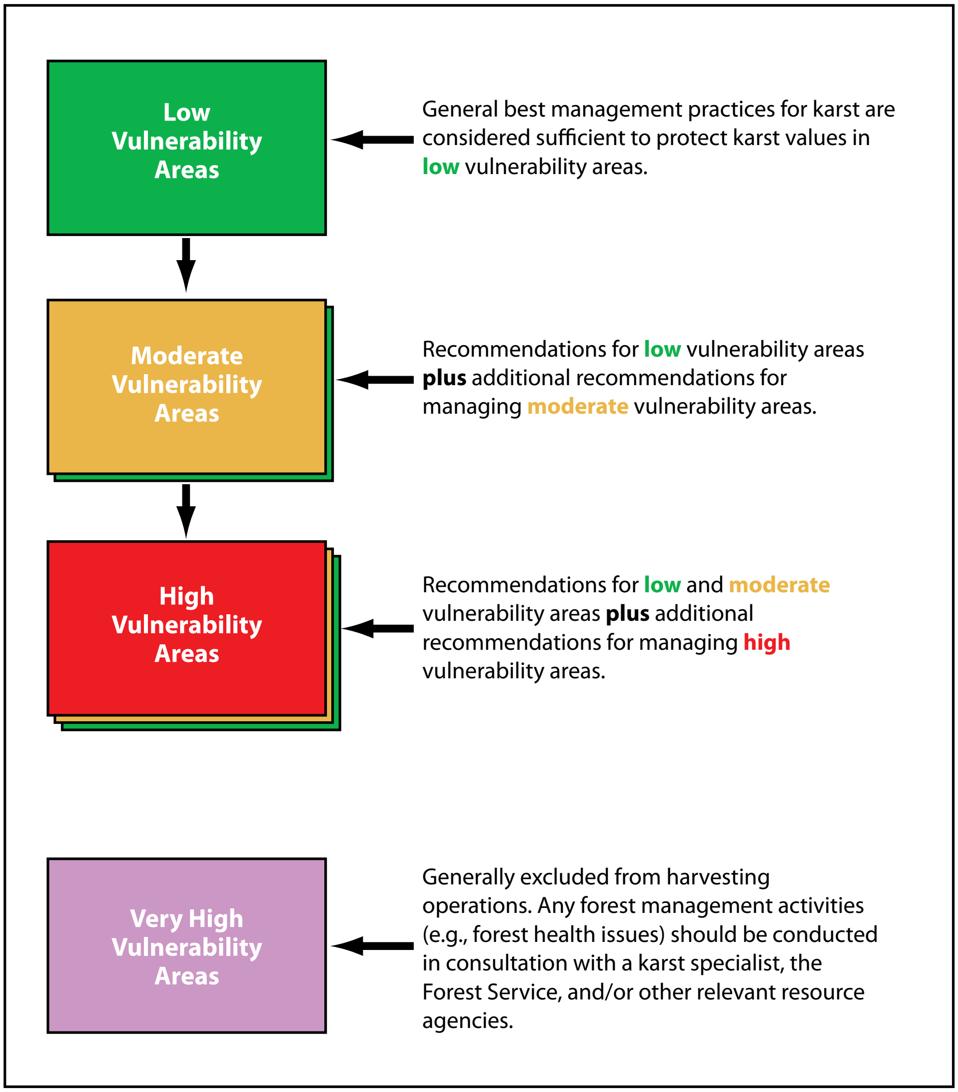


FIGURE 11.20 *Framework for best management practices and karst vulnerability ratings for broader karst landscapes (B.C. Ministry of Forests 2003b).*

Karst attributes A detailed KFA was carried out for this cutblock during 2005. The greater portion of the unit received a moderate karst vulnerability rating, with localized areas of high vulnerability. Approximately 10 karst features were encountered within or adjacent to the cutblock. These features consisted primarily of sinkholes of various sizes. Two larger sinkholes with cave entrances were found along the proposed access road leading to the cutblock, and two other large (15–20 m diameter) sinkholes were found within the cutblock near to the proposed road (see Figure 11.21).

