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APPLICATIONS OF GEOMORPHOLOGY
FOR INLAND WATERS DIRECTORATE STAFF:
WESTERN AND NORTHERN REGION

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Applications of Geomorphology for Inland Waters Directorate Staff: Western and Northern Region

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Abstract

This report provides a primer on geomorphology and river processes relevant to the environmental monitoring and assessment activities of Inland Waters Directorate of Environment Canada. An earth history of Canada is presented with an emphasis on the history of the Quaternary, the character and distribution of surficial materials laid down in Canada in the Pleistocene, and the nature of post-glacial processes acting on the landscape. The relevance of the geomorphic context to contemporary river process description and explanation is illustrated. Implications for sediment regimes, water quality and water use are described and the context for environmental impact assessment of channel developments, and planning of river monitoring sampling sites, are provided. Specific applications of the approach in basins and rivers of the Western and Northern Region of Canada are described.

Key Words: geomorphology, river processes, environmental monitoring, environmental impact assessment, Canada.

Résumé

Ce rapport constitue une introduction à la géomorphologie et aux processus fluviatiles dans le contexte des activités de surveillance de l'environnement et d'évaluation environnementale de la Direction générale des eaux intérieures d'Environnement Canada. L'histoire géologique du Canada y est présentée, avec une attention particulière portée à l'historique du quaternaire, aux caractéristiques et à la distribution des matériaux de surface qui se sont déposés au Canada durant le pléictocène, et à la nature des processus postglaciaires modifiant le paysage. On illustre le lien existant entre le contexte géomorphologique et la description et l'explication du processus fluviatile contemporain. De même, on y décrit les conséquences pour les régimes sédimentaires, ainsi que pour la qualité et la consommation de l'eau et l'on fournit le context nécessaire à l'évaluation des répercussions environnementales causées par des aménagements sur le cours d'une rivière et à la planification de sites pilotes estinés à la surveillance de rivières. On présente les applications particulières de cette méthode sur les bassins et les rivières de l'Ouest et du Nord du Canada.

Mots-clés: géomorphologie, processus fluviatile, surveillance de l'environnement, évaluation environnementale, Canada.p

PREFACE

This report provides a published document for staff of the Inland Waters Directorate (Western and Northern region) of Environment Canada, dealing with material previously presented to them in a two-day lecture course.

The purpose of that lecture course was to introduce IWD staff to the subject matter of geomorphology and to the approaches taken by geomorphologists in landscape investigations, and to show how this material could assist them in their own work dealing with surface water hydrological issues, e.g., identification of sediment sources in drainage basins, inter-relationships between water quality and sediment quality, planning the location of river sampling sites, interpretation of processes taking place in present-day river channels, and environmental impact assessment of channel development.

Evaluation of the course by its participants indicated a certain degree of "information overload" which, it was thought, might be alleviated by publishing the course material as a permanent reference document. This is the purpose of the present report.

The Geomorphology lecture course was conceived as a pilot initiative in the Surface Water Training Program of the Inland Waters Directorate, under the leadership of Dr T.J. Day, formerly Head of the Sediment Survey Section, Ottawa, and now Regional Chief, Water Resources Branch (Manitoba) in Winnipeg. It is proposed to follow up this pilot initiative with other introductory courses such as Hydrology, Remote Sensing, Statistics, and more advanced areas of these topics (such as river processes and sediment transport in relation to the Geomorphology course).

The course was given at centres throughout the country from Vancouver, British Columbia to Dartmouth, Nova Scotia. The framework for the course was the same at all centres, but regional examples were selected from each local area to indicate the application of geomorphology to IWD work. The present report is written exclusively for the Western & Northern Region and deals with the material covered in the Calgary, Yellowknife, Regina, Saskatoon and Winnipeg courses.

In the preparation and delivery of the course I have received assistance from numerous individuals which is gratefully acknowledged. In particular, I am indebted to Terry Day for his professional and personal perspective throughout the entire period, to Bob Fulton of the Geological Survey of Canada who made available a copy of "Quaternary Geology of Canada and Greenland" many months before its official publication in late 1989, and to Henry Hudson for management of the present contract and thorough review of the text.

Many of the figures included in the report are taken directly from texts acknowledged in their captions. Other figures are based on my own work, much of it undertaken in the last five years as part of contracts negotiated through Supply and Services Canada with different departments of IWD across the country. The text was formatted and prepared for publication by Water Resources Branch, Winnipeg.

Lastly, and certainly not least, I thank the participants at the various centres for their contribution, criticisms and suggestions.

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30 March, 1991

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1. INTRODUCTION

1.1 Purpose of report

Issues concerning surface water hydrology in Canada have traditionally been related largely to engineering problems, e.g. flood forecasting, river diversions, pipeline and bridge crossings, and sedimentation of reservoirs. To a large extent, the staff of the Inland Waters Directorate, especially the Water Resources Branch and the Water Planning and Management Branch, has its training primarily in engineering, whether this be at the planning level of office staff or at the technical level of field staff. Issues investigated by the Water Quality Branch have similarly been the domain of its own specialists - biologists, chemists and biochemists.

The environmental issues that have emerged in the 1970's and 1980's have shown no distinct subject matter boundaries, however, and involve linkages between many parts of the natural landscape. The enlargement of South Indian Lake by the impoundment of the Churchill River in Manitoba, together with the resultant acceleration of shoreline erosion in permafrosted silty-clay sediments, the increase in nutrient levels associated with this shoreline flooding, and the release and conversion of mercury into forms toxic to aquatic life, is but one example, among many, of this multifaceted character of surface water issues (CMMA, 1987).

With this perspective it might be argued that such specialists need at least some training in matters related to natural processes taking place in the landscape, and in particular in disciplines which show strong linkages to many different parts of environmental science. One such subject is geomorphology.

The nature of geomorphology is defined more carefully in the next section (1.2). It is sufficient here to note that it is usually regarded as part of the subject matter of both geology and physical geography. In Canadian universities it is generally taught in both departments. In scope, the present text could be regarded as a primer in geomorphology, equivalent to what would be taught in a first year course. Two important differences should be emphasized though: the present text does not cover the entire field of geomorphology, only those parts which, indirectly or directly, might be related to the work of IWD staff (and space and time preclude the coverage of all of this); and, secondly, emphasis throughout the text is directed to illustrating the application of geomorphology to IWD work, rather than spending time on more theoretical aspects of the subject.

This text therefore has two main purposes: to provide an awareness of the nature and usefulness of geomorphology to other earth scientists and civil engineers; and to provide a discussion of those specific aspects of geomorphology that directly bear upon the work of the Inland Waters Directorate. These two goals are not entirely separate: the best way to appreciate the geomorphologist's "perspective" is through specific examples.

1.2 The nature and spirit of geomorphology

The usual definition of geomorphology is the scientific study of landforms (e.g. river valleys, coastal beaches, etc.) and landscapes, where the term landscape refers to a suite of individual landforms. The term "glaciated landscape", for example, is used to refer to a host of individual glaciated landforms (e.g. eskers, drumlins, etc.) juxtaposed in a systematic manner over an area of the earth's surface.

There are many different types of landscapes on the earth's surface, and hence many different specialized fields of study in geomorphology. These are evident in the chapter headings of most textbooks on the subject, e.g. Earth's Changing Surface (Selby, 1985). Such landforms and landscapes tend to be subdivided into two main groups: those created by the internal energy of the earth, e.g. volcanoes and rift valleys; and those moulded by external energy at the earth's surface, e.g. processes of rock breakdown and sediment transport by rivers, wind, glaciers, etc. Emphasis in this text is directed to the latter group of processes and landscapes, because these have much more relevance to surface water hydrology.

It should be noted that the term "study", used in the definition above, means not only the description of the landforms, but also the explanation of the origin of the landforms in terms of the processes that created them. As an example, in the area of fluvial (river) geomorphology, a great deal of effort has gone into describing the form of river meanders, and quantifying their shapes in terms of wavelength, amplitude, sinuousity, etc. (Fig. 1.1), and into noting how these aspects of meander morphology differ among different river types. But equally important has been the contribution by geomorphologists to discerning the processes involved, e.g., why some rivers meander and others do not, why some meanders migrate down valley through bank erosion and others stay fixed in place, and, more generally, what causes deposition in one part of a bend and bank erosion in another.

With this perspective it is easy to see that there is considerable overlap between this "process" aspect of geomorphology and part of the subject of civil engineering. Indeed, often the only obvious difference is one of time scale. The engineer, for example, may look at meander processes in a laboratory flume for a few months, or in the field for a few years. The geomorphologist, on the other hand, is more likely to be interested in how a meandering river has evolved over a period of decades (using aerial photographs) or even centuries (by looking at the floodplain itself).

Yet the geomorphologist may also work at the same timescale as the engineer. Several meander studies by geomorphologists have involved short field seasons of intensive work such as measuring current speed and direction, sampling suspended sediment in different parts of the river channel, etc. But the goal is different. The geomorphogist wants to know what processes are at work, not to solve a 30-year problem, but to understand how river systems develop over centuries and millennia. In searching for that understanding, the geomorphologist is, nonetheless, often able to bring forward knowledge that will assist the engineer in short-term problems.

Having noted this apparent overlap between parts of geomorphology and parts of civil engineering, there is a fundamental difference in perspective between professionals in the two disciplines. This is the point of the next section.

1.3 Importance of geomorphology to civil engineering

Though the work of geomorphologists and civil engineers often overlaps in the study of landscape processes, it must be emphasized that geomorphologists are concerned not only with the processes that mould landforms, but also with the landforms themselves. An important component of the training of geomorpholgists, therefore, is in the recognition of landforms both directly in the field and indirectly from aerial photographs of the landscape. In particular, the geomorphologist is trained to look at the landscape and ask key questions. Why is the land like this? How was this feature formed? When was it formed? And, why was it formed over there and not here?

This general inquisitiveness regarding the landscape is the fundamental basis of the geomorpologist's perspective. Civil engineers, by the nature of their training, do not have this particular curiosity, nor instruction in the recognition of different types of landform. This is not a criticism of the civil engineer. The training of geomorphologists is also incomplete, to the extent that (unlike the engineer) they have usually had little instruction in applied physics. Yet an understanding of the stress distribution in any body, whether it be a glacier, a river or a hillside slope, is crucial to understanding how that body will behave. No scientist can be trained in all fields. What is important, however, is that they have an appreciation of related fields.

In view of the fact that a large portion of middle management in IWD staff has a training in civil engineering, and little familiarity with geomorphology, it seems important to emphasize this difference in perspective between the engineer and geomorphologist. The reason is quite simple. There will be times when engineers will need to call upon the assistance of the geomorphologist. Before they can do that, they need to know something of what geomorphology is all about, and how the geomorphological approach to landscape issues differs from his own approach.

The best examples to illustrate the importance of geomorphology in landscape issues come from past case studies where mistakes were made by not including a geomorphological perspective. The example used here refers to a landslide near Rigaud in the Ottawa Valley of southern Quebec in the late 1970s. Though the example falls beyond the geographic area of the Western & Northern region, it might be noted that similar slides do occur in parts of the Prairies. In fact, one occurred in 1990 near Rycroft in the Peace River lowlands of Alberta.

An oblique aerial photograph of the Rigaud landslide is given in Fig. 1.2. It occurred on the valley wall of a small creek that had cut down about 15 m into a deposit of clay laid down about 10,000 years ago, at a time when the area was part of a much larger Gulf of St. Lawrence (discussed in Chapter 3). As indicated in the photograph, the slide occurred at the site of a pylon base being built for a transmission line for electricity from James Bay to the United States. The pylons are built upon a set of eight friction piles driven deep into the clay, two at each of four corners. At the time of the slide, the crane operator was driving the third pile down as illustrated in Figure 1.3. The valley side then simply moved en bloc across the valley floor, and was followed in succession by subsidence of four wedge-shaped blocks, which assisted in driving the initial valley-side block across the valley floor. Inside the cavity formed by the subsidence of these blocks, the landmass was broken up into strips (corresponding to the tops of sunken wedges and intervening ridges) running parallel to the backscarp (Fig. 1.4). All this happened in about 90 seconds.

Several scientists, along with engineers from the agency responsible for the route selection of the transmission line and for the pylon construction procedures, visited the site the next day, and the dialogue between the two sets of professionals highlights the difference in perspective afforded by their training. The engineers maintained that the occurrence of this slide could not have been predicted; that, although similar slides were known from other parts of the Ottawa lowlands, there was no evidence to indicate that such a slide could have occurred near this site. The geomorphologists who were present were puzzled by this comment because, it was clear to them that, in walking from the road to the north side of the creek's valley (Fig. 1.2) to obtain a vantage of the slide, they had in fact been walking across fields laid out on the floor of a similar, but larger, slide, that had occurred several thousand years earlier. The last few steps climbing up to obtain a vantage point for the 1970s slide had in fact taken place up the backscarp of this much older slide. This was apparently not evident to the engineers who had not recognized the undulations in the fields more or less parallel to this old backscarp. To the untrained eye it must be admitted that the floor of this old slide looked much like a terrace (an old floodplain) on the side of the larger river to which the small creek was tributary.

This is the important point: the difference between the untrained eye and the one instructed in the recognition of such features. In areas where previous slides of this type have occurred in the recent past, their morphological distinctiveness is recognizable with little difficulty (Fig. 1.5), but after a few thousand years, the morphological attributes are no longer fresh enough to be recognizable to most people. Clearly, then, having an eye for the landscape can be a tremendous asset in landscape planning, and this point is increasingly being recognized. In Quebec, for example, it is now not unusual to have civil engineering firms hire geomorphologists to do aerial photograph interpretation of transmission line routes, and similar projects, so that a terrain map is produced before the field engineering work is done.

A similar example in the Prairies was provided by Mollard (1986) in connection with the investigation of potential damsites along the South Saskatchewan River in the 1950's. In that case the important task was to determine, at such sites, whether the hummocky terrain found along some valley slopes represented morainal debris from glacial deposition (as some supposed) or whether it was the morphology of former landslides, which could still be moving, albeit imperceptibly. The importance of this task in the selection of dam sites on the South Saskatchewan River hardly needs additional comment.

The previous examples have only limited relevance to IWD work, though the recognition of deposits from massive landslides that impinge upon channel courses and their floodplains might well be important in some parts of Water Planning and Management (WPM) work. The next section turns to examples that are more directly related to IWD work.

1.4 Importance of geomorphology to IWD work

A key aspect of Water Resources Branch (WRB) work is selection of appropriate sites for hydrometric and sediment sampling stations. The specific requirements of such sites for hydrometric purposes are well-known to WRB staff, e.g. a section control that changes little over time so that only few measurements are needed to obtain an accurate stage-discharge relationship. As a result, great care is taken in the selection of sites for hydrometric stations.

WRB sediment sampling sites and Water Quality Branch (WQB) water sampling sites are often located at these hydrometric stations so that discharge information is readily available. Sediment concentrations at these sites can thus be converted to sediment loadings through multiplication by water discharge and the appropriate constant to convert units:

$$L = 0.0864 \times Q \times c$$

where L is the sediment load (tonnes per day), Q is discharge (m^3/s) and c is concentration (mg/L). Availability of discharge data similarly allows dissolved concentrations (e.g. nutrients) to be converted into chemical loadings.

What is sometimes not realized is that sediment sampling sites can also have exacting requirements, and that to simply add on a sediment sampling program to an existing hydrometric station, without any assessment of the station's suitability from the standpoint of sediment sampling (as has often been done in the past), can produce major problems at a later date when someone has to interpret the data. The same problem can, to some extent, arise in the context of water quality sampling. The point being emphasized here (and which will be demonstrated in several places in this report) is that an understanding of geomorphology can assist significantly in the proper location of sediment and water quality sampling stations, and in the interpretation of data from such stations. An example is provided below.

A review was recently undertaken of the sediment data for WRB stations in the lower Athabasca River basin of northern Alberta (Fig. 1.6). Most of the stations were on tributaries of the Athabasca River. They were established in the 1970s to collect background data on sediment conditions in typical streams in the oilsands mining region, before major development took place. Clearly not all streams could have stations installed, and so sediment work was done on a sample of rivers; the ultimate intent was to allow inferences about other small basins in the region through interpolation of data from those basins that had been sampled.

Analysis of the sediment data showed that three of the sites on the tributary streams had high average concentrations of suspended sediment (using annual sediment load divided by annual runoff): these were the Ells, Hangingstone and Joslyn sites, with mean concentrations in the range 300-400 mg/L. In contrast, three sites had very low average concentrations: the Muskeg and Hartley (about 10 mg/L) and the Firebag River with less than 40 mg/L. The data seem to suggest, therefore, a distinction between left bank rivers (high sediment levels) and right bank rivers (low sediment levels). Such a conclusion would have been, to a large extent, wrong, however, because the results primarily reflect whereabouts on the river the sediment station was located, rather than differences between the rivers.

The important point in the interpretation of these data is that sediment is supplied to these streams primarily through bank erosion. The Athabasca River mainstem cut down deeply in late-glacial and post-glacial time (for reasons which will become evident in later chapters); this means that its tributaries, also, have cut down at their confluence with the mainstem; and this tributary downcutting has been slowly working its way headwards up the tributary courses during post-glacial time. Fig. 1.7, for example, depicts the long profile of the Hangingstone River (tributary to the Clearwater River) and shows the distinct break of slope near to Prairie Creek between the oversteepened lower reach (where the river has cut down) and the rest of the stream's course into which this downcutting process is working back but which, at the present time, is still unaffected.

In geomorphological parlance, this break of slope on the long profile is called a nickpoint; it is a diagnostic feature of progressive headward extension of downcutting along a reach (as will be discussed in Chapter 4). In the main part of the Hangingstone catchment, upstream of the nickpoint, the gentle valley slopes meet the stream through a wide valley floor. In contrast, in the lower reach, the stream has cut down into its old floodplain, so that it has new, steepsided valley slopes at this lower level (Fig. 1.7) Lateral undercutting by the stream maintains the steepness of these lower slopes, inducing shallow landslides and rill and gully erosion: it is not difficult to imagine that most of the suspended sediment at the WRB sampling station at the mouth of the Hangingstone comes from this short, incised, lower reach of the river, with much less coming from the main part of the catchment upstream.

Almost all the rivers in the oilsands area show this downstream change in character where they pass from the upper region over a nickpoint into the steeply-incised lower reach. What this means therefore is that the exact location of the sampling station, relative to the nickpoint, is an important control on sediment levels at that station. Stations located far downstream of the nickpoint (such as the Hangingstone) would be expected to have high sediment levels; stations located only slightly downstream, or even upstream, of the nickpoint, should have much lower sediment levels. This seems to explain much of the variance in sediment concentrations between the different stations in the lower Athabasca River basin: the Ells station, like the Hangingstone, is located at the confluence with the Athabasca, and thus has a long incised reach, and this is reflected in its high sediment levels. The Muskeg and Hartley Creek gauging stations, in contrast, are located upstream of the nickpoint (Fig. 1.8): they therefore collect sediment only from the gentler upper basin, and this is reflected in their low sediment levels.

The full interpretation of the sediment data is more complicated than this (and it will be pursued further in Chapter 3), but the example is sufficient to indicate the importance of an understanding of geomorphology in the location of sediment sampling stations, and in the analysis of sediment data. It should be emphasized that there is nothing wrong with the WRB data collected at these stations, merely that great care needs to be taken in the extrapolation of data from these stations to other sites on other rivers and, indeed, to sites on the same rivers further upstream.

Nor is the example intended as a criticism of the sediment program in the lower Athabasca. Hydrometric stations are usually established in a basin long before a sediment program is thought to be necessary. When the sediment program is eventually implemented, logistical and financial constraints then often dictate that the sediment operations be undertaken at existing hydrometric stations rather than establishing new ones. The intent in using this example was simply to indicate the usefulness of a geomorphological perspective in the interpretation of sediment data, and, where possible, in the selection of sediment sites. This is a theme which will emerge at different places throughout this report.

Similar comments can be made in the context of WQB sampling, because, increasingly, it is being recognized that many contaminants in rivers are attached to sediment, and are not just in the dissolved phase. Thus understanding the reasons for differences in sediment concentrations between rivers (and even within an individual cross-section on a river) is crucial in much WQB work as discussed later (Chapter 11).

Some indication of the relationship between geomorphology, sediment and water quality, in fact, is evident in the data from the Lower Athabasca basin. There is some concern, for example, as to the level of contaminants from the oilsands development in relation to background levels of contaminants in rivers draining basins in which oilsands strata outcrop naturally at the surface. The bedrock geology of the area is given in Fig. 1.9. The map indicates large areas of McMurray Formation (the oilsands strata) north of Fort MacKay, especially in the Muskeg River basin. In contrast, the Hangingstone River basin is underlain primarily by the Clearwater Formation, with just a small and narrow strip of McMurray Formation on both sides of the lower river.

From this map it might have been anticipated that background levels of oilsands contaminants would be highest in the Muskeg River rather than in the Hangingstone River. Yet this is probably not true, for reasons given in the earlier discussion: the contaminants are likely to be adsorbed to organic and mineral grains in the McMurray Formation sediment, and it is where this <u>sediment</u> is entrained by the river that the greatest contamination is likely. Thus the Muskeg River, which has not incised into the oilsands strata, is likely to have much lower background contamination than is the Hangingstone River which has cut down into the underlying McMurray Formation at the downstream end of its course.

The above discussion of the Lower Athabasca basin is directed primarily to staff in WRB and WQB. Some examples that are of more immediate concern to WPM work are provided in later chapters (especially Chapter 6 dealing with floodplain morphology).

1.5 Outline of report

The text that follows is organized into three parts: the *first part* deals with the basic framework of geomorphology; the second part deals with the usefulness of that framework in the conduct of IWD work; and the third part discusses the geomorphology of the main drainage basins in the Western and Northern region.

In effect, Part I constitutes a primer on geomorphology, but only that part of the subject that is likely to be of use to IWD staff in the course of their work. It focuses on the events of the last two million years, during which time alternating glacial and non-glacial episodes had dramatic effects on Canada's riverscapes as we see them today. The initial chapter (2) outlines the basic sequence of events during this period, focusing, in particular, on the emergence of Canada's rivers in the last 15,000 years or so, as the ice masses disappeared. This chapter is followed (3) by a discussion of the various types of sediments that were laid down by glacial ice and glacial meltwater across the country during this period. These deposits formed the platform on which the post-glacial rivers started to flow, into which many of them have now cut down, and from which numerous processes continue to deliver sediment to Canada's rivers today. The final chapter (4) in Part I discusses these various sediment-delivery processes and, in particular, examines the reasons why some of our rivers continue to cut down into this platform, whereas others, in contrast, experience massive deposition so that their floodplains are slowly built upwards over time.

Part II then builds upon this geomorphic foundation and attempts to show how an understanding of geomorphology can be useful to IWD staff across the country in many different ways. There are, however, no distinct geomorphological "tools" or "skills" that are applied. Rather what should have been gained from Part I is an awareness of landscape processes that provides a new "perspective" with which to confront surface water issues. Part II deals with two key features of this perspective: a historical element (5), that is how an understanding of past events can help in the interpretation of the present; and (6) an ability to infer much of what is taking place in the landscape today, especially river channel processes, by looking closely at the landforms (channel deposits, floodplain morphology) that these processes are producing.

Part III applies the basic geomorphic knowledge gained in Part I, together with the perspectives developed in Part II, to specific river basins of the Western and Northern region: the Assiniboine basin; the Saskatchewan River basin; the Peace-Athabasca river basins; and the Mackenzie River basin.

1.6 Additional reading

There are many basic textbooks on geomorphology, but none written specifically for Canada. Perhaps the best reference book, but one which includes little reference to this country, is that by Selby (1985) already mentioned. The text by Holmes (1965), though published twenty-five years ago, is still a classic, easily read, and well-illustrated. Though the "Natural Landscapes of Canada" by Bird (1972) provides a discussion of the scenery of the different regions of Canada, it was not intended as a systematic and analytical treatment of the subject matter of geomorphology.

One of the most useful publications for students of Canadian geomorphology is that by Mollard and Janes (1984). It is not so much a textbook, as a large collection of vertical aerial photographs illustrating different types of landscape across Canada. Each photograph is discussed in some detail, and, if used in conjunction with a basic text such as that by Selby, it offers an excellent approach to the landforms and landscapes of this country. Similar photographic collections exist for particular regions of the country, that by Smith (1987), "Landforms of Alberta", being particularly relevant in the present context.

An additional publication by Mollard (1977) is also worth examining, although it is not especially relevant to IWD work. It is a catalogue of the different regional types of landslide that are found in Canada in different tectonic, geologic and sedimentological settings. To the geotechnical engineer in this country, it is an excellent synoptic coverage pulling together many case studies and classifying them in a coherent manner. A similar work devoted to the different "riverscapes" of Canada, explaining their essential features in terms of appearance, river behaviour and engineering problems, etc., would be very useful to IWD staff. Unfortunately none exists. The nearest is the "atlas" of Alberta rivers by Kellerhals et al. (1972). It provides data, long profiles and aerial photographs for many channel reaches in Alberta.

A final publication that is drawn to your attention is a paper, presented by Church (1983) to the sixth Hydrotechnical Conference (Ottawa), dealing with the importance of geomorphology in hydraulic engineering. It echoes many of the points dealt with in the present report, although illustrated entirely with British Columbian examples.

PART I

THE GEOMORPHIC FRAMEWORK

Part I provides a short overview of the field of geomorphology, as it might be of interest to IWD staff in the conduct of their work. It is not a synopsis of the full field of geomorphology, therefore, and does not include items of marginal relevance, e.g. global plate tectonics, weathering processes, coastal processes and landforms, etc. Nor is it organized in the standard format of most geomorphological textbooks. Rather, it provides a basic framework of the subject, organized to address the perceived needs of the professional concerns of IWD staff.

This framework comprises three components:

- an understanding of the sequence of events in the recent past that have been responsible for the present-day riverscapes of this country (Chapter 2);
- a knowledge of the different types of sedimentary materials laid down in the Pleistocene and now being reworked by the streams and rivers of Canada today (Chapter 3);
- a brief discussion of some of the more important landscape processes currently acting to rework these sedimentary materials, together with some indication of the effects of these processes on stability of the channel long profile (Chapter 4).

Though occasional examples of the relevance of this geomorphic framework to IWD work will be provided in Part I, in the main this first part of the report is largely "academic" rather than applied in scope. The specific application of this body of knowledge to IWD work is more extensively illustrated in Parts II and III of the report.

2. EVENTS OF THE LAST TWO MILLION YEARS

2.1 Introduction

The landscape of Canada has evolved over thousands of millions of years. Some parts of it, in fact, have changed little since they were formed many millions of years ago. This is certainly true of the major terrain units at the scale of the country as a whole. Prior to outlining the events of the last few million years, it is worth looking at the events of earlier times, simply to put "recent" geomorphic activity in some perspective.

The core of Canada is the Shield (Fig. 2.1). It comprises mostly granite-type rocks, worn down over geologic time into a broad platform underlying the sediments and surface soils of most of Canada (and the United States too). The lighter area in Fig. 2.1 indicates the peripheral part of the Shield which was buried by sedimentary strata hundreds of million years ago, when these margins were either under the sea (as a continental shelf) or (as in parts of the west) a continental plain slowly buried under sediments washed onto it from the growing mountains of the Cordillera.

These continental and marine sediments - sands, muds, limy oozes, etc. - grew in thickness, were consolidated by the huge weight of sediments above them, and eventually became the sandstones, shales, and limestones, etc. that cover the Canadian Shield in the Prairies and in southern Ontario and Quebec. This sedimentary cover over the Shield platform was certainly more extensive in the past as ongoing erosion of these relatively soft rocks has broadened the area in which the Shield rocks outcrop at the surface. Today these sedimentary rocks form a capping on the Shield rocks (the surface of which dips towards its margins) only on the periphery. This border zone is extremely narrow in the eastern part of the country (where there has been much downfaulting of the Shield platform) compared to its southern (in the United States) and western (Prairies) margins.

Abutting the western and eastern margins of the (buried) Shield is mountainous terrain, the Cordillera in the west and the Appalachians in the east. The latter are much older (formed roughly 500-300 million years ago), have been worn down during much of the last few hundred million years, and hence are much lower and more subdued in their appearance than their western counterparts. The Cordillera, in fact, started to form only about 150 million years ago, and indeed uplift is still occurring in its mountain chains. In both cases the innermost zone of the mountain zone (adjacent to the Shield) is generally composed of the same sedimentary rock as that found covering the Shield, but heavily buckled and contorted when it was slowly thrust upwards. The outermost mountain zone tends to be dominated, not by sedimentary rocks, but by igneous rocks formed from the cooling of magma and volcanic lava, expelled from deep inside the earth. Fig. 2.2 is a cross-section through western Canada showing the east-to-west changeover from the exposed Shield, the Shield buried by sedimentary rocks, the buckled sedimentary rocks of the Rocky Mountains and the igneous rocks of the outermost ranges.

The origin and evolution of these rocks and landform units is part of the fascinating story of global plate tectonics (see, for example, Selby, 1985, Chap. 3) and cannot be pursued here. These basic geologic and landscape units across the country need to be emphasized, however, because they do have a definite effect on the quantity and quality of the country's water resources. In particular, there is a clear distinction between rivers draining the (exposed) Shield and those draining the Interior Plains formed by the sedimentary strata covering the Shield platform.

The Shield rocks are not easily dissolved by water. Their total dissolved solids count is generally very low, whereas the opposite is the case of the sedimentary rocks of the Plains. The surface water of the Eastern Slopes of the Rocky Mountains, for example, tends to be "hard" because of large amounts of calcium and magnesium dissolved from limestone and dolomite. Also in parts of the Prairies, much of the groundwater is saline because of solution of sodium chloride from the marine shales. Exploration for fresh groundwater sources for rural areas in the Prairies has therefore been a major challenge, and the best sources have often been found in recent drift deposits from the glacial period that overly these sedimentary rocks. Though the present report is not really concerned with subsurface hydrology, it should perhaps be emphasized that rock type is a major determinant of both groundwater quantity and quality.

The rivers draining the Shield area tend to have much lower sediment concentrations (as well as dissolved solids) than watercourses on the Interior Plains, and, to some extent, this also reflects the fact that the Shield rocks are strong (more resistant to breakdown and erosion) than the sedimentary rocks of the plains. Yet this point - the influence of rock type on sediment levels - should not be overemphasized, because the major control on the amount of sediment in stream water is not so much the erodibility of the underlying bedrock as the character of the soil debris overlying that rock.

It is true that, in some cases, the soil overburden has been derived from the breakdown of the underlying rock by weathering. But the most important control on the detailed pattern of loose surficial sediment across this country was the glacial period of the last two million years or so. In addition, though the major terrain units of this country just described - the Shield, the plains and the mountains - were moulded many millions of years ago, the detailed topography of the country (the locations of rivers, lakes etc.) are largely the result of events of the last few million years. This is the period to which the rest of this chapter is therefore devoted.

2.2 General chronology of the Quaternary in Canada

The last two million years of geologic time are usually referred to as the Quaternary period (Fig. 2.3). The major part of this period (in which glacial conditions were widespread in Canada) is called the Pleistonce epoch, the changeover to the post-glacial (or Holocene) occurring only about 15,000 to 10,000 years ago. The most up-to-date and comprehensive account of the development of Canada during the Quaternary is presented in a compendium compiled by the Geological Survey of

Canada, edited by Fulton (1989). This tome is too detailed for anyone interested in just an overview of Canada during the Quaternary, but provides a valuable reference document for those wishing to explore particular regions in some depth.

Though the term Pleistocene is almost synonymous with "glacial" in this country, it should be recognized at the outset that the Pleistocene was not, in fact, a period of continuous glacial ice cover in Canada. Rather, there were several episodes (lasting up to a few hundred thousand years) in which huge ice caps covered most of the country, separated by intervals of longer duration when, because of generally warmer conditions, most of Canada was actually free of glacial ice. The Pleistocene was thus made up of an alternating sequence of glacial and interglacial episodes.

The last major glacial episode started about 80,000 years ago and is called the Wisconsin(an) glaciation. But even it (and previous glacial episodes) should not be thought of as a single glacial event. There were, in fact, short periods of time when the glacial ice masses expanded (called glacial stades) separated by short intervals when parts of the country were mostly free of glacial ice (interstades).

The last major interstade in the Wisconsinan was about 30,000 years ago. At this time, for example, the Appalachians and Atlantic Canada were mostly free of glacial ice, the ice front in the east being in the St. Lawrence Lowlands. This interstade was followed by the last major expansion of Wisconsinan ice masses in this country, the ice reaching its maximum extent about 18,000 years ago (though this date varies by a few thousand years in different parts of the country).

The rest of this chapter focuses on the manner in which the Wisconsinan ice sheets retreated from their maximum extent - the so-called deglaciation of Canada. The processes of deglaciation had a major effect on the river systems of this country, as will be evident in most chapters of this report.

2.3 The late Wisconsinan deglaciation: some general comments

The Geological Survey of Canada has published a splendid map of the extent of glacial ice across the country at the end of the Wisconsinan and its subsequent shrinkage during the last 18,000 years (Dyke and Prest, 1987a). An earlier, less detailed, version (Prest, 1971) is shown as Fig. 2.4. It shows glacial ice covering virtually the whole country at the height of the late Wisconsinan, except for very small tracts of land in the Prairies near the United States border, and a larger area extending from Alaska, through the Yukon, and penetrating as a corridor through the Northwest Territories, just west of the course of the Mackenzie River.

The ice cover shown in Fig. 2.4 varied greatly in thickness, thinning not only towards the edges, but also towards points and lines within the ice cover. In the west, ice thickness decreased towards a trough in the ice surface along the eastern side of the Cordillera. As a result, though maximum retreat in the early stages of deglaciation took place along the southern margins (where conditions were warmer), relatively quickly a belt of land became exposed along the eastern flank of the Rockies, separating decaying ice in the Cordillera from the main body of glacial ice (known as the Laurentide Ice Sheet) in central and eastern Canada. There is evidence that points to a similar pattern of breakup of the ice in eastern Canada with separation of the decaying ice masses of the Appalachians from the eastern part of the Laurentide ice mass along the St. Lawrence Lowlands, this occurring about 12,000 years ago.

As the extent of glacial ice continued to shrink, the huge Laurentide ice mass itself began to split up into smaller components (Fig. 2.5): first of all a split between a Keewatin-Baffin mass and an Ungava ice mass, about 8,000 years ago in the area occupied today by Hudson Bay; and then a split of the first of these into a southern block over Keewatin and a northern block on Baffin Island, remnants of which exist today.

This pattern of deglaciation emphasizes one important point: the glacial cover over Canada did not decay, and therefore probably did not grow, as a single ice mass. Although most of the country was covered with the single Laurentide ice sheet, even that ice sheet had its own distinctive centres: Baffin, Keewatin and Ungava. In effect, the Laurentide ice mass grew from the coalescence of these three ice domes. It is estimated that, at their maximum, the main domes were more than 4,000 m thick in their central parts, with the ice cover gradually sloping away from the centre of the domes. In the east, there appear to have been local ice domes over Newfoundland, New Brunswick and Nova Scotia, constituting a separate Atlantic ice complex that merged (in an apparently confused manner) with the south eastern margin of the Laurentide mass. In the west, an even more impressive marginal ice mass formed from local domes in the mountains to create the Cordilleran ice sheet.

In contrast to eastern Canada, the contact between the local Cordilleran ice sheet and the southwestern margin of the Laurentide ice mass was very well-defined. The ice in the trough marking the contact zone thinned towards the warmer south and to the drier north, creating a narrow corridor of land between the two ice masses. This was particularly extensive in the Northwest Territories as noted earlier.

The corridor is shown, at the time of the late Wisconsinan maximum, in more detail in Fig. 2.6. The western margin of the Laurentide icesheet was pushing against (and being deflected northwards by) the base of the eastern flank of the Cordillera, this being the Mackenzie and Richardson Mountains in the Northwest Territories and the Yukon. In the western and central ranges of the Cordillera, proximity to the moist air masses of the Pacific allowed sufficient snow accumulation to extend the Cordilleran ice complex northward into the Yukon in the Selwyn Mountains with tongues penetrating east into the Mackenzie Range. Thus, land rivers in the nonglaciated Yukon, together with glacial meltwater from the Cordilleran, glaciers flowed down the eastern slopes only to have their courses blocked by the Laurentide ice front. This summertime melt (and from the Laurentide ice also) was thus impounded to form proglacial lakes in front of the Laurentide ice margin. These lakes built up in level until they were sufficiently high to spill over exposed land and drain away from the ice front.

These developments have been carefully documented by workers of the Geological Survey of Canada in the case of the Peel and Bonnet Plume rivers, the lake outflow taking place to the north into the Porcupine Basin (Fig. 2.6). With the eventual retreat of the Laurentide ice front to the east, the waters of the Peel River (and its Bonnet Plume tributary) were able to resume their former course to the Mackenzie Delta, though now cutting down through thick deposits of silt and clay. These materials had been deposited in the proglacial lake, and subsequently exposed as the ice front blockage was removed and the lake drained. On exposure to the cold atmosphere, these muds gradually became frozen into permafrost sediments, limiting, to some extent, the amount of muddy sediment now delivered to the rivers. Nonetheless, large areas of the permafrost are subjected to summertime thaw and spectacular mudflows occur releasing substantial quantities of fine sediment to the Peel. Recent analysis of the suspended sediment regime of the Peel River, at the WRB station upstream of Fort MacPherson, shows that the clay content of the suspended load abruptly increases during summer high flows, and this may well be related to sediment production from these thawing glacio-lacustrine sediments.

In the case of the Porcupine River, the impoundment had even more drastic consequences. Originally the Porcupine flowed through McDougall Pass into the Mackenzie Delta area. But this course was similarly blocked by the Laurentide ice front, impounding a lake in the mid-Porcupine basin, which spilled over to the west at Old Crow. The rapid downcutting associated with this spillway was so substantial that by the time that McDougall Pass was unblocked, the Old Crow outlet was then lower than the old course through the mountains. As a result, the Porcupine did not revert to its former course, and now continues to drain into Alaska instead of the Northwest Territories; the Rat River now flows in McDougall Pass as a "beheaded" river.

2.4 The late Wisconsinan deglaciation in western Canada

With this background, attention is now focused on events in the last 18,000 years or so in western Canada, looking in particular at what was happening at the margins of the Laurentide icesheet.

Under the Keewatin ice dome, the basal ice would have been under tremendous pressure from the weight of overlying ice. The deformation of this ice, because of this pressure, would have led to its being squeezed outwards away from the dome centre towards its margins, and a sinking of the ice surface at the dome centre. This process gives rise to the slow flowage of glacial ice schematically illustrated in Fig. 2.7. As more and more snow and ice accumulates in the central region, ice thickness in this region can be maintained while the ice margins continue to extend outwards. It should be noted, however, that even when the ice front is retreating (not expanding) this slow flowage of ice from the ice dome centre to the margins continues. The only reason for the retreat of the ice front during the deglaciation was that summer conditions were sufficiently warm to produce more retreat of the ice margin through melting than ice margin advance through glacial flowage.

The flow paths of the ice thus take the form of descending in the central regions of the ice dome towards the underlying ground surface, shifting towards a horizontal flow direction further away from the dome centre, frequently having a slight upward component at the ice margin, especially where actively moving ice from the centre is forced to ride over dead (stagnant) thin ice at the margins. In the central regions of the ice dome, the ice that gradually descends to the underlying ground surface acts to scrape away loose material and pluck away fractured bedrock, carrying it towards the margins, some of it being deposited en route, some of it being carried right to the ice front.

These different zones under the ice mass will be examined in the next chapter dealing with different types of glacial sediment and landscapes. The important point here is that, when the ice front lingers for decades or even centuries in roughly the same place (because melting and ice flowage are roughly in balance), the continual transport of debris by the ice to its margin results in extensive accumulation of sediment there. This often takes the form of a ridge flanking the ice front (called an end moraine). Accumulation of debris brought by the slowly flowing ice is not the only process depositing sediment at the ice front. Meltwater, percolating through the fractured ice in the outer part of the ice mass, also entrains sediment from the basal layers of the ice. As the meltwater flows off the ice cap onto lower-lying terrain, this sediment is deposited as outwash. The outwash sediment, just like the glacial debris, can form thick deposits at times when the ice margin is more or less fixed in position, and end moraines are typically juxtaposed with significant outwash plains on their distal (downglacier) side.

These processes are discussed in more detail in the next chapter, but it is not difficult to imagine that areas of end moraines and associated outwash plains (where the ice fronts paused during their general retreat towards the ice dome centres) constitute important areas of potential sediment supply to Canada's rivers in the post-glacial period, continuing in fact to the present day. This will be evident in many of the later chapters.

Apart from the rapidity of retreat (and its punctuation by halts or even small readvances), a second aspect of deglaciation affecting the character of the post-glacial landscape that emerges is the ease with which meltwater can drain away from the ice front. Where this is possible, outwash deposits are likely to be produced, but where it is not (because the land surface slopes toward the ice sheet rather than away from it, as in the case of the Mackenzie Mountains, for example), meltwater accumulates in ice-marginal lakes. The sediment carried by the meltwater then accumulates as lacustrine (lake) deposits and not as outwash deposits. Again, these different deposits are described more fully in the next chapter, but it is useful here to indicate here why glacio-lacustrine deposits are so widespread in the Prairies.

Whether the land surface slopes toward or away from the icesheet depends, clearly, to some extent on the preglacial topography. In the Cordillera, with ice accumulation in the high parts of the mountains, tongues of ice extended along preexisting troughs as valley glaciers, and the meltwater from such glaciers could usually escape along the lower courses of the same valley. Over the Canadian Shield, where the preglacial landsurface generally sloped away from the central part towards the Shield edge, meltwater from the receding ice dome was generally able to escape along preexisting valleys towards the Shield margin. In these circumstances, outwash deposits, rather than lacustrine deposits, would tend to prevail.

On the other hand where the preglacial regional slope is very gentle, the immediate post-glacial land slope tends to be towards the ice dome, favouring ice-marginal lakes. This situation was common in the prairies. The circumstances that lead to this situation are indicated in Fig. 2.8, and warrant some discussion.

A fundamental point to remember here is that, in effect, the thin solid crust at the Earth's surface "floats" on an underlying material (called sima) that is slightly denser and, not solid, but viscous. Thus, when a huge load, such as an ice cap, gradually builds up on the Earth's crust, the crust will tend to sink to some extent into the sima, until a new equilibrium is reached. This <u>isostatic depression</u> of the crust under an ice mass is not small. An ice cap 3,000 m thick might be expected to create about 1,000 m of subsidence of the crust, bearing in mind that the density of ice is about one-third that of the crust. As the ice mass thins towards its margins, so the amount of crustal depression also decreases (Fig. 2.8). This depression was made possible by the fact that the viscous sima under the crust was squeezed away from the area of loading to those parts of the crust flanking the ice mass. This buildup of sima in those parts of the Earth's surface surrounding the ice sheet produces a slight uplift (bulge) of the crust there. This means, then, that at the maximum glacial extent, the ground surface slopes towards the fronts of the ice mass and continues dipping underneath the ice towards its central region.

When the ice mass starts to shrink, so that the loading on the crust decreases, the sima previously squeezed out beyond the ice margins moves back towards its original location, thus gradually raising the crust under the thinning ice mass. <u>Isostatic rebound</u> begins as soon as the ice mass starts to shrink, and is relatively rapid. Fig. 2.9 shows the amount of rebound that has occurred, during the last 7,000 years, in the area formerly occupied by the Laurentide ice mass. By this time the rate of rebound had slowed appreciably, though over much of the country this rebound still continues even today particularly around Hudson Bay (Egginton and Andrews, 1989).

Isostatic rebound was not fast enough to keep pace with shrinkage of the Laurentide ice mass, however, which meant that over most of the western interior, the newly exposed ground surface sloped towards the ice front. Thus post-glacial rivers that began to emerge on this newly exposed prairie land surface drained towards the northeast, and became blocked by the retreating icesheet front. This river water became impounded as an ice marginal lake, supplemented, of course, by the much greater quantities of meltwater draining off the ice sheet. These lakes would build up until sufficiently high that they could spill out along the ice front and then eventually away from the ice.

Fig. 2.10 depicts conditions in the prairies about 11,000 years ago. The newly formed North and South Saskatchewan rivers found their course blocked by a huge proglacial lake, Glacial Lake Agassiz, the water of which eventually spilled south in the Mississippi River system. At earlier periods, with the ice front further to the southwest, other ice-marginal lakes existed, similarly draining eventually to the south. These will be discussed in some of the regional chapters later. The period about 11,000 years ago represents something of a turning point because, by then, sufficient retreat of the Laurentide ice mass had occurred further north to allow drainage to escape to the Arctic Ocean, along the Mackenzie system. Thus, the Beaver River is seen flowing east into Glacial Meadow Lake (Fig. 2.10), then spilling out to the northwest along the present course of the Clearwater-Athabasca rivers into another proglacial lake, Glacial Lake McConnell.

These glacial lakes eventually disappeared for two reasons: (1) the ice front retreated sufficiently to expose a new outlet at a lower elevation; (2) the colossal quantities of meltwater pouring out of these lakes caused extensive erosion, deepening the spillways and thus lowering lake levels. It is interesting to note, for example, where the Athabasca and the Clearwater Rivers join at Fort McMurray, that the valley of the Clearwater is much deeper, wider and more spectacular than that of the middle Athabasca, even though it is by far the smaller of the two rivers. The reason is that it is flowing in a valley cut largely by a much bigger meltwater river in the immediate post-glacial (Smith, 1989). Such rivers are frequently called <u>misfits</u> (specifically <u>underfit streams</u>) because the dimensions of the present day river are inconsistent with those of its valley.

The discussion above was not intended to be detailed. Its purpose has been merely to highlight the major events in the deglaciation of the western interior. More detailed discussion will be provided in the regional chapters of Part III. In summary, three points are worth emphasizing.

- Glaciolacustrine deposits are widespread in the prairies (Fig. 3.13), because the ice front retreated down the regional landslope to the northeast. These deposits vary greatly in thickness, of course, being thickest over low spots of the emergent land surface. Over large areas, these deposits are very thin. In general the deposits are primarily silty-clay (representing offshore conditions), but locally, large thicknesses of sand occur, where rivers produced deltas extending into the lakes.

- Lowland areas that were not submerged by lakes did in fact exist, and in some cases carried large quantities of meltwater away from the ice sheet. This meltwater deposited outwash deposits on the emerged landsurface - gravels and sands, the finer material being washed into streams and lakes. The significance of these outwash deposits will become evident in the next chapter.

- Spillways, the outflows from the old proglacial lakes, are numerous in the prairie landscape. In some cases they exist today as essentially dry valleys; in other instances they are occupied by small rivers, e.g. the Battle in Alberta, the Qu'Appelle in Saskatchewan, and the Pembina in Manitoba, where the present flood discharges are clearly much smaller than the flows that carved out these valleys. This itself is important in understanding the present-day behaviour of these rivers: this will become very apparent in Chapter 6, and in some of the regional sections.

2.5 Deglaciation in coastal areas

So far, little has been mentioned of the deglaciation in the northern part of the region. One reason for this is that much less work has been done in the north compared to the studies in the prairies. In some respects, the same principles apply in the north as in the southern area (this will be evident in Chapter 10, dealing with the Mackenzie basin). On the other hand, a considerable part of the northern region, is different in the sense that its post-glacial emergence represents a balance between ice, land, and also the sea. This is also true of much of eastern Canada, and in bringing this chapter to a close, brief mention is made of deglaciation in the east, especially in relation to marine effects. Similar effects are then to be expected in the north, even if they haven't been studied to the same extent.

Fig. 2.11 shows three stages in the deglaciation of southern Ontario and Quebec, between about 10,000 and 7,500 years ago. The retreat of the ice front was accompanied by some proglacial lakes here (notably Glacial Lakes Barlow and Antevs), but ice marginal lakes were not at all as extensive as in the prairies because the regional slope of the Shield in this region is to the south away from the ice front. More importantly, perhaps, large parts of this region were under sea water in the early post-glacial period. Thus the St. Lawrence Lowlands were submerged beneath a water body known as Champlain Sea, lasting between about 12,500 and 10,000 years ago. Since then the water body has shrunk in aerial extent so that today, the marine inlet is confined to the St. Lawrence estuary, and small lakes in the lowlands are, in effect, remnants of the early post-glacial water body. Huge quantities of sediment were poured into the Champlain Sea from the retreating front of the Laurentide ice mass. Some of this took the form of sandy deltas, where the meltwater streams draining from the Shield entered the sea, but more extensive were the deposits of mud along the old sea floor, and now exposed as the surface sediments in the Ottawa-St. Lawrence Lowlands (as noted in the Rigaud slide).

At a later date, as the Laurentide ice cap split into its Keewatin-Baffin and its Ungava components, about 8,000 years ago, the North Atlantic ocean was able to penetrate back into the Hudson Bay area. At that time, however, its geographical extent was much greater than today (Fig. 2.11c), the name Tyrrell Sea being given to this water body. As with the Champlain Sea, it has gradually become smaller, so that sediments laid down on its bed are now exposed around the shores of the present James and Hudson bays. These marine muds, incidentally, are prone to the same flowslides as the Champlain Sea sediments.

What has caused this gradual shrinkage of the water bodies in the St Lawrence Lowlands and Hudson Bay? The answer is that the water levels have (or have the appearance of having) gone down as the land itself has risen through isostatic rebound. This is a characteristic feature of much of coastal Canada: land areas that were previously depressed beneath sea level by glacial loading (and which were buried by glacial sediments deposited in the coastal waters) have subsequently emerged from the sea, exposing these glacio-marine sediments to reworking by present day processes such as landslides, river erosion and so forth. These same deposits can be expected along the northern coast of Canada, though today they will be permafrosted and more likely to suffer repetitive minor slumping each summer, rather than the catastrophic flowslides found in the south.

Not all of Canada's northern coastal area, however, has undergone this progressive emergence because of isostatic rebound. The Mackenzie Delta, for example, appears to be slowly submerging beneath the Beaufort Sea, and has, in fact, been doing so for at least 10,000 years (Hill et al., 1985). Evidence for this exists in layers of peat (formed in the terrestrial part of the Mackenzie Delta) that now exist under the Beaufort Sea, buried by more recent sediments. These peat layers have been dated by radiocarbon methods to provide an indication of the rate of submergence of the deltaic deposits by the locally rising sea level: over the last 10,000 years, this submergence has averaged about 5 mm per year.

The immediate question, however, is how it is possible for the local sea level around the Mackenzie Delta to have been rising, while almost everywhere else along the Canadian coastline, local sea level has been falling in post-glacial times. The answer is complicated, but two points are worth emphasizing. Firstly, whether local sea level rises or falls, depends not only on whether isostatic rebound is occurring, but also what is happening to the true sea level in the world as a whole. True sea level changes are referred to as eustatic sea level changes, and the important point here is that worldwide sea level (ignoring local landmass instability) was rising throughout virtually all of post-glacial time. This is indicated in Fig. 2.12. The reason for this is that, at the height of the late Wisconsinan (about 20,000 years ago), the oceans contained much less water than today, because large quantities had evaporated and fallen as snow to form the expanding ice caps. When these ice caps began to melt, the water was returned to the oceans allowing true sea level to rise.

Thus, whether or not local sea level has been rising or falling at a given place depends on the rate of isostatic rebound at that place compared to the rate of eustatic sea level rise. Over most of the Canadian coastline, the rate of isostatic rebound outstripped the rate of eustatic rise. This has been true even during the last 6,000 years, because though rebound has slowed down appreciably, eustatic sea level rise itself came to a halt about 6000 years ago. In general, therefore, sea level has had the appearance of falling over most of Canada's coastal areas in post-glacial times. The exceptions are where isostatic depression was minimal (at the margins of the icesheets) so that isostatic rebound was less than the eustatic rise in sea-level. This was true, for example, along much of the coast of Nova Scotia.

Such an explanation, however, is inadequate to explain the continued rise in local sea level during the last 6,000 years in the Mackenzie Delta area, given that eustatic sea level has been essentially stable in that time. There are several components to this problem and it is beyond the scope of this report to discuss them all. The major reason for the ongoing submergence of the Mackenzie Delta, however, appears to rest in a point noted earlier in this chapter. During the loading of the Earth's crust by an ice cap, sima under the crust is squeezed out to areas beyond the margins of the ice sheet. This produces a bulge (a rise in the crust) in these peripheral areas. As the ice cap shrinks, this sima moves back to produce isostatic rebound in the areas under the ice cap, the loss of sima from the peripheral bulge area producing a fall in the level of the crust in these areas. One such bulge area (just beyond the margins of the Laurentide ice cap) was the Beaufort Sea (Fig. 2.4). Thus the crust level there would have been expected to rise during the buildup to the maximum extent of the ice sheet (and there is some evidence to indicate that this did happen) to be followed again by crustal subsidence as the ice masses subsequently shrank.

The discussion above has again been primarily on the basic geomorphological events of the last 15,000 years or so in the Beaufort Sea area, and no attempt has been made to "apply" the usefulness of this knowledge to IWD work. Yet, clearly, this observation has rather important implications for interpretation of sediment accumulation in the present-day delta, and for any plans to develop the area for oil and gas, and bring these fuels onshore across the Delta's rivers. Particularly in the face of a possible rise in eustatic sea level as a consequence of global warming (and the related thaw of permafrost in the Mackenzie Delta sediments), this post-glacial rise in local sea level in the Delta area may accelerate, drastically affecting many aspects of the natural environment, including the stability of river channels.

2.6 Endnote

Attention has been focused above on the last stage of the Wisconsin glaciation and upon its deglaciation. The justification is that the surface deposits of Canada today were largely formed at this time (Chapter 3). The basic sequence of events in the last 30,000 years must, however, have been repeated many times during the Pleistocene, and this should not be forgotten. Indeed, in many cases, downcutting by Canada's rivers (Chapter 4) has fully penetrated through the deposits of the late Wisconsin, into those of early stades, interstades, glaciations and interglacial periods.