Pre-harvest recommendations It was recommended that harvesting and road construction practices over the broader karst landscape follow those outlined for karst areas with moderate karst vulnerability ratings. For the two sinkholes with cave entrances and associated shallow caves situated outside of the cutblock, minor changes to the alignment of the access road were recommended to avoid these features. For the two larger, significant sinkholes within the cutblock, the preferred option was to develop suitable windfirm reserves with surrounding management areas to protect the structural, functional, and ecological integrity of these features. Development of these reserves and surrounding management areas would require some adjustment of the road alignment within the cutblock.

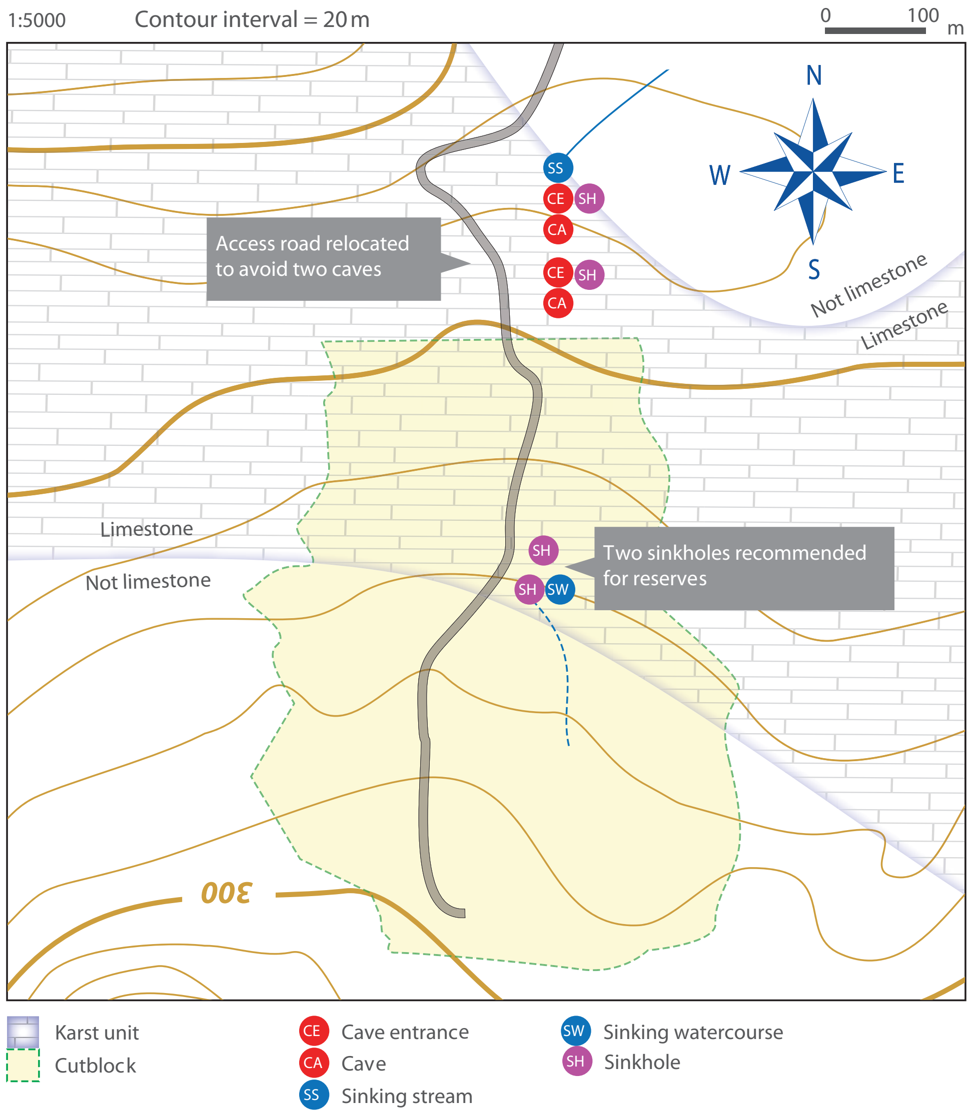


FIGURE 11.21 *Case study 1: Cutblock with large sinkholes and small caves along access road and potentially significant sinkholes within cutblock (P. Griffiths).*

Results of harvesting and road construction A follow-up visit to the site confirmed that the alignment of the access road to the cutblock had been adjusted to avoid the sinkholes and associated cave entrances. Within the cutblock, however, no reserves or management areas had been left around the larger sinkholes. Falling and yarding away from these sinkholes had been carried out with no notable disturbance to these features' sidewalls. From subsequent discussions with the licensee, it was apparent that windthrow hazard near the proposed reserves and management areas around the two sinkholes was considered high, and that the licensee was unable to realign the road at this location.