3. SEDIMENTS FORMED IN THE PLEISTOCENE

In the last chapter, the historical geomorphology of Canada during the last two million years, particularly the period between about 20,000 and 8,000 years ago, was examined. It was emphasized that, during this late glacial period, different sediments were laid down in different parts of the country and at different times. It was also indicated that these sediments have played an important part in determining the sediment regime of Canada's rivers today. This last point is pursued more fully in subsequent chapters. Before undertaking that task, however, it seems worthwhile spending a little more time looking at these different types of sediment: why they occur in certain areas and not in others; and the properties of the sediments as they affect landscape development.

To the engineer, these materials would be termed "overburden"; the geologist would call the same materials surficial deposits. The different types of sediment discussed in this section belong to a special class of surficial deposits generally grouped under the term "drift". This refers to the accumulation of debris laid down directly or indirectly by processes related to glacial ice and its associated meltwater. In post-glacial (Holocene) time, much of this Pleistocene drift has been reworked by other processes, such as landslides, wind deflation, stream erosion, transport and deposition into other types of overburden. These processes are examined in the next chapter.

A detailed account of glacially-related processes and sediments can be found in specialized texts of glacial geomorphology such as that by Sugden and John (1976), Glaciers and Landscape. A somewhat dated, but extremely useful treatment of drift deposits in Canada, is given in the book edited by Leggett (1961). Particular chapters of interest to the western and northern region include: glacial deposits of Alberta, and soils in the Lake Agassiz region.

Deposits of drift can be subdivided into many different categories. In the present context, four main groupings seem adequate:

glacial: sediment deposited by the glacial ice mass itself;

glaciofluvial (or fluvioglacial): sediments deposited by meltwaters from the glacial ice mass on the exposed landsurface (also known as outwash);

glaciolacustrine: sediments deposited by glacial ice or its meltwater in and around the margins of lakes;

glaciomarine: sediments deposited by ice and meltwater in the sea.

These different deposits are discussed in turn below.

3.1 Glacial deposits

Glaciers are often compared to bulldozers: they scrape and erode material at their base, and the debris so produced is incorporated into the basal parts of the ice sheet and moved downglacier where it is eventually deposited - sometimes only a few kilometres away, sometimes hundreds of kilometres.

There are two key features to note regarding glacial deposits. They can be - and usually are - very poorly sorted with respect to particle size. Material in the sediment can range in size from clay to boulders, and, in fact, in Britain the term "boulder clay" is still commonly used to describe glacial drift. More generally, though, the sediment is called till, sometimes "glacial till", but the prefix "glacial" is redundant because, by definition, there is no such thing as non-glacial till. Till is not necessarily so poorly sorted though. In areas where the till has been derived solely from one particular area it may be quite homogenous, though that is relatively unusual. Nonetheless it must be appreciated that the textural character of till does vary widely according to the rock or sediment of the source area. Many tills in the Prairies, for example, are very clayey, because they have been derived primarily from the scour of soft shaley bedrock.

A second key feature is that, usually, tills lack any well-defined bedding or stratification. In other words, it is generally hard to find evidence that the sediment was deposited in distinct layers, one on top of the other. This is not surprising since till is rarely laid down in this way.

Apart from these two attributes, however, till can vary considerably in character, depending not only on where the debris in the till originated, but also how it was laid down by the ice mass. In particular, it is important to differentiate between till that was plastered down on the ground beneath the moving ice (called lodgment till) and sediment from dirty parts of the glacier that simply accumulated on the ice surface as the glacier wasted away and was gradually laid to rest on the ground surface as the ice mass disappeared. The latter is termed ablation till, the term ablation referring to loss of ice, primarily (but not only) by melting.

There are important distinctions between these two types of till. Lodgment till, because it has been under the full load of the ice sheet, is well-consolidated, and can be very strong. Ablation till, in contrast, having accumulated on, not under the ice, is poorly-consolidated, often almost loose. In addition, during the process of deposition of ablation till, the initial sediment accumulating on the ice may be reworked by various processes, especially by meltwater that washes away the finest sediment particles. As a result, ablation till is often coarser-grained than lodgment till.

These distinctions are obviously important: ablation till is usually more erodible than lodgment till, and more susceptible to landsliding on slopes of medium steepness. On the other hand, ablation till would be expected to produce less suspended sediment in streams than lodgement till, because of the partial depletion of fines by meltwater during its deposition.

In addition, there is some indication that the two types of till tend to prevail in quite different parts of the country. Ablation till tends to dominate in areas that made up the peripheral zone of the ice cap, where (as schematically shown in Fig. 3.1), dirty basal ice tends to move obliquely upwards to the ice surface. Lodgment till, in contrast, tends to dominate in the mid-area between the centre and margins of the ice cap. This positional difference between conditions under different parts of a large icesheet (where ice moves radially from the centre to the margins) is shown schematically in plan form in Fig. 3.2: the innermost (right) zone is one where glacial erosion usually dominates; the next zone downglacier is the zone of active ice under which lodgment till accumulates; next downglacier is the wastage zone (ice being brought to the surface and wasting in place) with ablation till; and finally, at the outer margin, till that accumulates in end moraines.

Fig. 3.3 is an early map from the 1960's showing, in a highly generalized way, the distribution of the different zones under most of Canada that was occupied by the Laurentide ice sheet. It is notable that most of the country belongs to the zone of ice wastage and ablation till. Of course, not all of the country, has a surface veneer of till, even though it probably mantles the bedrock in almost all areas of the country. In low-lying bedrock areas, the till mantle has usually been covered by other types of drift, such as glaciolacustrine sediment. Often, it is only the high points of the preglacial landscape where till is exposed today.

Although some criteria have been presented above for distinguishing the two main types of till in the field, a simpler method is often provided by looking at the landforms and landscapes that the tills produce, rather than by focusing detailed attention on the sediment itself. These till landscapes are called <u>moraine</u>. The term has already been used in connection with ridges formed at the outer margins of an ice sheet (end moraine), but in the present context it is being used to denote the kind of landscape that is produced by till having accumulated in the area occupied by the ice sheet, rather than at the ice sheet margins. This type of landform (landscape) is referred to as <u>ground moraine</u>.

It must be acknowledged that the terms till and moraine are sometimes used interchangeably, but this is strictly incorrect: the term till refers to the sediment; the term moraine refers to the landform produced by the sediment.

Ablation till and lodgement till typically produce very different types of ground moraine. Areas dominated by ablation till are generally characterized by mounds of loose till debris, a landform assemblage usually referred to as <u>disintegration</u> ground moraine, or, if the mounds are quite big, <u>hummocky</u> ground moraine. The detailed processes responsible for producing these mounds and depressions are discussed in standard texts such as those by Selby (1985) and Sugden and John (1976); the general pattern is indicated in Fig. 3.4. Seen from the air, disintegration ground moraine has a very characteristic appearance (Fig. 3.5): it commonly lacks any sense of preferred orientation of the hummocks, though there may be a weak banding at right angles to the direction of ice flow. Other types of ground moraine have more distinctive corrugated surface appearance, but this is not important in the present context.

What is more important here, is to note the marked contrast in ground moraine formed in the peripheral wastage zone, with that formed by lodgment till in the area of active ice. In the latter case, while under the ice, the till has been moulded by the flow of the basal layers, and therefore usually shows ground moraine with lineations parallel to the direction of ice flow. This streamlined landscape may be dominated by linear depressions (grooves) in the till (tens of metres wide and many kilometres long) producing a fluted landscape (Fig. 3.6). In other areas, the ground moraine may be dominated by linear mounds, called drumlins, producing the drumlinoid landscape shown in Fig. 3.7. It is usually easier to identify which direction of the streamlined ground corresponds to the path of ice flowage in the case of drumlinoid landscapes: compare Figs. 3.6 and 3.7.

The details of these various landscapes may not seem particularly relevant to IWD work, but they do provide one useful way of assessing the probable sediment in an area prior to undertaking an expensive borehole program.

In addition, an appreciation of the morphology of a landscape, and the way in which it evolved, can be useful in water resources studies, sometimes in subtle ways. For example, stream pollution from soil erosion on farmland (both the overall sediment levels and the contaminants, such as pesticides, tagged onto that sediment) has recently attracted attention in North America. Studies in southern Ontario (Coote, 1980) documented its severity in the land areas draining to the Great Lakes. It was anticipated that similar findings would emerge in the prairies too. Yet, the prairie landscape is quite different from that of southern Ontario, and, in particular, is dominated in many areas by hummocky ground moraine (Fig. 5.15). This is the landscape colloquially known as the "pothole" district, where a considerable proportion of the sediment moved from farmland simply doesn't reach the main stream network, because it is trapped en route by the depressions of internal drainage provided by the disintegration ground moraine. In these areas, sediment and the associated contaminants are more likely to be found in the sloughs (an important wetland habitat for nesting ducks) than in the main watercourses.

There is insufficient space, and probably little need, to discuss here the complex types of sediment that can make up end moraines. Certainly, end moraines constitute a much smaller area of the landscape than ground moraine. On the other hand, they can add distinctive features to the watercourses of a region. The belt of end moraine sediment frequently contains large numbers of boulders in the drift. Where such moraines cross rivers, the flow may winnow away the fines, but experience much difficulty in moving the boulders. Thus, downcutting is held up at the site of the moraine crossing, whereas upstream and downstream it may continue much more easily. The result is frequently a break of slope (nickpoint) in the long profile, typically in the form of rapids. Some of the major end moraines in central Canada are shown on the map of Fig. 3.8.

The real importance of end moraines is the information that they provide regarding the evolution of that part of the landscape. They denote a line which previously corresponded to the ice margin, at a time when the ice front lingered for quite some time in its overall course of retreat. This is the reason why the accumulation of till occurs along this line. At the same time, however, it implies that meltwater pouring off the ice margin, would have been depositing its sediment in the region immediately in front of the moraine for an unusually long period of time also. This means that on the distal side of an end moraine it would be expected that large thicknesses of meltwater- transported sediment would occur, either in the form of outwash (if meltwater discharged onto land), or glaciolacustrine or glaciomarine sediments. These deposits are discussed next.

3.2 Glaciofluvial deposits

The diagnostic features of glaciofluvial deposits, in contrast to till, is that they are stratified, and, perhaps more important in the present context, they are much better sorted in terms of grain size.

As meltwater pours off the ice cap, the coarser material (boulders and then gravels) become deposited as channel bed sediment almost immediately, though the material may be reworked many times in flood flows, being swept further away from the ice front by the meltwater in the form of a braided outwash stream. In front of valley glaciers, the width of this outwash deposit is restricted by the valley walls, but in front of continental ice masses, numerous meltwater streams along the ice front may produce coalescing deposits known as outwash plains (Fig. 3.9), which, where the ice front has lingered for tens or even hundreds of years, may constitute thick accumulations of gravel. Sometimes (where a forest cover does not exist) the braided pattern may still be detectable in the outwash surface, notwithstanding the passage of thousands of years. In other cases, however, the outwash gravels may have been covered by sand and silt (loess) blown over the plain as the sediments dried out, and before vegetation recolonized the area.

Further downstream, away from the ice front, gravel becomes less and less abundant, and the dominant component of the outwash is sand, typically the coarse and medium sands, capable of transport by the meltwater over much further distances than the gravel. The finest sediment - the finer sands, silt and clay - are carried far in front of the ice margin, and only settle out in much quieter environments, such as lakes, the sea, or gentle floodplains. In the case of the prairies, most of the silt and clay settled out in proglacial lakes, as noted in the previous chapter. The spatial sorting of glaciofluvial sediment by grain size is thus a major feature of the deglacial landscape, and represents a marked contrast to till deposits. It also can have significant implications in the subsequent development of that landscape under processes of fluvial erosion. An example of this is provided below.

The previous chapter included a map of the ice front in the prairies at the time of a halt in glacier retreat about 11,000 years ago. Another, even more significant halt, occurred about a thousand years later (Fig. 3.10). The end moraine formed at this time is known through most of the prairies as the Cree Lake moraine.

In the northeastern part of Alberta, this moraine fronts upon the Lower Athabasca area mentioned in Section 1.4. It was noted in that earlier discussion that the major control on sediment levels in tributaries to the mainstem appeared to be the siting of the sampling station relative to the incised lower reaches on these streams. This is not the only factor though. It was noted, for example, that whereas the left bank Ells River had high mean sediment levels (400 mg/L), those at the station on the right bank Firebag River (Fig. 1.6) had mean levels an order of magnitude lower (40 mg/L). It was noted that the Ells station was downstream of a long incised lower reach; yet in fact the same comment applies to the Firebag. The question therefore arises as to why sediment levels at the Firebag sampling station are so much lower. The answer rests primarily in the sediment deposits laid down in front of the Cree Lake moraine.

The environmental setting of the Lower Athabasca basin about 10,000 years ago is shown in Fig. 3.11. The ice front directly abutted Glacial Lake McConnell (in the area occupied today by the west end of Lake Athabasca), this lake submerging the area that is today the Peace River lowlands and the lower Athabasca valley. The map includes the locations of the sediment stations on the Ells and the Firebag rivers. Meltwater pouring off the icefront in the northern part debauched immediately into the lake depositing its coarse sediment (gravels and sand) at the lake margin in contact with the end moraine. The finer sediment (silts and clays) dispersed throughout the lake to eventually rest on the lake bed as bottom mud deposits. These are the deposits that were laid down in the Ells area. In contrast, meltwater in the southern area poured off the ice front onto higher terrain sloping away from the moraine to the lake, and here thick deposits of sand and gravel (up to 60 m thick near the moraine) were laid down over an extensive area on the east side of the Athabasca inlet, including the basin now drained by the Firebag River.

Thus in contrast to the Ells River, which has cut down through relatively thin silty clay lacustrine sediments, into a thin covering of clayey till, and then into the fine grained sediments of the upper McMurray Formation (silts), the Firebag River has had to cut down through a coarse deposit of sand and gravel, largely devoid of fines. As noted previously, present day sediment being delivered to the rivers is primarily from erosion of bank sediments in these downcutting reaches. In these circumstances it is hardly surprising that the Firebag has such small amounts of suspended sediment. Most of the material into which it is cutting, both downward and laterally, is presumably so coarse that it must be moved as bed load, rather than in suspension.

The reference to Glacial Lake McConnell conveniently leads into the discussion of glaciolacustrine sediments, the third category of drift deposits listed at the start of the chapter. The next section therefore attempts to highlight the major characteristics of these sediments, and, in particular, to contrast them with drift deposits laid down in proglacial marine water bodies.

3.3 Glaciolacustrine and glaciomarine sediments

3.3.1 Introduction

Sediments laid down in lake and coastal areas today can range from boulders to clay: boulders and cobbles occurring, example, on some beaches where wave action undercuts cliffs composed of coarse rubble; and clays (and fine silts) settling out of suspension in the quiet water well away from lake and sea margins. In between these two extremes, sands (and sometimes coarse silts) are carried into lakes and seas by rivers penetrating partially into the waterbody, depositing these sediments as deltaic material.

Sediments laid down in proglacial lakes and seas during the Pleistocene would have displayed the same basic spatial pattern of sorting: coarse sediment around the water's margins; sandy sediment extending short distances into the water as deltaic plumes; and fine sediments settling offshore. It should be remembered, however, that, in many cases, the actual rates of glacial sedimentation in Pleistocene times were much higher than in non-glacial areas today.

Beaches of coarse sediment are probably of little direct interest in the context of the present report, and, in any case, constitute only a small part of the landscape. They are, however, often extremely useful in indicating the margins of former lakes and seas which have subsequently shrunk in extent or disappeared.

Glaciofluvial sandy deltas similarly occupy only a small part of the post-glacial landscape, although, where they do occur, they can play a major role in the fluvial sediment regime of the local area. Attention will be drawn in the next chapter, for example, to the spectacular braided pattern of the lower William River of northern Saskatchewan, where it has cut down through high-level late-glacial sandy glaciolacustrine delta deposits laid down in Glacial Lake McConnell, and now exposed on the southern margin of Lake Athabasca. A similar large sandy delta was formed in the western margins of Glacial Lake Agassiz (in the vicinity of Brandon, Manitoba: Fig. 3.12) where meltwater spilling down the Assiniboine valley deposited its coarser sedimentary grains. Analysis of WRB sediment station data in the Assiniboine River basin indicates that a considerable portion of the sediment load of the lower Assiniboine appears to originate from bank scour of these deltaic deposits (Ashmore, 1990).

The most widespread glaciolacustrine sediments in the interior plains, however, are the silts and clays that settled out as bottom sediments over large areas formerly occupied by proglacial lakes. It is true that over large areas, where the water was relatively shallow, these bottom sediments were very thin and not radically different in appearance from other types of wave-washed glacial drift (Fig. 3.12), but thick deposits of clays, silts and very fine sands, are nonetheless very extensive in prairie areas formerly submerged by proglacial lake water. In the rest of this section, then, attention is concentrated on fine-grained glaciolacustrine sediments, and their differences with glaciomarine fine-grained deposits.

Fine-grained glaciolacustrine sediments cover large parts of western and northern Canada (Fig. 3.13); fine-grained glacio-marine sediments also occur in the northern area (along the coast east of the Melville Hills), but are far more extensive in the Hudson Bay region and the St. Lawrence Lowlands where isostatic depression was much greater than in the peripheral coastal regions.

3.3.2 Glaciolacustrine fine-grained sediments

Sediment inputs to lakes (and marine basins) from glacial meltwater are (and were in the late-glacial) largely a summer time phenomenon, when the marginal ice areas are melting. The actual sedimentation of the fine-grained sediments carried into the central areas of these water bodies is, however, quite different in the two cases.

In the case of proglacial lakes, the silt (and very fine sand), being coarser and with a higher settling velocity, can settle out relatively quickly. Most of this silt is therefore likely to have settled out to the lake bed by the end of summer. The clay size sediment, in contrast, with a much lower settling velocity, takes much longer to settle. This is especially the case where lakes are exposed to internal mixing due to cross-lake winds, as well as to water movements due to inputs of large quantities of meltwater. In fact, in such situations, it would not have been until winter (when inputs of meltwater had stopped, and when an ice cover over the lake impeded wave development) that the water body would have become sufficiently quiescent for the clay size sediment to have settled out. In this way, then, the annual input of fine-grained sediment to a proglacial lake takes on the form of a double band: a lower layer of silt, grading upwards into a layer of clay. This couplet, with a gradual fining in texture upwards, and representing the full sediment accumulation in one year, is called a varve.

Glaciolacustrine bottom sediments are typically varved, showing a banded pattern, produced by the abrupt change in texture (and often colour) from the clay at the top of a varve to the coarse silt above it at the bottom of the following year's varve. Indeed, the single most characteristic feature of glaciolacustrine bottom sediments (in comparison with glaciomarine silty clays) is usually this varved appearance, though, as might be expected, it becomes less apparent in the deposits of shallow lakes. These varves can vary greatly in thickness depending on the rate of sedimentation and the depth of water in which the sediment accumulates. Often they are no more than a few centimetres, but varves tens of centimetres thick have been encountered in some areas.

Varved lake bottom deposits are not necessarily easily erodible sediments in today's environment. Much depends on the age of the sediments. The older deposits, for example, have usually been overridden by subsequent ice advances, and have therefore been overconsolidated by the loading of these later ice masses, by squeezing out the interstitial water and producing closer bonding between the particles, particularly within the clay sediment. Post-glacial lake clays, without this consolidation mechanism, are generally much weaker. Landslide activity, for example, is common in the silty clays of the former Glacial Lakes Barlow and Antevs in the Shield area of the Ontario-Quebec border.

The strength of glaciolacustrine bottom sediments also depends on the mineralogical makeup of the clay size sediment. In much of central and eastern Canada, the clay particles have been derived from the grinding of Shield bedrock and are not truly clay minerals. They tend to be made up, rather, of tiny particles of quartz, mica, feldspars, amphiboles, hornblendes and other primary silicate minerals, often called rock-flour, with only very weak bonds between the particles. Thus the sediments tend to be more susceptible to landslippage and erosion by water. In contrast, in much of the Prairies, the fine-grained sediment in the lacustrine bottom sediments originated, not from abrasion of Shield rock, but from scour of the shaley sedimentary strata that underlies much of the area (Chapter 2). This shale sedimentary rock is usually made up of true clay minerals, such as illites and montmorillonites, in which the close-range inter-particle bonds are much stronger. These clays therefore tend to be somewhat more resistant to normal processes of erosion.

It seems likely, however, that there is considerable variation in the strength of glaciolacustrine silty-clays in the Prairies. It is assumed, for example, that the rapid and short-lived 1990 flowslide in the Peace River lowlands near Rycroft (mentioned in Chapter 1) took place in glaciolacustrine silty clays, though no details of the deposits at this site have been seen. Haug et al. (1977) provided geotechnical logs of boreholes through a large retrogressive failure on the right bank of the south Saskatchewan River upstream of Saskatoon: their data indicated overconsolidated conditions (a degree of consolidation in excess of that consistent with the present overburden loading) in the silty clay lying beneath the surficial sands. The origin of this increased strength remains unknown.

In any case it is important to remember that the ease of erosion of a sediment will also depend on the type of erosional process taking place. Many of these prairie glaciolacustrine bottom sediments are relatively difficult to erode by stream downcutting processes, because though the silty part may be easily scoured, the top of the next underlying varve (the clay) is usually relatively stiff. On the other hand, these deposits are quite prone to bank erosion, streams tending to undercut into the more erodible silty part of a varve (especially where it also contains large amounts of very fine sand), resulting in the collapse of the overlying bank sediment. Because of this, therefore, rivers cut into glaciolacustrine silty clays can have quite high sediment levels, even though the channel itself is quite stable in terms of downcutting (this point is developed further in connection with channel cutoffs in the Pembina valley of Alberta: Chapter 9).

3.3.3 Glaciomarine fine-grained sediments

Glaciomarine muds are quite different from glaciolacustrine silty clays: they rarely, if ever, show varves. They may, however, be banded, especially closer to the landward margins of the sea, where thin seams of sand may extend out from deltaic plumes, interbedding with the normal settling of the finer grained silt and clay. Many, if not most, deep-water glaciomarine sediments are massive, however, with little indication of any bedding at all.

This contrast between massive marine muds and lacustrine varved silty clays reflects the salty character of the water in which the muds were deposited. Under saline conditions, the individual silt and clay particles are able to flocculate (that is fine-grained particles stick to each other at their edges and form a larger floc), so that the sedimentation process is not one of individual grains (of different size, mass, and therefore settling velocity), but the settling of flocs made up of many discrete clay and some silt particles. The settling velocity of these flocs will vary to some extent (depending on the amount of open void space between the particles) but in general most of the sediment, being in flocs, will settle out at roughly the same rate. There is not, therefore, this distinction between rapid settling of silt and protracted settling of clay particles.

As just noted, these flocs contain voids (filled with seawater) between the touching grains, and thus the accumulating bottom sediments will tend to have a higher porosity, and hence higher water contents, than glaciolacustrine sediments in which no flocculation occurred. This has important implications for the behaviour of these sediments when they eventually became exposed by isostatic rebound of the crust in these glaciated areas. These mud sediments may be expected to be slightly weaker, and softer, than their glaciolacustrine sediments in terms of the ability to resist land slippage. The last comment may not always be true, however, at least in terms of the sediment in an undisturbed state. Much would depend on how much cementation occurred at the grain-to-grain contacts both while submerged in the sea and after exposure to the atmosphere.

On the other hand, such strength as these marine muds might have, if cemented, is a fragile one at best, given the amount of water they contain, and if the sediment is disturbed, so that the inter-grain bonds are broken, virtually all the strength is lost and the mud can collapse as if it were a fluid. Fig. 3.14 shows two samples of mud from the Champlain Sea sediments of the Ottawa Valley; the only difference between the intact core, and the slurry-like material in the beaker is that the latter is sediment from a core that has been disturbed. The water content of the two specimens is the same! These muds are described as sensitive muds, the degree of sensitivity being markedly increased when the original saline water in the interstitial spaces has been replaced (by infiltration of rain water for example) by fresh water. The reasons for this increase in sensitivity with removal of the salty water are complex (Williams, 1982, Section 9.8 provides a simplified explanation), but it is important to note that many of these post-glacial marine muds have, in the thousands of years since their emergence from beneath the sea, experienced almost complete loss of salts from their porefluid. Many of them are therefore highly unstable if disturbed, whether this be through the shock of an earthquake, or pile-driving (as in the Rigaud slide discussed in Chapter 1), or as a result of the vibrations set up in an ordinary landslide due, for example, to stream undercutting a valley side in this kind of sediment.

The potential catastrophic effects should be borne in mind in any planning or development of stream valleys that have cut through such sediments. This is true not merely of the St. Lawrence Lowlands, but also the exposed muds around James Bay and Hudson Bay, and in other areas of glaciomarine sedimentation. In the far north, however, these sediments have been subsequently permafrosted (discussed in next section), and are therefore much stronger (provided that the permafrost does not thaw).

The topic of fine-grained sedimentation is a complex one, and involves chemical as well as physical forces, and often microbiological effects too; the above discussion is therefore inevitably oversimplified. One point in particular deserves emphasis. Not all lacustrine sedimentation involves varves! Much of the fine-grained sediment carried by today's river actually moves, not as individual grains, but involves flocculation, a large part of the sediment classified as "silt" on a size basis, actually being flocs of clay particles (Droppo and Ongley, 1990). This means that sedimentation in many lakes today involves the settling, not of primary grains, but of flocs formed, it seems, as a result of microbiological action in stream and lake water (Rao, 1990). Such action is likely to be temperature dependent and is probably much less significant in cold-water lakes today, and in the past.

The last point is worth emphasizing here, however, because much of IWD work is based on the assumption that river sediments are primary particles; often they are not. Thus if suspended particulates of coarse silt and sand size, actually include flocs of clay particles, to which contaminants may be attached, the common attitude of ignoring sandy material in pollutant studies may be inadvisable. There is also the related question of what is meant by "grain size", irrespective of whether the grains are undisturbed flocs or primary particles. When a substantial portion of fine-grained fluvial sediment is actually flocculated material, there is likely to be appreciable differences between the grain size, depending upon whether the determination has been done on the basis of true size (as with a microscope) or whether it was done on the basis of inferred size, using settling velocities (hydrometer; bottom withdrawal tube) and an assumed particle specific gravity.

3.4 Permafrost sediments

One question that has yet to be addressed is: "What sediments were produced in the Pleistocene in Canada, immediately beyond the ice margins (the so-called periglacial area)? The question may seem to be of minor importance because at the height of the late Wisconsinan glaciation, there was only an insignificant amount of land that was not under glacial ice. On the other hand in most of Alaska, and large parts of the Yukon, the ground surface was not buried beneath glacial ice (which would have provided some insulation from the loss of ground heat in the winter time), but exposed to the cold atmosphere. More importantly, perhaps, in the northern part of Canada, as the ice fronts retreated, winter time conditions were so cold that permafrost sediments were formed in the late Pleistocene and Holocene. This topic needs to be discussed, therefore, even if only briefly, especially in the context of the Northwest Territories.

Permafrost, it should, be emphasized, does not refer to a particular kind of sediment. It describes, rather a thermal condition, in which the ground (whether it be rock or sediment, whether it contains ice or not) has a temperature which remains below the freezing point of water, throughout the annual cycle from one winter to the next. Technically, the condition of permafrost may last for a period as short as a few years, but over much of northern Canada it has existed for many thousands of years.

Fig. 3.15 shows the distribution of permafrost in Canada today. The discontinuous zone represents the southern part where some areas at the ground surface (especially where lakes or rivers that do not freeze throughout their full depth in the winter) are not underlain by permafrost, in contrast to the more northerly band where permafrost is continuous. Within the permafrost zone, as might be expected, the thickness of permanently frozen ground increases: at Hay River it is about 15 m; at Norman Wells it is about 40 m; while at Resolute (75° N) it is about 400 m.

On the other hand, while permafrost doesn't necessarily imply that sediments are rich in ice, it is true that in certain circumstances, processes do take place to produce thick ice masses in such sediments. This then assumes some importance to anyone concerned with landscape stability because thaw of these areas of permafrost (as might be expected in a forest fire, by removal of vegetation allowing solar insolation to warm the ground in the summer, or through global warming) will yield large amounts of water in the sediment, rendering the material soft and liable to slippage and more prone to erosion by rain or flowing water.

An important area of northern geomorphology is therefore concerned with understanding the processes that can produce ice-rich permafrost, and the landforms that result, the latter being an important clue, to the trained eye, that ice-rich sediments occur in a given area. Space limits the extent of discussion possible here and reference is made to texts by Washburn (1973) and others for additional information. It is important, however, that the major processes of ground ice formation are at least listed in this chapter. These are: (a) buried ice; (b) segregated ice; (c) ice wedges; and (d) pingo ice.

<u>Buried ice</u> is the most obvious, occurring where lake or river ice has become buried by sediment. It is possible that the massive ice beds along the shores of the Beaufort Sea (several metres thick and usually buried 10-20 m beneath the surface) result from burial in the past, though their exact origin is still debated. These areas are prone to rapid coastal erosion, since undermining of these cliffs produces little coarse debris to provide a protective beach.

<u>Segregated ice</u> forms in the soil during the process of annual freezing. As the frost line penetrates down into the sediment in the winter, there is a tendency, especially in silty soils, for water to be sucked up to the frost line (and past it into the "frozen" soil: Williams, 1982) where it freezes. The degree of lensing of ice will, to a large extent, also depend on the rate of advance of the frost line downwards. At times when the process is temporarily halted, large thickness of segregated ice lense can be formed at the frost line. Such ice is common in the permafrost of the Mackenzie Delta.

<u>Ice wedges</u> are typically associated with <u>patterned ground</u>, particularly fissure polygons (Fig. 3.16): these wedges can be up to several metres wide at ground level and penetrating more than 10 m into the permafrost. The formation of ice wedges requires very cold conditions (a mean annual air temperature of -10°C or less) with the top of the permafrost cooling down in winter to -

20°C or less. Under these conditions, the ground surface shrinks in winter, and, in the process, cracks well into the permafrost (Fig. 3.17). In the next summer, surface water seeps into the cracks and, within the permafrost zone beneath the thawed surface sediment (called the active layer), it changes to ice. In subsequent winters, cracking occurs in the same place, and water is added in the next summer, and the ice wedges gradually increase in width over time.

Pingo ice is generally thought to form in at least two ways, but those most widespread in Canada (in the Mackenzie Delta area) are believed to form as in Fig. 3.18. Under lakes which are deep enough not to freeze through the full depth, a talik (non-frozen zone) exists. This zone is enclosed by almost impermeable permafrost around the sides and, frequently, by some layer of impermeable sediment at its base. In situations in which the lake gets shallower (either because of drainage, or buildup of bottom sediments), the insulation provided by the lake in the winter diminishes, and the margins of the talik become converted to permafrost. The freezing of this water in the talik margins produces an excess volume, however, because there is an approximately 9 percent increase in the volume of ice compared to the original water. What happens, therefore, is that water in the remaining talik is squeezed out, and the only permeable route open to it is upwards into that part of the lake that is still, at depth, unfrozen. But eventually, the lake centre itself becomes sufficiently shallow that winter freezing extends down into the bottom sediments, ultimately producing a complete permafrost seal around the top of the conical talik. Continued conversion of the talik margins, in subsequent years, produces water pressures in the talik that force the top of the cone upwards into the bottom waters of the lake. At the same time, the water in the top of the talik, just beneath the frost line, changes to ice. In this way, an ice-cored hill is pushed and grows upward above the water surface. This is the closed-system pingo.

These various ice-cored landforms are of inherent interest to periglacial geomorphologists, but they are also important to anyone concerned with development of permafrost terrain because of the problems that occur if thawing occurs.

On slopes, for example, if the permafrost contains only pore ice (ice in voids between grains formed by in-situ freezing of water), thawing may be no problem. The resulting meltwater in the inter-grain spaces occupies voids that are slightly (ideally 9%) greater than the amount of water and therefore, in effect, the sediment mass is slightly unsaturated. There is, therefore, no appreciable reduction in the strength of the mass, and no major concern for landslippage.

On the other hand, thawing of slope material with ground ice produces large pockets of water in the soil which can soften the surrounding sediment, leading to slippage, and eventually, flowage, of the debris. Fig. 3.19 shows typical ground conditions in the discontinuous permafrost zone of the Mackenzie Valley. Slopes currently underlain by permafrost may presently be much more stable than those where the permafrost under the plateau does not extend completely to the valley slope. But under conditions of global warming (or removal of the vegetation cover exposing ground to the summer sun), these areas, where there is widespread segregated ground ice, are likely to experience dramatic slope instability.

To IWD staff concerned with sediment delivery to streams in the Mackenzie system, such processes are likely to increase sediment yields dramatically. This is something that must be borne in mind in WRB's planning of sediment operations. It is always tempting, because of budget restraints, to abandon sediment operations at a hydrometric station after a period of 10-15 years, allowing resources to be redeployed to another site. The assumption made is usually that sediment levels monitored over a 15-year period provide a reasonably reliable characterization of long-term levels. But clearly, in the scenario described above, this assumption would be invalid. And, without at least some longterm sediment stations, operating over many decades, this situation may not be apparent. Newer short-term stations might indicate higher levels than short-term stations in the past, but it would not be immediately evident whether this was due to different circumstances at the new sites, or whether it reflected widespread ongoing increase in sediment levels over time.

On flat, rather than sloping, land, thaw of permafrost can also have significant repercussions. Thaw of ground ice will lead to subsidence of overlying sediment to produce a landscape of pockets and craters commonly referred to as thermokarst. In low-lying areas such as the Mackenzie Delta, massive thermokarst development could have dramatic effects on the extent of wetlands and even on changes in the form and course by Delta distributary streams. Beaded drainage, where channel width is locally increased because of thaw of ice-rich banks along a short stretch of channel, and isolated thaw lakes (away from a channel) are illustrations of the effects on drainage networks (Fig. 3.20). The ability to recognize areas of current and future thermokarst is clearly a useful skill in the planning of such areas.

A problem of obvious direct concern to IWD staff involved with channel stability in permafrost areas is thermoerosion niching (Fig. 3.21). Here, the flow of (relatively) warm water against frozen stream banks can lead to preferential undercutting of banks along layers of massive ground ice. The resulting thermoerosion niche undermines the stability of the overlying floodplain sediment resulting in sudden, massive bank collapse, quite different from the slow and gradual bank erosion usually expected on meander bends. This needs to be borne in mind in planning floodplain instrumentation such as gauging stations. Areas of relatively slow bank scour at the present time could, at some future date, involve massive retreat over a long stretch of bank if the floodplain is underlain by, for example, a thick band of buried ice.

On the other hand, it is not obvious that, in the long term, the existence of massive ground ice in floodplain banks will automatically produce more rapid bank erosion rates. Much will depend on the size of the channel. Lapointe (1984), for example, notes that thermoniching, and exposure of ground ice is much more common on the mainstem of the Mackenzie Delta than in side distributaries and that, as might be expected, this is associated with much more rapid erosion rates on the mainstem. He points out, however, that this does not necessarily mean that bank erosion is more rapid along the mainstem because of the greater occurrence of ground ice along its course. On the contrary, it could be argued that exposures of ground ice are more common along the banks of the mainstem because the bank erosion rates are more rapid there (due to the greater velocities in the mainstem). The absence of ground ice exposures along the side channels may simply reflect the limited thaw and bank erosion along these smaller channels, the amount of retreat being insufficient to clean off the active layer sediment along the channel bank, thus preventing exposure of the ice.

Ground ice may also be implicated in a recently documented feature of Mackenzie Delta side distributaries, termed by Lapointe (1984) as "scour holes". These features are relatively deep, localized pockets in the beds of the side channels. In many cases they occur in the expected location in relation to meander geometry, i.e. near the outer bank of a bend where it would be expected that the thalweg would deepen. The depth of these pockets in relation to channel geometry is unusually great (tens of metres in a channel with mean depth of perhaps 5 m), however, and their sides unusually abrupt (Fig. 3.22). In some cases, their occurrence appears to be unrelated to meander bend geometry. The holes are frequently associated with embayments in the adjacent channel bank, suggestive of beaded drainage. The mechanism producing these scour holes remains to be discovered, but it seems likely to be related to pockets of ground ice. There is, of course, the separate question of why these pockets do not appear to fill in with bed sediment, another issue that remains to be addressed. The relevance of these features and their formative process to channel stability, especially in relation to pipeline crossings, is obvious.

The topic of permafrost in relation to channel processes in the north is an important one, though not widely documented. IWD staff may therefore find it difficult to obtain a great deal more understanding of the topic through published material (though the books by Washburn (1973) and Williams (1982) are useful starting points). Such staff should, at least, be aware of its importance, however, and recognize the need to call upon the services of geomorphologists who are specialists in the field where potential problems exist.

When that awareness is missing, tragic environmental consequences could ensue. The problems of massive sediment generation in Southern Indian Lake in the Churchill diversion area of Manitoba is an example in point. The area has been documented to be one of the few extensive areas of glaciolacustrine silt-clay deposits on the bed of the former Glacial Lake Agassiz (Fig. 3.12) and one located within the borders of the area of discontinuous permafrost. Widespread sediment production, through wave and thermal erosion along the margins of the expanded lake system, might have been anticipated in the area long before it actually took place.

4. POST-GLACIAL PROCESSES AFFECTING CANADA'S RIVERS

The previous section examined the main types of sediments formed in Canada in the last two million years. The present section looks at processes that have been taking place in the post-glacial period acting to rework these deposits. In particular, attention is focused upon those post-glacial processes that have had greatest impact on the river systems of this country, and the nature of these impacts.

Perhaps the most important single point in this context is that, throughout most of the post-glacial, and on most of Canada's rivers, the Holocene has been one dominated by the downcutting of rivers into their valley floors. This, in turn, has spawned related effects, such as the rejuvenation of tributaries, and active undercutting of channel banks and side slopes, which, as a consequence, emerges as a major sediment source in Canada today (Chapter 5).

The reasons for this post-glacial stream downcutting (or degradation) will be discussed shortly. Prior to that, brief mention is made of conditions immediately after deglaciation in which, for a period of a few thousands of years, the norm on many of Canada's rivers was for sediment deposition (or aggradation) rather than downcutting. The main body of this chapter then explores the subsequent period of downcutting, with, finally, some reference to possible conditions in the immediate future.

4.1 The paraglacial period

The name "paraglacial" period has been given to the short span of time immediately following the Wisconsin deglaciation (Church and Ryder, 1972). It denotes a period in which there was considerable reworking of Pleistocene sediments and, associated with it, extensive movement of sediment from drainage basins into the valley bottoms of the major streams, in many parts of the country. The term should not be confused with the "periglacial" period. As noted in the previous chapter, the term periglacial refers to cold-climate conditions with permafrost, as experienced near the margins of glaciers in either high-latitude or high-altitude areas, but not dependent upon the proximity of ice sheets.

Church and Ryder (1972) emphasize the latter point, noting that the term "periglacial" does not imply the necessity of glacial events occurring, or having previously occurred. In contrast, "paraglacial" was introduced by them to denote both proglacial conditions (i.e. in front of the margins of present day glacier ice) and conditions "occurring around and within the margins of a former glacier that are the direct result of the earlier presence of (glacial) ice".

The potential for confusion is, nonetheless great because, as already noted, the immediate post-glacial period in many parts of Canada, especially the north, did correspond to periglacial conditions. It is thus clear that periglacial conditions can exist in a paraglacial setting, even though the latter is not essential for the former, nor is the former a requirement for the latter! The least confusing approach to the paraglacial concept is perhaps that contained in Fulton (1989, p. 576):

> During and immediately following deglaciation, in each part of the country, there was a period of intense geomorphological activity, as the newly exposed, unvegetated surface was affected by the full range of geomorphic processes.

In many parts of the ice marginal area, especially in high-altitude and high-latitude conditions, periglacial processes played a pivotal role during this paraglacial transition period between glacial and post-glacial setting. In other areas, however, paraglacial activity involved processes unrelated to ground ice. It is these that are described in the present section. The nature of these paraglacial processes varied to some extent across the country.

In the B.C. part of the Cordillera, Ryder (1971) summarized the processes as follows, emphasizing the existence of a plentiful supply of drift in unstable settings on steep mountain slopes:

> Glacial drift supplies a form easily transported by streams, and particularly well suited to the development of mudflows. Further unconsolidated drift, particularly till, saturated with meltwater, is susceptible to slumping and blockage of drainage channels, again conducive to the formation of mudflows.

Jackson et al. (1982) have emphasized the same processes in the Alberta Cordillera and foothills. They pay particular attention to the fluvial terraces flanking the course of the Bow River for 100 km. from the eastern margin of the Rocky Mountain Front Ranges to Calgary and beyond. These <u>terraces</u> were formed by post-glacial stream downcutting into valley fill material, the downcutting removing most of the infill except for strips along the valley sides (Fig. 4.1). The structures in the gravelly terrace deposits indicate rapid deposition in a braided stream environment, and previous workers had assumed that the sediments were outwash deposits laid down in this, and other valleys of the east slopes, by glacial meltwater during the Wisconsin deglaciation. Radiocarbon dates on organic material in the terrace sediment indicate deposition about 11,500 to 10,000 years ago, however, two thousand years after the retreat of the Cordilleran ice front. They concluded that the valley infill was probably produced by debris-flows in the mountain region, delivering huge quantities of sediment to the valley bottoms, the sediment being reworked into braided <u>alluvium</u> (stream deposited sediment) by post-glacial flood processes. Similar dates in the highest (oldest) sand and gravel terrace deposits in the North Saskatchewan River near Edmonton (Rains and Welch, 1988) also indicate a paraglacial origin.

The extent of paraglacial landslippage on slopes in Prairie valleys (rather than in Cordilleran valleys) has attracted less investigation. It seems likely that downcutting of mainstem rivers into loose drift (such as sandy deltaic deposits along the South Saskatchewan River) would have produce large influx of sediment into stream channels and braided sedimentation, similar to that in the foothills, but involving finer bed material. Such processes would have been dependent upon prior downcutting of plains streams, however, and would represent the onset of present-day conditions, rather than short-lived paraglacial sedimentation.

On the other hand, most of the Prairie mainstem rivers were already incised into the general landsurface to some extent, either because they were flowing in spillways or because the post-glacial rivers still occupied pre-glacial valleys (though partially infilled with drift). Landslips and gullies, for example, are common on slopes of the Qu'Appelle spillway, and it seems likely that sediment supply from these sources to the Qu'Appelle valley bottom would have been much greater in immediate post-glacial times. The blockage of drainage by sideslope fans to impound the many lakes of the Qu'Appelle valley is assumed to have taken place at a relatively early date in the post-glacial (Chapter 7). The temporary absence of a protective vegetation cover would certainly have encouraged sheet, rill and gully erosion in immediate post-glacial times. Klassen (1975), in fact, documents post-glacial aggradation along the Qu'Appelle Valley many tens of metres thick (Fig. 4.2), the silty clay alluvium of the valley bottom having accumulated in phase with colluvium (sediment laid down by surface wash or slippage on side slopes).

These observations on the Qu'Appelle valley are likely to be applicable to many of the Prairie valleys, though they would be most apparent in spillways in which the present streams are underfit. The gentler valley gradients of these spillways were capable of transporting large quantities of sediment (and obviously even allowing downcutting) in the presence of huge meltwater flows, but smaller post-glacial water flows would have been (and still are) largely impotent in the face of sediment supply from side slopes and gullies.

Heavy supply rates of sedimentation to Prairie streams (and consequent valley bottom aggradation) were not just the result of mass movement and gullying. The bare landscape would have been susceptible to extensive wind erosion as well. Extensive tracts of sand dunes, currently inactive under a cloak of vegetation, testify to the importance of wind transportation of sediment in the immediate post-glacial (Fig. 4.3). Sediment particles finer than sand would have been moved in dust storms far away from the original source settling out as thin loess sheets across the plains and, perhaps thicker, in valley bottoms. Imposition of wind-blown sediment on Prairie streams may have led to aggradation in the early post-glacial. Even today, it certainly has dramatic effects in local areas, such as the William River of northern Saskatchewan, discussed in Section 4.2.

Gradually, however, as less and less of the drift remained in unstable high-slope positions, and as vegetation began to colonize the emergent post-glacial landsurface, sediment supply to streams decreased and valley infilling slowed down. In fact, over much of the Prairies, supply of sediment to streams decreased so much that not only could it all be transported by the existing stream, but sufficient energy was still available to allow additional stream bottom sediment to be picked up. In this way, then, stream regime changed from depositional to erosional, and stream downcutting became the major process in many (but not all) river reaches. Exactly when this changeover took place has not been established, and certainly it varied from place to place. The paraglacial probably lasted several thousand years after the onset of deglaciation at a given place.

4.2 Post-glacial processes and stream behaviour

4.2.1 Importance of channel stability

One of the points to emerge from the previous discussion is that whether or not a river reach is experiencing aggradation or degradation (or whether it is in roughly in equilibrium, with sediment inputs to the reach being balanced by sediment outputs from it to downstream) depends on the rate of supply of coarse material to the reach relative to the ability of the stream to transport that material. High relative supply rates imply aggradation; low relative supply rates ultimately allow degradation to take place. The reference to "coarse" sediment is deliberately vague, because what is coarse in one setting may be considered fine in another. As an example, excessive supply of sand to a Prairie stream may produce aggradation, whereas it would have much less effect in a Cordilleran river because the imposed sediment would be largely removed in suspension. In the jargon of sediment hydraulics, any discussion of the effects of "relative supply rates" on channel stability must be couched in terms of channel bed material load, not the wash load (the fine part of the suspended load that moves through a reach without significant deposition, even temporary, on the bed).

Whether a stream is aggrading, degrading, or in equilibrium, is a topic of major important in fluvial geomorphology. It may not be obvious that it is of concern to IWD staff, given the slow rates of channel bed change on a historical, rather than a geological, timescale. Yet when channel deposition is occurring at rates approaching a metre per century (unusually rapid rates to the geomorphologist, but nonetheless found in some river reaches in Canada today), it must be evident that fluvial engineers need to take this into account in planning stream crossings and floodplain development.

When issues of bridge pier scour, floodplain management, bank protection, channel navigation, etc. are tackled without addressing the issue of the prevailing stability of the channel, problems are likely to develop. As an example, in some parts of the world, gravel-bed streams are "mined" by bulldozers to provide coarse aggregate for highway construction. One of the justifications is that excavation of stream gravel slows down the build-up of the channel bed which, in some areas, would reduce channel capacity and ultimately cause overbank flooding. The logic is sound in areas where, in fact, aggradation is occurring, but in other areas, such mining could have disastrous effects downstream: the reduced supply of gravel to these downstream areas would be expected to initiate channel bed scour (restoring the gravel transport rate to one in equilibrium with the carrying capacity of the river) with the potential for undermining of bridge pier scours and other subsurface installations. In the Mackenzie Valley, for example, mining of gravel has been considered as a possible source of coarse granular material needed for insulation of gas and oil pipelines over the discontinuous permafrost. One obvious environmental concern is what effect removal of large quantities of gravel in the area upstream of Norman Wells would have on the stability of the artificial islands in the channel at Norman Wells and on the network of pipelines under the bed connecting these islands to the townsite.

The effects of inter-channel diversions of stream water (an issue likely to assume increased importance in the years ahead) will also vary in type and severity depending upon the present stability of the channel earmarked to receive the diverted water. Where the receiving channel is presently in a state of aggradation, the extra inflow would be expected to reduce or even eliminate the sediment buildup. Where degradation is currently occurring, the extra inflow would be likely to increase that downcutting perhaps with negative environmental repercussions.

Channel stability must be regarded as an issue that matters to IWD staff, especially to those in the WPM branch, and therefore an understanding of the principles controlling channel stability, as well as the ability to infer from field conditions whether aggradation or degradation is currently taking place (Chapter 6), is needed. The topic of channel stability has long been an area of overlap between hydraulic engineering and fluvial geomorphology (e.g., Lane, 1955). It is pursued in this section primarily from the longer term perspective of the geomorphologist, but the engineering implications should be apparent throughout.

4.2.2 Channel aggradation

Various processes continue to take place in some river reaches in the west and northwest to supply excess sediment maintaining aggradation: glacial meltwater and debris slides in the Cordillera; and badland gullying and wind transport of sand in the prairies.

Smith (1972) has described channel aggradation in the North Saskatchewan River in the Cordillera (Fig. 4.4). Huge quantities of gravel are transported by summer glacial melt down the steep North Saskatachewan River: the river gradient flattens abruptly downstream of the confluence of the tributary Alexandra River, however, and the reduced carrying capacity of the flow produces aggradation. The buildup of gravel at the confluence is so pronounced (about 7 metres in the last 2500 years) that the flow of water and sediment down the Alexandra tributary (much smaller sediment supply than on the mainstem) are, in effect, blocked, though there is no large lake. Sediment transport down the Alexandra valley has just been sufficient to keep aggradation in that valley at the same pace as at the confluence. This "backwater" reach upstream of the confluence is quite different in appearance from that of the braided gravel-bedded main valley to which it is tributary: sluggish channels on a silty alluvial floodplain with frequent ponds and small lakes (Fig. 4.5).

Similar contrasts take place along the length of the North Saskatchewan River just downstream of the confluence: debris torrents on the valley sides also supply large quantities of sediment to the channel, and, at these points, the river is unable to transport all the coarse sediment, and a new gravel fan is built up in the river. Upstream there is the similar backwater reach (but this time being the North Saskatchewan River itself), while downstream of the Rampart Creek junction, the steep gravel fan is dominated by laterally shifting courses of a braided valley bottom.

The processes at work at this site, and the attendant channel aggradation, are found at many sites in the Cordillera, and give some insight into what conditions must be have been like (only much more extensive) in the paraglacial.

Aggradation induced by channel gullying in the prairies is certainly not as spectacular as that induced by debris torrents in the Cordillera, but has been suggested by McPherson (1969) in the Red Deer valley downstream of Dinosaur National Park. Upstream, the river flows southeast along a spillway formed along the Laurentide ice front. This deep late-glacial incision into the soft sandy-shale bedrock of the region has been followed by spectacular gullying of the valley sides to produce a "badlands" type landscape (Fig. 4.6). Measurements of the annual erosion in the badlands along the full length of the river downstream of Red Deer itself indicates that the annual volume of sediment from this small part of the basin exceeds the total suspended sediment load of the Red Deer River at Bindloss downstream (Campbell, 1977).

The implication is that the excess sand sediment is building up on the lower valley slopes adjacent to the river and/or in the river bed itself. McPherson (1969) notes that, on passing through the main badlands reach in the national park, the Red Deer changes from a single-thread (single channel) stream to a multi-channel (almost braided) pattern. The changeover in river style coincides with the river flattening upstream of the braided reach, and then steepening, producing a hump in the long profile similar to the hump of the North Saskatchewan River at Ramparts Creek (Fig. 4.4) noted earlier. No attempts have been made to measure the rate of inferred aggradation in the river reach.

Even if aggradation is occurring at this particular site, it may be of no direct concern to IWD staff. There will, however, be reaches such as this in which aggradation, even slow sediment buildup, will have environmental significance. In some river reaches in the prairies, for example, stream sediment is derived from farmland erosion (rather than gullying of soft rock as here). Such sediment is often polluted with nutrients, metals and pesticides. Any accumulation of such sediment in side channels may produce contamination of benthic fauna. This is likely to be much more serious in the case of smaller streams flowing sluggishly on the gentle floors of old spillways than the Red Deer.

As an example of excess supply of sediment to streams through wind entrainment, the William River of northern Saskatchewan (Fig. 4.7), is quite spectacular. The river flows north into Lake Athabasca, and, over most of its course, is an irregularly sinuous, narrow, single-thread channel, carrying little bed sediment as it flows over a rock-controlled, muskeg-bordered bed. In the lowest 30 km of its course, however, it has cut down through a late-glacial sandy delta built out into the proglacial lake (McConnell) that was a forerunner to (but about 120 m higher than) the present Lake Athabasca. The delta surface is still largely bare of vegetation, and in fact constitutes the largest desert-like area of active sand dunes in Canada today (Fig. 4.8). Sand is blown from the west field to the river in large quantities (Carson and MacLean, 1986), building up on the left flank of the river and then avalanching down. Though no direct measurements of bed load transport rates have been made, it is inferred from the change in sand grain types along the reach, that there is about a 40-fold increase in bed load movement in the dunefield reach of the river compared to upstream (Smith and Smith, 1984)

This increase in sediment supply to the river is associated with a dramatic change in appearance, the river abruptly widening from about 80 m to 350 m, and becoming transformed into a braided sand-bed channel dominated by numerous mid-channel bars (Fig. 4.9). No direct evidence is available regarding aggradation in the reach, though visually this is evident in the impressive delta of the river built into the lake (Fig. 4.7). Again, it is not clear whether this transformation of river style and aggradation of sediment, all brought about by this localized input of sand from the dune field, is of any concern to IWD staff. The example is important, however, because it highlights the role that sand deflation can play in determining river regime (and certainly did play in the paraglacial).

There are many other parts of the prairies in which rivers have cut down into sand deposits and where, at least in the recent past, wind deflation of sand into the river must have been a significant process. At present, many of these old deserts remain covered with vegetation, but it is not difficult to envisage that, with global warming producing increased evaporation in the prairies, some of these desert surfaces may reappear. The William River provides an example of what may happen to river regime in such case. Aggradation and river metamorphosis are not purely academic matters. Intakes for irrigation water, for example, are likely to prove somewhat ineffective if they became blocked by a buildup of bed sediment, or, alternatively, became detached from the main channel of the river by the build up of braid bars. An appreciation of the potential changes imposed by increased sediment supply through wind erosion, and the locations in which they might take place, would seem to be a desirable part of the training of anyone concerned with river engineering.