Post-harvest issues for further consideration Although management of the larger sinkholes outside the cutblock appeared to have been successfully carried out by realigning the access road, tree retention around the two larger sinkholes within the cutblock had not occurred. In hindsight, it would have more efficient and effective to have completed a KFA before cutblock and road design so that the karst issues could be more readily addressed and alternatives for forest development considered more thoroughly.

Case study 2: Karst drainage linears

Site conditions A 7-ha cutblock is located on east-facing slopes at elevations of 100–150 m and is within a well-developed second-growth stand. Two well-defined karst drainage linears trending east to west occur immediately outside of the cutblock boundaries to the north and south, respectively.

Karst attributes A detailed KFA was not carried out for this cutblock; however, a field review of the cutblock and associated report were completed in 2007. The two karst drainage linears were both associated with interrupted streams, as well as a number sink points, springs, and small caves (Figure 11.22).

Approximately 1–1.5 km downslope from the cutblock, a number of other springs occur and a nearby homeowner uses one of these springs as a domestic water supply. A cluster of smaller sinkholes was encountered within the cutblock and along the proposed road alignment.

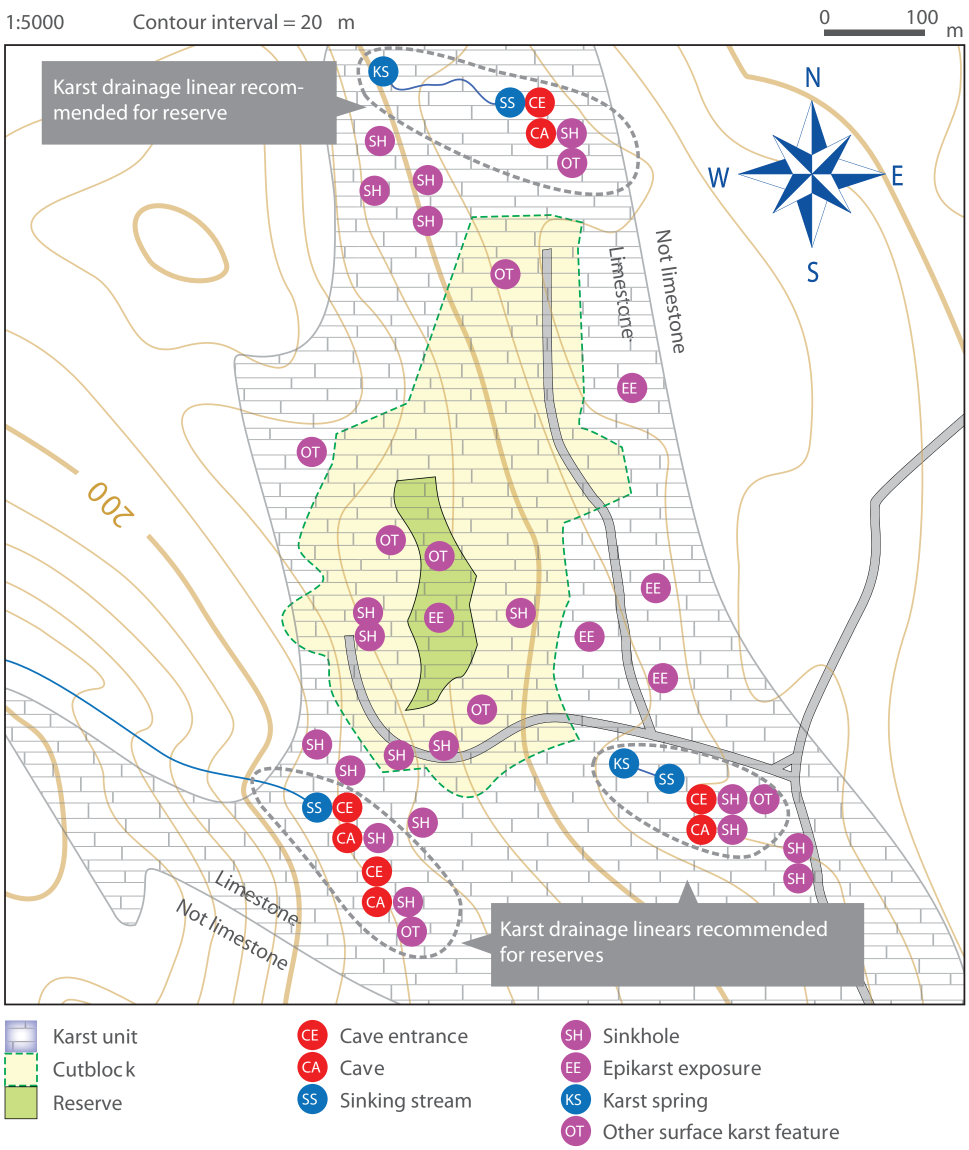


FIGURE 11.22 *Case study 2: Cutblock between two major and significant drainage elements (P. Griffiths).*

Pre-harvest recommendation Both drainage linears were identified as sinking streams with significant recipient features that warranted reserves. The recommended minimum buffers were 20 m from the drainage linears, and up to 80 m for some of the significant karst features associated with the linears. It was also recommended that road builders avoid a group of smaller sinkholes.

Results of harvesting and road construction Harvesting and road construction were carried out in 2008. Reserves were left around the drainage linears and associated significant karst features as recommended. Road construction did encroach on a number of the smaller sinkholes, some of which were infilled with road ballast and surfacing material. Significant sandy fines were also exposed along road cuts in proximity to karst features and drainage linears.