All the previous examples of channel aggradation have resulted from increased sediment supply to a river. Yet the same effect can materialize without any change in sediment supply if there is a reduction in the transporting capacity of the river to move sediment that is delivered to it. The classic case of this in western Canada is the Peace River downstream of Bennett Dam in British Columbia. The dam impounds water for hydroelectricity generation, and flow from upstream is regulated to provide increased flows in winter. There has thus been a substantial (about 50%) reduction in the magnitude of summer flood peaks in the Peace downstream. Yet tributary rivers, steep, deeply-incised into the plains rocks, continue to deliver bed material and wash load to the Peace. In some cases, the bed load supplied is relatively coarse gravel. Reduction in summer flood levels has rendered the Peace now incompetent to move this debris which is now accumulating in the mainstem at tributary junctions. This has been well-documented by Church (1983b) in the case of the Pine River junction, just upstream of Taylor Bridge (Fig. 4.10).

4.2.3 Adjustments of channel stability to changes in stream gradient

As already stressed, aggradation is the result of coarse sediment being supplied to a stream in excess of the long-term capacity of the stream to transport that material; degradation is the result of coarse sediment being supplied to the stream in insufficient quantities (less than stream capacity) permitting the stream to augment its bed material load by channel scour. The previous examples emphasized the role of particular landscape processes in choking streams with excess sediment, and the comparable situation where a decrease in discharge, in the face of unchanged sediment supply, achieves the same effect. Before dealing with degradation, it is useful to adopt a slightly more generalized approach to channel stability: the previous examples dealt with only one subset of controls.

The carrying capacity of a river is largely controlled by the product of its discharge and its velocity (or stream gradient). The first of these is determined by upstream controls: runoff conditions in the drainage basin. Stream gradient may be altered through changes in either upstream or downstream controls: deposition upstream or erosion downstream will increase the gradient, while the reverse changes will decrease channel gradient in the reach in question. Thus channel instability in one reach will clearly have repercussions both upstream and downstream of that reach: deposition in a river reach through increased sediment supply will decrease the gradient upstream (inducing aggradation) and increase it downstream (thus increasing the channel's ability to transport the increased sediment supply), as noted in the case of the upper parts of the North Saskatchewan River.

The effect of changes in downstream control of channel stability is indicated in Fig. 4.11. In geomorphological parlance, the downstream control is usually referred to as the base point of a reach; and the changes involve shifts in the horizontal location of the base point, or in its elevation (base level changes). In all cases, the shifts have the potential to affect the gradient of the reach, and thus the sediment transporting capacity of the stream flowing through it. The figure depicts five base point changes.

1. Progradation: advance of the downstream end of the reach without any change in elevation, as in the extension of a river delta into a lake.

2. Base level rise produced by upstream shifting of the base point, as would occur at the downstream end of a river meeting a water body (sea, lake, reservoir) which experiences a rise in water level.

3. A vertical rise in base level of the reach without any shift in location, as takes place, for example, during isostatic rebound of the land mass, or accumulation of debris at this point as from an alluvial fan.

4. Retrogradation of the base point: the opposite case to progradation. The situation is produced by horizontal erosion at the downstream end of a reach where, for example, coastal recession produces retreat of the mouths of streams draining towards that coastline. Many

parts of the Mackenzie Delta are currently experiencing rapid retrogradation (through thaw of ground ice in cliffs), though the process is also superimposed on a gradual base level rise due to isostatic crustal subsidence, as noted in Chapter 2.

5. Progradation of base point accompanied by base level fall, as would be experienced on streams draining to coastal areas during a period of sea level fall.

Theoretically, a sixth change (the opposite of case 3) might be envisaged, where isostatic subsidence occurs in inland areas, but such cases appear to be rare. The first three changes are all capable of producing aggradation. The last two usually (but not always) produce channel degradation.

The effect of <u>progradation</u> of the base point (Fig 4.12a) is to extend the downstream end of the stream course. This produces a quasi-horizontal river reach over which the flow is incapable of transporting all the sediment, thus initiating aggradation. Aggradation takes place not only downstream of the initial base point, but also upstream, as the flow attempts to produce a smooth long profile connecting the original reach with the prograding reach. Theoretically, if the base point stops prograding, aggradation will still continue extending its way upstream until the original long profile shape has been restored, but at a level meeting with the new base point. It should be clear that even a short progradation has the potential to produce massive aggradation over a huge length of river upstream from the base point. This is the kind of aggradation which is often of most concern in reservoirs: not so much the infilling of the reservoir by the progradation of the river into it, but the associated upstream aggradation induced by that progradation. Kuiper (1960) discussed the significance of this at length in the context of the Saskatchewan River upstream of its historic delta margins in Cedar Lake and its current delta margins in Cumberland Lake.

<u>Base level rise</u> resulting from water level change (Fig. 4.11: case 2) does not produce aggradation through new processes acting to reduce river reach gradient. Rather, it produces an upstream shift in location of the progradational environmental (Fig. 4.12b). In the long term, any point upstream of a progradational reach (such as A in Fig. 4.12a) would be expected to suffer eventual channel sedimentation as the aggradational wedge located on the old base point expands. It may, however, take thousands or millions of years for the aggradational wedge to extend this far upstream, by which time other changes (such as a fall in lake level) might have taken place to counter the effects of the initial progradation. The effect of base level rise, therefore, is to promote aggradation in such upstream reaches at a much earlier time, simply through the inward relocation of the channel base point. Again, this is a potential concern in areas where reservoirs have been impounded by dams, or raised in level. One of the few areas in Canada where the process is naturally taking place today is in the Mackenzie Delta where, as previously noted, local sea level seems to be rising at the rate of about 50 cm per century. The rate may seem slow, but in an environment where river gradients are extremely gentle, the rate of upstream shift in the base point can be deceptively rapid.

<u>Isostatic rebound</u> (Fig. 4.11: case 3) does not necessarily produce aggradation in river reaches: this will depend entirely on the spatial pattern of rebound. The point is that such rebound would occur throughout the whole land area affecting not a point on the river but a huge length of river. In areas where the rebound occurs more slowly at the base point than in the upper reaches, the effect of the rebound is in fact to produce a steeper gradient. On the other hand, where the base point is uplifted more rapidly than the upstream end of the river reach, the reach gradient will decrease, and aggradation is likely to occur. It seems likely, for example, that post-glacial prairie streams draining to the northeast, such as the South Saskatchewan River, experienced this effect, given that rebound amounts and rates increased from the outer (southwest) margins of the Laurentide ice cap to the inner (northeast) points (Kugler and St-Onge, 1973).

It should be noted that, though progradation is still evident in the lower reaches of many Canadian rivers today (and therefore conducive to aggradation in the reaches just upstream), and though some reaches may still be suffering from the effects of isostatic backtilting, the first three cases of Fig. 4.11 are not widespread across the country. On the contrary, especially in reference to base level change, post-glacial time has been one of lake lowering (through gradual erosional lowering of lake outlet elevations) and sea level fall (through isostatic rebound) (case 5). In other words, downstream controls of channel stability in post-glacial time have acted more to induce degradation than aggradation.

Many lake margins, moreover, have suffered ongoing recession through wave action undercutting cliffs and bluffs in soft drift deposits (Fig. 4.11: case 4). This is evident around both man-made lakes, e.g. Lake Diefenbaker, as well as natural lakes such as the north shores of Lake Erie and Ontario. The potential contribution of shoreline erosion to sediment infilling of lakes and the associated turbidity is certainly a point that should not be overlooked. In the present context, however, it is the effect on the stability of streams draining to such lakes that matters. This undercutting produces a steepening of the lower reach of such streams (Fig. 4.13) with the characteristic nickpoint separating the oversteepened reach from the established reach upstream: the result is that transporting capacity in the lower reach is increased beyond the level needed to transport the prevailing sediment supply from upstream and, thus, scour of the channel bed initiates degradation. The process continues by headwater retreat of the nickpoint until, theoretically, lowering of the entire length of stream bed has been attained restoring the original long profile to meet the new position of the base point.

Production (and subsequent recession) of nickpoints has also been promoted by base level fall (Fig. 4.11: case 5) which has been widespread in post-glacial time. Theoretically, nickpoint emergence takes place only (case 5b) when the newly-extended channel reach is steeper than the original lower reach of the river; when base level fall exposes a gentler slope (case 5a), aggradation would result. Usually case 5b applies (base level fall exposing the delta front slope of Fig 4.12) and degradation is indeed initiated.

To IWD staff, much of this discussion may seem academic because the rate of channel change is usually so slow that it poses no engineering or environmental problem. This is false reassurance. The rate of change in channel stability will depend largely on how close existing channel reaches are to some threshold condition. For example, the aggradation of tributary bed load at confluences on the Peace River was not an inevitable consequence of the reduction in flood discharges down the main stem. In situations where the size of bed material brought in by tributaries was well within the competence of the mainstem to transport it, this ability would have continued even after regulation. It was only where competence was already a somewhat marginal issue that problems have arisen.

The same point can be emphasized in terms of channel degradation. Increased channel energy (through reach steepening) does not automatically produce incision and degradation. Much depends on the prior stability of bed material in the reach. In some cases, bed material may be sufficiently coarse that, even under new conditions, it will remain immobile, thwarting degradation. Rains and Welch (1988), for example, comment on the oversteepened downstream reaches of tributaries to the North Saskatchewan River in the Edmonton area. Present channel stability is afforded by a gravel bed armour overlying the Cretaceous sedimentary strata into which these lower reaches have been incised.

It is clear then that understanding the effects of changes in drainage basins on channel stability requires not merely a knowledge of theory, but also a sound appreciation of the existing behavioral condition of the river reach, something that can only be gleaned by field inspection, something that is, in fact, at the heart of a fluvial geomorphologist's approach to channel stability (Chapter 6). The discussion of natural examples, even where slow-acting rather than spectacular effects occur, is therefore essential to providing field verification of channel stability theory. These comments merely repeat those made long ago by a respected American hydraulic engineer, E.W. Lane (1955), regarding the importance of fluvial geomorphology in channel stability studies.

4.2.4 Channel degradation

The previous sections have already provided reasons why it should be expected that channel degradation, rather than aggradation, would be the norm in Canada today: changes in downstream controls in post-glacial times have mostly tended to produce oversteepening of lower river reaches (which should then result in retrogressive degradation by upstream movement of nickpoints); and marked decreases in sediment supply to streams as paraglacial conditions disappeared have also made available more stream energy for channel erosion.

The effect of these changes has been to produce a marked rejuvenation over much, if not most, of the length of Canada's rivers today. It is because of this post-glacial rejuvenation, that river terraces (Fig. 4.1) are so common along river valleys, though such terraces are largely of historical interest. The existence of terraces implies the presence of nickpoints along river courses, however, and these must always be of interest, if not concern, to anyone involved with channel stability. Nickpoints represent a phase of channel instability that has not yet come to an end. The process of retrogressive degradation may have been temporarily slowed down or even halted (through, for example, exposure of a sill of resistant rock in the bed), and the period of temporary stability may last long enough on a human timescale to eliminate concerns regarding channel stability.

The true stability of nickpoints is, however, not always immediately obvious: a lag of gravel over soft sediment (as in an end moraine on lacustrine silts) may afford only a fragile stability. Once the lag has been breached, subsequent downcutting through the softer silts could be catastrophic. Under normal circumstances, such breaching may be highly improbable, the stream having already been subjected to extreme floods in the past. Nonetheless the fact needs to be recognized by riverscape planners that these nickpoints are the potential weak spots in the fluvial system. Diversions of water into a stream system are likely to be much safer (from a channel stability perspective) well downstream of nickpoints than upstream of them. Examples of catastrophic recession of man-made nickpoints are known in several parts of the west (East and West Prairie rivers, Alberta (Parker and Andres, 1976); Seine River, Manitoba (Galay et al., 1970) and should serve as a warning for all river channel alterations, not just channelization intended to alleviate flooding. Kellerhals et al. (1979) provided documentation on interbasin river diversions in Canada which similarly highlight the importance of valley bottom sediments and terrain in the prediction of morphological effects. Finally it should not be thought that channel stability problems such as these are confined to WPM staff: in small streams, sudden, catastrophic nickpoint recession may produce sediment loads far in excess of background levels, with possible repercussions on water quality too.

The role of nickpoints in controlling the response of streams to human interference is but one specific example of the legacy of post-glacial rejuvenation for present day river management. That legacy manifests itself in many other ways, but above all, in the point that in most parts of Canada, the major source of sediment transported in today's rivers is from the margins of these channels, where undercutting has produced steep, unstable channel banks. This theme, and its implications, will be developed further in Chapter 5. Prior to closing the present chapter, however, it seems appropriate to extend the perspectives just developed and ask what the future holds in terms of channel stability in Canadian rivers.

4.3 Sedimentary processes affecting Canada's rivers in a world of global warming

Though the impact of global climatic change in the years ahead is likely to vary in different parts of Canada, it is not difficult to argue that in many cases the response is likely to be a slowing down of degradation, an increase in the length of river reaches affected by aggradation, as well as an increase in wash load levels.

In maritime areas, the effect of global warming, if pronounced enough, will be a rise in world-wide sea level. This rise in eustatic sea level may be insufficient to counter ongoing isostatic rebound along the margins of Hudson Bay, but in areas where crustal movement is presently minimal (and especially where subsidence appears to be occurring), the result will be a rise in base level for those rivers that drain directly into the sea. Areas of present aggradation near coastal margins are therefore likely to shift further upstream (Fig. 4.12a) with possible problems (navigation difficulties, increased overbank flooding, etc.) in communities that are presently unaffected directly by aggradation. The relocation of the aggradational wedge in deltaic reaches such as those of the Fraser and Mackenzie rivers will be gradual, and may not produce noticeable effects until well into the next century, but the likely changes should certainly be borne in mind. The severity of the changes will also be dependent upon other components of the channel stability equation: changes in the supply rate of sediment to these lower channel reaches, and changes in the flow's sediment transporting capacity. These are likely to increase rates of channel sedimentation as noted below.

In northern areas, global warming may be expected to melt permafrost and increase the frequency of landslides in permafrosted sediments. Such slides already occur in the Mackenzie basin (triggered by removal of vegetation or ground warming through forest fires) especially in areas of glaciolacustrine fine-grained sediments. The processes will become much more widespread in the future. Theoretically the severity of the problem, and the areas most likely to succumb to such instability, could be predicted from the present pattern of thermal conditions and surficial deposits, and the likely change in temperatures.

The effect will certainly be to increase wash loads in such rivers (ground ice being most prominent in silty clay sediments), though whether there will be significant increase in channel sedimentation is less obvious. Instability of coarse-grained drift may not increase appreciably because of the reduced extent of ground ice in such material. However, many of the rivers of the Mackenzie basin have relatively fine bottom sediments: the Arctic Red River and most of the Mackenzie itself, downstream of the Ramparts, are sand-bedded; the Peel River, at least in the vicinity of Fort MacPherson, is mud-bottom in many parts of its channel. In rivers such as these, therefore, increased supply of silt and sand (from thawing permafrosted stratified sediments) is likely to lead to increased channel sedimentation.

Whether this is likely to constitute a problem is a separate question, but there are certainly grounds for taking the issue seriously. In terms of navigation, sand bars already pose problems in certain reaches of the middle Mackenzie, and the cost of dredging and channel maintenance may be expected to increase. Much would depend, however, on the source of the sand delivered to the Mackenzie. Most of the sand seems to come from west bank tributaries which have cut down and into old proglacial sandy deltas. Such deposits are unlikely to contain massive ground ice and the increase in sand delivery to the mainstem may, in fact, be insignificant.

In terms of the fishery resource of these rivers, the impact may be more severe. The environmental agenda in the Mackenzie system has always highlighted the possible negative impacts of dredging (increased turbidity, downstream sedimentation of fines in side-channels and sloughs, etc.), yet these concerns are likely to appear trivial in the context of the expected increase in sediment concentrations and in fine-grained sedimentation of backwater reaches from thaw of ice-rich glaciolacustrine sediments flanking these rivers. Again, it must be emphasized, that these changes will take place slowly, so gradually that bed-level changes may not be noticed for many decades, and increased sediment levels in stream water may not be noticed at all unless environmental monitoring is pursued on a longterm basis.

In southern parts of the country, the nature of climatic change is more difficult to define, but it is generally agreed that global warming will result in increased summer evapotranspiration, reduced soil moisture levels, and lower mean stream flows. The effects are likely to include deterioration of the vegetation cover so that soil erosion (from both windstorms and localized rainstorms) will probably become more severe. Sediment yields in prairie streams have already been noted to increase in years following a period of successive dry years (Kuiper, 1960); the increased frequency of such dry spells in the future suggests that soil erosion and sediment delivery to streams will likewise increase. These changes may not be simply the result of drier soil conditions and less protective vegetation: summer rainstorms, though probably less frequent in the future, may be more severe (a typical feature of semi-arid climates) and such high-return-period storms may accomplish more sediment delivery to rivers than several years of "normal" climate at the present time.

Again the severity of the resultant sedimentation is a separate issue. One of the ironic aspects of water and sediment management, however, it that the trend to drier conditions not only increases the demand for water-storage structure, but tends, at the same time, to reduce the viability of such structures because of the increased rate at which they infill with sediment. The impact is also likely to be different between mainstem rivers and their tributary courses. The bulk of runoff in the mainstem rivers of the Prairies originates with snowmelt in the Cordillera: mainstem river discharges (and carrying capacities) will therefore be determined also by climatic change in the mountains as well as in the plains.

Much of the above is, of course, speculative, but the last decade has shown increasing evidence of climatic change, and the need to address what the impacts of such change will be on sediment production, transportation and deposition in Canada's rivers is a real one. As noted earlier, the change in suspended sediment concentrations due to climatic change is likely to be gradual. Its severity will be documented in the data collected by WRB at its sediment sampling stations. It is essential that some of these stations be earmarked as longterm stations in order that these trends can be determined. It is therefore ironic that, at the very moment when the onset of climatic change may be initiating a new sediment regime in Canada's rivers, WRB is terminating the sediment program at many of its stations, on the grounds that sufficient data have now been collected to adequately represent the longterm condition! A strong case can be maintained for maintaining at least some of these stations (even if only as miscellaneous stations) over a much longer timeframe.

It should also be apparent that the sediment record compiled by WRB is not simply one of documenting the transition from one sediment regime to another. The data have the potential to be extremely useful in predicting what the sediment changes will be in different climatic change scenarios. It seems likely, for example, that some of the longterm stations in the western and northern region already have sufficient data to go beyond the construction of sediment rating curves, and to permit the examination of other aspects of climate on sediment levels. In the prairies, several such questions come to mind. For example, how important are high-return-period summer storms in the longterm sediment yield of prairie streams? How much increase in sediment concentrations is produced by one or more prior years of very low rainfall? The IWD engineer might turn to the geomorphologist for answers. Yet, in fact, the most accurate answers are likely to come from an analysis of the data that IWD itself collects as part of its longterm monitoring program.

One of the most important aspects of a "geomorphic perspective" as far as IWD staff is concerned should be the recognition that its own field programs can (and should be constructed to) provide valuable information regarding the operation of many sediment-related processes that impact on this country's rivers. In other words, the kind of geomorphic background that is being provided in this report, will hopefully not merely assist in the location of sediment stations, and in the analysis of sediment and water quality data, but, in addition, lead to a better recognition of the value of such data in the environmental agenda. Some insight into this utility is afforded by the chapters of Part II that follow.

PART II

USEFULNESS OF THE GEOMORPHIC FRAMEWORK

Part I provided a short overview of the field of geomorphology. This framework comprised three components: an understanding of the sequence of events in the recent past that has been responsible for present-day landscape form and process (Chapter 2); a knowledge of the different types of sedimentary materials laid down in the Pleistocene and now being reworked by streams and rivers of Canada today (Chapter 3); and a brief discussion of some of the more important landscape processes currently acting to rework these sedimentary materials, together with some indication of the effects of these processes on stability of the channel long profile (Chapter 4).

Parts II and III deal with the usefulness of this framework in the conduct of IWD work, noting, at the outset, that the various components often interact in the development of a proper understanding of present day channel behaviour. As an example, in several floodplain areas of the prairies (e.g. the Qu'Appelle), the tendency of normally sluggish, meandering rivers to overflow their banks has prompted programs of channel straightening in order to increase in-channel velocities and (through increased scour) cross-sectional flow area. These programs have often met with limited success. To understand why (and to have been able to predict why prior to the programs) requires an appreciation of all three components of the framework. Firstly, many of these valleys are low-gradient spillways, formed by glacial meltwater (Chap. 2), and now occupied by misfit streams that are only weakly capable of reworking the valley bottom material. Secondly, the floodplain sediment itself can be quite variable in its resistance to reworking depending on the late-glacial environment in which it was deposited (Chap. 3). Thirdly, the efficacy of sediment reworking by such streams depends also upon other processes affecting channel stability including supply of extra-channel sediment (Chap. 4).

It is therefore difficult to isolate the usefulness of particular components of the framework: the regional examples (Part III) emphasize their interdependence. Nonetheless it does seem important that two major hallmarks of the geomorphic perspective are highlighted prior to examining specific regions of the western and northern landscape. These are:

- an appreciation of the importance of historical conditions in understanding present-day riverine problems and issues;
- an ability to infer what is going on in the landscape and riverscapes of today by looking at the outcome of those processes, the landforms themselves.

The first of these is illustrated in Chapter 5. The role of the past in conditioning the present, and the importance of recognizing this role,is evident in all aspects of IWD work. Church (1983), for example, illustrated the importance of a sound understanding of the origin (and properties) of valley bottom materials in the context of bridge pier scour. In order to provide some cohesion in the present discussion, the examples used in Chapter 5 all relate to one topic: the quantity and quality of suspended sediment transport in streams, where that sediment originates, and where it is deposited. The chapter is therefore primarily of interest to IWD staff in the WRB and WQB areas.

The second topic (Chap. 6) will, in contrast, be much more relevant to those concerned with channel stability, floodplain management, etc. (WPM staff), though the significance for WRB staff will again be very apparent. The theme developed is quite simple: understanding the nature and rates of operation of present day river processes is rarely as simple as plugging values into engineering formulae. The measured values may be correct, the formulae may be mathematically sound, but if the processes represented by the formulae are not those that are at the heart of the issue in the real world, then the formulae are irrelevant, and the conclusions reached are, at best, invalid. But how are river engineers to know what the crucial processes taking place in channels are, especially when they may last for only short periods of time, and occur only infrequently? The answer put forward in Chapter 6 is that, the landscape itself usually shows what the key processes are, through the different landforms that it contains. All that is required is the ability to recognize the landforms and to relate them to their formative processes. This ability requires a special "eye" for the landscape and a special training: these are skills that are the hallmark of a geomorphological education.

5. THE HERITAGE OF PAST GEOMORPHIC PROCESSES: SEDIMENT LEVELS IN CANADIAN RIVERS

IWD's Water Resources Branch has been collecting data on suspended sediment levels in Canada's rivers for more than twenty years. Though often prompted by specific concerns (e.g. how quickly would a proposed reservoir infill with sediment deposits), a major goal has been to acquire background data throughout the country in order to determine natural sediment levels in different regions as they are affected by different regional conditions. These regional patterns (and the factors responsible for them) form the subject of the first section of this chapter.

The very notion that individual sediment stations can be used to represent regional sediment levels (expressed as specific sediment yield [tonnes per square kilometre of basin], or as mean sediment concentration [annual sediment load divided by annual runoff]) is itself, however, something that needs to be examined. It implicitly assumes certain characteristics in the sediment-generating processes that may not necessarily be valid in all environmental settings. This is the topic of the second section of the chapter.

In both sections, attention is drawn repeatedly to the need to understand the recent geomorphic past in order to investigate these issues properly.

5.1 Regional levels of suspended sediment across Canada

In 1972, the WRB undertook its first attempt to produce maps of sediment levels across the country. One was prepared for specific sediment yield; another for mean sediment concentration (or mean load/flow ratio). Discussion here is focused on the latter, though, of course, the two are closely related.

The 1972 map, prepared by Stichling (Fig. 5.1), was, it should be emphasized, a preliminary one only. It was based on short-term data in many cases, and there was considerable interpolation between stations in order to provide a complete coverage across the country. The map is certainly wrong in some areas, as updated information, discussed shortly, indicates. The overall impression conveyed by the map does seem however, to be largely valid. The basic pattern of sediment levels in Canada is shown to be influenced, not only by bedrock geology, but, in particular, and more importantly in the present context, by the Quaternary history of the country.

5.1.1 The Canadian Shield

The Canadian Shield is seen to stand out as the area of lowest sediment levels (< 50 mg/L). This would be expected from the low degree of erodibility of the igneous Shield rocks. The point was made in Part I, however, that river erosion of bedrock is a relatively slow process, and therefore regional differences in sediment levels are unlikely to be completely determined by regional differences in bedrock erodibility. The usual course of landscape evolution is one in which bedrock is broken down into a surficial cover of sediment. It is this surficial sediment - rather than its parent bedrock - that is the main source of sediment in streams. This is important in the context of the Shield: during the last two million years, the repeated glacial scour of the Shield areas has left only small amounts of surficial sedimentary debris there, the material having been transported to the peripheral areas of the former ice caps. Overall, the main reason for low sediment levels in the Shield area appears to be the general scarcity of unconsolidated sediment in the area. The existence of numerous small lakes throughout the Shield (in part also a reflection of events during the Quaternary) must also act to reduce sediment levels in streams, though the generally slow rate of sedimentation in such water bodies implies that this is a relatively minor point.

There are, of course, exceptions to the above generalization. The lower courses of most Shield rivers are shown on the map to have slightly higher concentrations (up to 200 mg/L and, locally slightly higher). The areas of higher sediment levels appear to coincide with tracts of fine-grained proglacial sediment that are currently being dissected by streams. These include the glaciomarine deposits laid down in the former Tyrrell Sea inlets of the Hudson Bay part of the Shield, and glaciolacustrine sediments laid down in the late Wisconsin proglacial lakes (Antevs, Barlow) of the Ontario-Quebec border noted in Chapter 2.

5.1.2 Areas of high sediment levels

At the other extreme, the highest sediment levels (> 1000 mg/L) are shown in the Red Deer Valley of Alberta and in the Milk River headwaters of the Missouri system; sediment levels almost as high are shown in the headwaters of the Peace River system, and in the presently-glaciated mountain basins of the Yukon - British Columbia border.

The Red Deer and Milk River regions appear to be somewhat unusual in that the extremely erodible Tertiary bedrock into which the mainstems have incised are gutted by spectacular badland areas, the gullies of which appear to supply the bulk of the sediment to these streams (Campbell, 1977). The role of the Quaternary seems, in this case, to be dwarfed by the unusual character of the soft bedrock. To some extent, this may be a misleading comment. The spatial distribution of badlands, in relation to the pattern of bedrock geology, has not received much investigation. While it is true that the areas of spectacular gullying occur in regions of unusually soft sandy shale, the converse (that all such bedrock areas have produced badland landscapes) has not been established. It should be borne in mind that the pre-glacial bedrock landscape was deeply buried by Quaternary drift, especially in valleys cut into the general plateau terrain. In areas where post-glacial channel downcutting has taken place along the courses of these preglacial valleys, badland development would not be expected, because the soft shales are often still buried by drift.

One of the striking features of the present-day drainage network of the Prairies, particularly in Alberta (Fig. 5.2), is the degree to which the post-glacial streams have, in their incision into the landscape, largely missed the courses of their pre-glacial valleys. One of the reasons for this is that many of the reaches of the post-glacial rivers were cut as ice-marginal spillways by meltwater flowing northwest-to-southeast along the retreating front of the Wisconsin Laurentide icecap, whereas, in contrast, the preglacial valleys drained more directly to the northeast (Fig. 5.2). Thus the post-glacial drainage network of southern Alberta is a composite one made up of NW-SE reaches that were formerly marginal to the icesheet front and SW-NE reaches (some of which are aligned with preglacial valleys). Some implications of this complex post-glacial drainage evolution are pursued in the regional discussion of the Saskatchewan River basin (Chap. 8). The important point here is that, in almost all cases, these former spillway reaches were cut in areas of relatively thin drift cover, and the present rivers (where they have not become underfit streams) have therefore exposed sedimentary bedrock in their valley walls. Badland gullying seems to be restricted to such areas.

The high sediment levels shown in the Peace River headwaters and the upper Smoky River (Fig. 5.3) appear to have been largely conjecture (there were no data available) and are now known to be in error: analysis of WRB data for the Peace River at Peace River suggests that, even prior to construction of the Bennett Dam (1968), sediment levels in the Peace, where it drains from the Cordillera, were extremely low (Carson, 1991). It is interesting that the Parsnip-Finlay-Peace area was originally thought to have been one of high sediment production. Whatever the basis for that assumption, it is consistent with the popular belief that high mountain terrain produces high sediment yields in streams. One of the conclusions to emerge from the present discussion is that, at least in western Canada, this view is essentially fallacious.

In fact, in examining the regional pattern of sediment levels in western Canada (Fig. 5.1), exactly the opposite conclusion emerges. Most of the Cordillera is shown with levels less than 400 mg/L, whereas the high concentration areas (> 400 mg/L) are predominantly in the interior plains. The only exceptions to this statement are the northern parts of Coast Mountains (400-700 mg/L): these are alpine catchments that are still occupied by active glaciers, and in which the high sediment levels appear to be the result of dirty glacial meltwater. Small glacierized basins in the Rocky Mountains of Alberta (e.g. Hilda Creek) have similarly high sediment levels (Hudson and Niekus, 1990).

5.1.3 The southern Cordillera

The general pattern conveyed by Fig. 5.1 would seem to reflect two main points: (1) much less fine-grained sediment is available for supply to streams in the Cordilleran basins, in comparison with Plains rivers; and (2) irrespective of supply rates to Cordilleran reaches, the lakes found in many of the valleys of the Cordillera trap sediment thus maintaining relatively low suspended sediment levels. Both points appear to be a direct result of the sequence of Quaternary events in the southern Cordillera and interior plains.

The argument that supply of fine-grained sediment to streams is much less in the Cordillera is based on qualitative observation of processes presently taking place. Over most of the Cordillera, unconsolidated debris is relatively coarse, sands and gravels, and much of it is likely to be moved as bed load rather than in suspension.

In part, the scarcity of fine-grained sediment available for delivery to Cordilleran streams presumably reflects the fact that the most unstable sources of Wisconsin drift in the Cordillera were mobilized in the immediate paraglacial period. Thus much of the potential sediment source in the high mountain areas has already been depleted. Hudson and Niekus (1990), in discussing sediment yields in Alberta, make the same observation: upland contributions from mountains and foothills are relatively small because of the limitations in fine sediment supply. Their observations were based primarily on the upper Oldman and Elbow rivers. In the latter area, present-day valley-side colluvium is derived from weathering of resistant limestone and dolomite, and little fine fragmental material is produced. In addition, in the foothills, upslope erosion of pockets of drift is limited by a protective forest cover. Hudson and Niekus (1990) conclude that the dominant sediment sources and processes in the Cordillera and foothills are channel erosion and sideslope undercutting. In much of the area, therefore, the strong bedrock and coarse lag drift found in valley bottoms is not conducive to producing high levels of suspended sediment in streams.

In contrast, fine-grained sediment is much more abundant in valley-bottom areas of plains streams. Whether the channel banks and lower slopes of these valleys is shale bedrock or drift (till and glaciolacustrine sediment ultimately derived from the shaley bedrock), the riparian sediment is generally not only relatively easily eroded, but, equally important, sufficiently fine-grained to be incorporated into the suspended load.

Sediment-trapping by lakes is also instrumental in holding sediment levels in the Cordillera to low values. Apart from sediment traps produced by debris fans from side slopes (e.g. the upper North Saskatchewan River, noted in Chapter 4), prior glacial erosion also played a major role in creating sediment traps in the mountains. As already noted (Chap. 2), glaciers in the Cordillera were typically confined to valleys, producing a distinctive pattern of bedrock erosion (Fig. 5.4). Glacial scour of the valley floor was concentrated under the thickest part of the glacier, gradually thinning towards the snout where debris accumulated in the end moraine zone. Thus, deglaciation produced rock basins, flanked by moraine at their lower ends, in which post-glacial lakes trapped sediment from the headwaters. Over time, many of these lakes have disappeared (through both lake infilling as well as downcutting into the moraine by the lake outflow), as noted in the case of the upper Bow valley in Chapter 8. Some of these lakes still remain to trap sediment, however, and, even where infilling has taken place, the gentler valley gradient in the aggradational zone of the old lake basin still encourages sedimentation of alluvium in floodplain areas during overbank flows.

5.1.4 <u>The southern Interior Plains</u>

The sediment data collected since 1972 indicate that, in detail, the spatial pattern of sediment levels shown in Fig. 5.1 is inaccurate. These new data do not, however, contradict the interpretation made above; on the contrary, to a great extent, the new data reinforce it. The new information emphasizes higher sediment levels in the southern interior plains than in the southern Cordillera

For example, the South Saskatchewan River upstream of the Red Deer confluence, was shown by the 1972 map (Fig. 5.3) to be in the 200-400 mg/L class. New data indicate the mean sediment level in the reach to be, in fact, about 600 mg/L. Though mean sediment levels monitored in the lower Athabasca are only about 400 mg/L, much of this sediment appears to come from the McLeod and Pembina rivers where these meandering streams are carving away at glaciolacustrine silts and clays. Sediment levels in the mid-Peace River are substantially higher than the 400-700 mg/L indicated by the 1972 map (Carson, 1991), and reflect active bank erosion by tributaries such as the Smoky River incised into fine-grained glaciolacustrine sediment overlying easily erodible shale. Finally, the Fort Nelson River of the interior plains of northern British Columbia, though shown on the 1972 map to have sediment levels in the 200-400 mg/L range, have mean concentrations in excess of 1000 mg/L (Carson, 1988c). The river, and particularly some of its tributaries such as the Muskwa, are incised in glaciolacustrine silty-clays overlying weak shales, just as in the Peace River valley. No attempt has been made to assess the relative importance of bedrock (shale) and glaciolacustrine deposits as sediment sources in these regions of high sediment levels.

These more recent data therefore confirm the importance of widespread deposits of fine grained glaciolacustrine sediments (and weak shale) in producing high sediment concentrations in the southern interior plains, and reinforce the contrast with the Cordillera. Although stream channel-bank erosion appears to be the major sediment-producing process throughout large areas of the interior plains, the levels of sediment so produced must be expected to be strongly controlled by the character of the sediment into which stream incision is occurring. The Highwood River of the western Alberta plains (Fig. 5.5), for example, is deeply incised into its landsurface, yet the mean load/flow ratio (based on data given by Day and Spitzer, 1985) is about 250 mg/L only. This is consistent with the character of the valley material into which the river has incised: the mainstem has cut down into resistant sandstones, while the tributary Sheep River seems to have downcut into more easily erodible (but coarse-grained) fill in the pre-glacial valley (Mollard and Janes, 1984). It seems likely that a detailed study of sediment sources in the western and northern region would indicate a close relationship between sediment levels and the pattern of Wisconsin deglaciation as hinted in Chapter 1.

5.1.5 The northern Cordillera and Plains

Recent data now available for the northern part of the region do not show this clear distinction between high sediment levels in the plains and low levels in the Cordilleran reaches. The 1972 map (Fig. 5.6) shows slightly higher sediment levels in west bank tributaries of the Mackenzie River, draining the Cordillera, than in the mainstem. The tributary levels indicated by the 1972 data generally range 200-400 mg/L, though up to 700 mg/L in the Arctic Red River basin. More recent data in fact indicate that these

Cordilleran rivers have much higher sediment levels (Fig. 5.7), with measured mean concentrations in the Arctic Red and Peel rivers at about 1000 mg/L, and comparable sediment levels implied for most of the west bank streams (Redstone, Keele, Mountain rivers, etc.). These northern Cordilleran rivers have much higher sediment levels than their southern counterparts, with levels comparable to, if not exceeding, the highest levels recorded in the southern interior plains.

The reasons for the high sediment levels in the northern Cordillera have not been firmly established. There is little doubt, however, that most of the sediment comes from glaciolacustrine sediments laid down in proglacial lakes impounded by the Wisconsin Laurentide ice front when it abutted the east flank of the Mackenzie Mountains (Chap. 2). In other words, once again, the late Wisconsin glacial setting seems to be very important: in contrast to the northern ice-corridor (Fig. 2.6), in the south, union between the Cordilleran glaciers (much larger than in the north because of the greater snowfall) and the Laurentide ice front resulted in ice covering almost the entire eastern slopes at the time of late-Wisconsin maximum. Proglacial lakes could only develop once icesheet shrinkage was well under way; and, any pro-glacial lakes that did occur seem to have been short-lived.

The central theme of this first section then is quite simple: interpretation of WRB sediment data clearly requires a sound appreciation of Quaternary events. By the same token, planning of new stations in the sediment network needs to be guided, in part, by a perspective that includes sediment sources in relation to the geography of Pleistocene drift.

5.2 Sediment yields and catchment size

It is a widely held belief that specific sediment yield (load/drainage area/time) decreases as one proceeds downstream in a drainage basin. And, indeed, in many parts of the world, especially parts of interior United States where much of the world's published sediment data originate, this belief is valid (Fig. 5.8).

The explanation of the pattern shown in Fig. 5.8 is quite simple. Stream sediment in these areas is derived primarily from the erosion of upland soils. In the headwaters, the terrain is steeper and thus soil erosion will be greater; as basins become larger, more and more of the catchment area is lowland of gentler slope, and the soils in these areas do not erode as much as in the steeper areas. In addition, as drainage basin size increases, an increasing percentage of the landscape acts as traps for any sediment that is eroded on upland slopes. Thus much of the eroded soil does not reach the main stream, but accumulates in tributary valleys, in alluvial fans and on the lower parts of slopes. Also, as one moves down the mainstem, more and more of the load that has reached the mainstem is deposited on the stream bottom and on the floodplain in overbank flows. The percentage of the soil lost in upland soil erosion (usually determined by formulae) that is actually delivered to the stream system (as determined at a sampling station at the mouth of a drainage basin) is called the sediment delivery ratio. The data of Fig. 5.9a indicate how, in the interior USA, this ratio decreases from almost 100% in small basins to less than 10% in large basins.

This pattern of decreasing sediment yield as basin size increases would be expected to prevail in almost any region in which stream incision has essentially ceased, and where most of the sediment carried by the river system is supplied to it from the interfluves, that is from the slopes between the streams. This is not the condition that prevails in most of Canada. On the contrary, during the last 6000 years at least, the main process involved in most Canadian river reaches has been stream downcutting (Chap. 4). This situation has resulted from (a) the depletion of upland sediment sources in the paraglacial period, and the protection of upland slopes against soil erosion by revegetation during the Holocene, and (b) the ongoing fall in stream base levels, arising from falling lake levels inland and isostatic rebound of coastal areas.

With this in mind - all of it a direct heritage of Quaternary history (and quite different from the Quaternary history of the interior USA) - it would be surprising if the pattern of decreasing sediment yield as basin size increases was also to be found in Canada. In most areas of Canada it does not seem to apply. On the contrary, the reverse scenario appears to be the norm: specific sediment yields increase as basin size increases, i.e. increases downstream. The reason is quite simple: most sediment is derived from undercutting of channel banks by the stream itself. Thus, the further downstream that a river flows, the greater will be the amount of fine-grained sediment (wash load) acquired from its channel margins (Fig. 5.9b).

This pattern was observed in eastern Canada as long ago as the early 1970's: in a small, 80% forest-covered drainage basin in the Quebec Appalachians (Fig. 5.10). The data were based on the spring flood period only, and for two years only; but most of the sediment load in the region is produced and moved out of stream basins during the spring. Smaller tributaries had much lower specific sediment yields than larger tributaries, and, in turn, these yields were significantly smaller than those at the basin mouth. Sediment was observed to originate primarily from scour of fine-grained till as the river network cut down through it. It is true that till thickness increased from the basin sides towards the basin axis, but there is nothing unusual in this, indeed it is a common characteristic of the deposition of drift. The point that such fine-grained drift is less widely-available in the stream headwaters simply reinforces a trend that would be expected on the basis of progressive bank scour along the stream network.

WRB sediment load estimates for British Columbia indicate a similar downstream increase in specific sediment yield, though the pattern is complicated by diverse basin characteristics and hydrologic regimes (Fig. 5.11a). In particular, those rivers receiving flows and sediment from present-day glaciers in their headwaters have much higher sediment yields, and a different downstream pattern of change. Disturbed stream basins affected by logging, mining, etc. also tend to have above-average sediment yields, while basins in which natural and man-made lakes trap sediment, plot below the main swarm. However, once these peculiar sites are identified and removed, a distinct trend emerges in the remaining data: sediment yields increase with increasing basin size up to about 30,000 sq. km, before decreasing slightly (Fig. 5.11b)

There is some indication that the same pattern exists in the streams of the prairies. As discussed in Chapter 5, downstream parts of the Assiniboine River have much higher yields than at stations further up the basin (Ashmore, 1990). Analysis of WRB data for two stations in the flat basin of the Roseau River of Manitoba (Fig. 5.12) indicates a similar conclusion. The specific sediment yield at Dominion City was twice that at the upstream station of Gardenton, much of the sediment being inferred to originate from bank erosion in the relatively steep channel reach between the two stations (Carson, 1990b). The same pattern was inferred in the lower Athabasca basin of Alberta in the discussion of Chapter 1.

This increase in sediment yield (and concentration) downstream is not an academic matter. It has obvious implications in the use of existing sediment data (on projects in areas where WRB data are limited), in the planning of sediment networks, as well as in sediment quality studies. The last point is worth expanding upon.

One of several aquatic environmental issues to emerge in the last decade has been the increase in sediment pollution (including the effects of contaminants bound up in sediment) arising from erosion of farmland soils. These off-farm impacts were highlighted in an American study (Clark et al., 1985), and also documented in the PLUARG investigation in southern Ontario (Coote, 1980). Attention has therefore recently been directed to exploring the importance of the issue in the prairies, where (Fig. 5.13) the on-farm costs of soil erosion are known to be almost as high as in southern Ontario.

Analysis of existing WRB and WQB data in the Saskatchewan River basin (Carson, 1990d) indicates, however, little evidence of off-farm pollution of stream sediment, at least in the mainstem prairie rivers. On the other hand, though only scant data are available for small prairie watercourses, there is some evidence to suggest that off-farm pollution of stream sediment is greater in small watersheds. (This is certainly apparent in the eutrophication of small prairie lakes in agricultural basins where much of the phosphorus originates directly from sediment washed into the lakes, or, indirectly, from the subsequent release of phosphorus from bottom sediments.)

The pattern of off-farm pollution just described is consistent with the pattern of sediment production in the prairies noted above. A large proportion of stream sediment originates from channel bank erosion (rather than from soil erosion on farmland fields), and this proportion appears to increase as one moves down through the stream network, as noted in Chapter 3. In other words, though off-farm sediment may be important in some of the small prairie basins, its relative importance decreases as one proceeds down the stream network because of the continuing additions of non-farm sediment from channel banks and undercut slopes (Fig. 5.14). In addition, of course, large parts of the prairie landscape that are subjected to cropland erosion do not actually contribute sediment to streams because of the poor integration of the landscape with the stream network. This is strikingly indicated in Fig. 5.15 showing the region of "potholes" in the interfluve area between the Pipestone and Assiniboine rivers in southwestern

Manitoba. Again, the importance of Quaternary events is highlighted: the pothole-region of the southern prairies corresponds to late-glacial disintegration ground moraine laid down on ground that was sufficiently high to avoid submergence under proglacial lakes.

The contrast in the severity of off-farm impacts between southern Ontario and the southern prairies therefore, in large measure, reflects contrasts in the typical riverscapes of the two areas. Sampling stations on prairie mainstems are located in large basins (in which stream sediment is dominated by inputs from bank scour and steep valley walls) and where, in any case, delivery of eroded sediment from farmland fields to the stream network is limited. This is not the case in southern Ontario: drainage basins are small (and therefore the proportion of sediment from bank erosion is also small), while the undulating land surface (the end-moraine zone of Fig. 3.3, rather than the ice-disintegration zone) produces a higher sediment delivery ratio on Ontario farmland fields.

The contrasts between the southern prairies and southern Ontario should serve as a note of caution, therefore, in making any generalizations regarding off-farm impacts on Canadian rivers. Though much of Canada corresponds to the prairie situation in terms of mechanisms of sediment supply to streams, other areas (with smaller basins and better integration of the land surface with its streams) are likely to correspond more to the Ontario scene; the river basins of Atlantic Canada immediately come to mind.

In conclusion, therefore, it should be clear that the interpretation of fluvial sediment data (and assessment of its implications) gains a great deal from an appreciation of the Quaternary history of the landscape. Moreover, the benefits provided by such a perspective are not restricted to sediment and water quality issues. A similar case could be provided in terms of those environmental issues that are related to river channel behaviour and stability. Some examples of this are provided, in fact, in the next chapter.

6. INFERENCES FROM LANDSCAPE MORPHOLOGY

Geomorphologists are concerned with how the present landscape has evolved over time. As emphasized in the previous chapter, therefore, one of the hallmarks of the geomorphologist's perspective, is the recognition that the landscape of today reflects processes that have been operating over thousands or millions of years. In some cases, present-day processes are quite different from those that operated in the past: and the associated landforms are, in a sense, fossil features.

On the other hand, in many situations, today's landscapes are being moulded by the same processes that have been operating for thousands of years or longer. This is an extremely important point, and one that is the foundation of much of geomorphology. Though the goal of geomorphology is to interpret present-day landforms in terms of landscape processes, the knowledge needed in that task provides the opportunity to operate the reasoning process in reverse: to infer the nature of processes operating at the earth's surface today from the character of the associated landforms.

This ability to infer landscape process from landscape morphology is an important asset, and one that differentiates geomorphologists from landscape engineers. This is true in all areas of the subject, but, given the context of the present report, it will be discussed here only in relation to fluvial geomorphology, i.e. the processes and landforms associated with flowing water.

The forte of the hydraulic engineer is his ability to model fluvial processes in a quantitative way, and thereby predict the outcome of a set of processes in terms of its impact on the landscape and constituent engineering structures. This is clearly a valuable skill. Some geomorphologists have this mathematical background also; but the scope of fluvial geomorphology is much less heavily focused on mathematical modelling. Instead of the numerical modelling of assumed processes, geomorphology is far more concerned with the identification of actual processes, even if this leads to a less rigorous process-response model. The two approaches are of course complementary. It should be emphasized that, just because a numerical model might not accurately predict nature, this does not necessarily mean that there are errors in the logic of the model: it may simply mean that the model is using the wrong processes!

It cannot be overemphasized that, long before the stage of mathematical modelling of landscape processes, the fundamental requirement is the correct identification of the processes actually operating in the landscape. There is no sense, for example, in designing protective measures at a particular site on a meander bend, if it can be shown that, in a matter of decades, the locus of bank attack will shift to another site on the bend; or if the natural behaviour of the river is such that it is likely to switch its course from one place in the floodplain to another through avulsion.

An overall appreciation of what the river has been doing over a period of many years is therefore obviously indispensable in planning any river project. Sometimes, this knowledge can be gleaned by examination of repeated aerial photographs of the river reach, but often such successive photography does not exist, or does not extend back far enough in time. How then do we establish what the river's normal behaviour is? One approach is through the morphology of the river system itself, this being its own way of showing what it has been doing over the years.

The field of fluvial geomorphology has been a popular one for many years and there are many textbooks (e.g. Richards, 1982), and thousands of scientific papers dealing with the subject. It is impossible to summarize the full scope of the subject in this single chapter. Instead, three main themes are developed:

(a) the use of floodplain morphology as an indicator of channel behaviour;

(b) the use of changes in channel morphology as a means of determining rates of operation of present channel processes;

(c) the use of channel morphology as an indicator of stream and river behaviour.

6.1 Floodplain morphology and channel behaviour

The existence of different types of alluvial rivers (rivers that flow in channels set in their own floodplain alluvium) is widely recognized in the literature, albeit often at a somewhat superficial level. Standard engineering texts, for example, distinguish between straight river reaches, meandering reaches and braided reaches, and descriptions are provided of the differences in river behaviour among the different types of reach. This is a beginning, but it is extremely simplified, and potentially dangerous. There are many types of straight reach, meandering reach, and multi-channel reach (generally grouped under the term braided). These different types of the main categories of alluvial river channel themselves correspond to different types of river behaviour.

It is impossible to discuss all the different types here, partly because of space restraints, and partly because not enough is yet known about river behaviour to produce the definitive classification. But progress is being made, and the category of floodplain types shown in Fig. 6.1 (from Lewin, 1978) is a useful starting point.

The 8-type floodplain classification includes three different types of so-called "meandering" (A,B,D) and "braided" or "multi-channel) (F,G,H) floodplains, but also two hybrid types which combine features of both (C,E).

6.1.1 Split-channel floodplains

The classic textbook image of a braided channel is the Type H floodplain of Fig. 6.1, termed by Lewin "a braid-bar plain with switching zone of active channels". Fig. 6.2, the braided channel of the Muddy River in Alaska, and used on the cover jacket of "Fluvial Processes in Geomorphology" (Leopold et al., 1964) is perhaps the most-widely seen photograph of this type of channel. The diagnostic feature of this type of floodplain is that the many channels are extremely unstable, being displaced through rapid bank erosion, and, more importantly, through blockage of channels by bed material and diversion (avulsion) to another part of the floodplain. Only a part of the floodplain channel network is occupied by flowing water (even in times of flood), the rest reflecting areas of former flow which have been temporarily abandoned by avulsion. This multichannel braiding is characteristic of present day glacial outwash environments, but is certainly not restricted to them. Typical floodplain sediment is sand and gravel.

The Type G floodplain is also multichannel, but radically different in character and process. The density of channels per unit area of floodplain is smaller, and, most importantly, the network of multiple channels seems to be much more stable over time. Typically, these floodplains are characteristic of much gentler valley floors than the Type H floodplain, and with finer sediment (sand, silt and clay). The alternation of "braided" and "backwater" reaches along the North Saskatchewan River (discussed in Chap. 4) shows the contrast in morphology and setting of the two types very well. The channel pattern associated with the Type G floodplain is generally referred to as anastomosing. They are characteristic of areas of aggradation associated with downstream controls, e.g. in reaches upstream of areas of locally rising base level (as on the North Saskatchewan River) and upstream of prograding deltas (Fig. 6.3), e.g. on the Mackenzie and Saskatchewan rivers. Avulsions do occur on anastomosed floodplains (as will be noted in connection with the Saskatchewan River delta: Chap. 8), but are separated by much longer periods of stability than in the case of the Type H floodplain.

A word of warning is needed here. The literature of fluvial geomorphology is still plagued by terminological confusion: the same term is sometimes applied to different floodplain types, and different terms are sometimes given to a particular channel. This confusion is especially true in the case of anastomosing channels. Leopold et al. (1964), for example, used the terms braided and anastomosing interchangeably, and it was only later through the efforts of Schumm (1977) in the United States and Smith (1983) in Canada, that the distinctive character of anastomosing floodplains emerged. Mollard and Janes (1984) use the term anastomosing to describe the Mountain River, tributary to the Mackenzie River (Fig. 6.4), when in fact the unstable character of its multiple channels is more indicative of braiding or "wandering" as defined below.

The Type F floodplain is much closer to the multichannel braided pattern than is the anastomosed pattern, but close inspection shows differences. Though characterized by channel splitting and rejoining, the appearance of the Type F floodplain is not so much one of multiple, narrow channels as of a single, broad channel in which the flow is continually divided around mid-channel bars. The term multi-bar braiding is therefore used here to describe it. Lewin cites the middle Brahmaputra as the type example, and points to the shifting character of these bars as one of the diagnostic features of this floodplain. In Canada, the braiding along the South Saskatchewan River (before impoundment of Lake Diefenbaker) and along the William River (noted in Chapter 4) would seem to fall into this category. Multi-bar braiding appears to be characteristic of sand-bed channels, whereas multi-channel braiding seems to be more common in gravel-bed channels.

In looking at split and multiple channels in western Canada, it is clear that in many cases the real world corresponds to none of the three categories described above. The classification by Kellerhals et al. (1976), for example, emphasizes that in many cases channels are not so much subdivided into multiple braids as punctuated, to varying degrees, by islands and channel bars. And it is certainly true that channel splitting around islands is the dominant characteristic of many western watercourses, e.g, the Yukon River, the Peace River. And even where floodplain appearance is dominated by channel, rather than island multiplicity, appearance often does not belong to the standard categories.

Indeed, the most common split-channel in western Canada is probably the Type E floodplain, and labelled by several workers (beginning with Neill, 1973) as wandering channels. The relatively low frequency of channel splitting provides some similarity with anastomosed channels, yet the rapidity of channel shifting indicates much greater resemblance to multi-channel braiding. Observations in other parts of the world indicate that the degree of anabranching (channel splitting) in wandering rivers may vary appreciably over a period of decades (e.g. Carson, 1984, 1986), these changes being controlled by changes in the magnitude of flood peaks and the degree of post-flood revegetation of the floodplain. At times, the wandering river system may have shrunk to a meandering single-thread pattern, yet to misinterpret that state as being the norm for the river system could lead to serious mistakes in floodplain planning. Such misinterpretation could be avoided by examination of the floodplain supporting the stream (Fig. 6.5): the microtopography of the floodplain of wandering rivers is quite distinctive, being dominated by anabranch scars formed at the time of shrinkage from a multichannel system. This topography is quite different from that of the floodplains of low-energy meandering rivers as discussed shortly.

The last comment indicates why the recognition of these flooplain types should be considered important to anyone involved in river management. The human mind subconsciously attempts to organize its knowledge by subdividing information into different categories. The practice is sound, but only to the extent that the allocation process is correct. To someone with little training in landscape observation, one meandering river system may look like another: yet to allow development of a floodplain that is occupied by a dormant wandering channel could prove disastrous.