Post-harvest issues for further consideration Retaining buffers along the drainage linears was considered important in this cutblock, particularly as possible hydrological connections existed to springs at lower elevations, one of which is used as a domestic water supply. The significant karst features associated with the drainage linears were adequately buffered, although windthrow along the buffer edges may be a future concern. Therefore, windfirming these boundaries may warrant consideration. A need also exists to ensure that fine sediment exposed along the roads does not enter the karst drainage linears. Infilling the smaller sinkholes is not a good practice as this will obviously affect the hydrology and function of the sinkholes. Knowing more about the water subsurface connections and flow paths between sink points and springs at this block is likely the key issue, particularly as a spring is used as a domestic water supply. A good pre-harvest option would have been the completion of a dye tracing study.

Case study 3: Sinkhole clusters and surface streams
Site Conditions: A 40-ha cutblock is located on a south-facing slope at an elevation of about 600 m above sea level and has slope gradients of 20% or less. The cutblock is blanketed by thick glacial sediments (mainly weathered till) and has dense stands of small, second-growth trees. The entire cutblock is underlain by karst but with only one limestone outcrop exposed. Small surface streams drain across the cutblock and flow into a small wetland, which in turn flows into a larger stream that eventually crosses other possible karst areas some distance downslope to the north.

Karst attributes A detailed KFA was carried out in 2007. Numerous sinkholes were found throughout this cutblock and appear to occur in clusters, most of which were aligned along a linear band extending across the cutblock. Some of the sinkholes were enclosed within broad and shallow depressions. A number of large (> 15 m diameter) and potentially significant sinkholes were present, some of them possibly large enough to sustain microclimates.[12] Most sinkholes were infilled with 40- to 50-year-old woody debris from previous logging. Some of the streams crossing the cutblock were considered permanent, whereas others were intermittent. Two minor sink points and two possible small springs were identified along or near these streams. An area with significant caves is located approximately 1 km to the northeast of the cutblock, but this area does not appear to have any direct hydrological connection to the cutblock or to the stream draining the cutblock.