Church (1983b) makes a similar point in his discussion of the Bella Coola River of British Columbia, where the floodplain alternates in appearance along its length from stable, single-thread reaches (with limited bed mobility and stable channel banks) to active wandering reaches (where bed load transport is much greater and, related to it, channel bank erosion and avulsion, and therefore anabranching, are more common place). The recognition of these different reaches is clearly important in any scientific monitoring program: data for bed load transport, for example, would depend entirely on where, within the length of the valley, a sampling site is located. An understanding of how these different reaches evolved is clearly important in floodplain management: this is especially the case given that these different reaches are themselves mobile, with wandering reaches shifting downvalley into presently stable single-thread reaches. The simple task of planning a bridge crossing in one such single-thread reach could have unfortunate longterm consequences if undertaken without this kind of knowledge.

Split or multiple channels are common in the west. In the descriptions above, some attention has been directed to the contrasting lateral stability of the channels in the different floodplain types. To anyone concerned with floodplain development, lateral stability of channels is usually a far more important topic than vertical stability (aggradation versus degradation) simply because the changes are more rapid. The different types of split-channel floodplain also seem to correspond to differences in vertical channel stability, however, and this is something that should also be borne in mind.

6.1.2 Meandering-channel floodplains

One of the most common questions posed to fluvial geomorphologists is why some rivers "braid" while others "meander" . No attempt is made here to answer that question, except indirectly in relating floodplain morphology to channel behaviour. The popular view that "braiding" is dependent upon high rates of channel aggradation is certainly oversimplified. In any case, as indicated in Fig. 6.1, there are different types of meandering-channel floodplain indicative of differences in behaviour even within the group of meandering channels.

There is a strong contrast, for example, between the Type A and Type D floodplains. The latter displays a much weaker sinuosity (defined in Fig. 1.1), related to the fact that channel migration is primarily downvalley with only limited extension of bends at right angles to the main flow direction. Type D floodplains comprise confined meanders in which the width of the meander belt (and the extent of across-valley bend migration) is restricted by relatively resistant valley walls. In the prairies, confined meanders are common on the valley floors of former glacial spillways, where the "valley bottom" is actually the old channel floor of a meltwater stream (and therefore relatively narrow). Under the present weak runoff regime, these rivers are flowing on floodplain gradients that are too gentle to allow rapid undercutting of the valley walls, and therefore the floodplain remains narrow. A particularly distinctive feature of confined meanders is their planform asymmetry (Fig. 6.6) in which the river flows towards the floodplain margin almost at right angles, and then is forced to turn abruptly downvalley. On the other hand it will be seen that this feature exists (but is less well developed) on wider floodplains too (Carson and Lapointe, 1983).

Over much of Canada, the Type A floodplain (in which there is no lateral confinement by the valley walls) is the most common of the three meander patterns shown in Fig. 6.1, and it is commonplace in the wider spillways of the prairies. The distinctive feature of these valley bottoms (as also found in Type D) is the scrolled floodplain appearance in which ridges of former "inner bank" locations indicate the way in which the meanders have moved across the floodplain. In the case of Fig. 6.6, the pattern of these scrolls indicates the dominant downvalley migration of meanders in contrast to the across-valley migration portrayed in Fig. 6.1(Type A); Page and Nanson (1982) discuss the evolution of scrolls in Type A floodplains, including reference to the Fort Nelson River, B.C. The scrolled character of the Type A and D floodplains is the diagnostic feature in contrasting them with the meandering Type B floodplain (Fig. 6.7) where such scrolls are absent.

The existence of meander scrolls on a floodplain is clearly helpful in understanding the pattern of evolution of the meander geometry. Its importance goes far beyond that though: it tells us something about the sediment regime and the vertical stability of the channel. The fact that scrolls exist indicates that overbank deposition cannot be very important on such floodplains: if it were, then the scrolls would soon become masked by a veneer of fine-grained alluvium from overbank floods. This weakness of overbank deposition might result from one of two factors: low sediment concentration in overbank floodwaters, and relatively infrequent spillage of floodwater over the floodplain.

There is certainly some indication that low sediment concentrations are a feature of scrolled floodplains. Many, if not most, Type A and D floodplains are associated with sand-bed rivers, in which the suspended load is also predominantly sand (fine sand) with relatively little silt and clay. As a result, during overbank flows, there is little fine material available to disperse over the floodplain away from the channel. In contrast, the sandy sediment in the main flow (usually deflected away from mid channel towards the inner bank because of "spiral" flow in a meander bend: Fig. 6.8) quickly settles out along the edge of the floodplain (adjacent to the inner bank) because of the reduced turbulence in the flow over the floodplain. The result is a levee along the inner bank, which, after subsidence of the water level, becomes colonized with vegetation. The ridge continues to grow in subsequent floods, but eventually sufficient scour along the opposite (outer) bank widens the channel, and the inner bank shifts away from the levee. The levee becomes part of the scroll ridge complex of the floodplain, while a new levee starts to build up along the new inner bank position. (The details of the process are explored by Hickin and Nanson (1975) and Nanson 1980).

The limited extent of overbank sediment deposition on scrolled floodplains is not, however, solely related to sediment levels. It is clear that scrolled floodplains rarely show widespread flooding: inner areas next to the channel may be overtopped, but much of the floodplain remains above water except in extreme floods. One common reason is that the scrolled floodplain is actually a mosaic of different floodplain levels, formed because the meandering river is slowly cutting down at the same time as it is migrating over the floodplain. In the field, this is evident in the fact that undercut banks are typically higher than the depositional banks on the other side of the channel. The different scroll areas in Fig. 6.1(A) are therefore likely to be at different levels, being higher away from the river as shown in the schematic cross-section above the plan sketch.

The fact that scrolled floodplains are usually associated with rivers that are downcutting is itself important in many different ways. In the context of floodplain management, for example, it raises the question as to exactly what is meant by the floodplain: the risk of flood inundation certainly varies appreciably across the valley bottom even though there may be no obvious terraces. It is also important in the context of bank stability: it means that the basal part of the channel bank is not formed by floodplain alluvium, but by the older sedimentary strata into which the river is downcutting. This older sediment may be much more variable than the overlying alluvium: in places it may be sand (easily scoured and therefore conducive to rapid bank erosion) while in other places it may be lacustrine clay (much more resistant). Thus the existence of scrolls automatically alerts the riverscape developer to the fact that the bottom sediment of the channel banks (not easily seen) is likely to be a factor in any attempt to predict sites of active bank erosion.

Not all meandering rivers have scrolled floodplains: the Type B floodplain does not (Fig. 6.7). The absence of scrolls on such floodplains again conveys important information about the sediment regime and stability of the meandering channel. Arguing in a direction opposite to the discussion above, it would be anticipated that: (a) the absence of scrolls on a meander floodplain indicates widespread overbank sedimentation; (b) this higher level of sedimentation reflects a greater abundance of fine sediment in channel flows; and (c) it also indicates more widespread flooding of the valley bottom, associated with the fact that the channel bed is aggrading. This reasoning seems to be valid in most cases. The floodplain in Fig. 6.7 is part of the Qu'Appelle valley, north of Regina, Saskatchewan: the valley is actually an old spillway, the floor of which has gradually been built up in post-glacial time. The dark patches in the photograph represent waterlogged low-lying areas away from the main channel, and correspond to the flood basins illustrated in Fig. 6.1(B). The floodplain plan morphology (so evident in aerial photographs) is a direct reflection of the less obvious cross-sectional geometry: contrast the valley floor "bulge" on cross-section B with the cross-section in the Type A floodplain of Fig. 6.1.

The interpretation above does not exhaust the inferences about the behaviour of meandering channels that can be made from floodplain morphology. Another logical inference is that scrolled floodplains involve higher rates of bank erosion than the Type B floodplain: that is, even if there is some floodplain sedimentation of fines, scrolls occupy a much larger part of the Type A and D floodplain because they are formed much more quickly. This itself is consistent with the previous reasoning: bank erosion is most rapid in areas of sandy channel bank material (typical of scrolled floodplains) and much slower where channel banks are made of clay (as inferred for non-scrolled floodplains).

The implications for floodplain development should be obvious. Channel straightening (by cutting of meanders), for example, is frequently advocated and implemented as a means of reducing flood hazard by increasing flow velocities through a reach and (arising from the higher velocities) activating channel scour and thus increasing cross-sectional capacity of flow. Whether or not such scour actually takes place will depend partly on floodplain gradient and partly on channel and floodplain sediment. It should be clear that Type B floodplains (those that may be the most severely affected by flooding) are less likely to respond to straightening than scrolled floodplains: the prevailing situation is already one of aggradation, rather than slow degradation; and channel sediment is usually fine-grained alluvium throughout (and at depth) - sediment that is much more likely to resist channel scour.

6.2 Measurement of process rates from changes in channel morphology

The conventional engineering approach to process measurement is through installation of appropriate instrumentation at specific sites for periods which, because of the urgency of the data, are usually relatively short. The approach is sound, but on its own, incomplete. How representative, for example, is the site used in terms of other sites in the reach? The expense of the instrumentation approach may not allow this question to be addressed. There is also the question of how representative the data are at that site in relation to periods longer than short-term monitoring.

The answers to these supplementary questions are frequently provided by the landscape itself: by changes in the river morphology over time. Two examples are given here to indicate the potential of the geomorphological approach: (a) rates of bend migration along meandering rivers; and (b) rates of bed material transport along straight channels. The general thesis of this section is, however, much broader than this. Anyone dealing with rivers needs to know how channels behave over time, before, during and after floods of different magnitudes. One of the best ways of ascertaining this behaviour (and the rate of operation) is by assembling repeated aerial photographs, and examining how the river has changed over time.

6.2.1 Rates of meander bend migration

A common approach used by geomorphologists interested in the behaviour of meandering rivers has been to search the archives of aerial photographs, and compare such photographs taken 20 or 30 years apart. The trace of one photograph can be superimposed on the trace of the other, adjusting for any distortion, and the shift in meander form mapped accordingly.

Figures 6.9 and 6.10 show an example from a bend on the Rouge River of Quebec. The comparison of 1969 and 1978 photographs reveals that, although the pattern of channel shifting during that period was essentially the same as in the 1928-1978 period (and in earlier centuries as represented by the scroll lines), the magnitude of channel migration was much greater in the last ten years. The latter period coincides with increased flood peaks in those years following curtailment of the flow regulation that had previously been used for driving log booms downstream. The air photograph comparisons afford a simple method of quantifying the effects of basin deregulation on channel stability.

Nanson and Hickin (1986) used the same approach on 18 rivers of the Interior Plains. Based on photographs taken 20 to 30 years apart, they were able to map the extent of bank erosion around the margins of selected bends, and then examine the factors controlling differences in the maximum migration rate (which varied from 0 to more than 10 m/yr) among the different bends. The information was utilized to develop a field-based equation for the prediction of mean migration rate in terms of height of channel bank, the 5-year flood discharge, channel slope (based on 1:25,000 maps) and median particle diameter of basal sediment in the outer bank. The equation predicted almost 70% of the variance of meander migration rate on these relatively large, sand- and gravel-bedded streams. The work provides another example of the usefulness of aerial photograph comparisons in the determination of rates of operation of meander processes.

6.2.2 Rates of bed material transport

The volume of bed material moved through a river reach is crucial information in many river problems, e.g. in establishing guidelines for the rate of extraction of bed sediment in areas of mining of granular sediment, in establishing the costs of ongoing dredging programs in areas where navigation is affected by shoaling, etc. But bed material movement is, for many reasons, not easily measured by conventional engineering procedures using bedload samplers. In addition, in sand-bedded rivers, there is the additional problem that much of the movement of bed material occurs intermittently and temporarily as part of the suspended load. Yet while bed material movement is difficult to measure directly, in many instances it can be inferred indirectly, through determination of the changes in river bed morphology over time.

One example of this approach is given here using data from the sand-bedded reach of the lower Mackenzie (Fig. 6.11), where the straight channel has a bed configuration dominated by pools and riffles (shallow areas), alternating from side to side. These bedforms are huge features, averaging about 10 km from the upstream end of a pool to the downstream end of its associated bar. This kind of bed geometry is well-known to fluvial geomorphologists (e.g. Thomson, 1986: Fig. 6.12), and in straight channels such as the lower Mackenzie, the whole bedform pattern is known to move gradually downriver. This occurs during floods when scour takes place on the upstream side of a bar, the entrained sandy sediment being carried over the bar front and settling out of suspension in the upstream part of the pool area ahead of the bar. Over time, what was once a pool area will become a riffle, and vice-versa. This process of downstream shifting of the pool-riffle sequence is well known to those with the responsibility of having to mark the navigation channel on the Mackenzie. The buoys (taken out of the river prior to freeze-up) need to be relocated each year after the freshet has advanced the bedforms.

The amount of movement of the pool-riffle system, together with the depth of the pools and the width of the river, is clearly a measure of the volume of sandy bed material moved past a given cross-section in a given amount of time. In the case of two-dimension pool systems (the geometry being uniform across the channel as in Fig. 6.13, rather than alternating) the bed material moved in a given time interval is readily given by the height of the bar front and its distance of movement downstream. In the case of alternating pool-riffles, the procedure is slightly more complex, but the same in principle. Thus repeated surveys of the channel bed each year should be capable of providing the necessary data to calculate the annual bed material load transport downstream. In the case of this reach of the Mackenzie, bathymetric maps produced by the Canadian Hydrographic Service in 1972 and 1978 allowed the approach to be used with minimal expense.

The results for one pair of pool-riffle systems are shown in Fig. 6.14: a map of point scour and fill along the river reach: the vertical changes are huge, up to 10 m, while the downstream shift in the bar front was in the range of 1500 metres. The resultant estimate of annual bed material load was of the order of 4.5 million cubic metres (about 8 million tonnes of sand), a colossal amount, and one that would have been very difficult to determine based on repeated sampling of either bedload or suspended load at some arbitrary point in the channel cross-section.

The morphologic approach to the computation of bed material transport is not restricted to straight channels with pools and riffles. In a more generalized form it has been applied to meandering rivers (Neill, 1971), wandering rivers (McLean and Mannerstrom, 1985), channels dominated by island-bars (Church et al., 1986) and to braided rivers (Carson and Griffiths, 1989), though additional assumptions are needed in all these cases.

6.3 Use of channel morphology as indicator of stream behaviour: application to WRB sediment program

WRB sediment station programs tend to follow a common sampling strategy on all large rivers, irrespective of river type. A fixed location in a river cross-section is designated as the so-called "single vertical" (formerly known as "daily site") and virtually all samplings are done at this vertical (usually comprising more than 95% of all samplings).

In recognition of the fact that the single vertical may not adequately represent sediment levels in the cross-section as a whole, occasional samplings are then made on multiple verticals across the section in order to compare sediment concentration on the single vertical (SV) with mean sediment concentration in the section. The ratio between mean sediment level and that found at the single vertical, at these times of comparative sampling, is termed the "k-factor"; and the value of this k-factor can then be used to adjust SV concentrations to those values which would be expected for the full river section.

There is, in principle, nothing wrong with this approach: indeed, it is a sensible strategy in the face of limited resources. In those situations where the fluctuations in the k-value over time are short-term and random, the appropriate adjustment can be made using the average of the individual k-values at the site. In those situations where a systematic annual pattern of scour and fill imposes a definite cyclical pattern in the k-values, then the adjustment process could take this cyclicity into consideration through the construction of a k-value time curve.

What is worrying from a geomorphological standpoint, however, is that in most cases WRB seems to pay little attention to the processes responsible for fluctuations in the SV k-factor values. As a result, it is sometimes assumed that if there is little change in the k-value in the first one or two years of the sediment program, this then justifies the use of the mean k-value for those two years on a long term basis, without any further multiple vertical sampling. Yet it is evident to fluvial geomorphologists that, though this assumption might be valid in some rivers, it would be totally inappropriate in others, and could lead to huge errors. Before commenting further upon why k-value changes can vary markedly between different types of river reach, it is necessary to briefly point out why the k-value approach is needed at all, i.e. to consider why it is that suspended sediment concentrations vary across a river channel.

The main reason for lateral changes in sediment concentrations in a river (at least at sites that are well downstream of major tributary inputs) is the fact that fine-grained sediment on the channel bed is picked up at high flows more easily at some parts of the cross-section than at others. As an example, if there is a bar of fine sand on the left side of a river upstream of the sampling cross-section (but not on the right), then any SV site on the left side of the section will show higher concentrations than the mean for the section because of the inclusion of larger amounts of bed material in suspension. Thus, if logistical constraints force selection of a SV site on a section downstream of a bend, frequent multiple sampling checks will be needed to establish the appropriate k-value.

As one example of intra-section variability in sediment levels downstream of a bend, Fig. 6.15 shows the location of the 1980s measurement section on the Mackenzie River upstream of Arctic Red River. The SV site is located slightly downstream of the section and, for logistical reasons, close to the village side of the river. It should be noted that the 1973 bathymetric map shows, as might be expected, a weak sand bar on the "inner" left bank of the bend, though this did not appear to extend as far as the measurement section at that time. Fig. 6.16(a) shows the general symmetry of the measurement section on a day of multiple vertical sampling in 1980, but it should be noted that, even then, there was some indication of an increase in sediment concentration towards the left bank, primarily (but not entirely) due to higher sand concentrations on that side.

By 1986, it was clear that the inner-bank sand bar had extended as far downstream as the measurement section (Fig. 6.16(b)). And multiple vertical sampling continued to show higher sediment concentrations on the left side of the river as well as in the bottom half of sampling verticals. The pattern is exactly what would be expected from the location of the measurement section: the plume of high sediment concentration (produced by higher levels of suspended bed material) is deflected towards the inner bank as flows move round the large bend, and, though some of the sand in this plume does settle out along the left side of the reach (which is why the bar extended downstream), above-average sediment levels persist along the left side through the measurement section, continuing, in fact, to the SV site further downstream.

Thus, during the 1980s, the comparative data for multiple-vertical and SV samplings at this station, showed an increasing tendency for the SV concentration to exceed the mean value for the measurement section; that is, the k-value for the site gradually decreased from values in the range 1.25-1.75 in 1980-81 to 0.86-0.92 in 1984-85 (Carson, 1988b). This change was directly associated with an increase in the percentage sand in samples from the SV site. In addition to affecting the accuracy of computed sediment loads at this site, the systematic change in the k-value brought out by the increased sand levels at the SV site, would also have had implications for any sediment quality sampling using the same SV site. The concentration of contaminants in suspended sediment is to a large extent affected by grain size: coarser sediment has much less capacity for contaminant adsorption. It seems likely, therefore, that had a sediment quality monitoring program been in effect at this station, the data would have indicated a gradual improvement in sediment quality during the 1980s, a pattern that (while true of the SV site) would have been totally unrepresentative of mean conditions through the full cross-sectional width of the river.

It should be clear from the above that the extent of change in k-values at a station (and hence in the frequency of multiple-vertical sampling needed) will depend in large part on the type of bed material in a river reach and upon the plan geometry of that channel reach. This is something that should be recognized at all levels in the sediment program: by management staff in planning the operations at a given site, and by field technical staff who (because of their frequent visits to the site) are best in a position to observe whether changes in bed sediment and bathymetry are taking place in the reach.

The most important point in this section, however, is simply the recognition that a qualitative understanding of what is happening to sediment conditions in a river reach is an essential prerequisite to any sediment monitoring program there. In addition, the degree of lateral variability in sediment levels in a river section (and hence the complexity of the sediment operations at a station in that reach) can usually be anticipated (before the onset of the sediment program) by a geomorphological assessment of the reach. In some river reaches, the k-value may change little over decades; in others it may change significantly, sometimes in distinct cycles.

As an example of the last point, consider the hypothetical case of a SV station being set up close to the left bank in the straight pool-riffle channel of Fig. 6.14. In 1973 it would have corresponded to a position immediately downstream of the left-side bar front: at that time, samplings at the SV site would have contained large quantities of sandy bed material being scoured from the upstream side bar (A) and transported in suspension to be deposited in the pool downstream. Sediment concentrations would have been much higher than on the opposite bank, where much of the sand scoured from the upstream flank of bar B would have settled out in the upper part of its associated pool. The k-value for the SV site would therefore have been substantially less than unity. In contrast, by 1977, the front of the left-bank bar (A) would have passed downstream of the SV site: there would therefore be less of the upstream flank of the bar available to contribute sediment to the sampling vertical. The "excess" concentration at the SV site would therefore be decreasing and the k-factor would be higher. Eventually, the entire length of bar A would have moved past the SV site, the location of which would then correspond to the downstream end of pool C. At this time there would be no excess sediment at all, a marked contrast to the right side of the channel which would then be receiving all the sand scoured from the upstream flank of bar B: the k-value at this time would then be well above unity.

This is not the place to pursue this topic at length: it is sufficient simply to emphasize that the inherent precision of a SV sampling program will be strongly affected by channel bed conditions. In fully gravel-bed rivers (but not mixed sand-gravel), there is likely to be little problem: the entire suspended load is "wash load" and the k-value should deviate little from unity. The same situation would also arise in channels cut in stiff-clay and lacking any mobile bed sediment. In sand-bedded channels (particularly those with abundant fine sand), the situation can be much more complex as just indicated.

6.4 Endnote

The three sections above provide examples of the usefulness of the qualitative understanding of river behaviour that can be gained from an appreciation of channel and floodplain morphology. The examples were chosen because of their potential importance in IWD work, but they were introduced simply to illustrate a much more general point: while it is a relatively simple matter to produce numbers (sediment transport rates, bank migration rates, etc.) in river channel studies, it requires much greater effort to ensure that those numbers are meaningful. The qualitative appreciation of landscape processes gained from assessment of landform morphology will never replace mathematical modelling of riverscape processes, but it represents an important adjunct to it.

The same comment was made by Kellerhals et al. (1976):

> The large potential benefits of successful river engineering works, combined with the dire consequences of failures, have provided one of the earliest and greatest challenges to the profession. Today this challenge continues to be met in research on river-related processes and by systematic collection of river data. Much progress has been made in both respects, but it is the writers' contention that the bias of most engineers towards readily quantifiable topics has lead to a serious gap in this work, the neglect of interpretive work on river-related landforms.

It is worth emphasizing that two of the three authors of that quotation are river engineers.

PART III

REGIONAL EXAMPLES

The third part of this report provides a summary of the regional geomorphology (together with some indication of its potential importance in the context of regional IWD issues) for four of the major river basins of the Western and Northern Region: the Assiniboine basin; the Saskatchewan River basin; the Peace-Athabasca basin; and the Mackenzie River basin.

It should be emphasized, however, that the chapters are not meant to constitute overviews of all regional issues. They are intended to simply highlight those issues in which an appreciation of the regional geomorphology and an understanding of general principles of geomorphology can be readily seen to offer IWD staff a useful perspective. Some of the examples have been touched upon briefly in Parts I and II.

In some instances, the same themes are common to all four drainage basins, but in other cases, there are substantial differences between the basins. There are, for example, pronounced differences in the way in which the southern drainage basins have evolved during the Holocene. The river network of southern Alberta plains (Fig. 5.2) constitutes an amalgam of truly fluvial SW-NE flowing reaches (often involving incision into drift laid down in preglacial valleys) and NW-SE trenches cut out by glacial meltwater and spillway outflows along the margins of the shrinking Laurentide icesheet. The two types of reach are also evident in the Saskatchewan part of the Saskatchewan River basin, but most of the drainage in that area corresponds to SW-NE watercourses that were formed, not by glacial meltwater draining away from the ice front, but by terrestrial runoff draining northeast towards Glacial Lake Saskatchewan and Glacial Lake Agassiz (Fig. 2.10 and, more completely, Christiansen, 1979, Figs. 12-19). In contrast, the Assiniboine drainage of southeast Saskatchewan and southwest Manitoba is dominated by glacial spillways (Figs. 7.1-7.4). Thus different basins in the prairies have radically different types of rivers depending on the exact nature of deglaciation in each case. This will become apparent in comparison of Chapters 7 and 8.

Finally, these regional discussions should not be regarded as definitive statements. In most cases, the commentary is based on limited geomorphological information, often gathered (usually by staff of the Geological Survey of Canada) for specific purposes and therefore incomplete in other contexts. The discussions are intended to indicate directions rather than to represent conclusions.

7. THE ASSINIBOINE RIVER BASIN

7.1 Overview of drainage evolution

As noted in the preamble to Part III, the Assiniboine drainage network largely corresponds to the pattern of glacial spillways cut during the immediate post-glacial period as the ice front of the Laurentide icesheet retreated towards the northeast. Fig. 7.1, for example, shows the drainage at an early stage of deglaciation (about 14,500 years before present [BP]) at which time meltwater draining from the ice front had already initiated the Souris channel and the Pipestone spillway, while outflows from Glacial Lake Souris, draining along the ice front to the southeast, were carving out the Pembina spillway of southern Manitoba.

By about 14,000 BP, additional retreat of the ice front had exposed low lying land near Regina (submerged as Glacial Lake Regina) and Glacial Lake Souris had shifted northeast along with the ice front to become the first stage of Glacial Lake Agassiz (Fig. 7.2). Drainage from Glacial Lake Regina and meltwater flows along the Qu'Appelle Channel poured into Glacial Lake Agassiz depositing sediment to form the "Assiniboine Delta" in the vicinity of Brandon.

By 12,500 BP, the ice front had retreated sufficiently far north (Fig. 7.3) that Glacial Lake Regina ceased to exist, the main ice-marginal lakes in Saskatchewan being Glacial Lake Saskatchewan and Last Mountain Lake. Thus meltwater drainage down the Souris valley had ceased by this time, though it continued down the Qu'Appelle-Assiniboine route.

A little later (about 11,700 BP), the northeasterly shift in Glacial Lake Saskatchewan (in contact with the retreating ice front) had isolated it from the Qu'Appelle route (Fig. 7.4), and all meltwater drainage was then via Glacial Lake Melfort (south of the Pasquia Hills in the Carrot and Red Deer basins) into the Assiniboine spillway. Ultimately (about 11,300 BP) the ice front had receded sufficiently far to the northeast that Glacial Lake Saskatchewan merged with Glacial Lake Agassiz, and the Assiniboine basin was completely isolated from meltwater drainage.

In short, the major valleys of the Assiniboine system were cut in a very short period (a few thousand years during the immediate post-glacial period) by huge discharges of glacial meltwater.

The present day rivers are considerably smaller and are distinctly "underfit" in relation to the size of valley that they occupy. The morphology of these valleys, and the subsequent river behaviour, reflects this distinctive mode of origin. As an example, meltwater channels with smaller flows (coming from only a small lobe on the ice front) and those in which meltwater flows lasted shorter periods did not cut down as much as the main channels. The result is that the floors of some of these smaller valleys are left "hanging" above the level of the main valley, their gradients steepening in the lower parts of their courses as the present rivers try to cut down to the lower base level of the main stem.

In addition, these old channels were cut to very gentle gradients (of the order of 10 m per 100 km or less): base level for the rivers was usually fixed by a downstream lake or rock outcrop, while the huge discharge of these meltwater rivers allowed degradation of the valley floor which (given the fixed base level) resulted in very gentle gradients (Fig. 8.8). Though the post-meltwater rivers have not had to carry anywhere near as much sediment as the meltwater streams, the combination of their very low discharge and low gradient has meant that in most cases the Holocene rivers have been unable to transport the sediment that is supplied to them and, as a result, aggradation has resulted in post-glacial time.

Alluvium in these valley bottoms is commonly 60 m or more thick, though it appears that in most cases the basal alluvium (and related sediment) is older than post-glacial: in other words the "trenches" in which present day rivers flow must have been cut during earlier periods of deglaciation (prior to the late Wisconsin episode), partly filled with drift at a later stage, and then re-excavated (but only partly so) during the Wisconsin deglaciation.

The sources of sediment that have been responsible for the aggradation of these rivers has not been identified with certainty; there is a general feeling that aggradation slowed down dramatically about 7,000 BP (i.e. it was a paraglacial phenomenon) though the evidence for this is not clear. It seems probable, though, that a distinction exists between true spillways (outflows from glacial lakes) and localized meltwater channels. In the former, there would be no bed material transport at the head of the reach (because of trapping in the proglacial lake), allowing ongoing downcutting as long as lakewater outflows occurred. In the latter case, the extent of degradation would depend on the amount of bed material carried by the meltwater as it left the ice front. At one extreme this might have been so little that the flow would act like spillway flow and produce degradation. At the other extreme, concentrations could be so high that instead of degradation, aggradation would occur. In particular, during the last stages of deglaciation, it seems probable that many non-spillway meltwater channels experienced a decrease in discharge (relative to bed material load), resulting in partial infilling of the channels which the flows had carved earlier in their history.

In any case, it should not be assumed that aggradation was necessarily synchronous through the Assiniboine basin. On the contrary, as the water level of Glacial Lake Agassiz fell (contrast the shoreline position near and east of Brandon in Figs. 7.2 and 7.4), and the lake itself eventually subdivided into the lakes of the present day (Lake Winnipeg etc.), so the lower Assiniboine River would have been rejuvenated: this is reflected in the cutting down of the river into the Assiniboine deltaic sands at Brandon. Theoretically, this rejuvenation would work itself backwards up the stream system, resulting in a progressive change from aggradation to degradation as the nickpoint moved upstream. The nickpoint is presently located at Brandon (Fig. 7.5) where its headward erosion has been slowed down by the exposure of bedrock outcrops in the river bed. Thus channel degradation has been taking place in the lower Assiniboine while aggradation has been occurring upriver of the nickpoint.

With this background in mind, it is now appropriate to examine certain parts of the basin in more detail.

7.2 The Qu'Appelle and Souris river basins

The Qu'Appelle River is a classic underfit stream, meandering sluggishly across the floodplain in a meander belt that rarely impinges on the valley walls. In its upper parts (between Elbow and Craven, north of Regina) it flows over the lake bed of former Glacial Lake Regina, and, slightly to the east, around Ft. Qu'Appelle, the channel is underlain by sediments of Glacial Lake Indian Head (Fig. 7.1, 7.2). Over most of its length, the river occupies the floor of a trench cut as a spillway by meltwater pouring out of the ice-marginal lakes.

Sediment from landslides and alluvial fans on the walls of the spillway have, during post-glacial time, contributed to a thick valley infill averaging about 45 m in thickness (Fig. 4.2). The thickness of Holocene sediment on the old spillway floor is, however, very variable. As indicated in the schematic long profile of the valley (Fig. 7.6), localized inputs of colluvial sediment (from gully-fans on the side walls) have impounded lakes over much of the valley floor, the present depths of the lakes having been increased by various control structures. At the upstream end of such lakes the Qu'Appelle delivers its own sediment, and develops a tortuous course as it extends this deltaic deposit into the lake. It seems likely that much of the floodplain of the Qu'Appelle has developed in this way (Fig. 7.7). In other words, much of the length of the floodplain is probably neither Type A nor Type B (Fig. 6.1), but formed by the progradation of the river into a lake which eventually disappeared as it was systematically infilled from upstream.

This inferred pattern of floodplain evolution may well be relevant in the context of present-day floodplain problems. The numerous channel bends, in combination with the overall low valley bottom gradient, severely limit the ability of the channel to convey water downstream at times of medium and high flows. The result is overbank flooding, an issue of some importance given the extent of cultivation of the valley floor. The problems are typical of many so-called "mature" valley bottoms, where the standard engineering solution is to straighten channel alignment thus increasing velocities and (hopefully) initiating sufficient channel degradation to increase the cross-sectional area of water flow. Yet the floodplain morphology of the Qu'Appelle valley is not that of a typical "mature" valley bottom: the Qu'Appelle channel has all the problems associated with Type B floodplains (fine-grained sediment that is not necessarily easily eroded), and valley gradients that are unusually gentle because of floodplain construction by progradation into level lakes. Initial reports (Miles, 1990, pers. comm.) suggests that channel straightening has not been particularly successful in alleviating flooding on the Qu'Appelle: nor should it have been expected to be!

Though sediment supply to the Qu'Appelle is believed to have been reduced in the late Holocene, it is clear that even today, where the stream abuts the valley side, there can be local buildup of bed levels, and hence flattening of channel slope upstream. The photograph in Fig. 7.8, for example, appears to show an area where blocks of landslide material (presumably the same marine clay shale described by Mollard, (1986) have, at some time in the past, slid sufficiently far into the floodplain to have partially blocked the channel course (see bulge in long profile at bridge site). Klassen (1975) comments upon charcoal deposits dated at 4300 BP (by C-14 dating) that were found in the floodplain buried by more than two metres of alluvial silt: this is indicative of aggradation, at least locally, of the order of 5 cm per century.

The speculative observations on the formation of the Qu'Appelle floodplain put forward above may be relevant in riverscape planning. Aggradation that is due to localized inputs of sediment from sidewalls may be easier to control than aggradation that is due to the overall channel gradient being too gentle to transport bed sediment supplied from upstream. In addition, it provides a new perspective to any valley side slope or top slope development: there are likely to be immediate floodplain impacts if sediment from these areas accumulates in the sluggish river channel.

There appears to have been less collection of geological data in the Souris valley than that of the Qu'Appelle, though at first glance there is considerable resemblance between the two old spillways. The trench-like character of the former Souris spillway is particularly well-developed south east of the town of Souris in the form of the Pembina valley (Fig. 7.9). Throughout this reach, the meltwater course was a true spillway (draining from Glacial Lake Souris (Fig. 7.1) in contrast to the Saskatchewan part of the Souris which at times (Fig. 7.1) was simply a meltwater channel, while at other times (Fig. 7.2) was a spillway. It is not clear at what stage the Souris River broke out of the lower spillway course to flow into the Assiniboine. The Pembina valley - like the Qu'Appelle - contains numerous lakes where landslips on the side walls have blocked the valley floor with debris.

7.3 The Assiniboine main stem

The course of the Assiniboine, about 11,700 BP, is shown in Fig. 7.4. The upper part of the spillway out of Glacial Lake Melfort is, in fact, no longer connected with the Assiniboine main stem: the channelized runoff from the Porcupine Hills now swings abruptly east as the Swan River. The reasons for this post-glacial diversion are not clear, but, in any case, are probably primarily of academic interest. At a slightly earlier stage in the deglaciation, spillway meltwater originated from Glacial Lake Assiniboine located just north of Yorkton.

The present-day upper Assiniboine drains eastwards from the uplands north of the Touchwood Hills (Fig. 7.10), entering the old spillway at Kamsack, and then following the spillway south (and then east) to Brandon. Upstream and downstream of the Shell River confluence there was (until submergence by the Shellmouth flood-control reservoir in the late 1960's) a significant change in the character of the Assiniboine River. In the first place, the Assiniboine valley had been cut much more deeply than the Shell valley: the lower reaches of the Shell River, therefore, were oversteepened and the river had been attempting to regrade its course during most of the Holocene, while, at the same time, the Assiniboine valley had been infilling with alluvium.

Associated with the increased flux of sediment into the mainstem valley from the lower Shell River, deposition had been particularly pronounced at the confluence (in effect the Shell provided a tributary fan into the mainstem), so that the valley floor gradient upstream of the confluence was flattened while that downstream of the confluence was steepened (analogous to, but not as spectacular as, the effect of the Rampart Creek fan in the North Saskatchewan River: Fig. 4.4). In turn, the effects of this aggradation were evident in floodplain contrasts above and below the confluence, representing the Type A (downstream) and Type B (upstream) floodplains discussed in Chapter 6.

Upstream of the confluence, the channel was relatively shallow (< 2m) and the sluggish river was only weakly sinuous, which is a marked contrast to further downstream (Fig. 7.11) where pronounced scrolls attest to the rapidity of bank erosion by the strongly meandering and deeper (about 5 m) river. Whether the contrast is solely due to the change in gradient imposed by the Shell River fan is unclear. To the west of the Assiniboine valley in the downstream reach, there are fossil hairpin dunes (Fig. 7.11) indicative of movement of windblown sand to the south east. The greater activity of the Assiniboine river downstream of the Shell confluence may be related, therefore, to a sandier (more erodible) alluvium. The Type A floodplain is somewhat unusual in the aggradating valley bottoms of these former spillways.

The overall appearance of the Assiniboine valley changes little downstream of Shellmouth until the nickpoint upstream of the old Agassiz delta at Brandon is reached (Fig. 7.5). Downstream of this point, the valley floor gradient increases appreciably (more than 100 m per 100 km) and the valley becomes much deeper as the river has cut down through the old delta. Within the deltaic reach, the river has maintained its meandering pattern as it has cut down into silty-sandy sediment, and there is active erosion of outer bank walls even today.

7.4 Sediment production in the Assiniboine basin

The varying regional character of the Assiniboine River basin, as described above, exerts a major influence on the present day sediment balance of the Assiniboine River network. It would be expected, for example, that sediment loads from the Qu'Appelle and Souris rivers would be negligible: there is little bank erosion by the sluggishly meandering rivers and much of the sediment from upland erosion is deposited as fans along the edges of the wide floodplains. Most of the sediment from the upper Assiniboine is presumed to be trapped in Shellmouth Reservoir. Meander bend erosion in the more active floodplain around St-Lazare (downstream of Shellmouth) may be a more important source of sediment, but the major sediment source is likely to be undercutting of valley walls in the rejuvenated reach of the old Assiniboine Delta as suggested by Ashmore (1990).

This interpretation of sediment sources is consistent with the suspended sediment data collected at WRB sites along the mainstem (Ashmore, 1990) and listed below (note that specific sediment yields are based on "effective drainage area" listed by PFRA, 1989):

- at Kamsack, the mean annual load is about 13,000 tonnes, equal to a specific sediment yield of about 3.0 $t/km^2O/yr$;

- at Russell (downstream of Shellmouth Dam), the post-dam annual load averages about 16,000 tonnes or about 16 $t/km^2O/yr$,assuming that all sediment is derived from downstream of the dam;

- on the Qu'Appelle (near the mouth) the annual load averages about 34,000 tonnes, or about 2 $t/km^2O/yr$;

- on the Souris (near the mouth), the mean annual load is about 29,000 tonnes, or about 1.5 $t/km^2O/yr$;

- and on the Assiniboine River near Holland (downstream of Souris confluence), the mean annual load is about 660,000 tonnes, or slightly more than 11 $t/km^2O/yr$.

The main sediment source in the Assiniboine system is thus located where the river crosses the Manitoba Scarp. It is likely that the scarp is also a major source of sediment where it is drained by other rivers, but the marked importance of this sediment source on the Assiniboine is not simply due to the oversteepened reach of the river in this location, but also a reflection of the availability of an abundant supply of easily scoured silty sand.

The sediment sources that control the regional pattern indicated above should also be borne in mind in any program to assess sediment and sediment-contaminants contributed by farmland soil erosion. Stations in the lower reach (part of the Red River lowland), for example, would show insignificant proportions of sediment from farmland sources because of the huge influx from bank erosion and slumping in the Brandon reach. Stations upstream may be more sensitive to off-farm effects, but again, this would depend on actual location. It seems likely, for instance, that much of the sediment load sampled on the Souris near the mouth is due to channel downcutting and bank erosion in the oversteepened lower reach of the river - the effect of rejuvenation from the downcutting of the Assiniboine in its delta reach (much like the pattern of sediment supply on tributaries to the Lower Athabasca noted in Chapter 1).

Suspended sediment in the Qu'Appelle River (which has not been affected by this rejuvenation) may show greater evidence of off-farm sediment. The same comment probably applies to the Souris upstream of its oversteepened lower reach. A corollary of this is that, while sedimentation issues (in an engineering sense) are likely to be minor in underfit prairie streams such as the Qu'Appelle and Souris, sediment quality issues are likely to be more evident in these areas.

8. THE SASKATCHEWAN RIVER BASIN

The Saskatchewan River basin comprises five main physiographic areas (Fig. 8.1): the eastern slopes of the Rocky Mountains, the high Alberta Plains, the slightly lower Saskatchewan Plains, the Manitoba lowlands (largely the Saskatchewan River delta area) and a small area of Canadian Shield in the northeast.

The boundary between the Alberta and Saskatchewan Plains (the Missouri Coteau) is not as clearly defined as that between the Saskatchewan Plains and the Manitoba Lowlands (the Manitoba Escarpment), and, more importantly in the present context, there is no real change in riverscape at this boundary, merely a slight decrease in the depth of valleys. The real distinction in riverscape in the Alberta and Saskatchewan Plains is in the changeover from gravel-bed rivers to sand-bed streams, this transition tending to occur within the eastern part of the Alberta Plains (Shaw and Kellerhals, 1982). The regional discussion in this chapter is therefore subdivided into the following sections: the eastern slopes of the Rocky Mountains; the gravel-bedded reaches of the Alberta Plains; the sand-bedded reaches of the central Prairies; and the Saskatchewan River delta.

8.1 <u>The Rocky Mountains eastern slopes</u>

A fairly common view in Canadian hydrology is that high mountain areas such as the Rockies are areas of heavy sediment production and high suspended sediment levels in streams. Yet, as indicated in Chapter 5, except for the upper parts of basins that are currently glacierized, this view is largely incorrect. One reason for low sediment yields is that streams draining out of the mountains are frequently blocked resulting in the impoundment of lakes which act to trap sediment.

One example of such sediment trapping was provided in Chapter 4: localized areas of debris slides and flows building up alluvial (and colluvial) fans in the valley bottom, as noted in the case of the North Saskatchewan River. Mollard (1973) and Smith (1987) describe the role of similar alluvial fans in the impoundment of Jasper Lake in the upper Athabasca Valley. Another reason, as indicated in Chapter 5, is that many mountain valleys were over-deepened by valley glacier erosion in the Pleistocene (Fig.5.4), the rock-rimmed lakes still not having completely filled with sediment.

An example of sediment trapping in a rock-rimmed basin has been described by Kostaschuk and Smith (1983) in the upper Bow valley immediately upstream of Banff (Fig. 8.2). Though deposition of sediment along the Bow floodplain may not be obvious at the present time, a clue as to its existence is provided by small lakes (sometimes called the Vermilion Lakes) throughout the floodplain (Fig. 8.2, 8.3). These are relics of a much larger early-mid-Holocene lake in the valley which has gradually infilled with sediment since deglaciation. The evolution of the Banff floodplain is discussed in some detail below, not because of its intrinsic local importance, but because it probably represents a sequence of events that has been common (and still prevails) throughout the Rockies.

A section along the river axis of the upper Bow floodplain is provided in the bottom diagram of Fig. 8.3. It shows three main types of sediment: laminated glaciolacustrine clay (at great depth); bedded fine sands, silts and organic-rich clays which overlie the clay and which, in turn, are overlain by poorly sorted silt and medium sands. The interpretation placed on these borehole logs is as follows. Between about 13,000 and 10,000 BP, fine-grained sedimentation from glacial meltwater prevailed over most of the rock-basin lake bottom, producing the laminated clays found at depth in borehole BH2 (Fig. 8.3). The level of this initial lake appears to have been 1432 m above present sea level, but rapid degradation of the outflow channel through drift lowered this to the level of the present valley floor (1383 m) about 8,000 years ago.

With the stabilization of this "Lake Vermilion" since that time, there has been a systematic easterly progradation of the Bow River delta into the lake at its western end. The thick deposits of bedded clays and silts (coarsening upwards into sands) that form the main body of valley fill are interpreted as the subaqueous part of this prograding delta. A tephra (volcanic ash) layer, dated at about 6,600 BP (the Mazama volcanic event), confirms the sequential progradation of the Bow delta into this lake body: in the west the ash fell upon sandy subaerial sediment (auger holes 5 and 8), while slightly further east (splitspoon hole 4) it settled through shallow lake water onto the subaqueous delta surface; and further east (SS3) the ash settled through much deeper water in advance of the delta front slope. In fact the Mazama tephra deposit was not reached at SS3, the only ash layer found being the much more recent (2500 BP) Bridge River tephra.

Since the Mazama and Bridge River tephra periods, lake-infilling has primarily taken the form of distributaries penetrating from the southern part of the floodplain into the northern part, and thus subdividing the once-continuous lake into the small residual features that are apparent today. Kostaschuk and Smith (1983) comment:

> Eventually, with continued sedimentation and peat development in the Vermilion Lakes, this area would become a flat, relatively dry alluvial plain. The natural development of the reach, however, is being impeded by the Canadian Pacific Railway, which separates the Vermilion Lakes from the Bow River. The railway acts as an unbreachable levee that restricts most of the overbank discharge of the Bow River to the south side of the railway, retarding modern deposition in the Vermilion lakes.

This last comment indicates possible repercussions arising from floodplain development, and also the value of a geomorphic perspective in such development. The shrinkage of the floodplain area by the CPR berm presumably increases the magnitude of flood levels in the remaining part of the floodplain, though no assessment of severity of the change is known. Similarly, by preventing floodplain sedimentation north of the railroad berm, it seems probable that sedimentation rates on the remaining part of the floodplain will have been increased. Again, no data have been found, and it may well be that the sedimentation rates are sufficiently slow that they are of no significance in the immediate future.

The Banff example affords some illustration of the way in which a geomorphic enquiry into floodplain construction could offer useful insights into development of valley bottoms. The information may be of little immediate importance in Banff, and especially in other east-slope valleys where floodplain development is minimal, but as human occupance and development of floodplains and valley bottom areas increases, these insights will begin to assume significance in forestalling environmental and property damage. The other reason for this reference to the Banff case is that it probably represents a common condition in many of the east slope valleys: these streams cannot be major contributors of sediment to the plains simply because so much of the sediment is trapped before it reaches that area.

8.2 The gravel-bed reaches of the Saskatchewan River in Alberta

Discussion of the Assiniboine drainage network in the previous chapter included a step-by-step account of the withdrawal of the ice front and emergence of the ice-marginal lakes and spillways. No attempt is made here to repeat this account for the Saskatchewan River. The pattern has been documented in various GSC publications by St-Onge (in the north) and by Stalker (in the south) listed by Maurice (1988); e.g. Fig. 8.4 shows the proglacial environment in the northern part of the region about 13,000 BP. It was noted in the introduction to Part III, however, how the drainage pattern of southern Alberta differed from that of the Assiniboine valley in comprising alternating reaches of SW-NE flow (often in drift-filled preglacial valleys) and of NW-SE flow in valleys carved by meltwater pouring along the icefront to the southeast. The difference in morphology of these two types of reach is marked, particularly in connection with the abundance of valley bottom gravel, and it is this point that is developed in this section.

Shaw and Kellerhals (1982) discuss the alluvial gravels of Alberta in some detail. They comment that little of the gravelly alluvium originated from the underlying bedrock of the prairies (largely shale which would quickly wear down to fine sediment as it moved along channel beds), but is composed of limestones and quartzites that ultimately originate from a narrow belt of Proterozoic and Palaeozoic rock outcrops in the Rocky Mountains. They add: "Thus it might be thought that, in terms of provenance, the Albertan rivers are a simple case, with bed material supply being confined to the headwater reaches; unfortunately, this is not so." One of the reasons for this is that some of the limestones and quartzites were eroded from these old strata in the Rocky Mountains millions of years ago and deposited, locally, in some of the late Tertiary (Fig. 2.3) Plains sedimentary strata. A second is that the succession of Cordilleran and Continental glaciations in the region has also incorporated much of this resistant gravel into drift that mantles much of the plains, and, in particular, infills the preglacial valleys.

In other words, where rivers have cut down through large thicknesses of Quaternary drift, then the valley floor is likely to be covered with thick accumulations of gravel; where rivers have cut down into soft Plains Cretaceous shale strata (i.e. where post-glacial incision has been away from the course of the preglacial valley), the accumulation of gravelly alluvium is likely to be meagre. Thus it would be expected that NW-SE ice-marginal spillways, having cut post-glacial courses largely in prairie bedrock, would show valley bottoms quite different from SW-NE river reaches where, in most cases, the latter have cut down through drift-filled preglacial valleys (Fig. 5.2).

Examination of valley bottom morphology along the gravel-bed reaches of the Saskatchewan River system shows a pattern of alternating reaches of thick alluvial deposits and clean valley bottoms that does, in general, correspond to the pattern expected above. Fig. 5.2 shows that the North Saskatchewan River follows its pre-glacial valley over a considerable distance upstream of Edmonton. This is especially the case in the SW-NE reach between Drayton Valley and Genessee Ferry. In this reach the floodplain is dominated by a multi-thread "wandering" river, typical of valley bottoms in which relatively large amounts of coarse sediment are delivered to the stream. The abrupt change in course of the river to the southeast corresponds to a marked change in morphology, the river taking on the form of a sinuous single incised channel in a trench-like post-glacial valley (Fig. 8.5). It seems likely that changes in the coarseness of gravelly alluvium in these rivers (e.g. the finer gravel between Rocky Mountain House and Edmonton on the North Saskatchewan River: Fig. 8.6) also reflects whether or not the reach is incising old valley fill, but little research has been undertaken into this matter.

A similar contrast between some of the SW-NE flowing tributaries to the NW-SE flowing Bow River also exists in many places (e.g. the anabranching Sheep River tributary to the single-thread Highwood: Fig. 5.5). The same juxtaposition of floodplain types occurs on the Oldman River further south: Hudson and Askin (1987), for example, emphasize the meagre amount of gravelly alluvium on the NW-SE reach of the Oldman upstream of Brocket, a sharp contrast to conditions further downstream where the SW-NE flowing river has incised itself into its drift-filled preglacial valley (Fig. 5.2). The long NW-SE reaches of the Bow River (especially between Bassano and the Oldman confluence) and the Red Deer River (downstream of Red Deer townsite), in contrast, follow post-glacial spillways cut mostly in weak Plains shale: the channel is largely single-thread with frequent bedrock exposures (indicative of the limited amount of alluvium).

The valley bottom character of the Alberta rivers is therefore strongly conditioned by the nature of post-glacial evolution. To many, this may still seem to be a largely academic matter, but in any investigation of riparian engineering structures, fisheries habitat etc., it is difficult to understand how such a perspective can be ignored.

8.3 <u>The sand-bed reaches of the central Prairies</u>

As indicated in Fig. 8.6, there is an abrupt change in character of bed material on the North Saskatchewan River from gravel-bed to sand-bed, the changeover occurring about 100 km above the Alberta/Saskatchewan border. A similar abrupt changeover in bottom conditions occurs on the other mainstems (somewhat prematurely on the Red Deer River) as depicted in Fig. 8.7.

The cause of the abrupt changeover to sandbed conditions along the mainstems does not appear to have been investigated, but there is clearly some relationship to the long profiles of these rivers (Fig. 8.7). On both the North Saskatchewan River (Fig. 8.6: at about km 1400, just upstream from Prince Albert) and on the South Saskatchewan River (at km 1100, upstream from Saskatoon) there is a marked levelling off in the stream gradient followed by an increase in slope downstream. The implication seems to be that some feature in the valley floor at these points is acting as a local base level to which the river upstream has gradually adjusted its gradient. In the case of the south branch, observations reported by Kugler and St-Onge (1973) indicate rock outcrops and till in the valley floor between Elbow and Saskatoon that would have achieved this effect. The resistance to downcutting afforded by this material (in contrast to valley bottom sediment upstream) means that degradation in the long reach upstream would have produced a flattening of the channel profile (Fig. 8.8). In turn this means that movement of bed material downstream would eventually reach a point (point P: Fig. 8.8) at which the gradient is too gentle to allow further movement.

Whether or not this is the correct explanation for the changeover to the sand-bed reaches of the central Prairies is uncertain, but it is consistent with the change in character of the valley bottom from extensive sand-braided channels upstream of the break of slope to irregularly sinuous single-thread rivers downstream of it in the vicinity of Prince Albert and Saskatoon. Indeed, by the time of the confluence of the north and south branches, the two river reaches show well-defined meanders entrenched into lacustrine silts (the bottom deposits of Glacial Lake Saskatchewan) overlying till (Fig. 8.9). The entrenched character of the Saskatchewan River at the confluence of the north and south branches is a marked contrast to the heavily braided sandy bottom that occurs in the vicinity of North Battleford and which prevailed on the South Saskatchewan River before its submergence by Gardiner Dam (Fig. 8.10).

No explanation of the origin of the multibar-sand braiding in these reaches of the north and south branches has been found in the literature, but in appearance, they show striking resemblance to the lower braided reach of the William River noted in Chapter 4. Borehole data at potential dam sites along the reach indicate that the alluvial sand deposits overlying the shale bedrock valley floor are more than 30 m thick in many places, indicative of Holocene aggradation. The source of the sand and its mode of supply to the river have yet to be determined.

Though the discussion above may seem largely like a collection of isolated geological observations and speculation, it should be apparent that this kind of knowledge is directly relevant in the management of these sand-bed prairie rivers. The existence of thick water-saturated sand bodies under the valley floor of the South Saskatchewan River, for example, could have been anticipated from its channel morphology. This braided morphology is also indicative of highly unstable channel cross-sections as bars move downstream or as one braid clogs and flows shift to another part of the channel: this is something that channel bank developments need to take into consideration. Smith and Wigham (1990), for example, chronicle the sedimentation problems at the water intake for the Chesterfield Flats Irrigation Project on the north side of the South Saskatchewan River 15 km east of the Alberta border (Fig. 8.11). Historically a point of bank erosion (being the outer bank of a gentle meander), the pump-side bank area became part of a sandbar complex as the thalweg shifted to the opposite bank. Again, this is something that might have been anticipated from the channel morphology.

8.4 The Saskatchewan River delta complex

It was previously noted that, downstream of Prince Albert and Saskatoon, non-alluvial materials become increasingly commonplace in the valley bottom of the main stems. In fact at several sites downstream of the confluence of the branches, bedrock exposures produce rapids, the most notable being at Squaw Rapids (Fig. 8.12), the site (since 1963) of the E.B. Campbell Dam and power house which now impounds Tobin Reservoir.

The valley bottom downstream of this site (as far as Cedar Lake) constitutes the delta complex of the Saskatchewan River, including (to the south) the Carrot River and (to the north) the network of channels leading into and out of Cumberland Lake. The delta drainage network converges to the single Saskatchewan River channel at The Pas (where the river cuts through The Pas end moraine) before splaying out again in distributaries entering Cedar Lake. The floodplain-deltaic deposits of the area are built on the floor of the former Glacial Lake Agassiz which extended into this region about 11,000 years ago.

Though the laminated silty clays of Glacial Lake Agassiz underlie virtually the entire region, they do not, however, form the surficial deposits. The lacustrine clays are overlain by sand and mud deposits, interbedded with peat layers (Fig. 8.13), representing post-glacial accumulation of alluvium. This ongoing buildup of alluvium over the Aggasiz lake floor corresponds exactly to Fig. 4.12 (case 1) in which aggradation is produced upriver of a prograding delta: as the Saskatchewan River has extended its course further into Cedar Lake, the associated decrease in gradient (and hence channel velocities) has forced deposition of its sand load along the bed of the river downstream of Squaw Rapids. This buildup in channel bottom deposits lowers the cross-sectional capacity for flows, increasing overbank flooding and thus producing sediment build up along the channel banks (levees) and away from the river (in flood basins) in tandem with the in-channel aggradation.

Radiocarbon dating of old floodplain surfaces buried by this sediment (and now represented by peat layers: Fig. 8.13) permits determination of the rate of aggradation. It must be stressed, however, that sediment buildup over the floodplain has varied considerably in recent time depending on location as noted below. Fig. 8.14 is a map of the anastomosed channel complex of the lower Saskatchewan River. The multiplicity of channels is somewhat misleading: most of the channels carry minimal flow, the floodwater of the Saskatchewan River tending to follow just one of these water courses. At the present time, the dominant channel is one of those to the north of Old Channel that feeds into Cumberland Lake. As recently as 1875, however, the main flow followed Old Channel, thus bypassing Cumberland Lake. The changeover apparently occurred during ice breakup in the late 1870's when ice-jamming of the Old Channel forced a diversion (an avulsion) to the north.

Such avulsions are common on anastomosed floodplains and, indeed, are the mechanism by which the multiplicity of channels is produced. Each of the Saskatchewan River distributaries was at some time in the past the main artery through the delta; but during that time, the large buildup of sediment along the floor of the main channel would gradually raise the channel above the level of the floodplain until it became so precariously perched that a breakout occurred. A cross-section over the delta floodplain (along the line shown in Fig. 8.16) is provided in Fig. 8.15 and shows how these various levee-bound channels are perched above the general floodplain surface, based on work by PFRA (Kuiper, 1960).

At earlier stages in the last thousand years, the Birch River and then the Sipanok Channel carried the floodwater of the Saskatchewan River. This raised floodplain levels in the southern part of the delta complex, eventually producing a breakout to the north. Over most of the last 500 years it seems that Old Channel was the dominant flow route until the 1880s avulsion, leading to New Channel (Fig. 8.16). The New Channel route itself has shifted in the last hundred years though: at first the main path of water and sediment flow was Steamboat Channel (Fig. 8.15; 8.16); subsequently (at some time during the 1920s) the flowpath shifted to that corresponding to Centre Angling Channel.

The floodplain cross-section (Fig. 8.15) emphasizes the dramatic increase in aggradation (in the northern area) since 1875 compared to the slower, centuries-long, buildup of sediment along Old Channel and, before that, Sipanok Channel. The explanation for this is quite simple. Prior to the 1875 avulsion, the full length of the floodplain down to The Pas was available for sedimentation, hence vertical accumulation was relatively slow. Since 1875, however, the delta front has, in effect, shifted from Cedar Lake to Cumberland Lake, and the area of rapid sediment buildup is therefore now confined to the upper part of the floodplain. The change is analogous to the effects of a rise in base level for the river (Fig. 4.12, case 2).

The 1880s avulsion from Old Channel to the north has been intensively studied by Smith et al. (1989) through a detailed program of valley bottom coring. Their report provides an excellent account of the mechanics of floodplain sedimentation following an avulsion. Part of the new (post-1920s) course of the Saskatchewan River is shown in Fig. 8.17. The extensive tract of light-coloured floodplain in the bottom of the photograph represents the Centre Angling Channel where it is being built up (by mobile and unvegetated sand deposits) in a wetland basin bounded by older channels with treed levees to the south and north. At point D, the water from this still-forming channel spills into pre-avulsion North Angling Channel. The marked channel widening produced by bank erosion along this downstream part of the old North Angling Channel (downstream of point D: Fig. 8.17) is evident even on such a small-scale photograph.

One of the points to emerge from this quick review of the Saskatchewan River delta floodplain is that, like all floodplains, it is highly active. The effects of channel switching (avulsions) are felt long after the actual event: heavy sedimentation may occur locally in part of the floodplain where the new channel is being built up; and severe bank scour may occur along old channels now forced to carry much larger quantities of flow. This needs to be borne in mind in any assessment of floodplain changes associated with engineering works in such a region.

For example, a sediment budget of the Cumberland Lake area by Northwest Hydraulics Consultants Ltd. (1986) noted that, prior to the impoundment of Tobin Reservoir, the Saskatchewan River discharged about 9 million tonnes (Mt) of sediment annually into the delta complex, of which only 4 Mt was exported downstream at The Pas towards Cumberland Lake. Since the damming of flow at Squaw Rapids, the input of sediment to the delta has essentially stopped, yet more than 2 Mt of sediment is still transported past The Pas (ignoring the Carrot River input) in the average year. The upper delta area has thus been transformed from a region of net deposition to one of net erosion.

The changes just noted seem to conform with what would be expected on the basis of engineering theory: the depleted sediment influx downstream of Tobin Reservoir would be expected to produce channel degradation and an increase in sediment entrainment in the Delta through this new process. Yet, in fact, this may be a highly oversimplified picture, because the morphological examination of the delta by Smith et al. (1989) indicated that considerable bank erosion was occurring along the main flow route through the delta even before the construction of Tobin Reservoir. In other words, instead of interpreting the pre-Dam sediment budget as deposition of 5 Mt of the initial 9 Mt annual load on the delta above The Pas with the remaining 4 Mt moving past The Pas, it is possible that almost all of the original 9 Mt of sediment were being deposited in the upper delta, and that the 4 Mt moving past The Pas were derived from bank scour within the Delta complex.

The important point is that delta floodplains are areas of huge exchange of sediment, an exchange that is not at all evident in the simple balance of "sediment in minus sediment out", but one that is clearly demonstrated by the morphological appearance of the delta surface. Such sediment exchange may be important in sediment quality studies. For example, if reservoirs had not been built at Squaw Rapids (and at Nipawin and on the South Saskatchewan River upstream), it is likely that the contamination of floodplain alluvium in the Delta by metals and pesticides absorbed to the wash load of the river would have emerged by now as a potential environmental issue. In that case, whether all the sediment load at Squaw Rapids is being deposited upstream of (and in) Cumberland Lake, or only half (5/9) of it, might have been relevant. Similarly in determining the effect of sediment trapping in Tobin Lake on the nutrient flux to the wetlands upstream of The Pas, any nutrient budget would need to know not just the difference in sediment inputs and outputs, but how much of that sediment outflow at The Pas was organic rich sediment from the upper basin in contrast to nutrient-poor sediment derived from bank scour within the delta.

Such issues may seem somewhat trivial in comparison with traditional sediment concerns dealing with the rate at which reservoirs might infill with sediment, or with the potential for channel scour at a bridge site, but the increased environmental awareness of the 1990's suggests that such issues will not remain ignored for long. In the Mackenzie River delta, for example, there has been considerable effort mustered by B.C. Hydro (e.g. Applied Ecology Consultants, 1987) into documenting the nutrient flux to the delta wetlands, and how this flux might decrease in the event of dam construction (and sediment trapping) on the Liard River far upstream in the basin. There are now also questions being asked regarding the contamination of Mackenzie River wash load with hydrocarbons (mostly natural seepage) and the effects of this pollutant in delta alluvium and in offshore sediments. In the case of the Mackenzie Delta, there is no clear picture yet of even the net input-output sediment balance, but the point is that, even when that is available, such sediment quality issues will also need data on how much of the sediment output to the Beaufort Sea is actually from bank scour along the main Delta channels as distinct from sediment throughflow from upbasin.

Again in the context of the Mackenzie Basin, if intermittent avulsion is the normal pattern for anastomosed floodplains (as borne out by the Saskatchewan River delta), how probable is such an avulsion in the Mackenzie Delta in the next century (and where is it likely to occur?)? The question is one which surely has relevance to development in the lower Mackenzie Delta, including pipeline emplacement, and yet it seems to be one that has yet to be asked, let alone answered.

In concluding this chapter, therefore, the point needs to be stressed again (as in the other chapters) that successful river and riverscape planning - be it the planning of a monitoring program or the planning of development - needs far more than just data collection. It requires an appreciation of what processes are taking place in order that the correct information is being gathered and interpreted. It is perhaps worth emphasizing that the large body of information collected (and the understanding of delta processes gained) on the lower Saskatchewan River by the Prairie Farm Rehabilitation Administration, was gathered for a very real practical problem, viz. to assess the impact of reclaiming large tracts of the delta floodplain surface for agriculture on the sediment balance and processes on those parts of the floodplain that are left (Kuiper, 1960):

> Reclamation of part of the delta surface will result in a smaller area being available for deposition of sediment and, hence, in a faster rise of the surface level. This in turn will increase the flood elevations and thus endanger the projected dikes.

The problem confronted by PFRA has largely disappeared as a result of sediment trapping in the subsequent reservoirs upstream, though new problems now exist (Carson, 1990c). Without question, however, any approach to future sedimentation issues in the Saskatchewan River delta complex will benefit considerably from the body of information collected, and perspective developed by this earlier PFRA program, and from the later geomorphological studies by Smith et al. (1989).

9. THE PEACE-ATHABASCA BASIN

In comparing the riverscapes, valley morphology and fluvial processes of the Athabasca and Peace River systems, there are many similarities, but also important differences between the two basins. One important difference, of course, is that since 1968 the discharge of the Peace River, downstream of W.A.C. Bennett Dam in the Rocky Mountains of British Columbia, has been regulated for power production (whereas the Athabasca flow is not), and some effects of this on the sediment regime will be examined shortly. Before that, however, it is useful to highlight the similarities and differences in the post-glacial evolution of the two drainage system.

9.1 Post-glacial drainage evolution

A major contrast in the mainstem valleys of these two Albertan rivers is that the Peace River is much more deeply incised into the surrounding plateau terrain than is the Athabasca River. This is evident in the long profiles of valley bottom and valley tops (Fig. 9.1). Between Hudson Hope (just downstream of the Bennett Dam) and Peace River settlement, the Peace River is incised more than 200 m into the interior plains, whereas, in contrast, the Athabasca River valley in its middle reach (at Athabasca settlement) is less than 50 m deep, as it is in fact throughout most of its length (except near Ft McMurray).

The significance of this contrast in depth of incision of the main stems should not be overlooked. It means also that the tributaries to the Peace River are much more deeply incised than those of the Athabasca. In turn, this has important implications for sediment production in the two basins. In Chapter 1 it was noted that sediment production in the lower Athabasca basin (downstream of Fort McMurray) was primarily from the lower courses of tributaries where they had been rejuvenated by the post-glacial downcutting of the mainstem (the base level of the tributaries). To a large extent (assuming similar underlying sediments and bedrock) the intensity of sediment production from undercutting and erosion of these valley side walls should be proportional to the height of the walls: it would be anticipated therefore that the much more deeply incised tributaries of the Peace River would deliver far more sediment to the main stem than in the case of the Athabasca. This conclusion is consistent with the sediment data collected on the two rivers. The Peace River at Peace River has a specific sediment yield of about 350 t/km^2/yr (ignoring the land area upstream of Bennett Dam: Hudson and Niekus, 1990); this is about seven times the specific yield of the Athabasca River at Fort McMurray.

The reason for this contrast in depth of incision between the two systems lies in the pattern of post-glacial drainage in relation to pre-glacial bedrock valleys (Fig. 9.2). The post-glacial Peace River followed essentially the pre-glacial path so that river incision has been largely into Quaternary drift laid down in that bedrock valley. This was not the case with the Athabasca River. The pre-glacial upper Athabasca River flowed eastwards linking either with the North Saskatchewan River or the Churchill River. To the north a small river drained along the present course of the Lower Athabasca, but this was not connected with the upper Athabasca course.

Emergence of the land area during deglaciation is indicated by the schematic positions of the retreating Laurentide ice front between 14,000 and 10,000 BP (Fig. 9.3). In both basins, landward drainage from the west was impounded by the ice front and (along with glacial meltwater) formed ice-marginal lakes. Figure 8.4 indicates typical conditions when the ice front was still west of Edmonton: water from the higher northwestern lakes (blocked by ice to the north) spilled to lower level lakes to the southeast, ultimately connecting to the Missouri drainage. As the ice front retreated to the northeast, these ice-marginal lakes shifted in the same direction.

During deglaciation of the area, any pre-glacial outflow route of the upper Athabasca River towards the bedrock Churchill valley would have been blocked by Laurentide Ice. It might have been expected, therefore, that the Athabasca drainage would spill out from the ice-blocked lakes towards the southeast, joining the North Saskatchewan River (in much the same way as the Red Deer spilled out to the southeast to join the South Saskatchewan River). This, in fact, did happen: Fig. 8.4 shows Athabasca drainage spilling out of Glacial Lake Windfall into ice marginal lakes occupying the mid-McLeod and mid-Pembina valleys, which in turned spilled into Glacial Lake Red Deer. Why this drainage to the southeast did not persist is unclear: as Fig. 9.3 indicates the withdrawal of the front of the Laurentide ice exposed much more land to the southeast, while drainage to the north still remained blocked. Presumably bedrock exposures in the successive spillways to the southeast impeded downcutting so much that Athabascan drainage remained impounded as a proglacial lake until the northwest route to the Mackenzie system opened up (Chap. 10).

The exact reasons for this failure of the upper Athabasca drainage to reintegrate with the plains drainage to the east are, however, probably unimportant here: what matters more is the fact that the post-glacial connector route (from Athabasca township to McMurray) between the upper and new lower Athabasca reaches cuts across a bedrock high. Post-glacial incision has thus been more difficult than along the Peace River (cutting into drift), and tributaries to the upper main stem (e.g. the Pembina and McLeod rivers) remain only weakly incised into the plains in contrast to their counterparts in the Peace basin (the Smoky, Beatton, Pine and other tributaries).

In both basins, valley depth becomes minimal in the lower reaches. Both the lower Peace and Athabasca are only weakly incised into the old lake floor of Glacial Lake McConnell, which was exposed as lake level gradually fell to produce the remnant lakes of today, Great Bear and Great Slave Lake to the north, and Lake Athabasca in the south (Chap. 10). The ability to cut down further into the old lake bed is, in the case of the Peace River, impeded by the base level control of the Slave River Rapids where the Slave is held up on resistant Shield rock.

Indeed, it would be expected that, following an initial post-glacial period of downcutting into the bed of Glacial Lake Agassiz, aggradation, rather than slow degradation, would have prevailed. In the case of the lower Athabasca, such aggradation would be triggered by the progradation of the river's delta into the now-stabilized Lake Athabasca. And, in both cases, ongoing Holocene aggradation of the lower reaches would have been assisted by isostatic backtilting: in the case of the lower Peace, for example, it seems that this land surface tilting has amounted to about 12 m per 100 km upvalley in the last 9,000 years (Fig. 10.6; Vanderburgh and Smith, 1988).

9.2 The Athabasca basin

Surficial deposits over much of the Peace-Athabasca basins therefore are fine-grained sediment laid down in these proglacial lake bodies; along the Athabasca valley itself, a staircase of sandy deltas occurs representing the intermittent progradation of the Athabasca River into ice-marginal water that decreased in elevation as it shifted northeast with the ice front (St-Onge, 1972, Fig 3). It is to be expected (based on previous chapters) that scour of channel banks and valley walls, where rivers have cut down into these sediments, would be important sources of sediment. Indeed, this seems to be the case in both the Peace and Athabasca basins.

In the Athabasca basin, for example, there is a distinctive pattern in the sediment regime at both Fort McMurray and, further downstream, at Embarras. The McMurray sediment rating diagram is shown in Fig. 9.4: it is clear that sediment concentrations are greater than expected on the basis of discharge at McMurray during the months of April and May. The "best-fit" (solid) line to the full data set is somewhat misleading because it appears to be too high at low discharges, being biased by the group of winter samplings. The sediment rating computed without the winter samples is shown as a dashed line; and residuals from this dashed line (i.e. actual concentration minus the concentration predicted by the sediment rating) for McMurray are shown in Fig. 9.5a. These residuals illustrate in a different way the tendency for sediment concentrations to be "above-average" (relative to discharge) during April and May (and below average during October). In fact, April sediment concentrations are five times greater than would be predicted by discharge. Why should this be so?

An obvious thought is that frost action on streambanks, along with ice scour during breakup, contributes sediment to the flow at rates which are higher than under normal summer time flow conditions. There may be some validity to this claim: the argument is frequently advanced in other parts of the country as well. Yet if that argument is valid for the Athabasca, it would presumably be valid for the Clearwater River (tributary to the Athabasca at McMurray) as well. But, as indicated by Fig. 9.5b, the same pattern does not hold on the Clearwater. The implication is that the excess sediment concentrations during April-May on the Athabasca are related to some other process. An alternative explanation emerges in terms of the geographic source of the sediment.

It was earlier hypothesized that a major source of suspended sediment in the Athabasca system would be areas where there is active bank erosion of fine-grained lacustrine sediments. Such sediments are found particularly in large areas of the McLeod and Pembina basins (Fig. 9.3; Fig. 9.6). It seems significant that it is precisely in the months of April and May that discharge from these two basins dominates the Athabasca flow: during these two months the combined flow of the McLeod (at Whitecourt) and the Pembina (at Jarvie) is 59% (April) and 40% (May) of the combined flows of the upper Athabasca (near Windfall) and the McLeod and Pembina; during the rest of the year it is less than 25 percent. Interestingly, a similar phenomenon occurs on the Liard River at Fort Simpson where sediment concentrations peak very early during breakup as a result of early snowmelt in the Muskwa - Fort Nelson River basins, where the rivers are similarly incised into glaciolacustrine (and soft shale) sediments.

The geographic distribution of glaciolacustrine sediments in the Athabasca basin is therefore something that has direct relevance to WRB operations. It is something that should also be borne in mind in WPM work as well. The reach of the Pembina between the Paddle River and Jarvie, for example, is an area of extensive summertime flooding (Fig. 9.6). In large part this flooding is a direct reflection of the limited amount of incision that has been possible into the old lake plain, reflecting the inability of the mainstem Athabasca River to cut down (and hence lower base level for the Pembina) in the rock-controlled post-glacial section between Athabasca and Fort McMurray. The problem is compounded, however, by the tortuous course of the Pembina itself. In the mid-50s it was proposed to alleviate this flooding through a process of selective cutoffs of some of the meander loops. The rationale was partly that the increased river gradient would cause an increase in velocity and thus carry a given discharge of water at lower stage; and partly that the increased velocities in the steepened reaches would lead to scour of the bed, with the degrading reaches propagating upstream like a nickpoint. It was thus hoped that the resultant channel degradation would increase channel cross-sectional capacity over a reasonable length of river between the cutoffs.

The theory is sound, but it obviously needs to be tied in with the actual conditions in the field. In the present case, since only a small number of cutoffs were being made, the ability to degrade the bed and to propagate bed erosion upstream to the next cutoff was crucial to the success of the project. In fact, minimal degradation occurred, severely limiting the success of the project.

Although it is easy to be wise after the event, it should be emphasized that there were certain aspects of the regional geomorphology that should have led to scepticism regarding the practicability of the approach at the outset. The first point is that, though the floodplain sediment itself is largely silty sandy alluvium (reflected in the meander scrolls of Fig. 9.6), this alluvium is underlain by more cohesive clayey sediment from the original lake bed. This material is relatively stiff and not easily eroded. In other words, degradation of the Pembina's bed in this type of material is not easily achieved unless flow velocities are relatively high. Most design charts show that critical velocities for the scour of such sediment are in excess of 1 metre per second. Herein lies the second problem: the area is part of an old lake bed on which gradients are extremely gentle and where, even in the 5-year flood, velocities would be barely above critical. It seems that the importance of this sub-channel clay (which was largely covered by thin deposits of channel silty-sand but which could have been anticipated from the geomorphological setting) was not fully appreciated.

9.3 The Peace River basin

One of the central issues in the Peace River basin must be the effects of the construction of Bennett Dam, coupled with flow-regulation, on the river network (including the Peace-Athabasca delta) downstream. In particular, it seems appropriate to ask whether any of the impacts of Bennett Dam could have been anticipated, prior to alteration of the flow regime, on the basis of the geomorphology of the region. While this is not the place to pursue a detailed enquiry (and such hydrological issues as the effect of lower summer flows on lake levels in the Peace-Athabasca Delta are beyond the scope of this report), certain observations are warranted.

One of the questions to have been raised during the planning of Bennett Dam would have been the impact on sediment levels in the downstream river reaches as a consequence of trapping of both suspended sediment and bed material in the dam. Specifically, some held the belief that trapping of <u>suspended sediment</u> in Williston Lake behind the dam would reduce sediment inputs (and hence nutrient loadings) to the Delta area, and that trapping of <u>bed material</u> would produce degradation downstream of the dam. Yet, in fact, aggradation rather than degradation has occurred in the reach immediately downstream of the dam; and while wash load inputs to the Delta may have been reduced as a consequence of less frequent overbank flooding (arising from the lower summer flows), there is no evidence that suspended load inputs to the lower Peace and Slave rivers from upstream have decreased appreciably. Both of these "impacts", while contrary to first impressions, are consistent with the regional geomorphology of the Peace River basin, and could have been anticipated.

9.3.1 Suspended sediment levels in the Peace River

The key question with regard to trapping of suspended sediment in Lake Williston is: "How much of the pre-Dam sediment supply to the lower Peace River originated from the headwaters upstream of the dam site, i.e. in the basins of the Finlay and Parsnip rivers?" The answer seems to be "very little". This conflicts with the general view of mountainous terrain as major suppliers of sediment to rivers (and with the preliminary WRB map of regional sediment concentrations: Fig. 5.1), but is consistent with various geomorphological observations put forward in previous chapters.

Sediment levels in rivers leaving the Cordillera (except those that are currently glaciated and involve significant glacial meltwater) tend to have low suspended sediment concentrations arising from various circumstances including the inherently limited availability of fine sediment in such areas, and the trapping of sediment in lakes and low-gradient floodplains. In contrast, where such rivers leave the mountains and cut down into fine-grained glaciolacustrine sediments, sediment supply from bank erosion and sidewall gullying have been observed to constitute a major source of sediment. In the Liard Basin, for example, sediment concentrations are low in the mountainous reach, increasing abruptly at the confluence of the Fort Nelson River. The same pattern of sediment supply would have been expected in the Peace basin, with far more sediment from the deeply-incised tributaries of the plains (entering the Peace downstream of Bennett Dam) and the Peace itself in the plains reach, than from the headwaters of the Peace above the dam site.

It is true that the late Wisconsin glaciolacustrine sediments account for only the uppermost deposits dissected by the Peace and its tributaries in the plains area. On the other hand, similar pre-Wisconsin glaciolacustrine sediments are also found in the valley fill (as would be expected since the Wisconsin pattern of deglaciation must have been repeated several times during the Pleistocene) so that, where the Peace and its tributaries have cut down along the course of their pre-glacial valleys, large thicknesses of fine-grained sediment have been (and continue to be) available for transport by the rivers resulting from slides and gullying. In addition, the underlying Cretaceous bedrock is primarily soft shale and fine sand, providing an additional source of sediment even where the overlying drift deposits are relatively thin.

The relative importance of soft shale "bedrock" and fine-grained Quaternary sediments as sediment suppliers to the Peace River and its tributaries has not been assessed. Irrespective of which is the dominant source, the important point here is that, within the plains reach, i.e. downstream of Bennett Dam, large quantities of available fine sediment result in sediment inputs by the plains tributaries being much more significant than sediment supply (even before Bennett Dam) from the upper Peace network. With this perspective, "sediment trapping" by Bennett Dam should have been expected to constitute only a minor effect on the downstream sediment budget.

However, even though, as the argument above claims, there has been no significant decrease in the sediment <u>load</u> delivered to the lower Peace and Slave Rivers, there has been appreciable change in sediment concentrations. This arises from the changed flow régime downstream of the dam (Fig. 9.7) in which, even downstream of the Smoky confluence, summer flows have been reduced dramatically, amounting to a 50% decrease in June. This means that, if there has been no decrease in sediment inputs to the Peace River (the tributary supply being unchanged), and if the tributary sediment is largely washload and unaffected by the reduced flow (as the largely silt and clay load is in the upper reaches), sediment concentrations should have increased following regulation. Indeed during the high flow month of June, they should have doubled at Peace River.

Fig. 9.8 shows the pre-Bennett sediment rating (dashed) at Peace River (based on 1966 and 1967 data) and the post-1968 sediment rating line, indicating an increase in concentrations at all flows greater than 1000 $m^3$0/s.The solid squares in the sediment rating represent the post-Bennett monthly flows and sediment concentrations for May-July that would be expected on the basis of the decrease in discharge and assuming that the sediment input to the river is the same as before construction of Bennett Dam. It is clear that the actual sediment levels experienced with regulated flow (as represented by the solid rating) are essentially identical to those expected (the solid squares) arising from the reduced "dilution" effects of clear water from the upper Peace now that summer flows from the headwaters have been reduced.

In parenthesis, the existence of just two years of data at Peace River (1966 and 1967) demonstrates the importance of having "background" monitoring programs in place before major riverscape developments occur. Data for the station at Peace Point are limited but tend to indicate the same conclusion: the impacts of decreased summer flows on increased sediment concentrations are felt far down the Peace River.

9.3.2 Bed material transport under regulated flows

Instead of channel degradation downstream of Bennett Dam, the opposite has taken place, at least in certain reaches of the river. This has been particularly well documented at the Pine confluence near Taylor Bridge (Fig. 4.10). As in the case of suspended sediment, tributaries to the Peace mainstem continue to deliver their gravel load to the channel, but in the case of this bed load, the reduced flood flows of the mainstem have meant that the river is no longer competent enough to sweep this load downstream. The result is aggradation at the confluences, the rise in base level at these points implying additional aggradation upstream. Whether this aggradation is sufficiently rapid to constitute a threat to floodplain development and structures in the years ahead does not appear to have been addressed.

The point that is emphasized here, however, is that this aggradation should not have come as a surprise. The channel morphology of the Peace River even prior to flow regulation would have suggested that even if aggradation was not already occurring in the Peace River channel, conditions for movement of the bed gravel were already marginal. The basis of this statement is the riverscape shown in Fig. 9.9, a view down the Peace River towards the confluence of the Moberly entering from the right (south) side.

One of the most striking features of the photograph is the terraces developed along the sides of the valley. These mid-level terraces are particularly well-developed at outer bend sites where the present channel has a much less sinuous flow line than the valley itself that was carved out in the earlier stages of downcutting prior to terrace formation. In other words, in the initial phases of incision, the Peace River channel had a much more strongly developed meandering course than it has at the present time. A second point related to the terraces is that, especially along the left side, they are blanketed by debris from slippage of the upper valley walls. In contrast, the lower parts of the undercut left bank seem to be more stable with respect to landslipping. This contrast between the morphology of the upper walls and the lower banks suggests that the valley walls are made up of different materials: soft sediments (more amenable to meander belt development and slope instability) in the upper part and more "resistant" material nearer river level.

This interpretation is consistent with the stratigraphy indicated in Fig. 9.10, at a section slightly downstream, near the Pine River confluence. The uppermost sediments are glaciolacustrine clays from the last deglaciation, overlying a thin layer of till (late Wisconsin) which, in turn, overlies thick glaciolacustrine silts laid down in a proglacial lake environment at the time of the westward advance of the late Wisconsin Laurentide ice front. These are the sediments in which the wide meandering style of the early valley was developed (much like the sinuous course of the Pembina River today: Fig. 9.6) and which are prone to retrogressive slumping (as occurred recently at Rycroft close to this site).

The bottom part of the valley walls is made up of gravels (more than 50 m thick) overlying weak shales. The exact origin of these older Pleistocene gravels is not important here, but their existence has controlled subsequent river development. As the river (and its tributaries) cut down into this basal gravel, the input of coarse bed material to the channel suddenly increased. Ongoing bank erosion today continues to add more gravel to the channel. The response of the river was to change from a meandering single-thread channel characteristic of fine-grained sediments, to a laterally-constrained wandering river in which the huge quantities of gravel were moulded into the pool-riffle planform shown in Fig. 6.12 (bottom sketch). The photograph of Fig. 9.9 shows the foreground part of the river being shunted from a right-bank pool over the diagonal bar front into a left-bank pool along which the inner bank of the first bend is experiencing massive bank erosion. Indeed, one of the most striking features of the photograph is the discordance between the present-day plan geometry of the channel and the inherited plan morphology of the valley: the river thalweg hugs the inner bank of the first bend in a manner that is not typical of normal meander geometry.

With progressive lowering of the Peace into these basal gravels, the greater would have been the load of gravelly bed material supplied to the main stem by its own bank undercutting and by inputs from the incising tributaries. This would certainly have slowed down the rate of incision, and perhaps initiated aggradation, equivalent to the "complex response" described by Schumm (1973) in the case of laboratory drainage networks. Such aggradation would have been particularly pronounced where tributaries (such as the Pine) delivered large quantities of coarse gravel to the mainstem.

Some indication of the importance of this tributary input of gravel, long before any change in the flow regime imposed by Bennett Dam, is provided by Fig. 9.11 denoting the downstream change in bed material along the main stem of the Peace River between Hudson Hope (just downstream of Bennett Dam) and the end of the gravel-bed reach upstream of Fort Vermilion. There is a marked increase in gravel size where the Pine and then the Kiskatinaw rivers enter the main stem; an abrupt decrease upstream of Dunvegan Bridge; and a strong decrease again where the Smoky River delivers its large supply of finer gravel to the reach. This pattern of increase and decrease in gravel bed material size is further evidence of the dominance of tributary inputs long before present day changes in flow regime. In other words, localized present-day aggradation is quite consistent with what the bed material character of the river indicates has been going on for several hundreds, if not thousands, of years.

9.3.3 Sedimentation of fines in Peace River valley bottom

The decrease in summer discharges along the Peace River since the operation of Bennett Dam means that higher parts of the channel bed are now abandoned, colonized with primary vegetation (poplar on the gravelly bar tops and willow in the finer-grained bar edge sediments), and converted to a new, lower-level floodplain (Kellerhals and Church, 1989). In addition, about half the length of side channels in the B.C. reach has disappeared, and is being incorporated into the new floodplain. Thus the width of the Peace has shrunk since flow regulation was implemented, and an increased part of the valley bottom is mantled by rough vegetation that is capable of trapping suspended sediment.

It is to be expected that - in contrast to the insignificant sediment trapping behind Bennett Dam itself - sedimentation associated with the new flow regime is decreasing the supply of suspended sediment to the lower Peace and Slave. Comparison of sediment data from the WRB stations at Peace Point and Peace River should indicate the magnitude of this sediment trapping. It might also be expected that, once these new floodplain areas build up to a level that is consistent with the new flow regime, this "overbank" sedimentation will decrease. This may not, however, be the case: where in-channel aggradation of gravel takes place, the gradual reduction in channel flow capacity may be expected to increase overbank flooding in the decades ahead and thus increase overbank sedimentation.

Apart from the reduction of suspended sediment supply to the lower Peace and Slave rivers, the valley bottom sedimentation in the reaches downstream of Bennett Dam also directly affects riverine habitat there: shallow flow areas adjacent to bar margins and some bank edges are being mantled with fine-grained sediment and primary vegetation; and there has been extensive loss of side channel habitat. Both changes may be expected to have some impact on fish.

Similar, but less obvious changes may be taking place in the lower reaches of the Peace River where the bottom sediment is sand rather than gravel. Tributary streams such as the Wabasca continue to supply sand to the main stem just as they did before flow regulation, and the reduced transporting capacity of summer flows along the main stem may be initiating sedimentation of sand downstream of these confluences.

Perhaps more importantly, the increased suspended load concentrations may be increasing sedimentation in side channels and sloughs in these lower reaches. These will be the preferred areas for fine-grained sedimentation (silt flocs and clays) and might well be the first places to reveal bottom-sedimentation contamination in the years ahead as pollutant loadings from farmland in the Smoky River basin, and pulp and paper (and other industrial) effluents in the middle basin, increase. On the other hand, the enormous quantities of "background" sediment from undercutting of channel banks and valley walls would seem to suggest that here, as in the Saskatchewan River basin (Carson, 1990d), levels of contaminants in stream sediments will be appreciably diluted, except, perhaps, in localized areas immediately downstream of point sources.

10. THE MACKENZIE BASIN

The Mackenzie Basin, strictly speaking, includes the basins of the Athabasca and Peace rivers. In the present chapter, attention is focused on the drainage network that drains to the Mackenzie River downstream of Great Slave Lake. The chapter begins with a brief outline of the bedrock geology, but the bulk of the material deals with the pattern of deglaciation and the implications of this for the Mackenzie River.

10.1 Pattern of bedrock geology

The Mackenzie basin drains, primarily, the Interior Plains, and is underlain by essentially flat-lying sedimentary rocks (shales, limestones, etc.) not radically different from those of the prairies. There are two exceptions to this statement.

Along that part of its course between Camsell Bend and St. Sault Rapids, the Mackenzie River strays into a lowland belt within the Cordillera (Fig. 10.1), where the width of the Cordillera increases appreciably compared to the narrow Rocky Mountains to the south.

Secondly, the northeastern rim of the basin (to the northeast of a line between the McTavish Arm of Great Bear Lake and the North Arm of Great Slave Lake) belongs to the Canadian Shield province (Fig. 10.2). The proportion of runoff from the Shield is, however, minor. The contact zone between the Shield and plains rocks in the vicinity of Great Slave Lake appears to be a zone of weakness. This is reflected in the etching out, along the contact, of the North Arm and a south arm to the lake, the latter being buried in Holocene times by deltaic infill from the Slave River (Vanderburgh and Smith, 1988).

The errant course of the Mackenzie into the trough between the Franklin Mountains and the Mackenzie Mountains, away from the plains, seems to be largely a post-glacial feature (Fig. 10.3). The location of the preglacial route of the Mackenzie is not known with certainty, but it appears to be east of the Franklin Mountains. There is a deep bedrock trench (largely buried by drift) along the east flanks of those mountains (traced upriver as far as Great Bear River) connecting with a similar bedrock trench now occupied by the small Hare Indian River. Drift deposits in the Hare Indian trench (Fig. 10.3) are older than the last glacial phase, in contrast to the drift along the present course of the Mackenzie which dates back only to the last deglaciation. It is unclear how many other pre-Holocene Mackenzie courses exist beneath the surface drift, and the preglacial Mackenzie route between Great Slave Lake and Fort Good Hope remains unknown. It is not difficult to imagine several old south-north spillways or meltwater channels formed at times of former ice front positions in this area, but now buried by drift.

10.2 Deglaciation of Mackenzie Basin

As noted in previous chapters, the Mackenzie Lowlands were occupied by the western margin of the Laurentide ice cap; the high parts of the Mackenzie Mountains on the western divide of the basin included eastern lobes of ice from the Cordilleran ice cap (Fig. 10.2); and, between these two ice fronts, a narrow band of ice-free terrain existed even during the maximum extent of the Laurentide ice mass (see also Fig. 2.6).

The pattern of retreat of the Laurentide ice front is indicated in Fig. 10.2. The ice front had been at its terminal position for about 6,000 years, before embarking on its relatively rapid retreat to the east. During the long period at its maximum position, the Laurentide ice mass built up massive end moraines along its length (especially in the flat terrain southwest of the Delta).

Similarly, meltwater streams from both Cordilleran and Laurentide ice deposited substantial thicknesses of sands and gravels (as deltas) and silts and clays (bottom deposits) in lakes adjacent to the Laurentide ice front, especially in the west bank tributaries such as the North Nahanni, Redstone, Keele, Mountain, Arctic Red and Peel rivers. Several maps and publications dealing with the surficial deposits of the Mackenzie corridor were prepared by the Geological Survey of Canada in the 1970's (listed by Maurice, 1988), but there does not seem to be any summary publication dealing with the deposits of the west bank rivers. Any attempt to understand the sediment yields of these rivers must certainly refer to the existing GSC information base.

The west bank tributaries were not the only areas where glaciolacustrine sedimentation occurred. Large parts of the present-day Mackenzie valley were also under proglacial lake water. The reason is that, between about 12,000 and 11,500, the ice front was essentially aligned with the present river course, so that isolated lobes would have impounded any meltwater from upvalley. The most spectacular occurrence (Fig. 10.3) of this was in the area between Fort Good Hope and Norman Wells (Mackay and Mathews, 1973).

The tentative ice front position of Laurentide Ice at about this time is indicated in Fig. 10.3. Ice occupied the lowlands upvalley of Norman Wells while a lobe blocked the valley at Fort Good Hope. Meltwater from the ice margins (including flow from upvalley funnelled down the Mountain River channel) was blocked in the ice-free reach, accumulating as a lake that eventually built up sufficiently high to spill out down the Ontaratue channel from the northwest arm of the lake. The lake area exists today as a large expanse of fine-grained lacustrine sediments, with a major sandy-gravelly delta where the Mountain River entered the lake.

As the ice lobe gradually retreated to the east, the lower land area around Fort Good Hope was exposed, and lake water started to drain through the present (Fort Good Hope) outlet rather than down the Ontaratue channel. The area remained a lake for several thousand years, however, being blocked by the bedrock ridge at Fort Good Hope (this not being the pre-glacial course of the river), even after withdrawal of the ice. As the outflow cut more deeply through this ridge to form the present-day Ramparts Canyon, so the lake shrank to become just the Mackenzie River. As the river continued to cut down in the Ramparts, it did so upstream as well, but, there, cutting down through soft lacustrine muds. In this way the section through these lake deposits at km 1041 (Fig. 10.3) was formed. It should be remembered that much of this marked incision into both rock and sedimentary infill was achieved by an early post-glacial Mackenzie swollen with meltwater from the retreating ice front, and with discharges much greater than those of the present.

It seems likely that substantial quantities of the Mackenzie's sediment load originates from undercutting and slumping of these lacustrine sediments along its margins in this area. The sediment supply is not restricted to just the area upstream of Fort Good Hope. D.G. Smith (pers. comm., 1991) has mapped similar lacustrine deposits upstream as far as Fort Simpson, and sees these as part of one long (but narrow) lake (Glacial Lake Mackenzie: Fig. 10.4) for which the lake identified by Mackay and Mathews was simply an early phase. The huge old delta of the Liard River at Fort Simpson is seen as having formed into this water body (Fig. 10.4). The difference in elevation between the tops of the Mountain River delta (92 m) and the Liard delta (189 m) is attributed to the much greater isostatic rebound in the southern part of the basin.

As the ice front retreated upstream from Fort Simpson, beginning about 11,000 BP, the depressed lowlands around the present Great Slave Lake were also submerged (the outflow down the present Mackenzie being blocked by both Glacial Lake Mackenzie and the Liard Delta). By about 10,000 BP this large glacial lake (McConnell) included embayments up the Slave, Peace and Lake Athabasca valleys as well as Great Bear Lake (Fig. 10.5): water spilled out down both the Hare Indian and Great Bear River outlets at early stages in the lake's history before the Mackenzie valley outlet became dominant.

At about 10,000 BP, Lake McConnell is believed to have experienced a catastrophic influx of floodwater from Glacial Lake Agassiz, which had been temporarily blocked to the southeast (Smith, 1989). The outflow discharge must have accelerated downcutting at the Fort Simpson outlet of Glacial Lake McConnell and the Ramparts outlet of Glacial Lake Mackenzie. As rebound continued, the extent of Glacial Lake McConnell diminished, the emergence of higher parts of the lake floor resulting in its eventual separation into Great Bear Lake, Great Slave Lake and Lake Athabasca, the latter separation occurring about 8,500 BP, by which time little, if any, glacial meltwater now reached the basin from the Laurentide ice mass (which had shrunk to a dome over Keewatin). The Peace-Athabasca delta surface formed in Glacial Lake McConnell is almost 50 m higher than the Liard Delta as a result of the increased amount of isostatic rebound towards the east. The actual tilt in rebound is about 12 m per 100 km to the northeast (Fig. 10.6).

10.3 The Mackenzie Valley

The changing character of the Mackenzie River (the long profile as far as Fort Good Hope being given in Fig. 10.7) and its valley are to a large extent the result of these events in the last deglaciation of the region. A history of individual reaches (and their sedimentological and environmental significance) is provided below.

Great Slave Lake - Fort Simpson

The location of the pre-glacial outlet from Great Slave Lake is unknown. The present outlet and course of the Mackenzie River reflect downcutting by meltwater from the receding Glacial Lake McConnell as it spilled out across the Liard Delta (up to about 8,500 BP) and normal terrestrial drainage since then. This water has cut down through lacustrine silts and clays and then into the underlying till. In some places along the new course, bedrock was encountered at relatively shallow depth, and thus further erosion by the river was slowed down there (and upstream, for which the rock outcrop acted as base level). Downstream of the rock exposure, downcutting would have been able to proceed more rapidly. In this way the pattern of alternating rapids and deeply-entrenched reaches was formed.

Present day entrainment of suspended sediment in this part of the Mackenzie's course would be expected to be minimal for two main reasons. First, the river has cut through into bouldery till along most parts of its course, so that bank erosion of the overlying lacustrine deposits is impeded. Secondly, the regulating effect of Great Slave Lake reduces the magnitude of flood flows in the spring. How much suspended sediment is actually entrained along this reach is not known. But, based on sampling undertaken in the early 1970's by WRB, the Mackenzie at the Liard confluence carries about 2.5 Mt in the open water season, substantially smaller than the Liard's mean annual load of about 42 Mt. How much of this sediment is outflow from Great Slave Lake is unknown. Bed load transport is also assumed to be minimal along the reach because the bed is armoured with large cobbles from the till in many places.

<u>Fort Simpson - Camsell Bend - McGern Island</u>

The armouring noted in the reach upstream continues below Fort Simpson, but only on the right side of the channel. The left side is now sandbedded reflecting the large quantities of sand brought in by the Liard River.

Downstream of Camsell Bend (as far as McGern Island) the character of the channel changes, with sand and gravel bed material from bank-to-bank, and multiple islands in the reach. It seems likely that the change is due to large quantities of coarse bed material being brought in by the North Nahanni and Root Rivers. These two rivers, especially the Nahanni, are cutting down in their lower reaches through sands and gravels laid down as deltaic sediment in Glacial Lake Mackenzie (Fig. 10.4).

This part of the river is one in which navigation is sometimes difficult and where dredging has been recommended. On the assumption that most of the bed material is coming from the North Nahanni, this reach of the Mackenzie may well be aggrading, and dredging is not likely to produce those adverse environmental effects downstream that would result from a sudden reduction in bed material supply. Dredging in this area might, in fact, be beneficial in other ways, in particular as a supply of coarse sand and gravel needed for permafrost blankets for roads and pipelines. The possibility of using Mackenzie River sediments for this purpose has already been put forward (EBA, 1987); areas of aggradation constitute a relatively safe place to implement this mining.

<u>The Wrigley reach</u>

The reach downstream of McGern Island as far as Blackwater River (upstream of Redstone River) is quite different in character, though little published information is available on the bed material. The valley is much narrower and this, along with the absence of islands, would seem to imply that bed load transport rates are smaller than upstream. This would be consistent with the view that aggradation is occurring at the North Nahanni confluence. No significant sediment-transporting tributary joins the Mackenzie in this reach. Fig. 10.4 indicates that the Wrigley River itself dissects a Glacial Lake Mackenzie Delta in its lower reach, but it is assumed that the total bed load from this source is minimal.

On the other hand, suspended sediment may be entering the Mackenzie in large quantities, because the valley flanks are tall, and mostly silt and clay down to water level. No data are available on suspended sediment in this reach

Redstone River - Seagull Island - Sans Sault Rapids

An abrupt change occurs in the character of the river just upstream of the Redstone River: the bed becomes gravelly and cobbled throughout; the gradient increases (Fig. 10.7); and the valley bottom widens appreciably, with the river swinging around large islands (Fig. 10.8). This is the "fan" of the Redstone and Keele rivers, imparting to the Mackenzie a "wandering" appearance, much like the braided character of the North Saskatchewan River downstream of the Ramparts Creek fan (Figs. 4.4, 4.5). The reach is similar to that downstream of the North Nahanni, but longer and more spectacular, indicative of the large quantities of coarse sediment brought in by the two rivers. The two tributaries themselves are also braided in their lower reaches in response to the large amounts of gravel that they are forced to carry as they have cut down into proglacial lacustrine deltaic deposits.

Again, the reach is one in which dredging has been proposed to improve navigation, and where mining of bed sediment has been suggested to provide granular material for construction on land. Any mining of bed material would raise questions regarding possible degradation downstream in the Norman Wells area (where pipelines exist beneath the bed), but it seems likely that the effects would be small. Inspection of the long profile indicates that most of the gravel bed material doesn't reach the Norman Wells area (Fig. 10.7). The reach between Seagull Island and the Mountain River is sand-bedded in the main; and though much of this sand is probably coming from the Redstone and Keele rivers, most of it would be moved through the steep gravel-bed reach as suspended load and is unlikely to be affected much by dredging.

Sans Sault Rapids to the Ramparts

This reach is in some ways similar to the gravel-bed reach between the Redstone and Seagull Island, though the limited bed material data available suggest that sand is more abundant than gravel. The channel morphology is consistent with a mobile bed and navigation difficulties are reported at several spots in the reach, particularly at Hume Crossing shallows. All these features appear to be related to the large sand and gravel load brought in by the Mountain River as it cuts down into its old delta.

The lower Mackenzie River

Downstream of Fort Good Hope, the Mackenzie flows within its pre-glacial trench, in a reach beyond the northern limit of Glacial Lake Mackenzie. The contrast between conditions upstream and downstream of the Ramparts is emphasized by Mackay and Mathews 1973):

> Upstream .. tributaries are graded near their mouths, whereas downstream, the tributaries are aggraded and may have drowned or oversized mouths. Upstream, some of the islands in Mackenzie River extend high above the limits of modern flooding; downstream the islands are lower, subject to flooding.

Thus, whereas the Holocene history of the upstream reach has been one of downcutting of a post-glacial channel, the history downstream has been one of aggradation (of sediment supplied from upstream) in a pre-glacial trench. Unfortunately little information is available on the gentle gradients of the reaches downstream of Fort Good Hope.

The only major tributary in the reach is the Arctic Red River, quite different from other west bank tributaries upstream in its lack of a coarse bed load. On the other hand its suspended load is huge in relation to its annual runoff, equivalent to about 1350 mg/L, which is more than twice the load/flow ratio of the Liard River. The source of this fine-grained sediment is unknown, though it was earlier suggested that, as in the case of the Peel, it reflects bank erosion of lacustrine sediments upstream.

10.4 Sources of suspended sediment

Tentative data presented in Chapter 5, together with the qualitative observations in this chapter, seem to indicate that the west bank tributaries of the Mackenzie are major sources of suspended sediment (much of which probably is washed through to the Delta) as well as bed load (most of which accumulates locally downstream of confluences). With this perspective, it is unfortunate that there are no sediment stations between those on the Liard near its mouth and the Mackenzie River upstream of Arctic Red River.

The traditional view has always been that the Liard is the major source of sediment in the Mackenzie. Yet in fact the Liard accounts for about 40% only of the Mackenzie's overall suspended sediment load upstream of Arctic Red River; and in terms of the clay fraction (the host of most nutrients as well as contaminants) the proportion is only about one-third. Addition of the clay loads of the Arctic Red and Peel Rivers lowers this amount further.

Any attempt to partition the sediment influx to the Mackenzie downstream of Fort Simpson more precisely is no small task as the above account should indicate. Theoretically, the inputs of groups of tributaries (such as Keele and Redstone) could be determined by load increase between successive mainstem stations (e.g. Wrigley and Norman Wells), but additions to the suspended load along the mainstem from instability of mud slopes must cause some uncertainty in that approach. Sampling stations on the tributaries themselves would seem to be the logical solution, but because of backwater effects from the main stem, these would have to be well upstream. Because sediment sources in the tributaries seem to be in the downstream reaches, such stations would likely convey misleading data. A more meaningful approach would be to have sediment sampling done near the mouth, with hydrometric monitoring upstream (as was the case in the Arctic Red River sampling in the 1970's), provided that no significant side stream flows enter the tributaries in the intervening reach.

This brief overview of sediment sources provides no obvious solution, but it should be sufficient to warn managers that any attempt to partition the Mackenzie Delta sediment load into different tributary parts of the basin requires a great deal of careful prior investigation in order to produce meaningful data.

10.5 The Delta sediment balance

One of the items of ecological interest in the Delta area is how much suspended sediment from the Mackenzie Basin is deposited in the Delta (hence providing inputs of nutrients, but also possibly hydrocarbon contaminants) and how much is delivered to the Beaufort Sea. Though much speculation has been offered regarding this balance, there are no conclusive data. This reflects the fact that while inputs to the Delta have been monitored on the Mackenzie, Arctic Red River and the Peel River, there has been little outflow sampling of sediment and monitoring of discharge, at the delta front. Proposals are currently being considered to address the problem.

The earlier discussion of the Saskatchewan River delta (Chap. 8) provides a geomorphic perspective, however, that indicates uncertainty will still prevail even when adequate sediment outflow data have been gathered. The problem is that differences between input and output loads merely indicate the net deposition within a delta environment, which is a minimum estimate of the actual amount. As an example, an annual input of 100 Mt of sediment and an annual outflow of 50 Mt cannot be taken to indicate that half of the load from upbasin is deposited in the Delta and half reaches the Beaufort Sea. The reason is that considerable bank scour occurs within the Delta, primarily along the Middle Channel, so that only part of the load delivered to the Beaufort Sea originates from upbasin. In turn, this would mean that the sediment load from upbasin that is deposited in the Delta must exceed the 50 Mt indicated by the simple inflow-outflow balance.

Thus the nutrient flux to the Delta must be in excess of that indicated by an inflow-outflow balance. In addition, the nutrient and contaminant flux to the Beaufort Sea must also be different from that computed on a simple balance. It is one thing to know that 50 Mt of sediment (say) are delivered offshore; the chemical quality of this sediment will differ, however, depending upon whether it is largely from upbasin, or largely from internal erosion within the Delta.

In short, prior to any numerical modelling of inflow-outflow balances, some attention must be directed to geomorphic processes operating to produce sediment within the Delta itself, and attempting to assess these through, for example, aerial photograph comparison (as done by Lapointe, 1984).

11. CONCLUDING REMARKS

Two points appear to be particularly worth emphasizing in conclusion:

- the usefulness of geomorphology to IWD staff are not restricted to any one branch; on the contrary, the perspective offered by geomorphology affords some insight into the interrelationships between the work of the different branches;

- the benefits of a geomorphic perspective is not restricted to any one level of staff within the Inland Waters Directorate; it is equally important to managers, supervisors and technical staff.

Two examples are offered in this last chapter in an attempt to reinforce these claims.

11.1 Short-term channel bed stability

At different places in the preceding chapters, attention has been directed to the importance of a proper understanding of channel bed behaviour in alluvial rivers. Changes in channel bed bathymetry, for example, were shown to be useful items of data in the determination of bed material transport rates through a reach, especially where collected over multiple cross-sections. Yet even data regarding channel bed changes at a single cross-section can provide useful qualitative information regarding the mobility of bed material: non-changing cross-sectional geometry implies a stability that is probably indicative of limited bed material transport.

Changes in bed geometry at a cross-section also provide some estimation of the depth of scour and fill possible in floods, especially when surveys are available before and immediately after floods. Such information is surely relevant to WPM staff concerned with matters such as the stability of bridge piers, the possibility of exposure of buried pipelines, or the likely burial of intake/outfall bank structures, etc.

Bed geometry - and its changes - have been shown to play a major role in determining the cross-sectional distribution of suspended sediment in a river, something of concern to WRB staff in assessing the suitability of a measurement section for sediment sampling, and in planning the frequency of multiple vertical sampling work for the computation of single-vertical k-factor values.

In other words, repeated cross-sectional surveys at a measurement section on a river provide valuable information to both WPM and WRB alike. And such information is already routinely gathered by WRB staff at measurement sections when conducting hydrometric surveys to determine the discharge through the section at times when checks are being made on the stage-discharge rating curve.

What is somewhat surprising to a geomorphologist is that such data are so rarely used for other purposes. One of the reasons for this is that the basic survey data collected are seen as simply a means to a single end (computation of cross-sectional area of flow), and are not stored in a form where they could be used for these other purposes.

WRB, for example, publishes data on bed material size at its measurement sections. Sometimes, however, these published values are extremely hard to interpret by potential users. In the past, one reason for this was the fact that the published grain size distribution curve was based on the simple means for multiple verticals across the section, so that (as an extreme case) a section with a non-alluvial clay thalweg and a sandy lateral bar might be reported as essentially a silty channel bed! But even where separate grain size curves for the different multiple verticals in a section are reported, interpretation can sometimes be very difficult because of marked changes over time.

As an illustration, consider the bed material data for the Mackenzie River upstream of Arctic Red collected at the 1970's measurement section. The published data showed marked differences between the samplings especially between 1972 and 1974 (Fig. 11.1: top). One of the left-centre verticals in 1974 had much coarser sand than in 1972; while right-side verticals showed a marked increase in the amount of silt-clay sediment. These differences provide some difficulty in any attempt to separate the Mackenzie's suspended load in this reach into "bed material load" and "wash load". What exactly is the lower size limit of bed material? The more general questions that need addressing, however, are why should these changes occur, and how many samplings are needed to determine some long-term averaged bed material composition?