Pre-harvest recommendations The large sinkholes and areas of closely spaced sinkholes were grouped into four clusters and were recommended as retention areas, with the retention boundaries to be located 15–20 m away from the rims of the sinkholes or shallow depressions enclosing the sinkholes (Figure 11.23). Windfirm treatment of the retention boundaries was recommended, if the licensee considered the site to have a risk of windthrow. Careful harvesting of the smaller isolated sinkholes was considered acceptable. Minimizing the input of fine sediment and logging debris was recommended for the two larger streams in the cutblock, particularly as these are connected to a swamp and larger stream that flows toward other potential karst areas to the north, of which little is known.

Results of harvesting and road construction No harvesting has been carried out as yet.

12 Large karst sinkholes with microclimates include those that exhibit a distinctive temperature and relative humidity gradient from the rim to base along the sideslopes, and may include higher biodiversity or habitat values (B.C. Ministry of Forests 2003b).

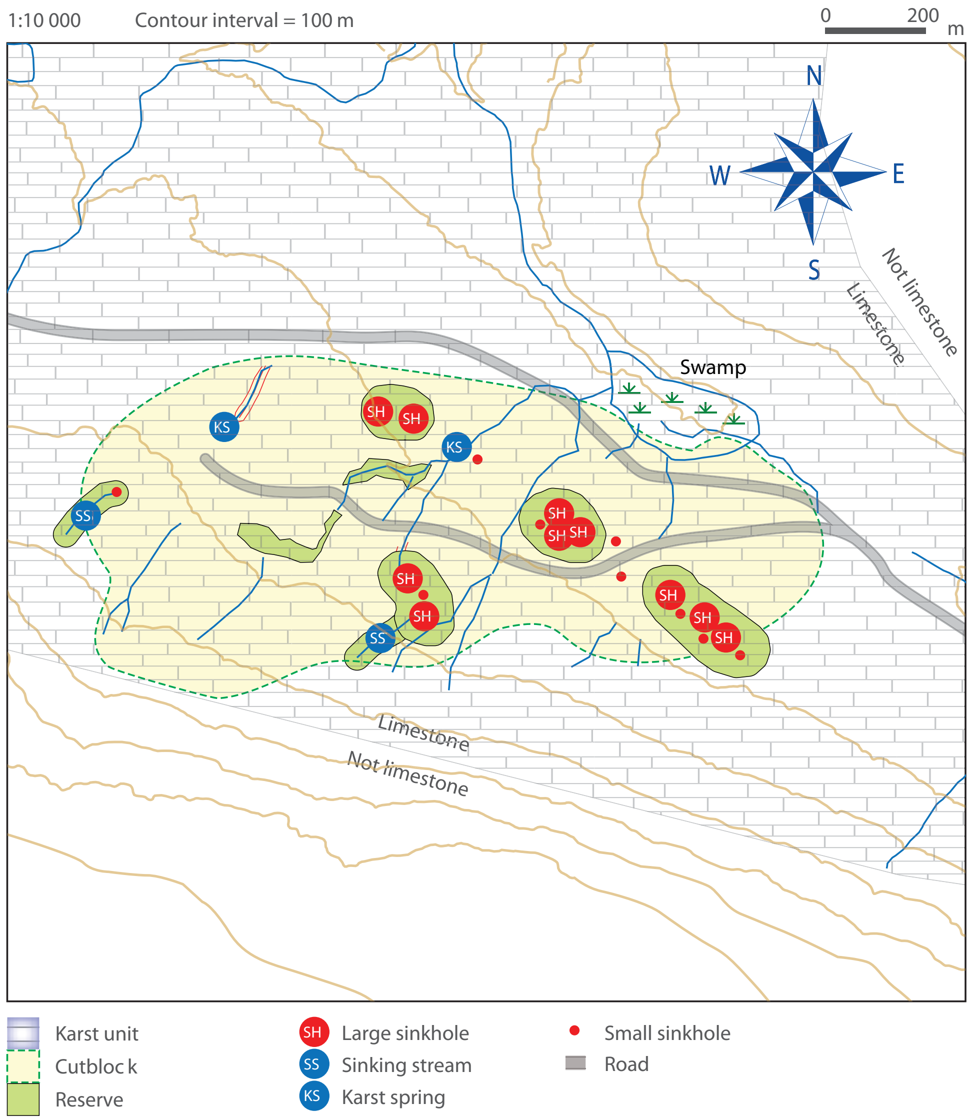


FIGURE 11.23 *Case study 3: Cutblock with clusters of sinkholes and development of retention areas (T. Stokes and P. Griffiths).*

Other issues for consideration The retention areas should be of sufficient size to retain the structure, function, and ecological integrity of many of the identified sinkholes. Though the significant caves to the northeast do not appear to have an obvious hydrological connection to the cutblock area, other karst features of concern (e.g., sink points with cave entrances) may exist along the larger stream to the north; therefore, significant care should be taken when harvesting in and around the streams within the cutblock to limit sediment and debris input.

British Columbia has a wide range of karst landscapes, many of which are forested and occur in areas with ongoing forestry activities. Consequently, a real need exists to manage the province's forested karst resources carefully with the understanding that karst landscapes function as integrated systems and are valuable resources for biodiversity, water, scientific research, and recreation.

Since 2004, when the *Forest and Range Practices Act* (FRPA) was adopted in British Columbia, karst has become a subset of FRPA's "resource features" value—one of 11 specified forest and environmental values that must be maintained. Under FRPA's Government Actions Regulation (GAR), the surface and subsurface elements of a karst system can now be legally established by government orders as resource features by type or category (e.g., caves, significant surface karst features, high or very high vulnerability karst terrain) and may be restricted to a specified geographic location (see Griffiths et al. 2005). Currently, six GAR orders have been established for karst in six of British Columbia's coastal forest districts—North Island–Central Coast, South Island, Campbell River, Chilliwack, Sunshine Coast, and Haida Gwaii.

The Forest Resource Evaluation Program (FREP) was established in 2003 to assess the effectiveness of FRPA in meeting government's objectives for each of the 11 resource values. The objectives of the FREP karst program are to evaluate the condition of karst resource features and the adjacent management areas following forestry activities of harvesting and road construction, and to determine whether FRPA standards and practices have achieved the desired level of protection for these features. The development of a checklist and detailed protocol for Karst Routine Evaluations has been initiated (see www.for.gov.bc.ca/hfp/frep/indicators/table.htm).

The Earth Science Department of Vancouver Island University is currently carrying out several research projects to investigate the characteristics of karst landscapes in the forested environment. Projects include: examining the microclimate of large sinkholes,[13] investigating soil development and hydrological processes in and around sinkholes, and evaluating and monitoring karst springs and the associated recharge areas and aquifers (Stokes et al. 2008).

An important first step in karst management is the completion of a careful inventory and evaluation, without which it is not possible to consider suitable management strategies or apply best management practices. The existing *Karst Inventory Standards and Vulnerability Assessment Procedures for British Columbia* (B.C. Ministry of Forests 2003a) and *Karst Management Handbook for British Columbia* (B.C. Ministry of Forests 2003b) provide guidance for completing karst inventories and carrying out karst management using best management practices. However, standards in forested karst management are evolving (e.g., ecosystem-based karst management; see Griffiths and Ramsey 2009) and therefore these publications are becoming outdated. Revisions to these documents should incorporate the important lessons learned in karst management over the last decade.

13 Stokes, T.R., P. Griffiths, and C. Ramsey. 2007. Preliminary microclimate study of forested karst sinkholes, Nimpkish River Area, Northern Vancouver Island, British Columbia, Canada. Poster presented at 17th Australasian Conference on Cave and Karst Management, Buchan, Victoria.

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Click here to go to Compendium Volume 2 index for pages 401–790

A

B

Click here to go to Compendium Volume 2 index for pages 401–790

C

Click here to go to Compendium Volume 2 index for pages 401–790

Click here to go to Compendium Volume 2 index for pages 401–790

G

Click here to go to Compendium Volume 2 index for pages 401–790

H

I

J

Click here to go to Compendium Volume 2 index for pages 401–790

K

L

M

Click here to go to Compendium Volume 2 index for pages 401–790

Click here to go to Compendium Volume 2 index for pages 401–790

Q

R

Click here to go to Compendium Volume 2 index for pages 401–790

S

Click here to go to Compendium Volume 2 index for pages 401–790

T

U

Click here to go to Compendium Volume 2 index for pages 401–790

V

W

Click here to go to Compendium Volume 2 index for pages 401–790

Y