Some answer to the first of these questions is provided by looking at the unpublished hydrometric survey data collected at the time of the bed material samplings. The sounding data for 1972 and 1974 are plotted in the bottom of Fig. 11.1. They indicate the large changes in bed geometry between the two samplings. The left bank thalweg was deepened, presumably by preferential removal of finer sandy sediment which would explain the coarsening of the bed in that part of the section. The right bank shallows (actually a minor slough just downstream of the right bank bar: see Fig. 6.15, bottom of map) has been built up with sediment, finer sediment able to settle out in this "backwater" setting. Much of this scour and fill - and hence alteration of the bed material composition - presumably occurred in the record floods of August, 1974, just before the second survey. The point being made here, however, is that it is only through the availability of successive cross-sectional data that any realistic interpretation of the bed sediment changes can be made.

The bathymetric information on measurement section geometry collected by regular hydrometric surveys of WRB is clearly invaluable, and yet its importance beyond the realm of calculating discharge does not seem to have been widely appreciated. For anyone with a geomorphic perspective, it is one of the first requirements in any attempt to understand the behaviour of a reach. The failure to appreciate its importance is reflected in several facets of WRB's sediment sampling program.

In many cases, for example, the routine hydrometric surveys are not made with respect to a fixed zero-point for the survey line, but simply some arbitrary point which will change from one survey visit to another. This is of no concern to the hydrometric program: the calculation of flow cross-sectional area does not require starting from the same origin point each time a survey is done. But, if these survey data are to be compared between different survey times to produce patterns of bed geometry change (as in Fig. 11.1), this can only be done if the origin reference point is fixed (or at least known relative to previous ones).

A second manifestation of this "neglect" of bed geometry from the perspective of sediment operations arises at those sites where the sampling station is located several kilometres downstream of the gauging site. In such situations, typically 99% of the suspended sampling is done at a single vertical (SV) at the downstream site, To understand what causes changes in the k-value of that SV site over time requires an appreciation of how the setting of the SV location changes relative to the cross-section in which it is located; e.g. on some occasions when it is being compared with multiple vertical sampling at the upstream measurement section it may be located over a bar, at other times there may have been a shift in channel geometry and the fixed SV site corresponds to a thalweg location. In other words, to correctly interpret the k-values derived from downstream SV samples and upstream MV samples requires cross-sectional surveys to be established, at the time of comparative sampling, at the downstream SV cross-section. Yet, in fact, this is never done: available cross-sections are always for the upstream measurement section, where discharge is being monitored.

A geomorphic perspective within IWD, and particularly within WRB, would certainly have led to a greater recognition of the importance of the hydrometric work that is already being done, beyond the realm of simply flow data.

11.2 <u>Sediment sampling for sediment quality analyses</u>

The recognition that many river-borne contaminants are preferentially found within the wash load sediment phase, rather than simply dissolved in the water, has led to increased attention being paid to sediment quality as part of the general mandate of the WQB of Inland Waters. Inevitably this means closer liaison between the two branches, and once again, a geomorphic perspective might prove to be beneficial in many cases.

In particular, whereas the concentration of contaminants dissolved in water may be relatively uniform across the river at a measurement section, it must be recognized that concentrations of contaminants within the suspended sediment are likely to be highly variable in some cases. Specifically, since these contaminants are preferentially incorporated in the clay and fine silt size fractions, cross-channel variability in grain size of suspended sediment can produce marked changes in the concentration of contaminants expressed as ng (contaminant) per gm of suspended sediment. Thus the issue of the representativeness of the single-vertical sampling site (discussed previously in the context of WRB sediment sampling) looms as a major concern in WQB sampling for suspended sediment.

As an example of the importance of the same "geomorphic" principles applied to WRB sediment work in the context of this new WQB sediment-quality sampling, reference is made to one of the pioneer studies of hydrocarbon contamination in a Canadian drainage network. The program was operated by the National Hydrology Research Institute on the Mackenzie River during June, 1986 (Fig. 11.2). Sediment samples were taken from a research boat, beginning upstream at the Liard confluence on June 19, and ending at Arctic Red River on June 29. The study was an "attempt to use ... downstream trends to investigate sources, transport and fate of hydrocarbons in the Mackenzie River" (Carey, et al., in press.). Comparison of one-point-in-time hydrocarbon loadings at the seven stations sampled would allow some estimate of the magnitude of contaminant influxes (or losses) between successive stations.

Though this is a sound approach to the problem in principle, the data collected might also be used by others to "characterize" the different reaches in terms of different levels of hydrocarbon pollution, and this could produce severe errors, because it implicitly assumes that the samples taken at each site were indeed representative of the full cross-section at that site. Since sampling was done at a single vertical (usually, but not always, over the thalweg), there is no assurance that this assumption is valid. On the contrary, reference to available WRB data at some of these sites, as well as a geomorphic interpretation of the reaches in question, suggest that the assumption is not valid, and the results are subject to misinterpretation.

Fig. 11.3 shows the sediment sampling site on the Liard at its mouth. One sample was taken at this site (over a period of several hours using a pump sampler at about 30 cm below the water surface) on June 19. The concentration of selected "priority" polycyclic aromatic hydrocarbons (PAHs) at this site was reported as 264 ng per gm of sediment. In comparison, the concentration at Halfway Island (downstream of Great Bear River) was only 52 ng/gm, while just upstream of Arctic Red River it was 391 ng/gm, somewhat higher than in the Liard.

The decrease from the Liard site to the Halfway Island site seems realistic: huge quantities of sand are delivered into the Mackenzie upstream of Halfway Island by the Keele and Redstone Rivers, and this large influx of sand (presumably with little hydrocarbon material) would have diluted the pollutant levels in the sediment from upstream. The increase in levels by the time water reached Arctic Red River also seems realistic, in view of natural seepages of hydrocarbons in the Norman Wells area, though it is not appreciably higher than in the Liard. Clearly the kind of data presented here lends itself to definite statements regarding the "severity" of hydrocarbon pollution in particular reaches of the Mackenzie River. But how accurate are these data?

It seems likely, in fact, that the reported Liard pollutant levels are in excess of what was the average for the reach at the time of sampling; and that the downstream pollutant levels at Arctic Red River were underestimates of the true pollutant level. This speculation is based simply on the location of the SV sampling sites in the cross-section in relation to the morphology of the channel in the two reaches.

In the case of the Liard reach, sampling was done in the right side thalweg: the left side of the river is much shallower, reflecting the deposition of sand along the left bank. It might be inferred from this geomorphic appraisal of the reach that, during high flows, sand concentrations in the suspended sediment are higher (and hence total suspended sediment concentrations are higher) along the left side of the lower Liard. Thus, hydrocarbons (incorporated in the silt-clay fraction) would be found at higher levels in the sediment at the sampling site: towards the left side of the channel, these levels would decrease as additional sand was incorporated into the suspended sediment.

Though there are no multiple vertical samplings at this section to confirm this speculation, data from the WRB sediment station on the Liard (about 10 km upstream) appear to support the view. Past multiple vertical samplings at the WRB section have consistently shown higher sediment levels towards the left side of the channel, largely due to increased amounts of sand in suspension. Moreover, on the day of NHRI sampling, the estimated WRB suspended sediment concentration for the Liard station was 473 mg/L, substantially higher than the sediment concentration sampled by NHRI of 87 mg/L. In part, the lower NHRI sediment level will be due to sampling only the near-surface water; in part it is due to sampling in the thalweg. In any case, what it means is that the bias in the NHRI sampling to sand-poor sediment produces a PAH level in the sediment that must be too high for the reach as a whole.

In contrast, the NHRI sampling in the Mackenzie at Arctic Red River was done essentially at the regular WRB single-vertical site (Fig. 6.15), downstream of the left bank bar, and well-away from the right-bank thalweg. This implies that the NHRI sample at Arctic Red River would have been enriched in sand relative to the cross-section as a whole (Fig. 6.16); thus PAH levels in the sediment would have been underestimates of those found in the thalweg and the mean for the reach.

The discussion above is not intended as a criticism of the NHRI study. Their goal was not so much to characterize reaches of the Mackenzie according to pollutant levels in the sediment as to allow inferences regarding sediment loadings (less affected by the criticisms noted above). But the goal of attempting to classify river reaches in terms of pollutant levels in sediment is certainly one that is likely to emerge in this new era of environmental awareness. To do so meaningfully, requires careful assessment of where water-quality sampling is undertaken, just as is required by WRB in terms of its regular sediment sampling program. In both cases, an appreciation of fluvial geomorphology can assist in this planning.

11.3 <u>Geomorphic awareness: who needs it?</u>

Throughout this report, the comment has been made that a geomorphic perspective can help in the planning of monitoring programs, in the construction of river and floodplain works, and in the interpretation of data from such programs. The implication seems to be that the IWD staff to benefit most from such a perspective are the professional staff, those responsible for program design and evaluation.

The same conclusion was reached several years ago, by Northwest Hydraulics Consultants Ltd. (1985) in their review of sediment issues in the prairies. They noted in their conclusion:

> Perhaps the most significant change required is a philosophical one whereby at least senior professional staff in the program would be expected to take a broader view of the program role, and be more concerned with data interpretation, and less with the day-to-day mechanics of data collection and processing.
>
> This implies a wide-ranging interest in the complexities of sediment erosion, transportation and deposition in relation to water resources engineering, land management and environmental quality.

This statement is much like the one put forward at length in the present report, with the difference that it was written by a consulting engineer, rather than a geomorphologist. But there is a second difference. NHCL (1985) went on in a slightly different tone:

> It should be noted, however, that in a program that requires careful routine data collection and record keeping, it would not be appropriate to require all staff to take a broad scientific view of their role, since the type of personnel who are suited to the latter viewpoint are generally unsuited to routine operations.

There is certainly some truth in the latter comment, but this should not be interpreted to mean that those who <u>are</u> suited to routine operations (the technical staff) are not suited to thinking about the scientific aspects of their work. On the contrary, at a qualitative level at least, the technical staff usually has a far better understanding of what is taking place in the drainage basin, because these are the individuals who spend most of their time in the field. In a real sense, the technical staff represents the eyes of the Inland Waters Directorate, and they, just as much as the professional staff, need the kind of perspective that has been put forward in this report.

Indeed, it is quite clear, that some of the IWD technical staff already have a good appreciation of many parts of geomorphology, but the importance of this perspective seems not to have been realized by them, nor by the office staff to whom they report. In talking with technical staff across the region, it is evident that many of them have a good knowledge of the history of the river reaches in which they work; it is not always obvious, however, that they appreciate the significance of what they know. While the movement of a major bar towards a hydrometric measurement section is likely to be noted with some concern, resulting in increased frequency of gaugings or the search for a new hydrometric section, similar (but less severe) changes in the bed morphology (which could have important repercussions for the sediment sampling program), while noted, may not be reported to the district office because its significance is unclear. Yet those in the office (who might perceive its significance) may be unaware of the situation for years!

It is insufficient to have only the professional staff trained in geomorphology, if this staff rarely sees the field in which the data collection is undertaken. By the same token, it is unrealistic to expect the technical staff in the field to voice their observations if they have never been instructed as to the potential significance of what they see in the landscape. Hopefully, this report will foster an increased geomorphic perspective at all levels in IWD, and, through it, an increased communication between field and office staff. There is no question that, at the technical level, accurate, routine data collection is the foundation of IWD work, but this does not mean that it should not be supplemented by qualitative observations in the field that also bear on the problem.

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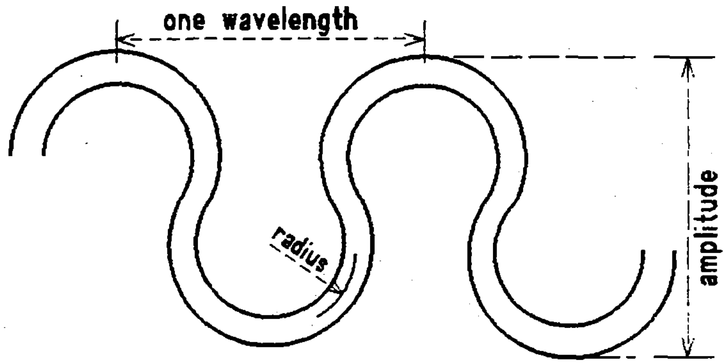


sinuosity (P) = channel distance (solid) / axial distance (dashed)

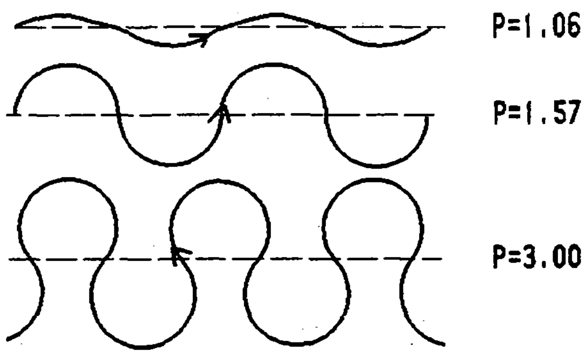


Figure 1.1 Attributes of idealized meander plan geometry.

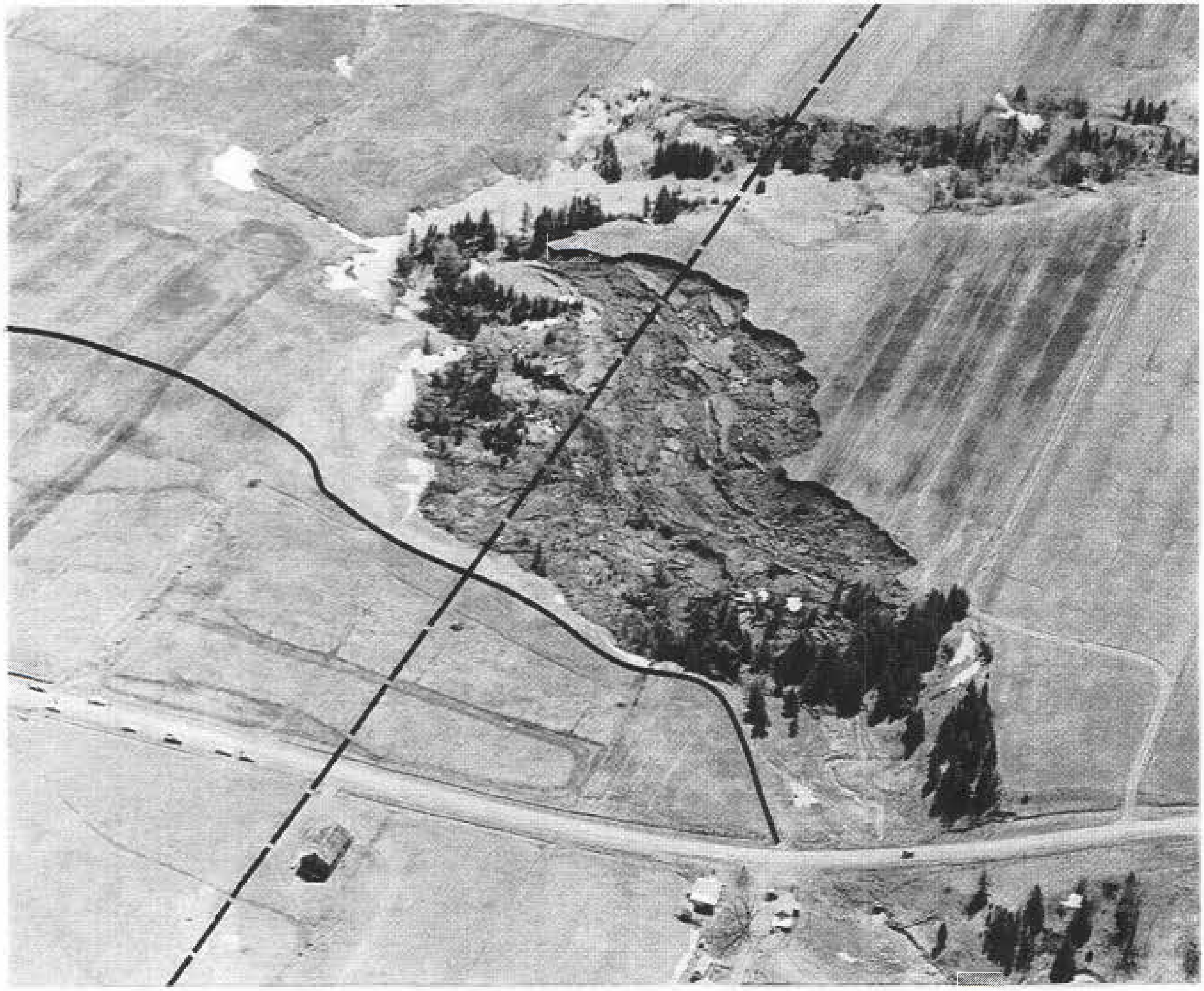


Figure 1.2 Oblique view of 1978 Rigaud landslide bowl, Quebec (from Carson, 1979). The centre-line of the planned pylon line is marked on the photograph. The cars of visitors to the slidebowl are parked in an older, larger slide bowl, part of the backscarp of which is marked on the photo.

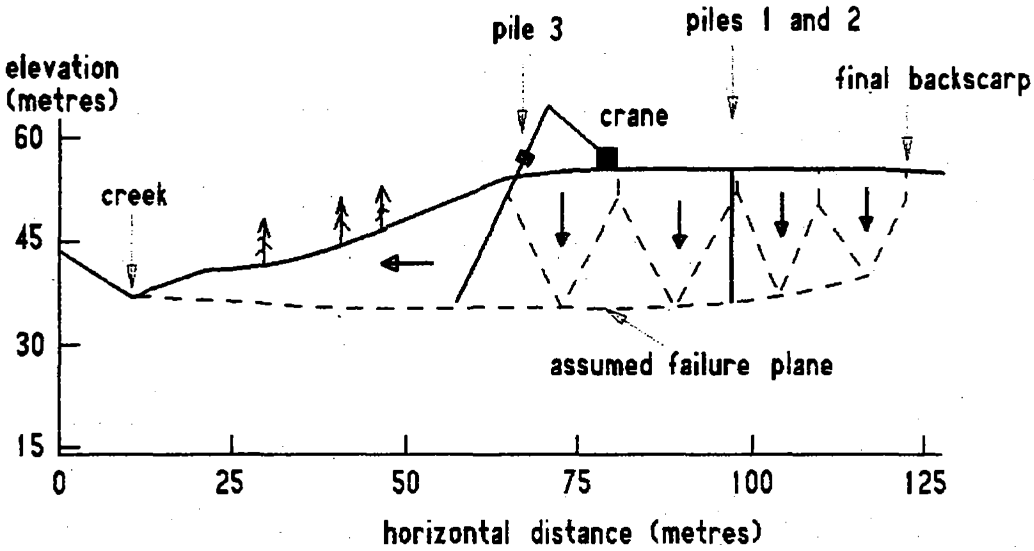


Figure 1.3 Mode of failure of Rigaud landslide.

Figure 1.4 Sunken wedge strips inside Rigaud landslide bowl (after Carson, 1979). Dashed line emphasize topography along three cross-sections.

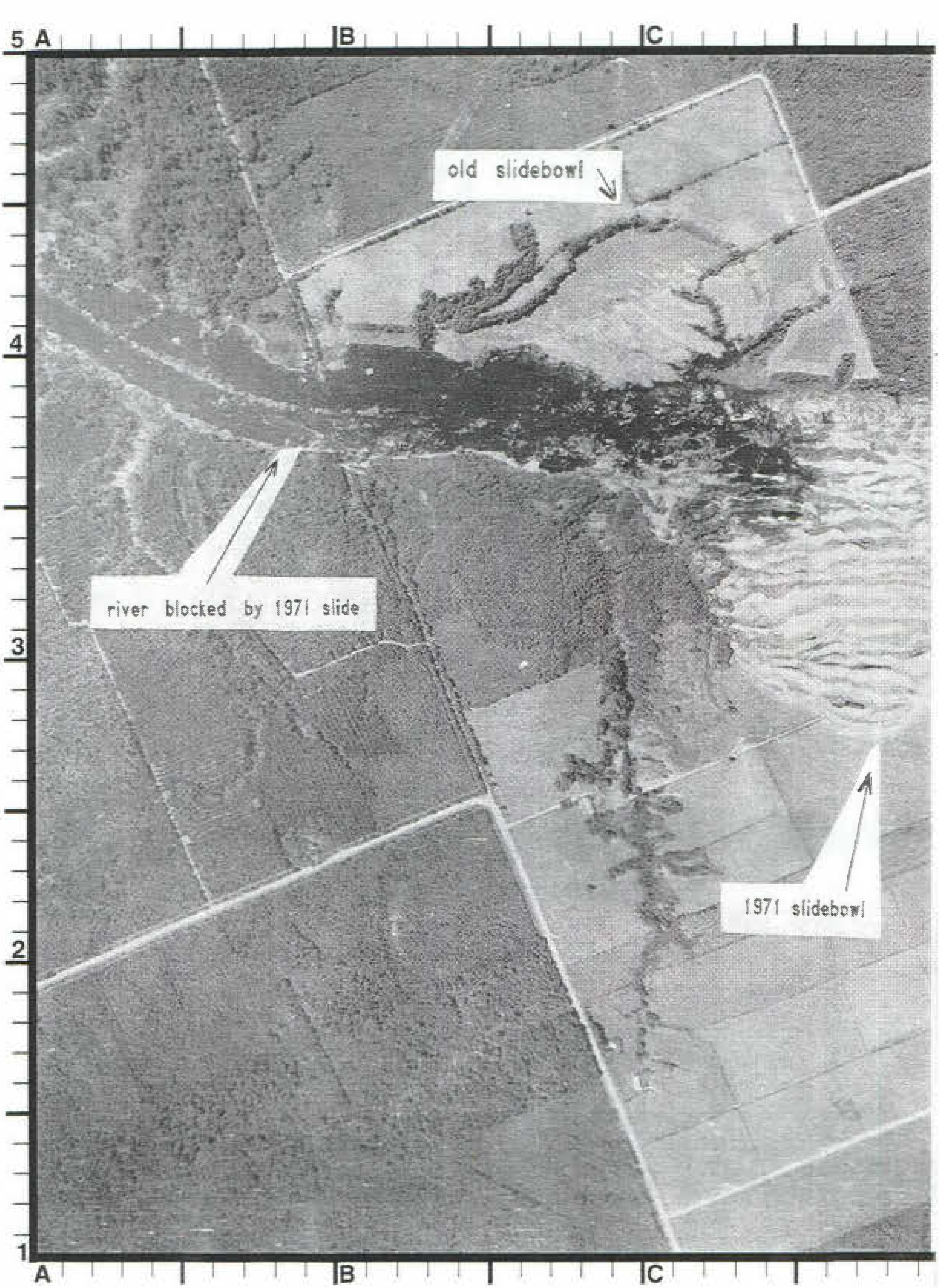


Figure 1.5 Landslide bowls in South Nation River valley, Ontario (from Mollard and Janes, 1984).

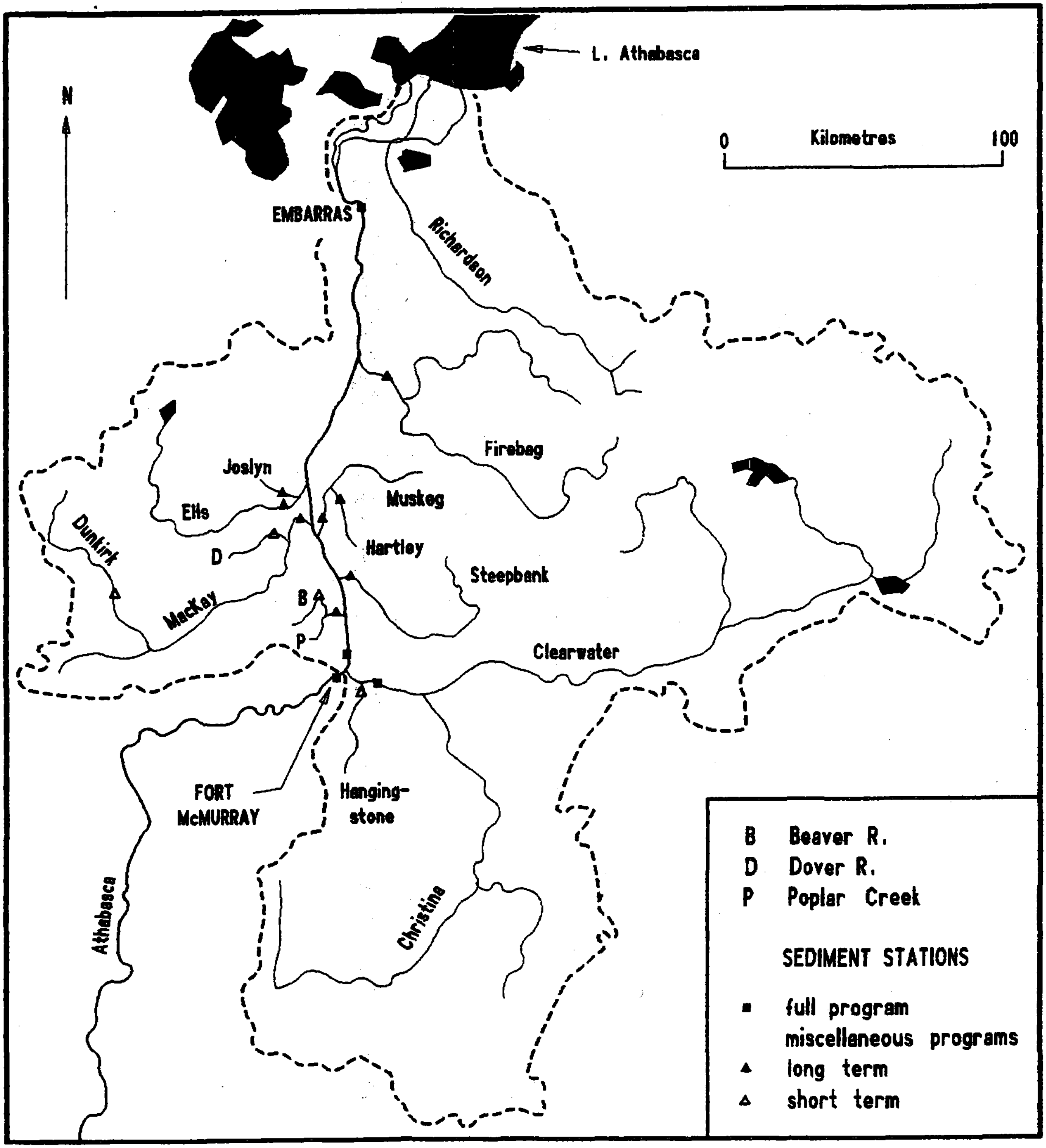


Figure 1.6 Sediment sampling stations of the Lower Athabasca basin, Alberta.

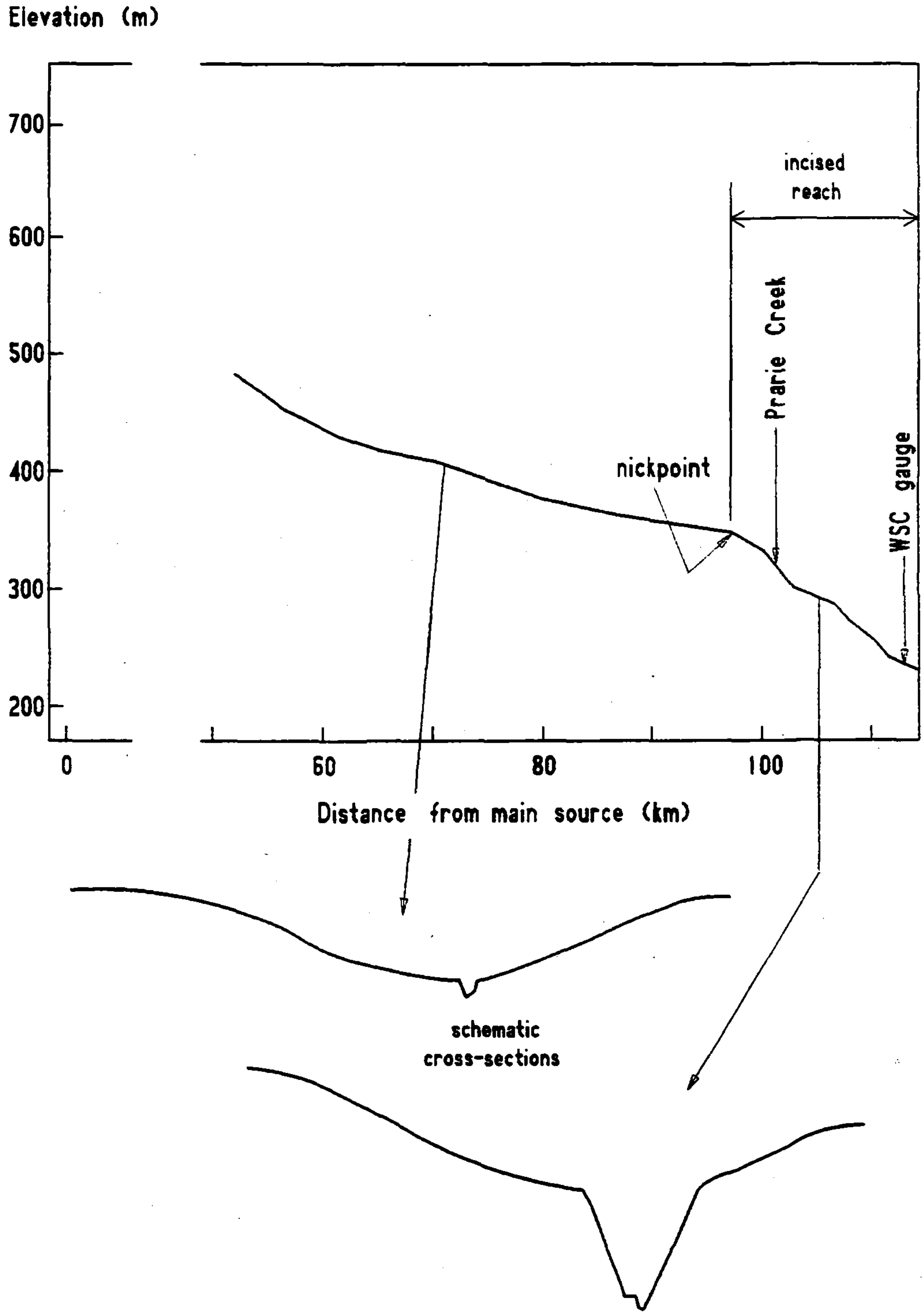


Figure 1.7 Long profile of Hangingstone River.

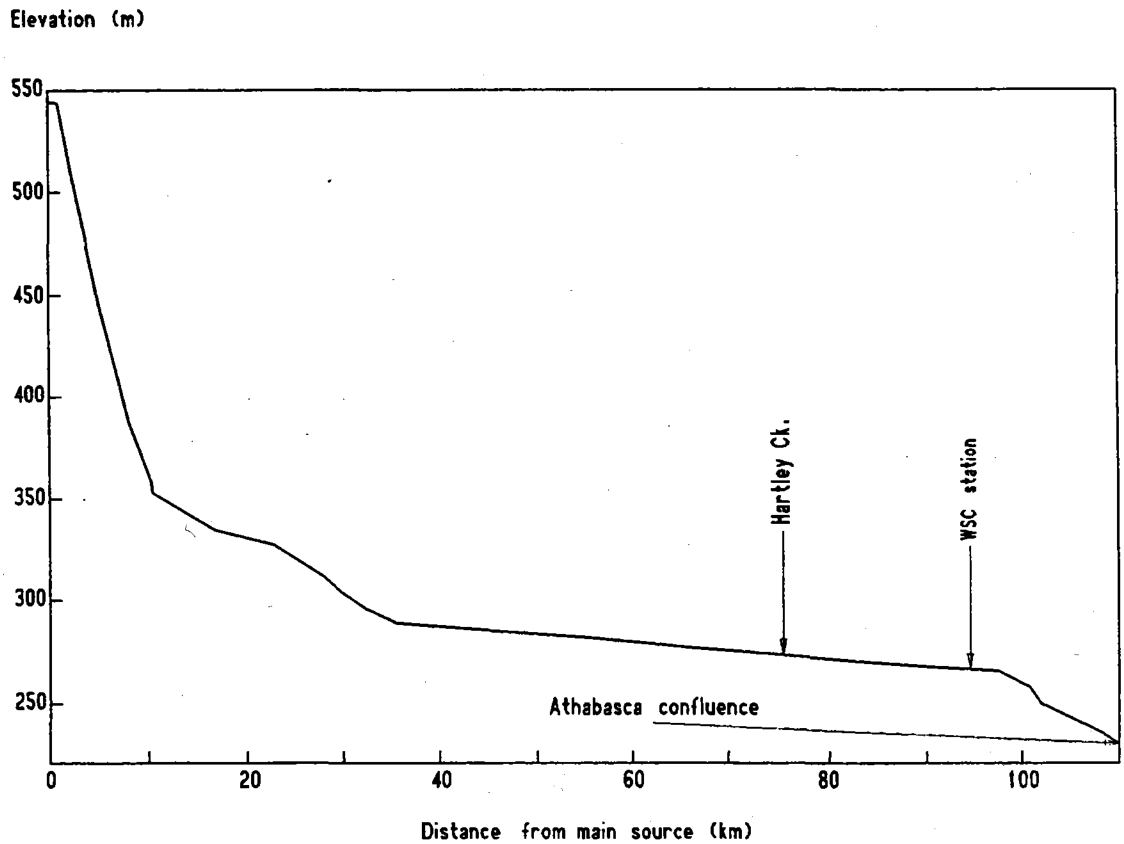


Figure 1.8 Long Profile of Muskeg River.

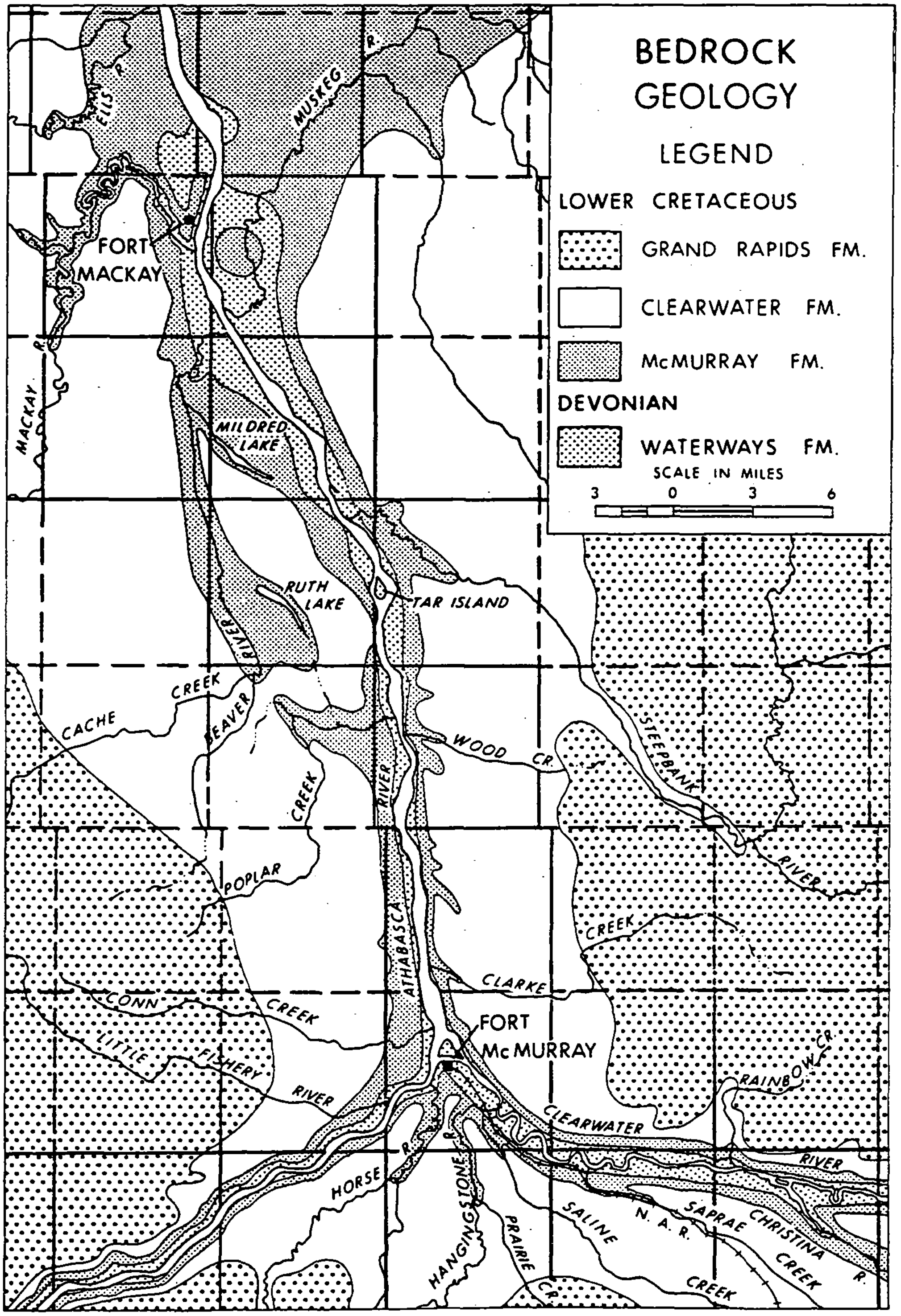


Figure 1.9 Bedrock geology of Fort Mcmurray area, Alberta (from Hamilton and Mellon, 1973).

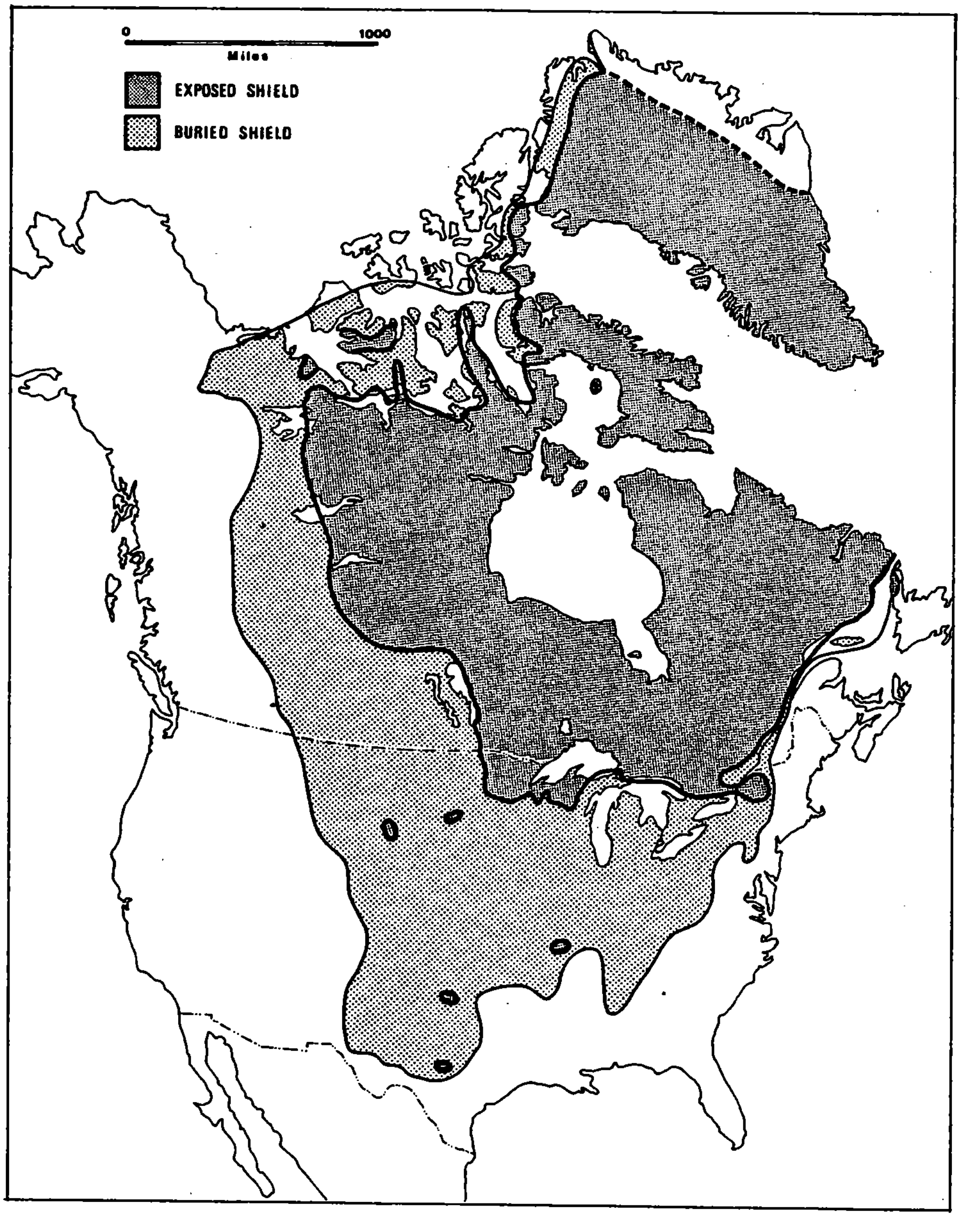


Figure 2.1 The Canadian Shield: exposed and buried areas (from Bird, 1972).

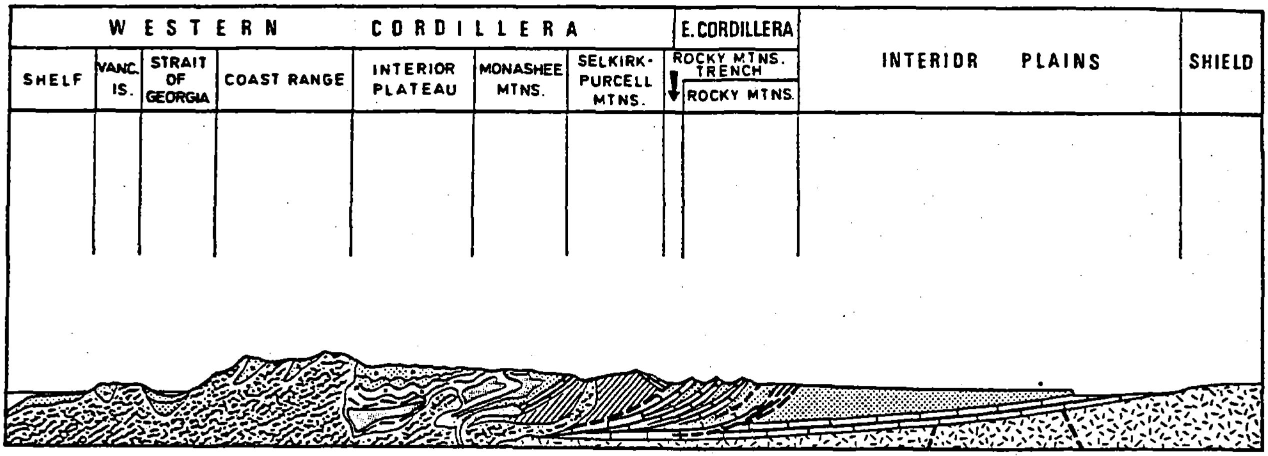


Figure 2.2 Topographic and geologic cross-section of Western Canada (from Bird, 1972).

EON	ERA	PERIOD / SYSTEM		EPOCH / SERIES	DURATION OF TIME UNIT IN 10^6 YEARS Ma (MEGA ANNUM)	AGE OF BOUNDARY IN 10^6 YEARS = YEARS BEFORE PRESENT Ma (MEGA ANNUM)	DISTINCTIVE FEATURES AND EVENTS
PHANEROZOIC	CENOZOIC 65 Ma	QUATERNARY	NEOGENE	HOLOCENE		0.005-0.02	Modern climate Historic man 6,000 years ago
				PLEISTOCENE	2.5 ±0.5	2.5	Extensive glaciation in northern hemisphere Modern man 40,000 years ago Stone age man 600,000 years ago?
		TERTIARY		PLIOCENE	2 ±0.5	4.5	Man (fire and tools) 3 Ma ago Near-man 5 Ma ago
				MIOCENE	18.5 ±1	23	Numerous mammals and flowering plants (including grasses, grains, etc.)
			PALEOGENE	OLIGOCENE	12 ±1	35	
				EOCENE	19 ±1	54	
				PALEOCENE	11 ±1	65	Building of Rocky Mountains at beginning of Tertiary
	MESOZOIC 160 Ma	CRETACEOUS		UPPER	30 ±5	95	Extinction of dinosaurs and many other animals at end of Cretaceous
				LOWER	42 ±5	137	First flowering plants First deciduous trees
		JURASSIC			53 ±5	190	Numerous reptiles; first true mammal First birds
		TRIASSIC			35 ±5	225	First dinosaurs; amphibians attain maximum size Forests mainly of conifers
	PALEOZOIC 360 Ma	PERMIAN			67 ±5	292	Extinction of many animals Numerous conifers (evergreens) Extensive glaciation southern hemisphere Building of Appalachian Mountains
		PENNSYLVANIAN	CARBONIFEROUS		28 ±10	320	First reptiles; abundant insects Maximum extent of primitive forests "coal forests"
		MISSISSIPPIAN			20 ±10	340	Extensive limestone deposits composed of remains of sea lilies and Foraminifera
		DEVONIAN			65 ±10	405	First amphibians and lungfishes Numerous fishes, corals, brachiopods First forests of seed ferns, scale trees, etc.
		SILURIAN			25 ±15	430	First land plants (mosses, rushes) and air-breathing animals (scorpions)
		ORDOVICIAN			55 ±15	485	Numerous graptolites and other invertebrates
		CAMBRIAN			100 ±15	585	First abundant marine fossils, including numerous trilobites
PRECAMBRIAN	PROTEROZOIC 1915 Ma	HADRYNIAN			365 ±50	950	Very rare, primitive marine invertebrates 670 Ma ago
		HELIKIAN		NEOHELIKIAN PALEOHELIKIAN	750 ±100	1700	Greenvillian Orogeny 1000 Ma ago Hudsonian Orogeny 1750 Ma ago
		APHEBIAN			800	2500	Kenoran Orogeny 2500 Ma ago
	ARCHEAN 2000 Ma				2000	4500	Oldest algae (unicellular, aquatic plant) 3000 Ma ago Birth of Earth as solid body 4500 Ma ago?

Figure 2.3 The geologic timescale (from Mollard and Janes, 1984).

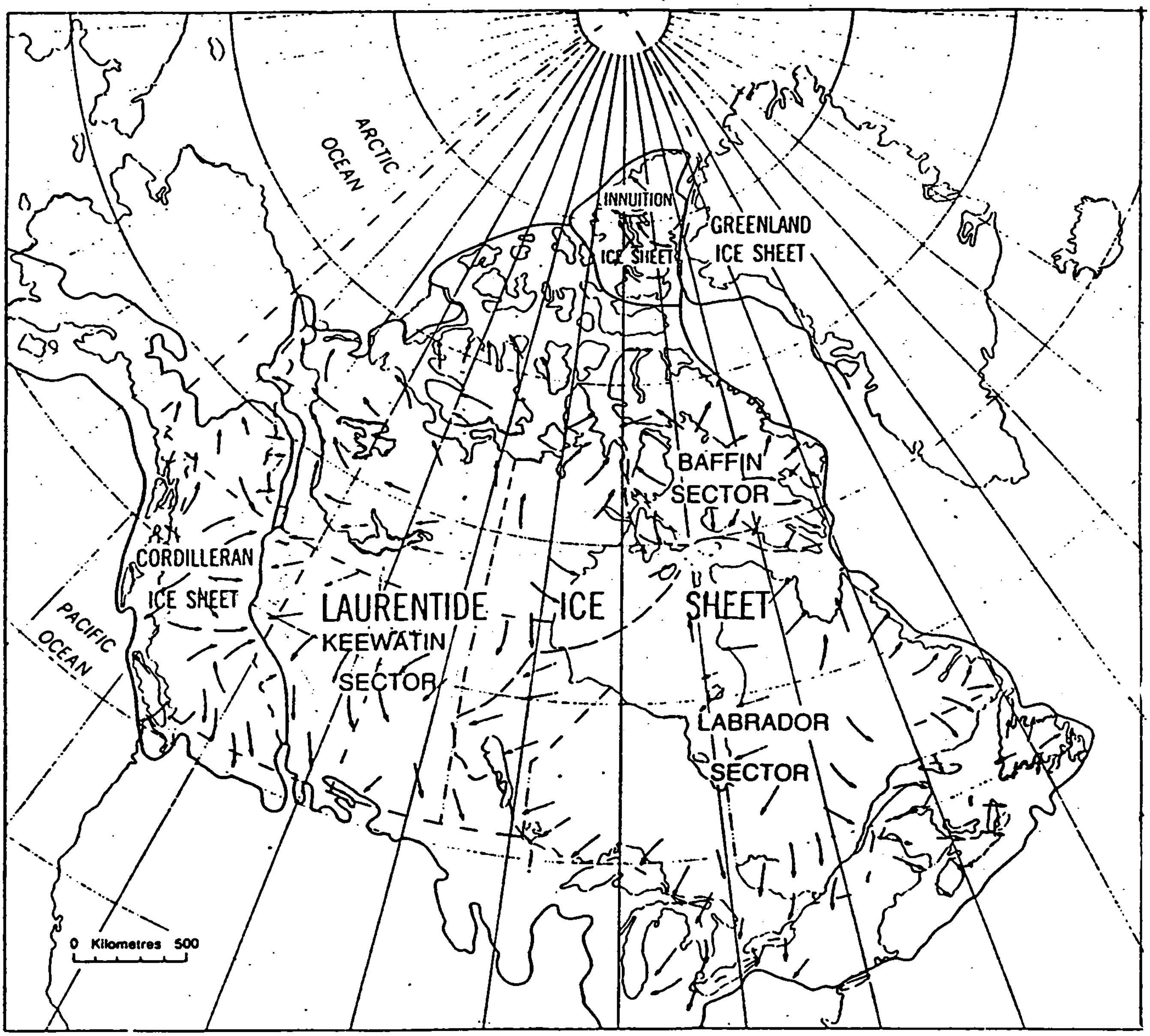


Figure 2.4 Approximate maximum extent of Wisconsin glaciation in Canada (from Prest, 1971).

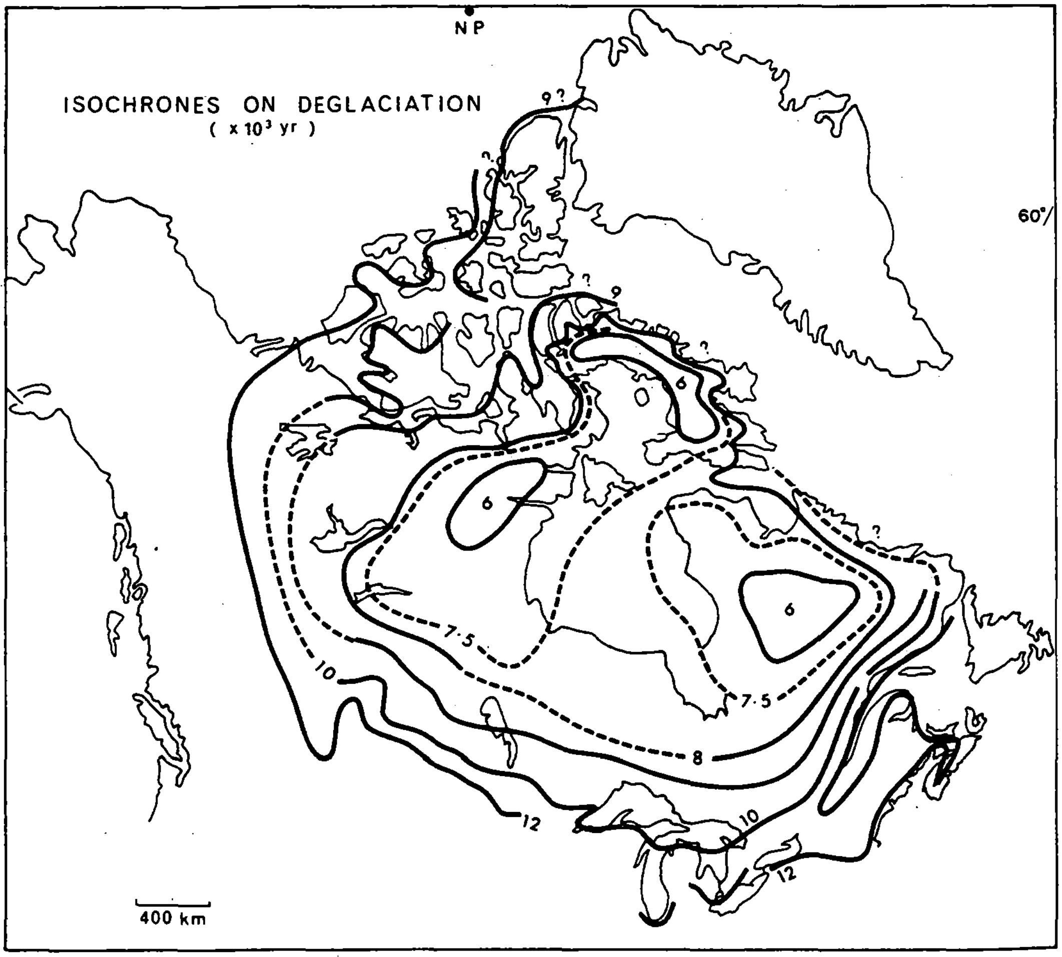


Figure 2.5 Retreat of the Wisconsin Lauretide ice fronts (from Bryson et al., 1969).

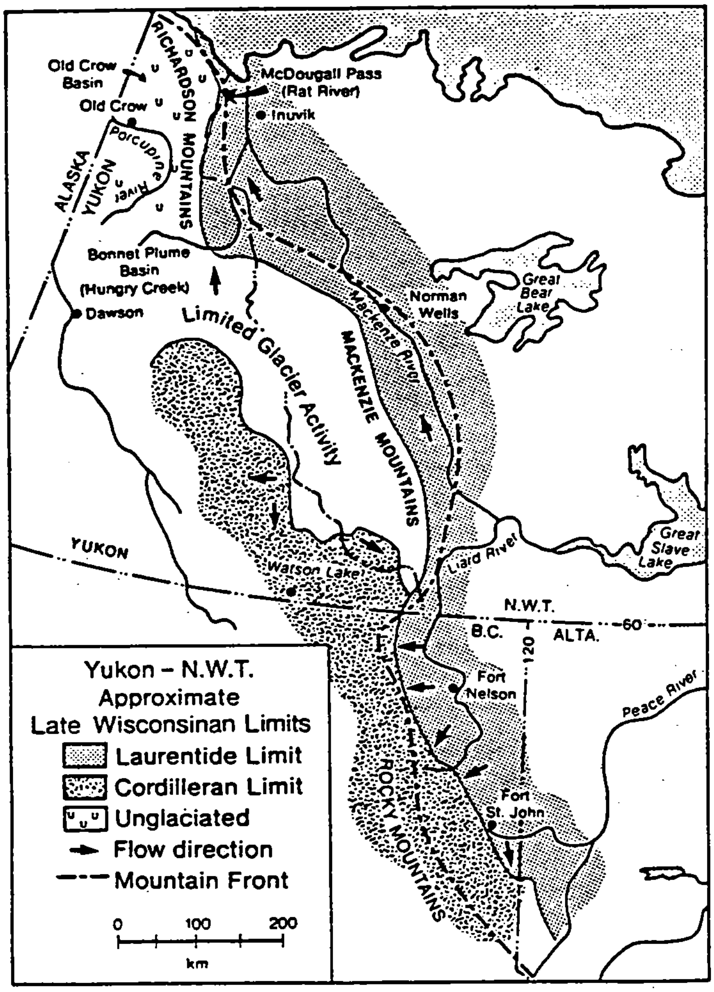


Figure 2.6 Unglaciated area of Yukon and NWT in late Wisconsin times (from Rutter, 1980).

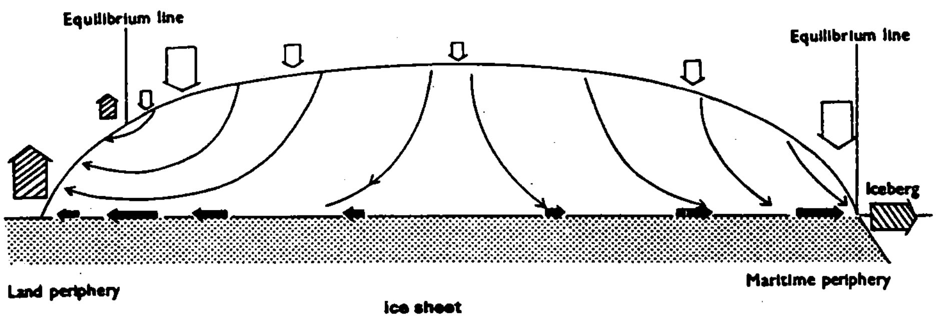


Figure 2.7 Model of ice flowage in a large ice sheet (from Sugden and John, 1976).

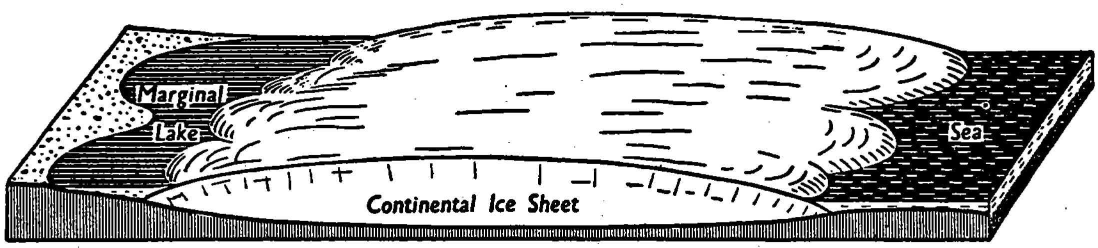


Figure 2.8 Pattern of isostatic depression under a large ice sheet (from Holmes, 1965).

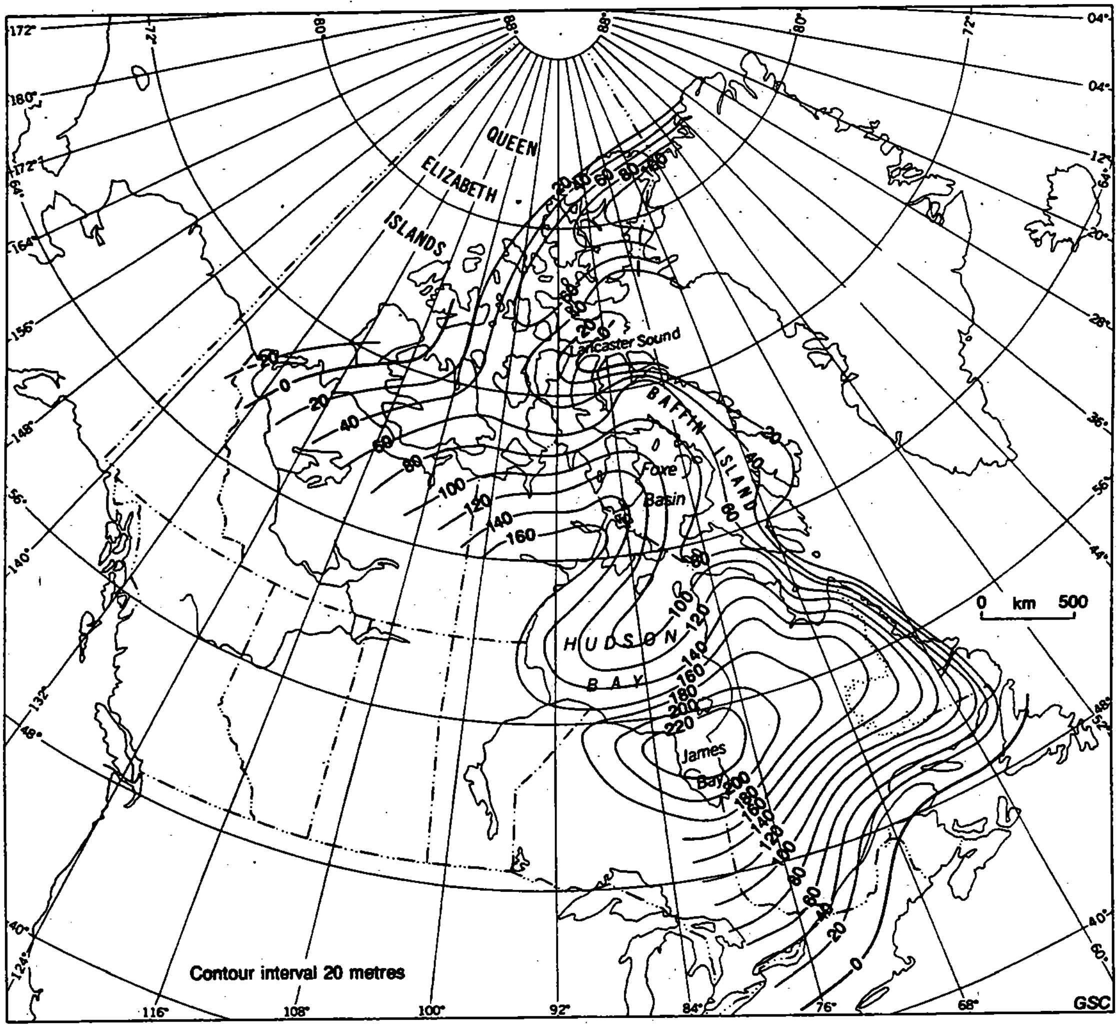


Figure 2.9 Pattern of isostatic rebound in Canada in last 7,000 years (from Fulton, 1989).

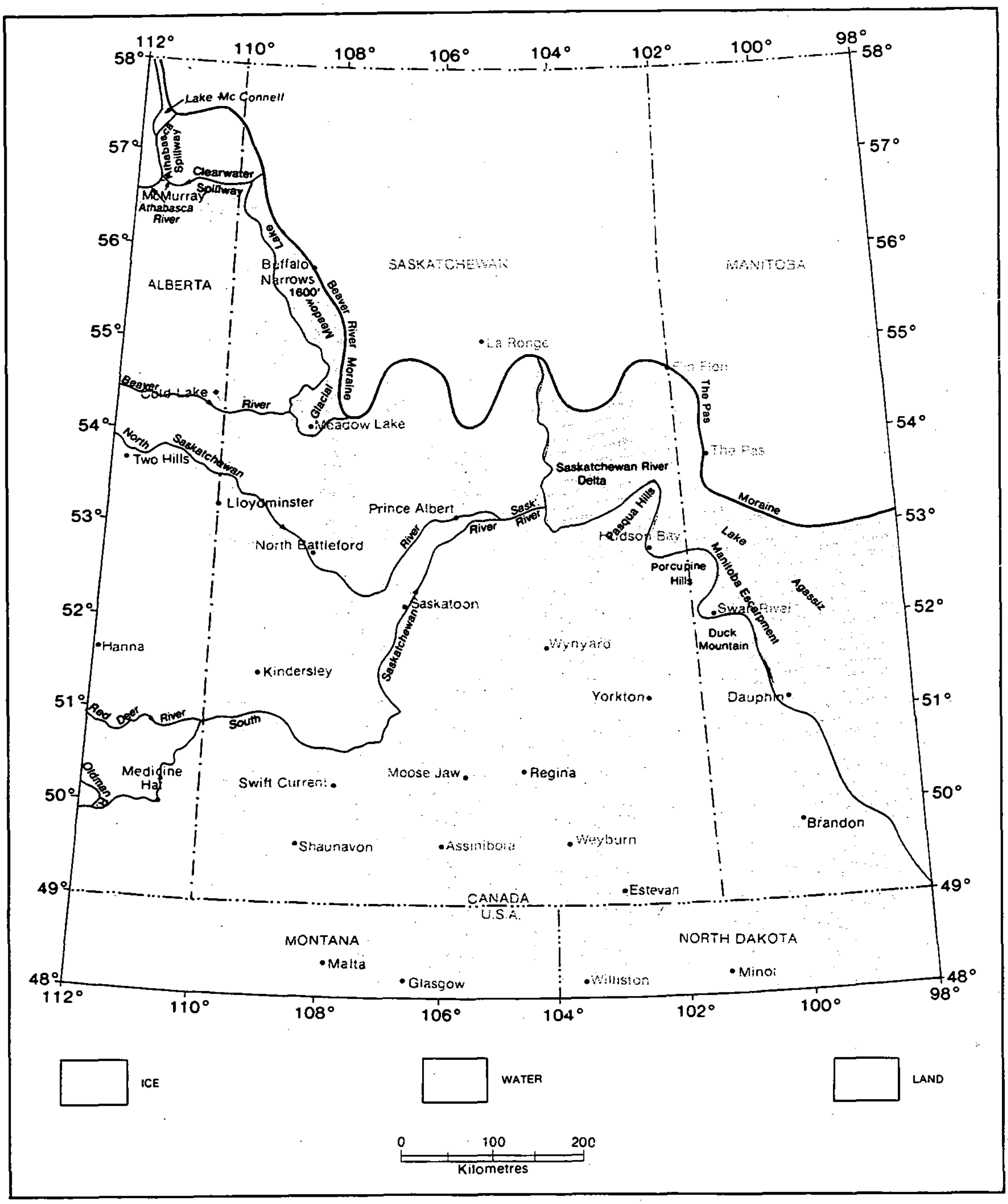


Figure 2.10 Drainage pattern of Central Prairies about 11,000 years ago (from Christiansen, 1979).

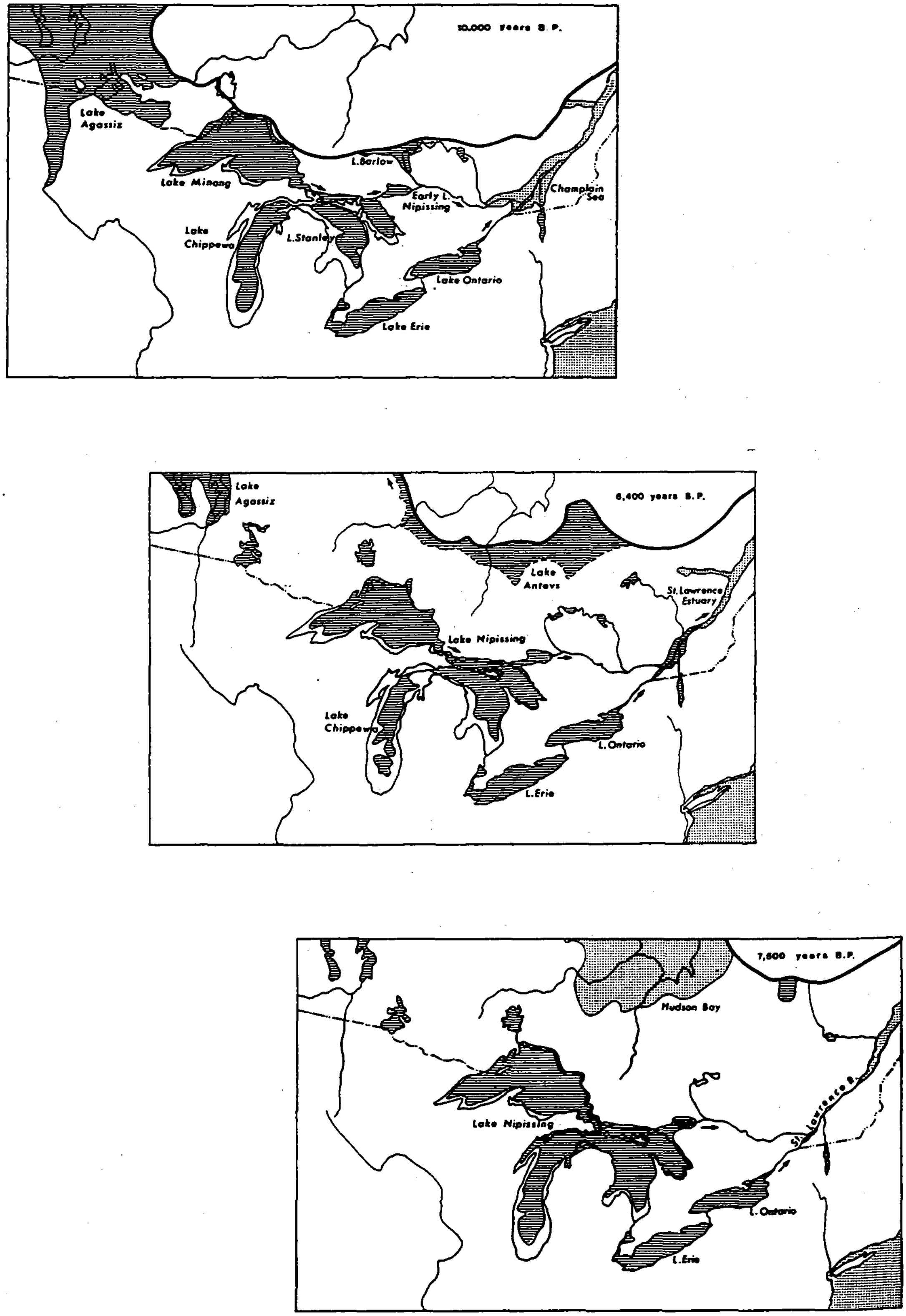


Figure 2.11 Pattern of Wisconsin deglaciation of Central Canada (from Bird, 1972).

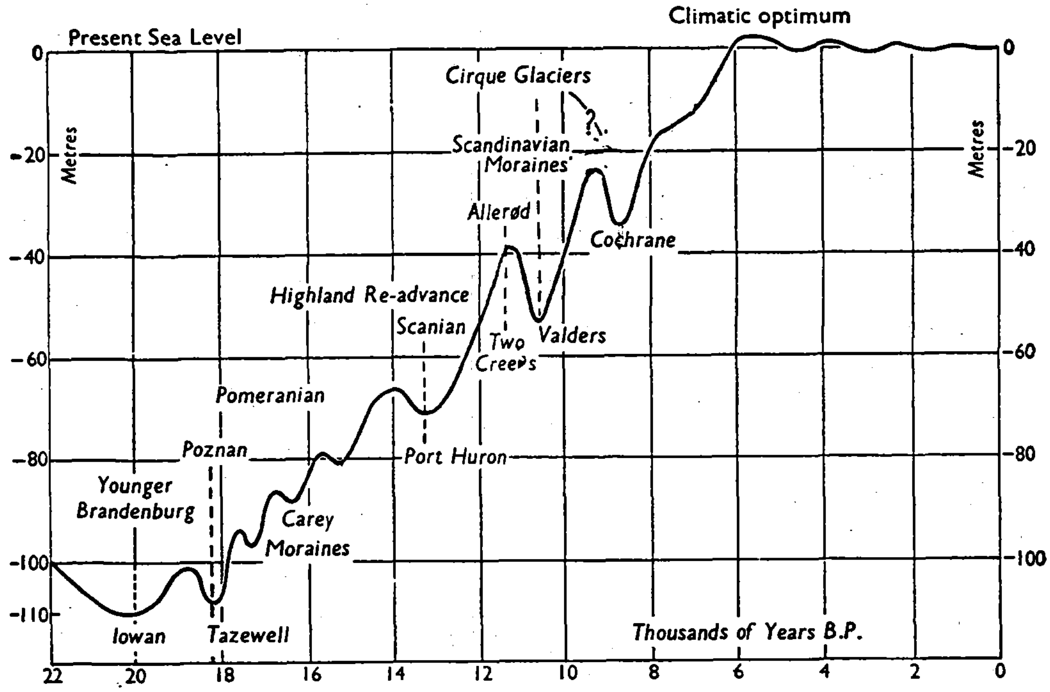


Figure 2.12 Change in eustatic sea level in the last 22,000 years (from Holmes, 1965).

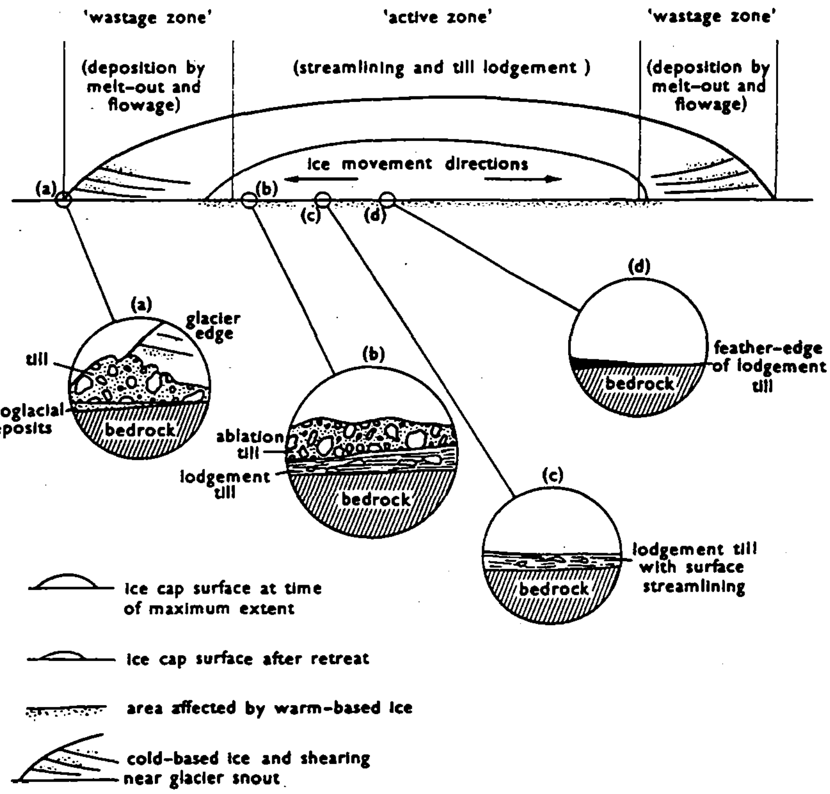


Figure 3.1 Model showing different zones of deposition under an ice cap (from Sugden and John, 1976)

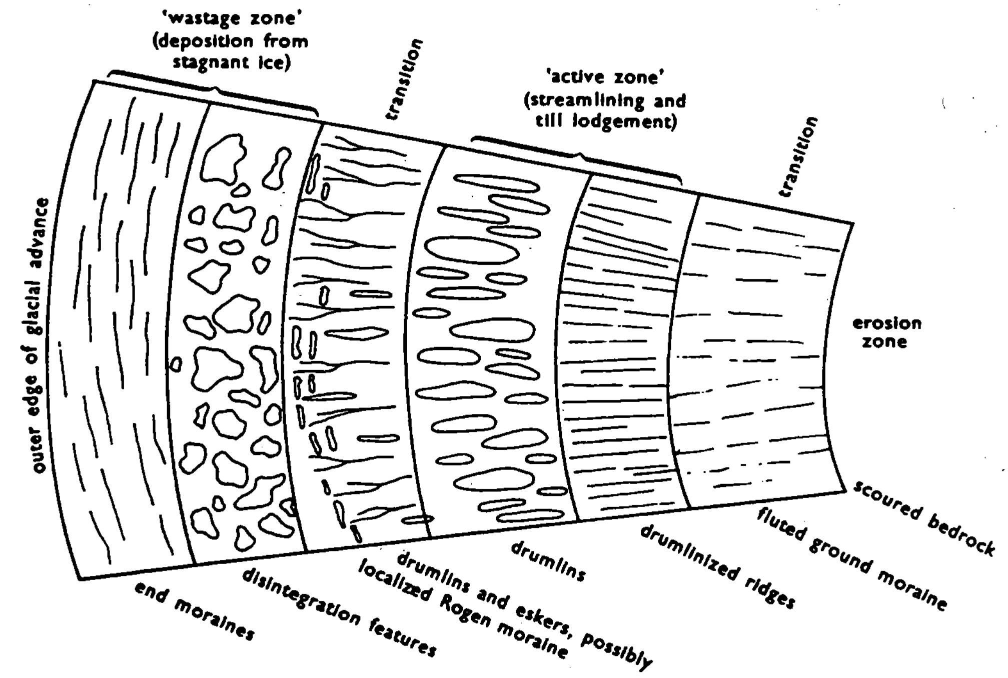


Figure 3.2 Idealized change in depositional zones under an ice cap (from Sugden and John, 1976

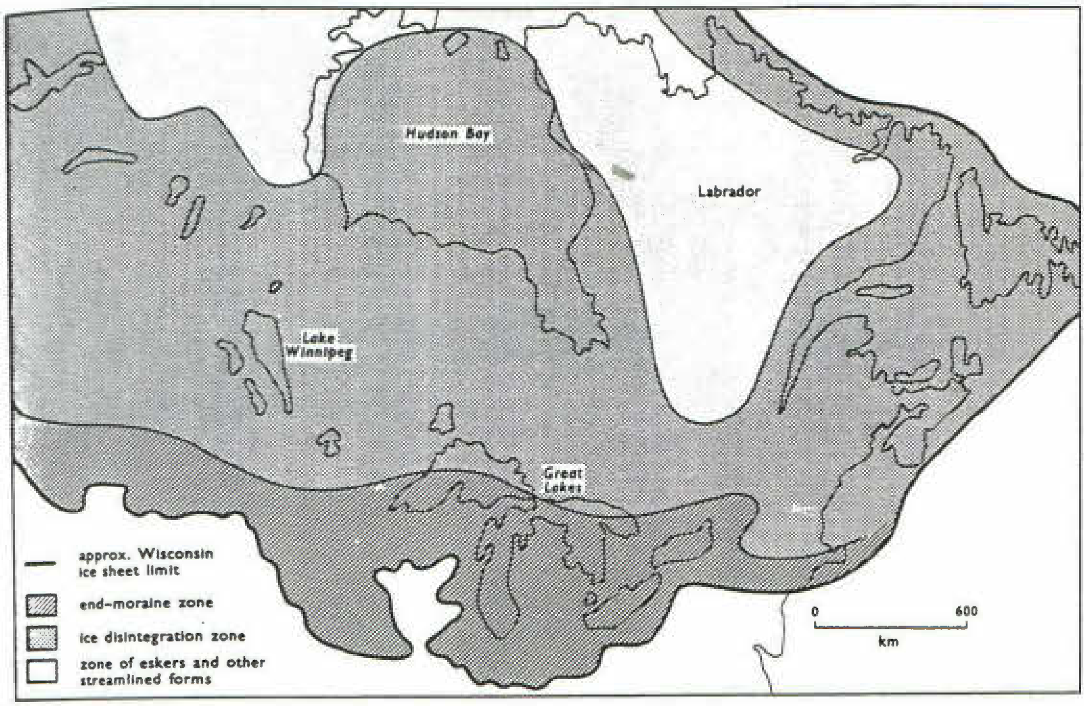


Figure 3.3 Generalized map of landscapes formed under southern part of Laurentide ice cap (from Flint, 1971).

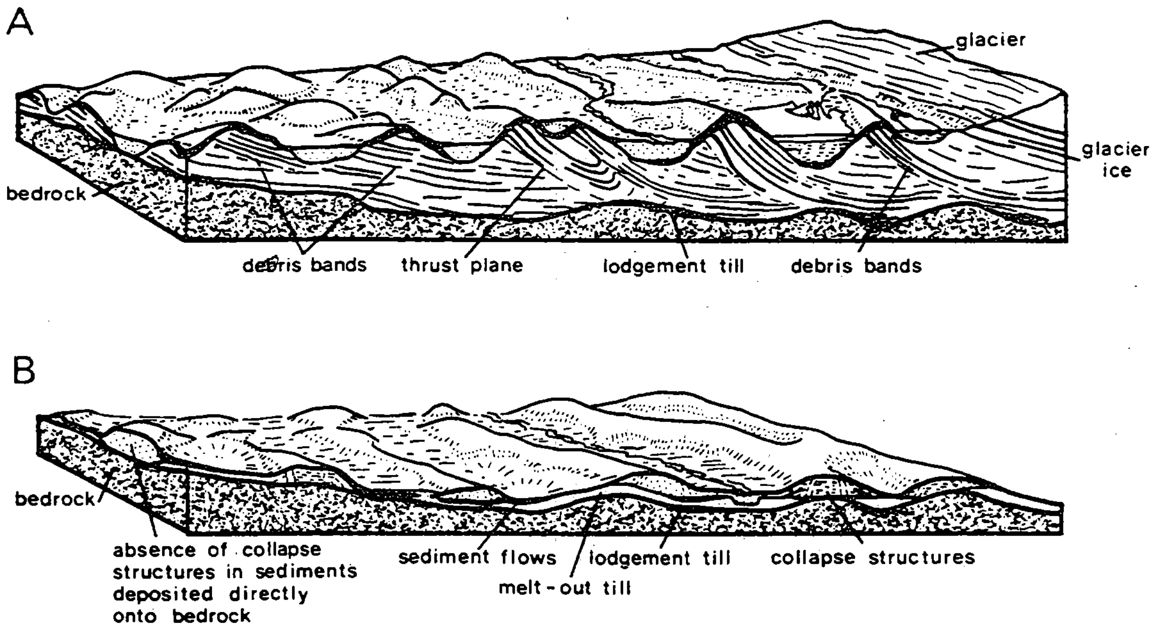


Figure 3.4 Schematic evelution of disintegration ground moraine (from Boulton, 1972).

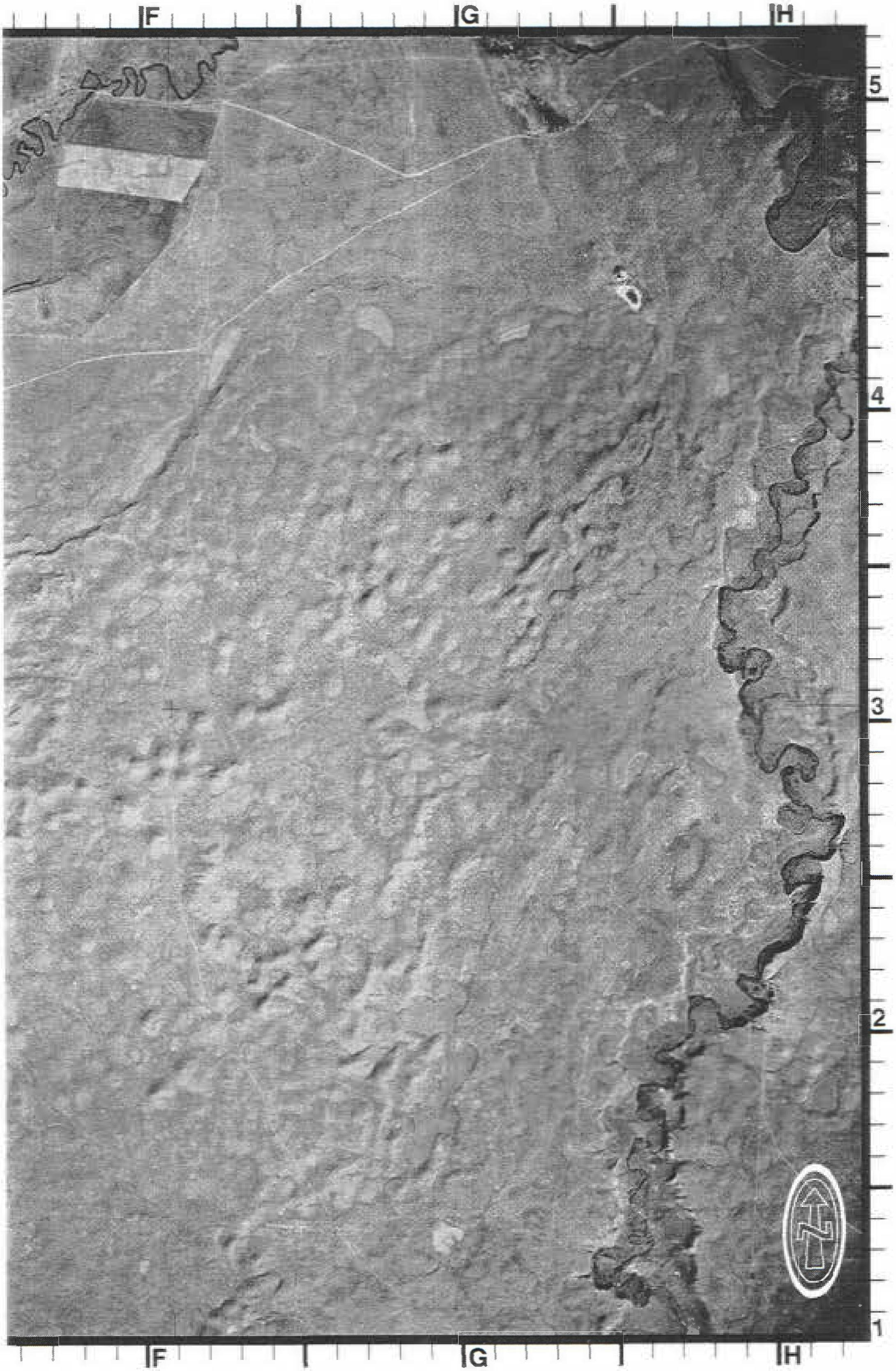


Figure 3.5 Hummocky disintegration moraine, southwestern Saskatchewan (from Mollad and janes, 1984).

Figure 3.6 Glacial grooves, Cameron Hills, Northwest Teritories (from Mollard and Janes, 1984).

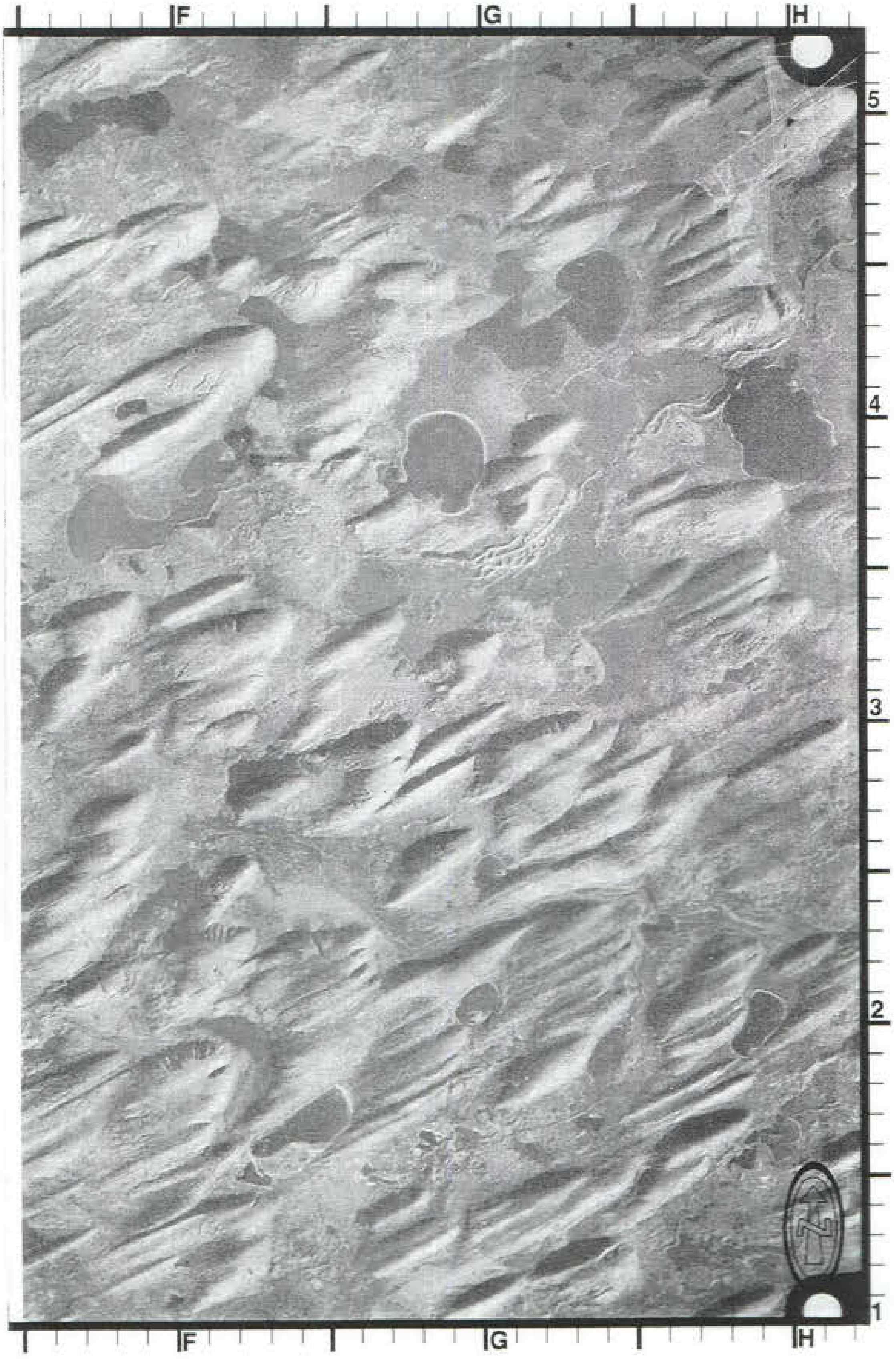


Figure 3.7 Drumlinoid landscape, Snare Lake, Sasktchewan (from Mollard and Janes, 1984).

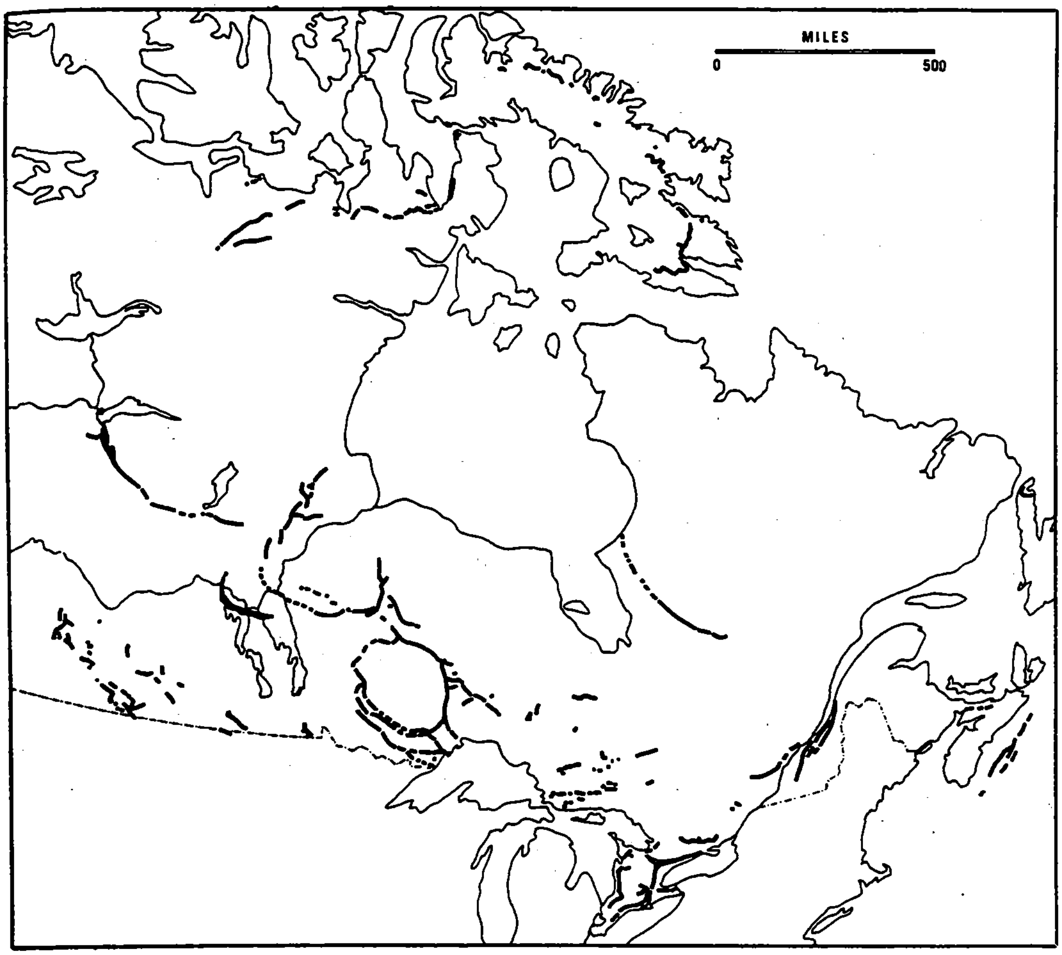


Figure 3.8 Map of main end moraines in Central and Eastern Canada (from Bird, 1972).

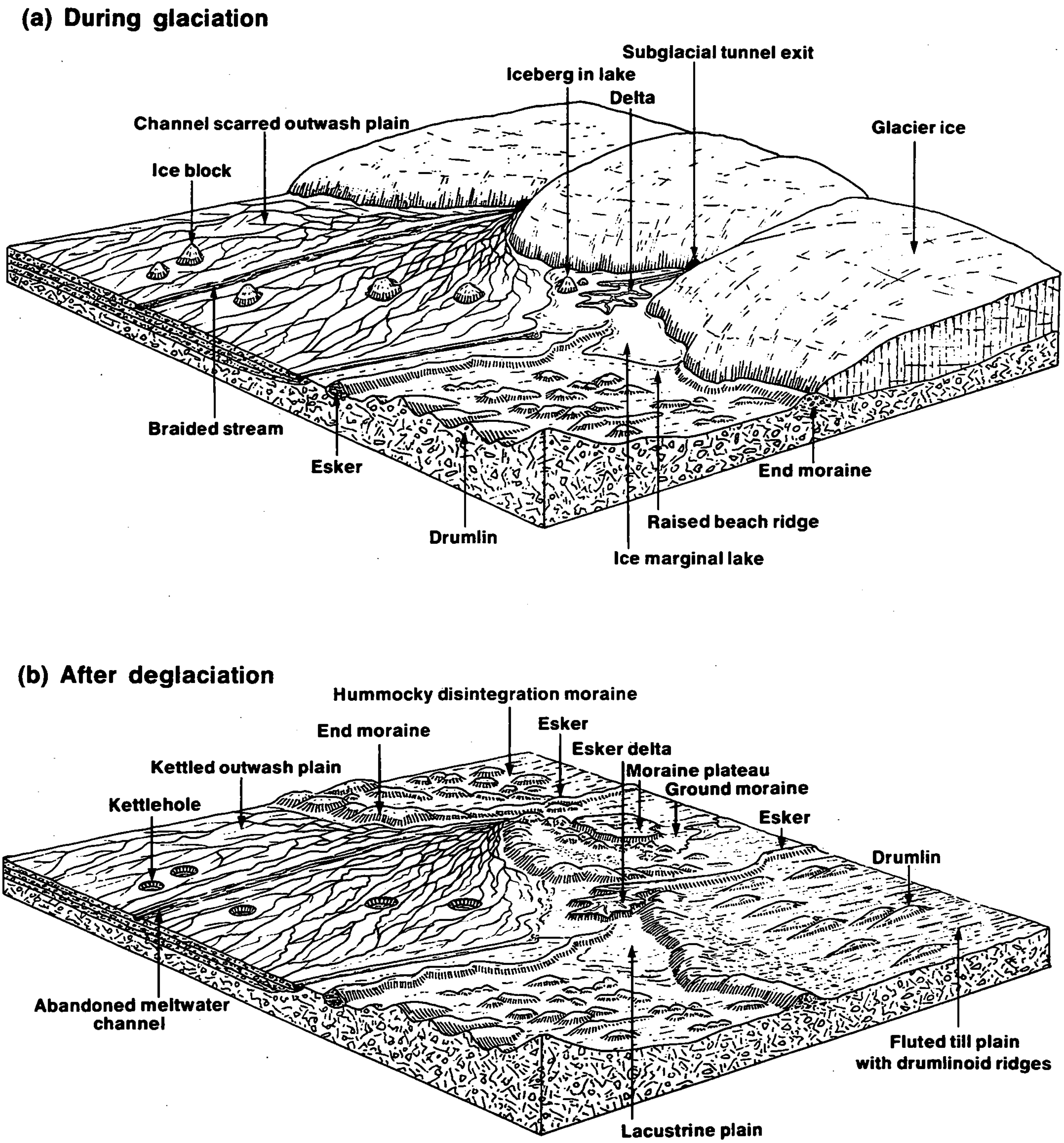


Figure 3.9 Landscapes formed by proglacial deposition (from Mollard and Janes, 1984).

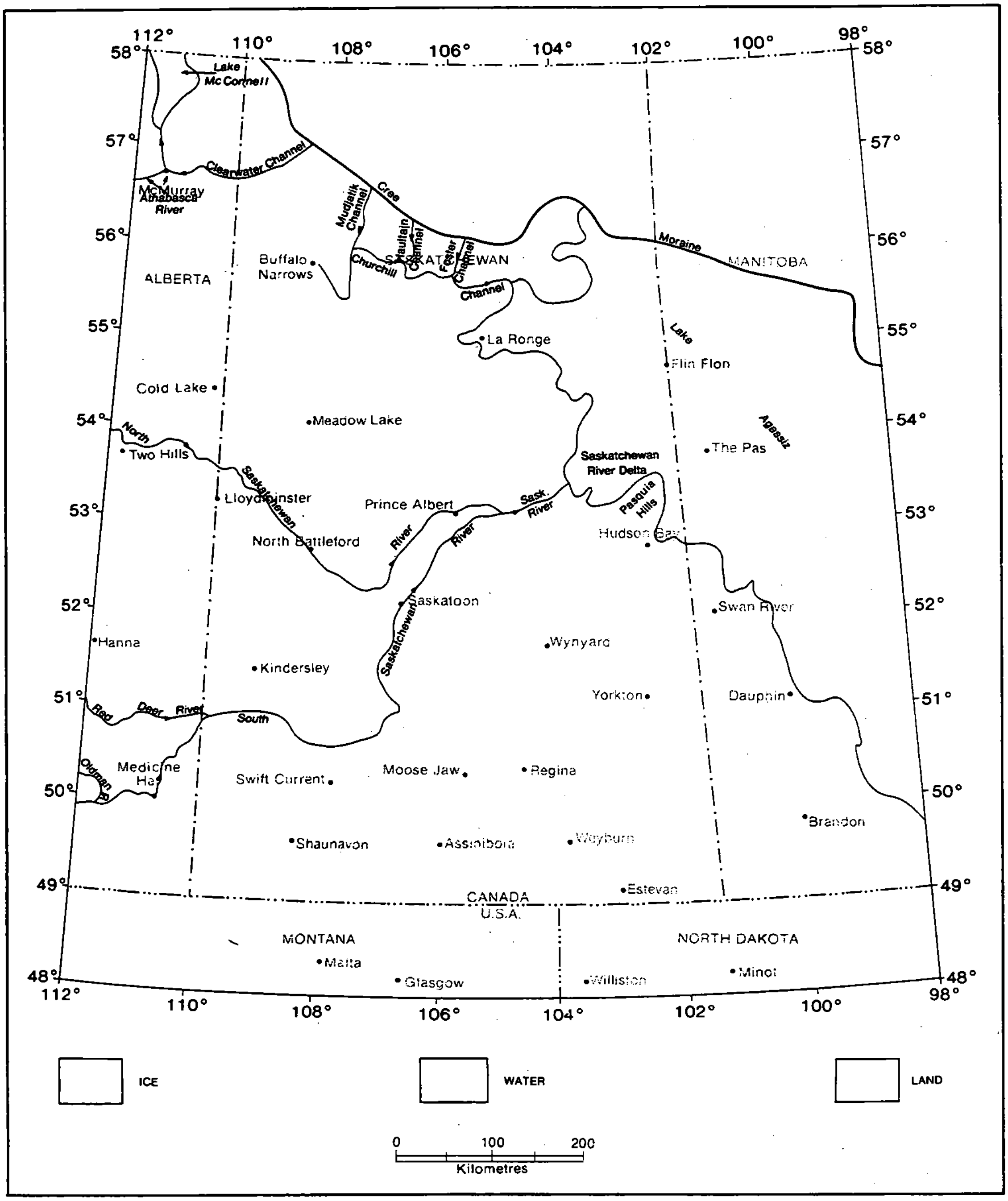


Figure 3.10 The Cree lake moraine of the north-central prairies (from Christiansen, 1979).

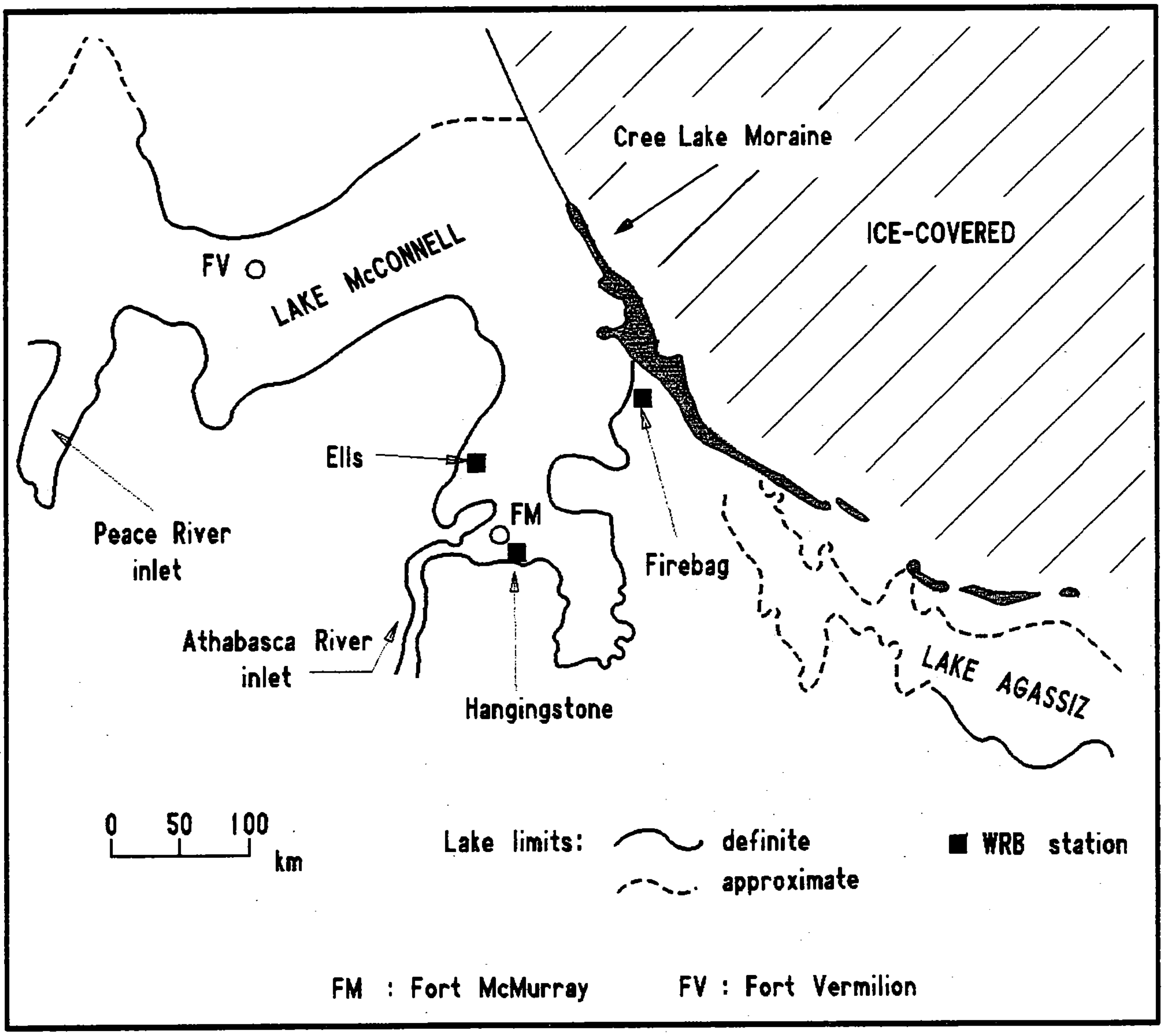


Figure 3.11 Late glacial conditions in the lower Athabasca basin.

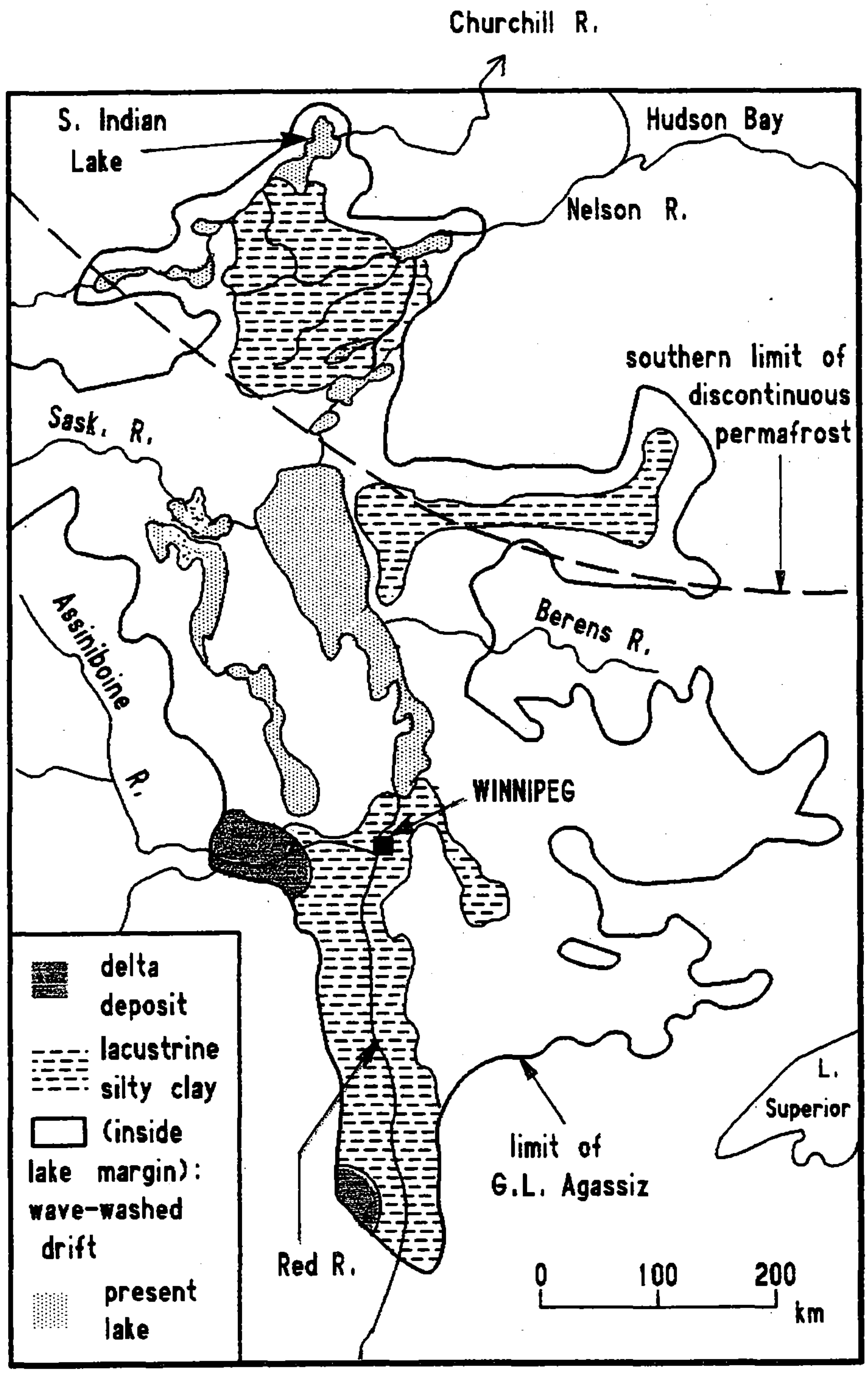


Figure 3.12 Surficial deposits in the area of Glacial Lake Agassiz (after Elson, 1961).

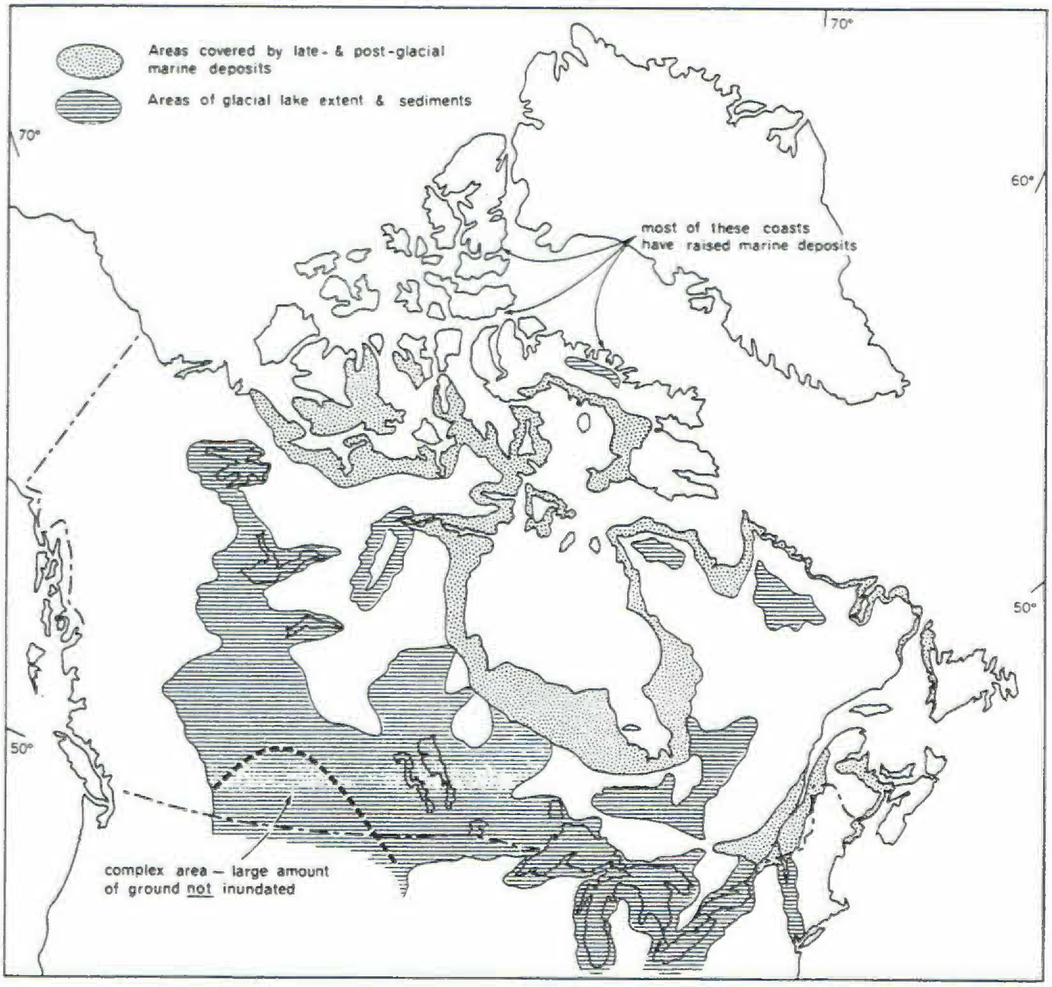


Figure 3.13 Map of areas covered by lakes and seas during the Wiscaonsin laurentide deglaciation of Canada (from Prest et al., 1968).

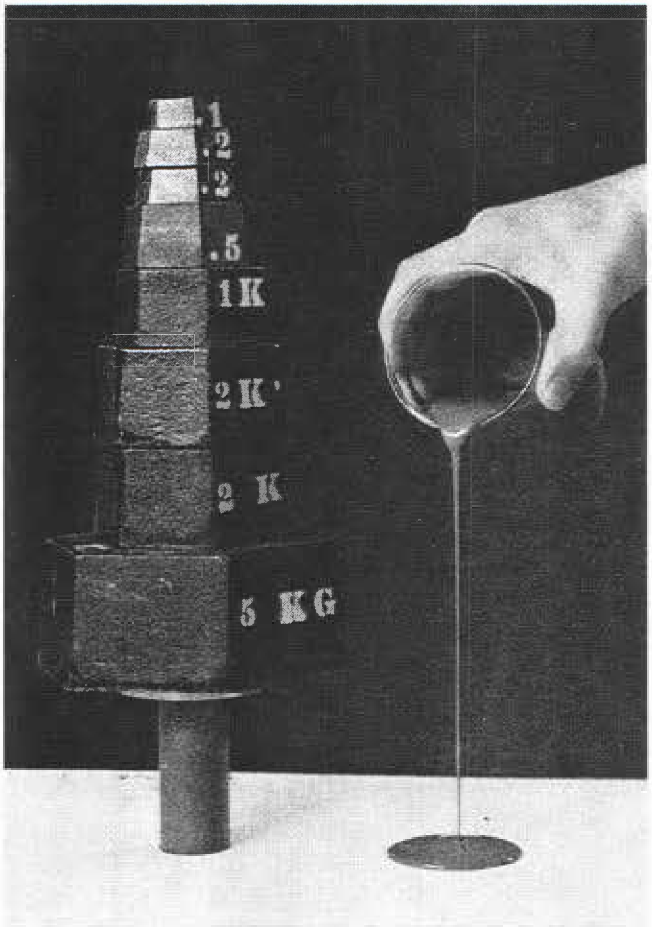


Figure 3.14 Two core samples of Champlain Sea in undisturbed and disturbed states (from Crawford, 1968).

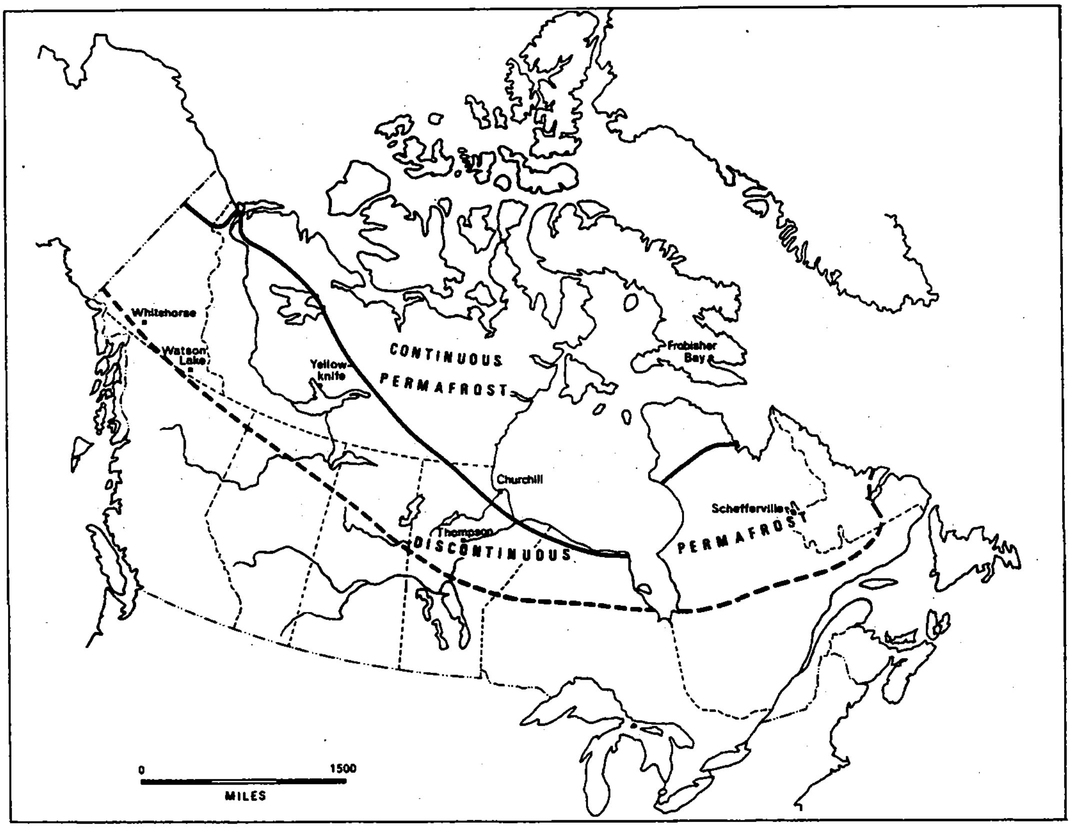


Figure 3.15 Map showing extent of permafrost in canada (from Bird, 1972).

Figure 3.16 Ice-wedge fissure ploygons, Ellef Ringnes Island, NWT (from Bird, 1972).

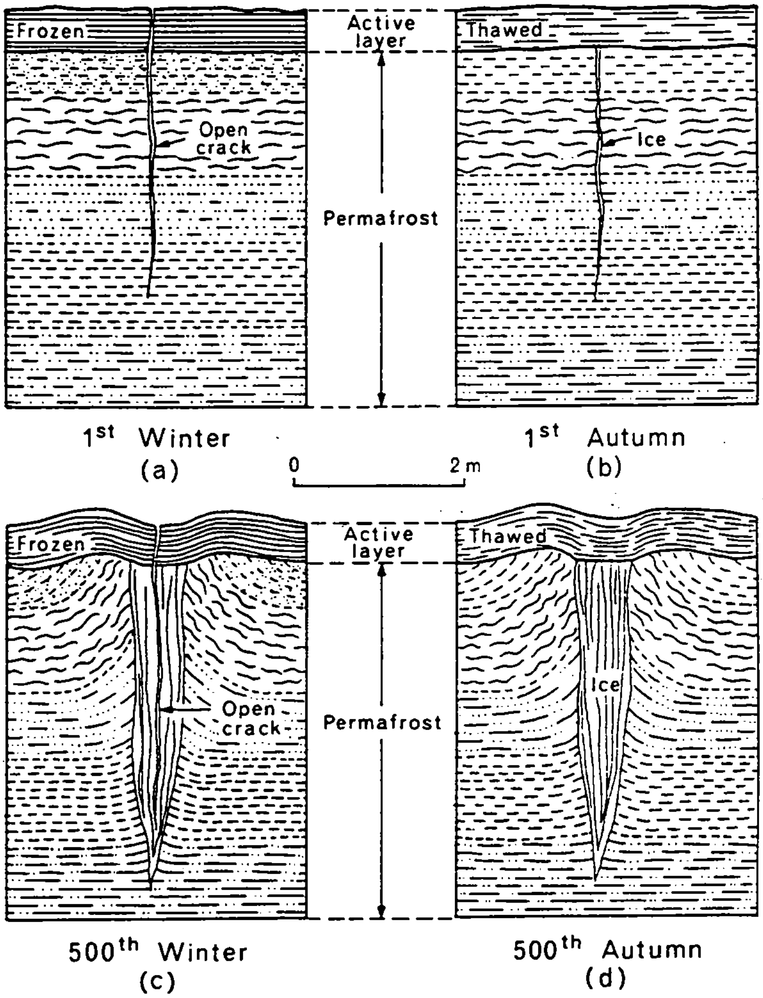


Figure 3.17 Schematic evolution of ice-wedges in surface permafrost (from Selby, 1985).

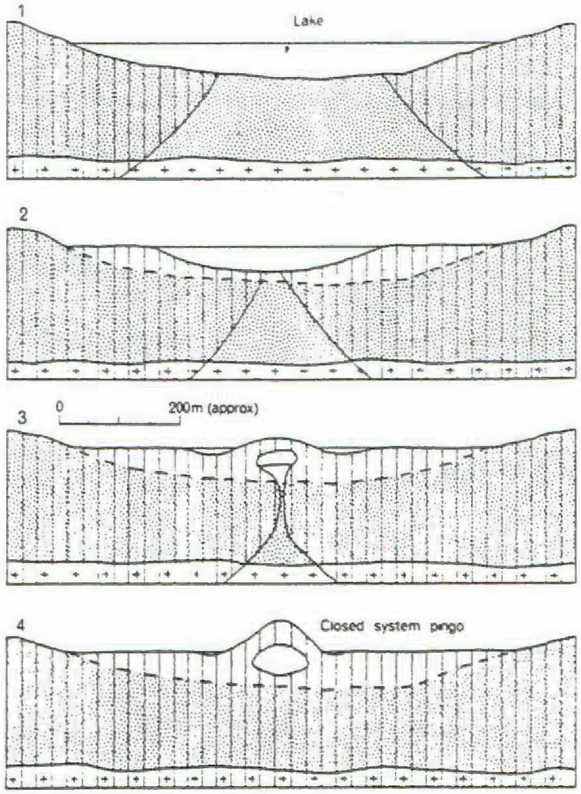


Figure 3.18 Model of pingo development in Mackenzie delta area (from Selby, 1985).

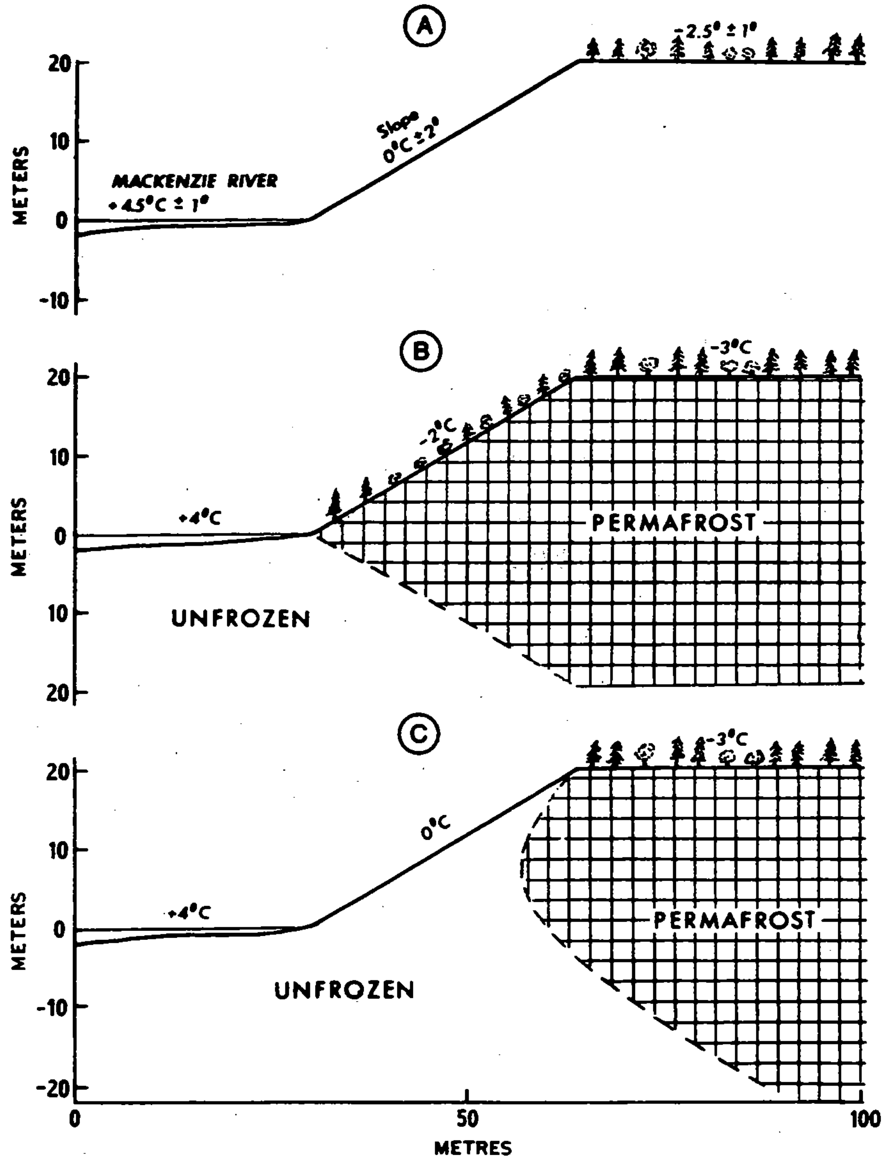


Figure 3.19 Schematic cross-sections through Mackenzie valley, NWT, Foot Good Hope region (from Mackay and Mathews, 1973).

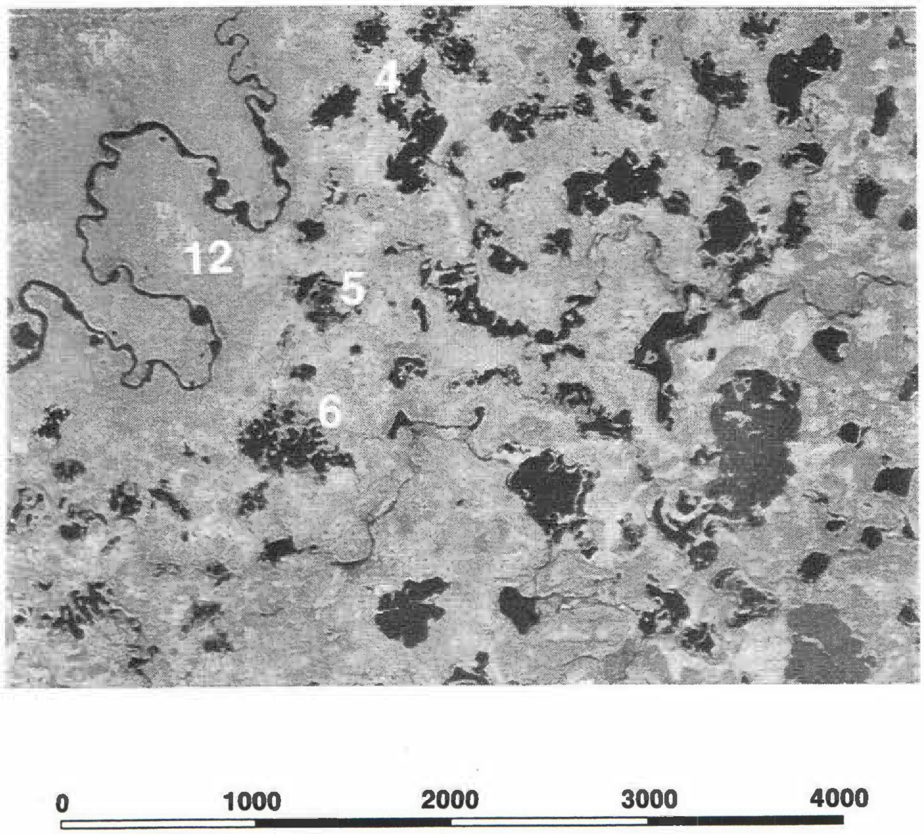


Figure 3.20 Thaw lakes and beaded drainage near Ramparts R., NWT (from Mollard and Janes, 1984).

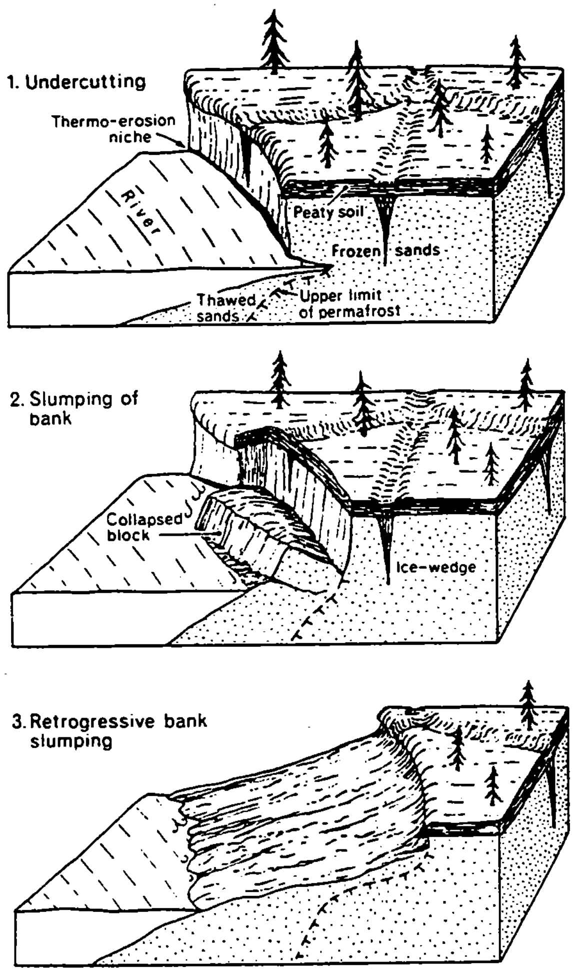


Figure 3.21 Idealized sequence of bank collapse through thermal erosion (from Selby, 1985).

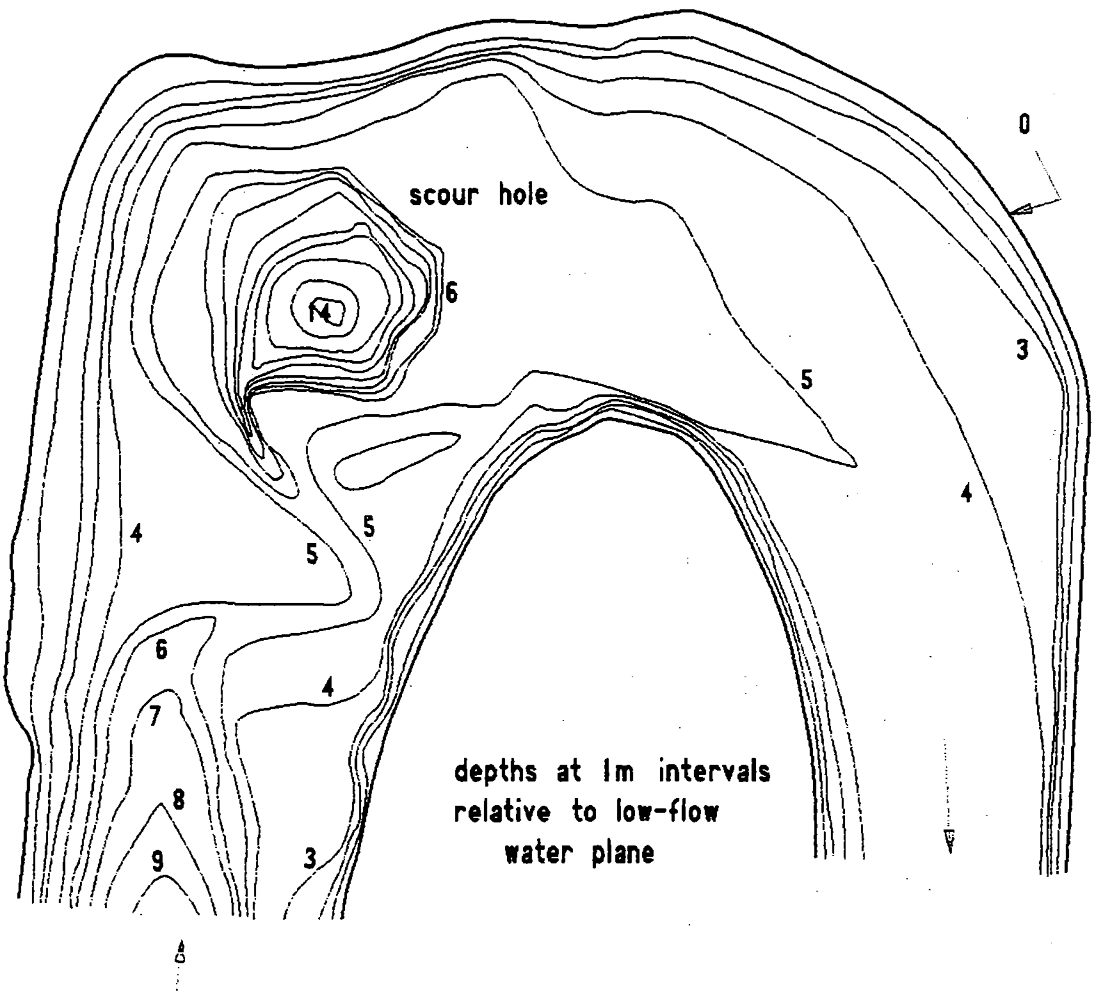


Figure 3.22 Bathymetry of meander bend and scour hole in small channel on Mackenzie Delta near Inuvik, NWT (after Lapointe, 1986).

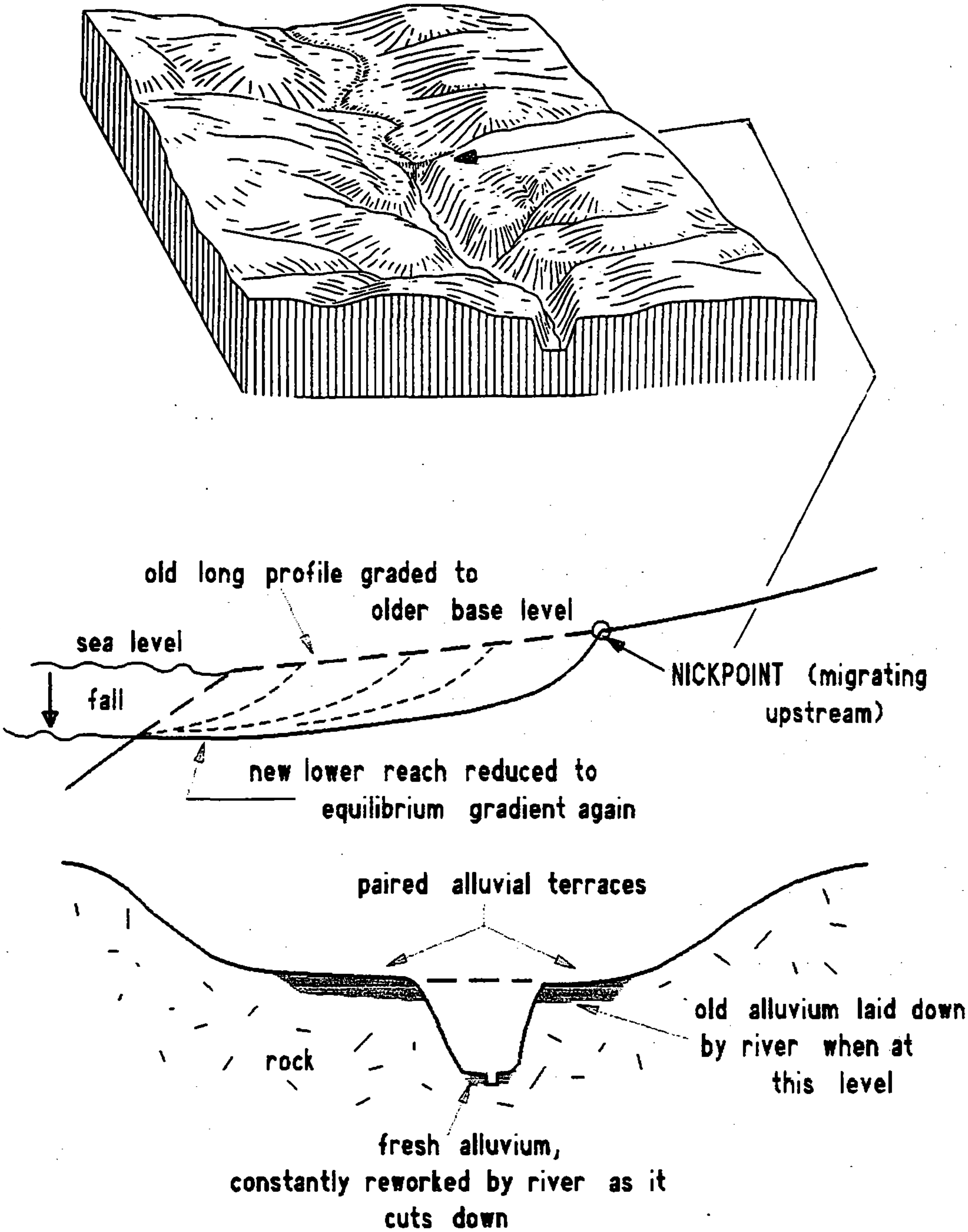


Figure 4.1 Schematic evolution of fluvial terraces formed by rejuvenation (block diagram from Thornbury, 1954).

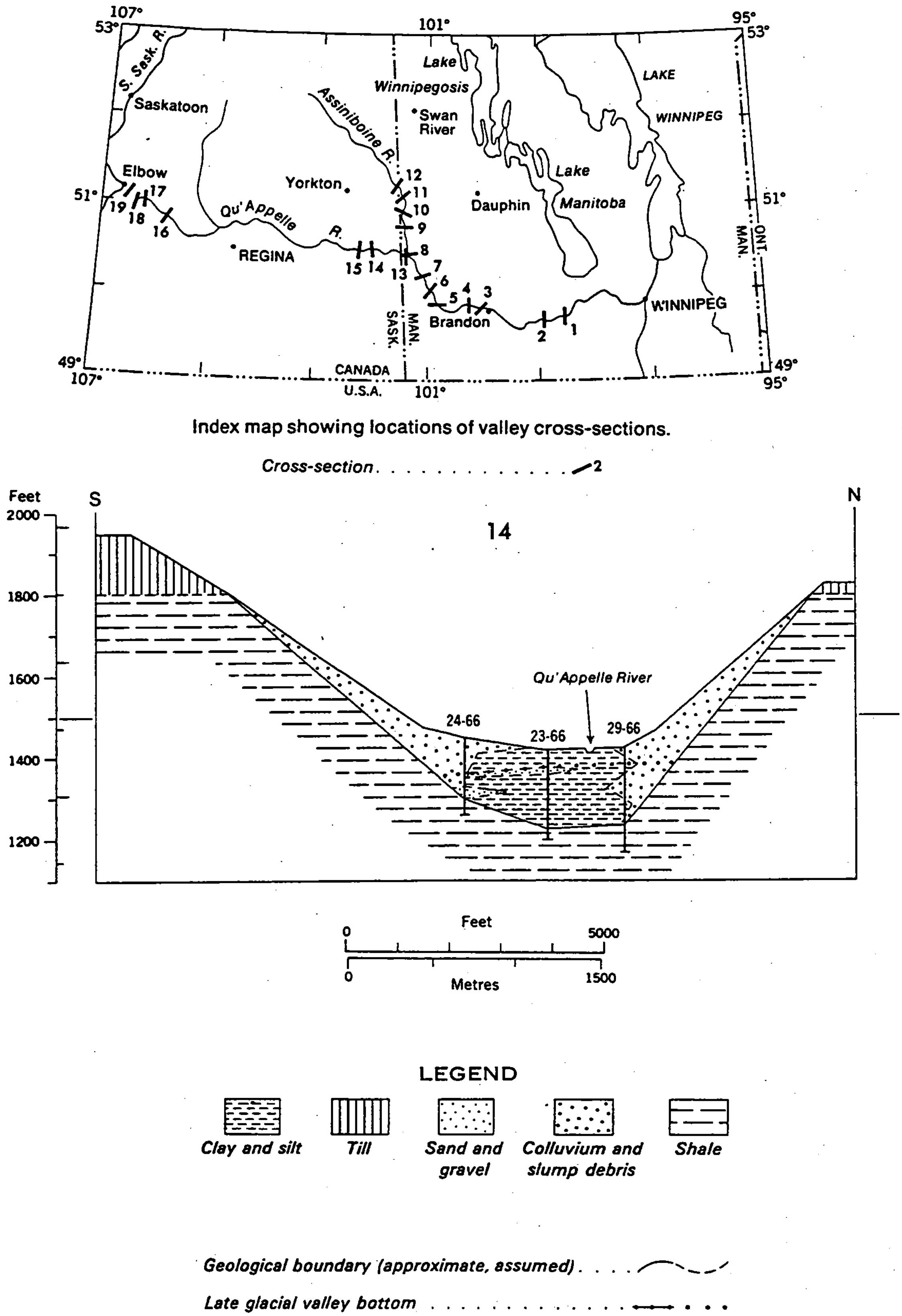


Figure 4.2 Holocene valley infill, Qu'Appelle River, Saskatchewan (from Klassen, 1975).

Figure 4.3 Photograph of stabilized aeolian sand dunes in vicinity of south shore of lake Athabasca, Saskatchewan (Saskatchewan Dept. of Tourism & Renewable Resources photo YC-7623-108).

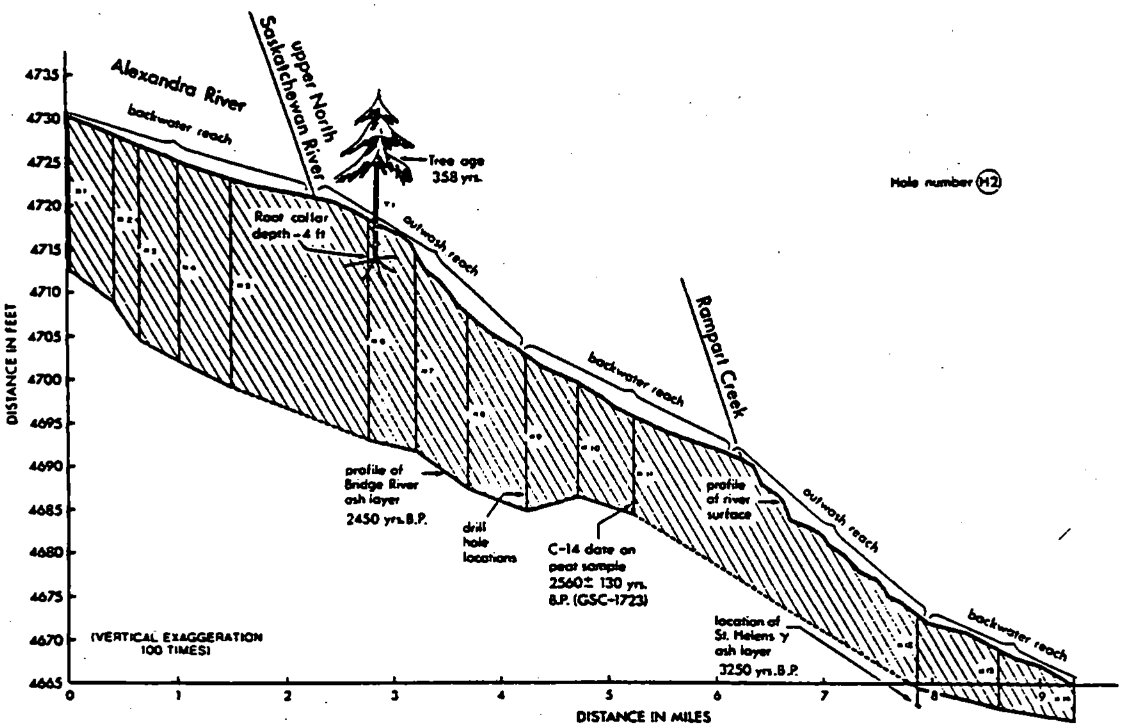


Figure 4.4 Alexandra and N. Saskatchewan R. confluence area, Alberta: long profile and upper alluvium (from Smith, 1972).

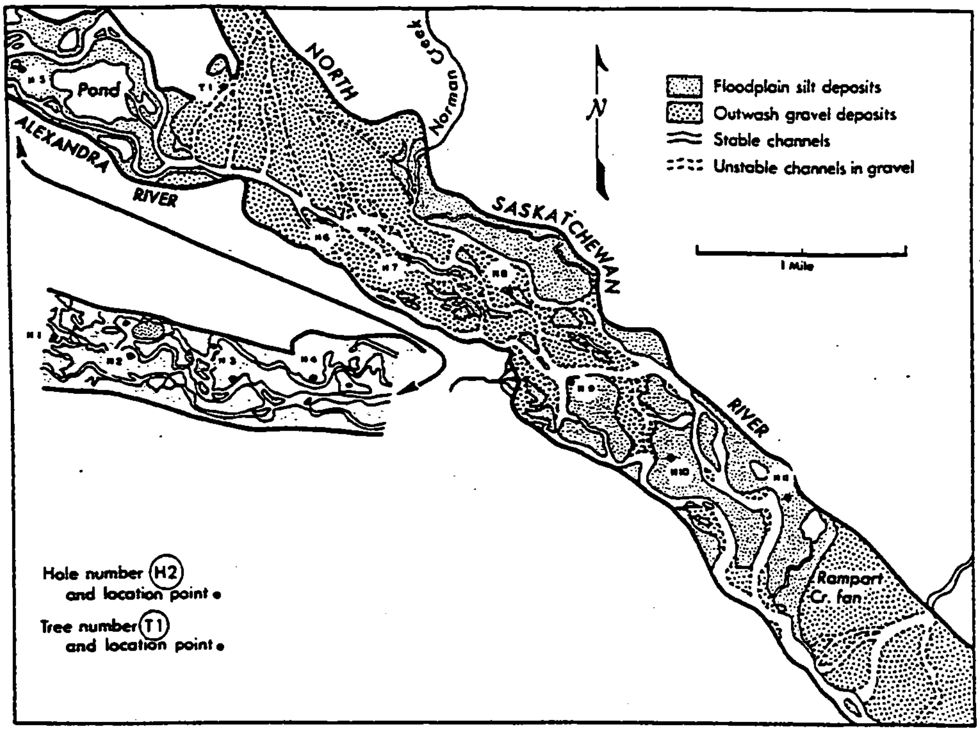


Figure 4.5 Alexandra and N. Saskatchewan R. confluence area, Alberta: plan morphology (from Smith, 1972).

Figure 4.6 Badland gullying in the southern prairies.

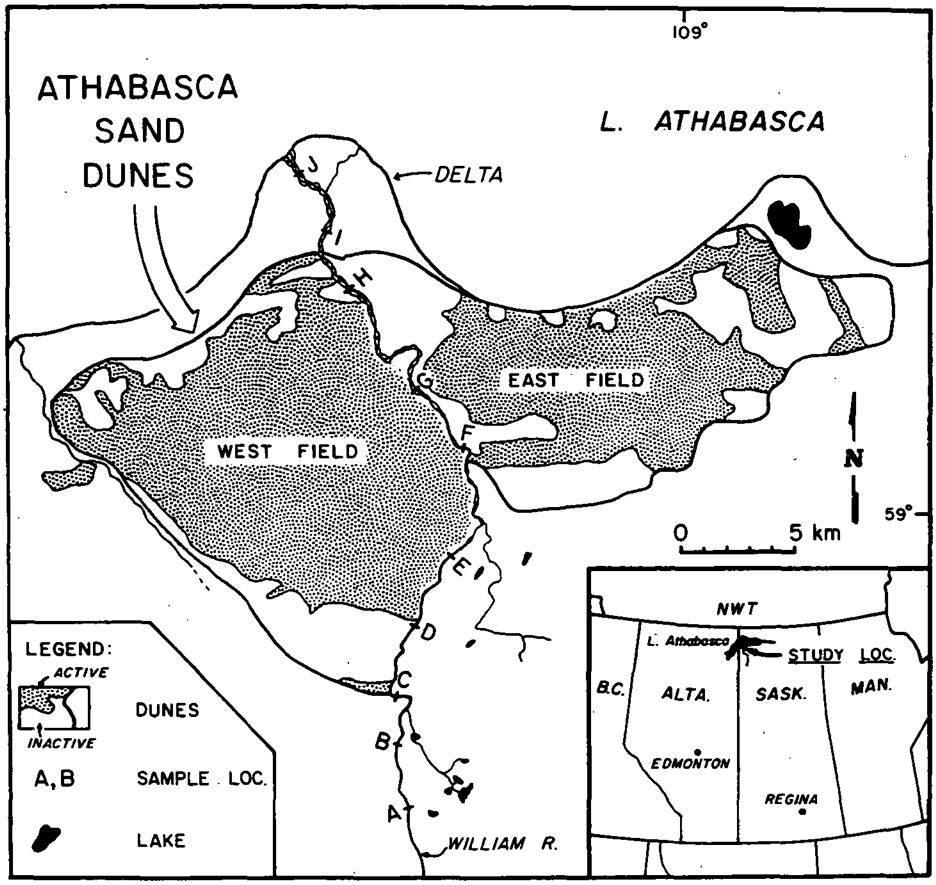


Figure 4.7 William River and Athabasca dunefield, Saskatchewan: location map (from Smith and smith, 1984).

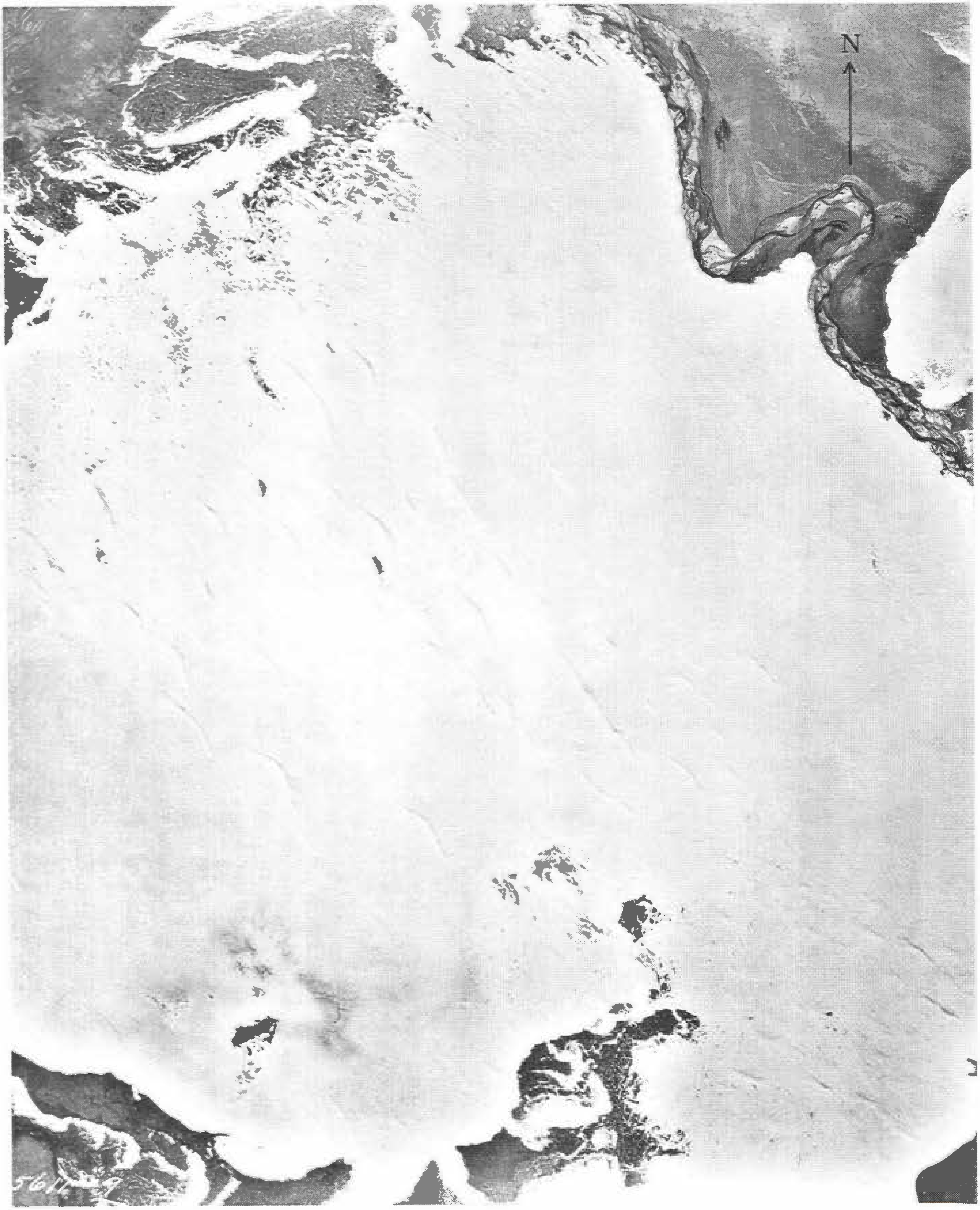


Figure 4.8 Aerial photograph of west William River dunefield. National Air Photo Library photo A-15611-9.

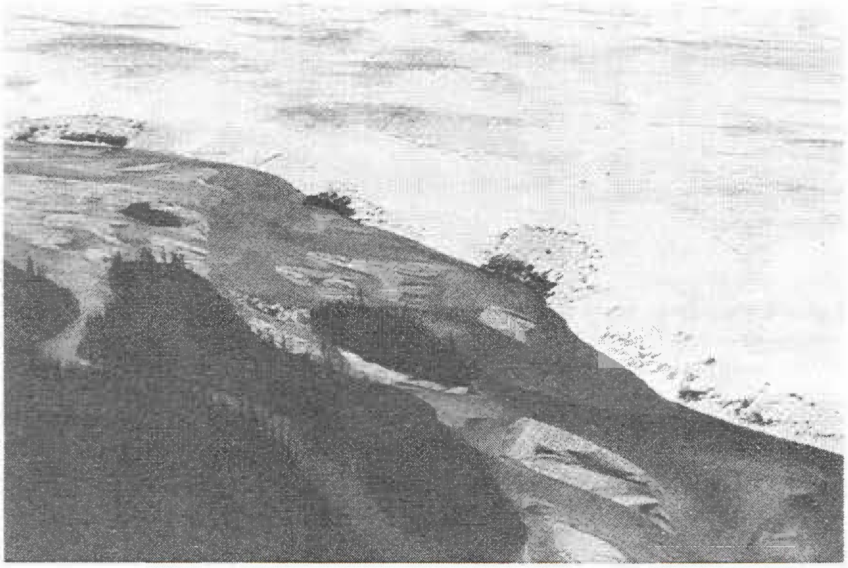


Figure 4.9 Oblique view across braided William River to west dunefield (from Smith and Smith, 1984).

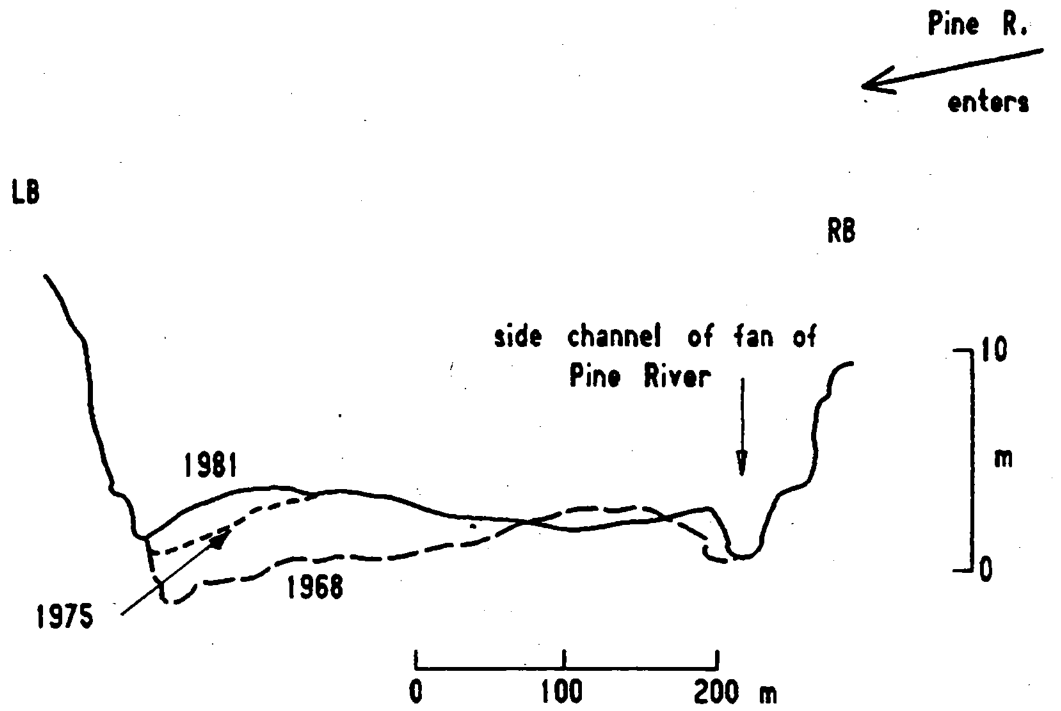


Figure 4.10 Aggradation on Peace River, upstream of Taylor Bridge, B.C. (after Church, 1983a).

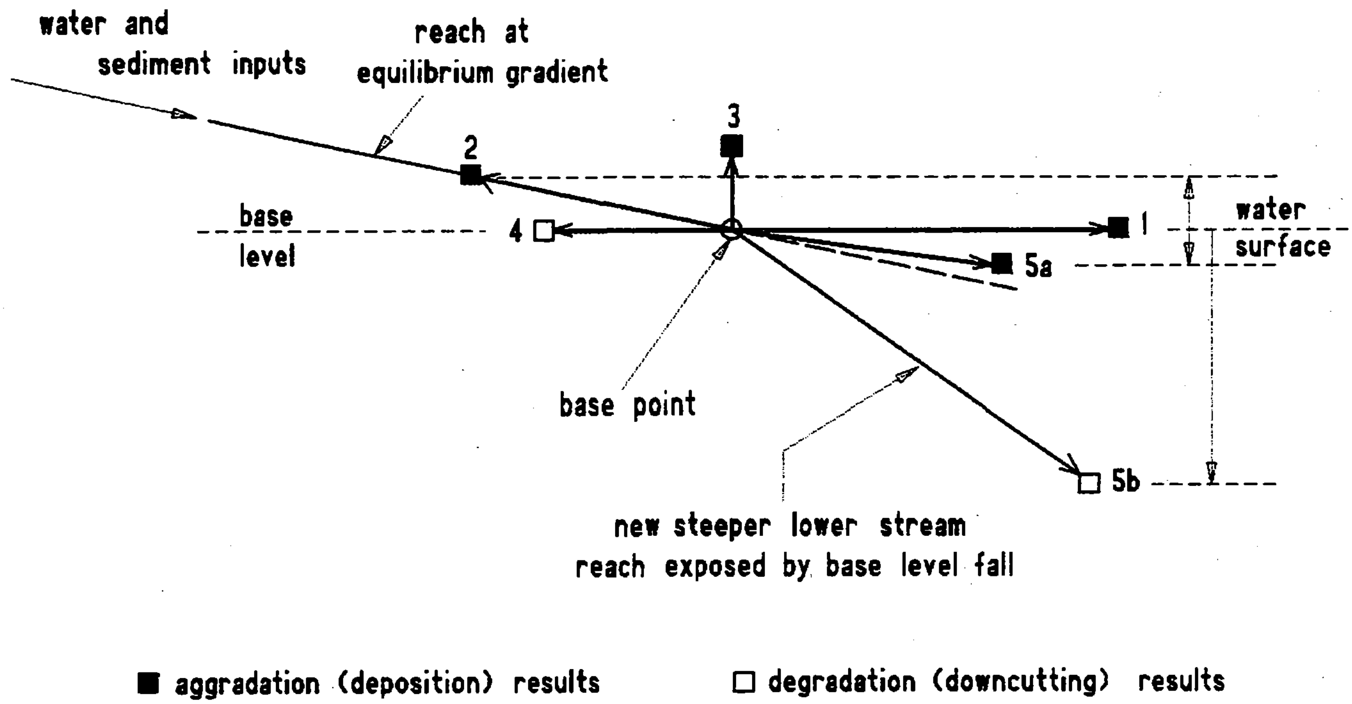


Figure 4.11 Changes in downstream (base point) control on a river reach (see discussion of vairous cases in text).

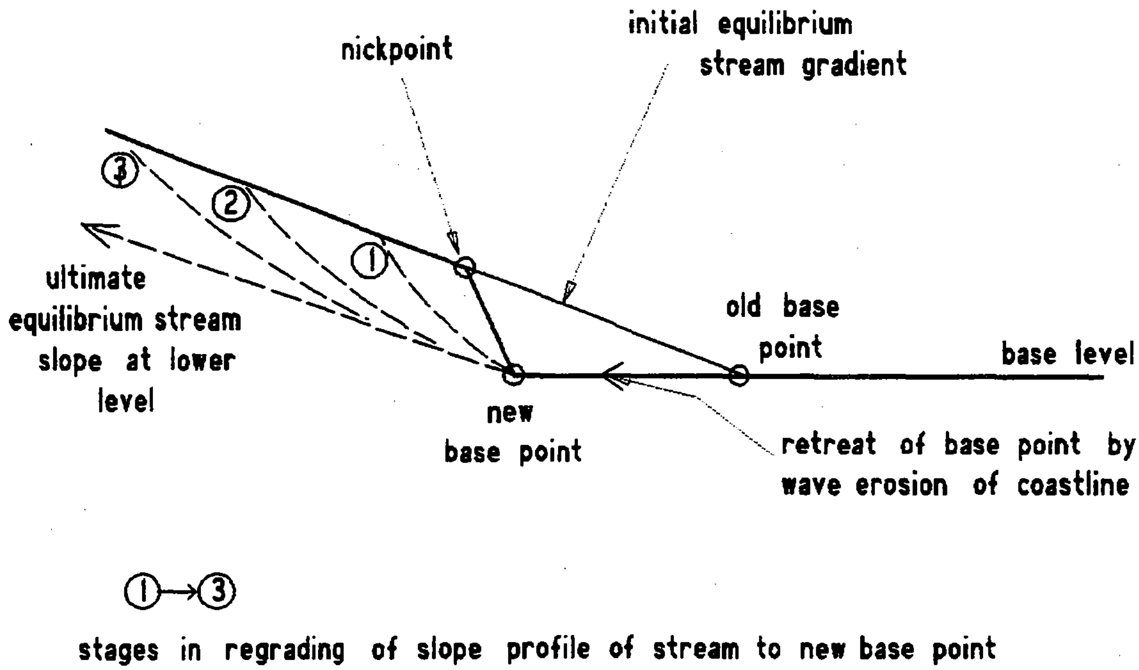


Figure 4.13 Initiation of rejuvenation by base point retrogradation.

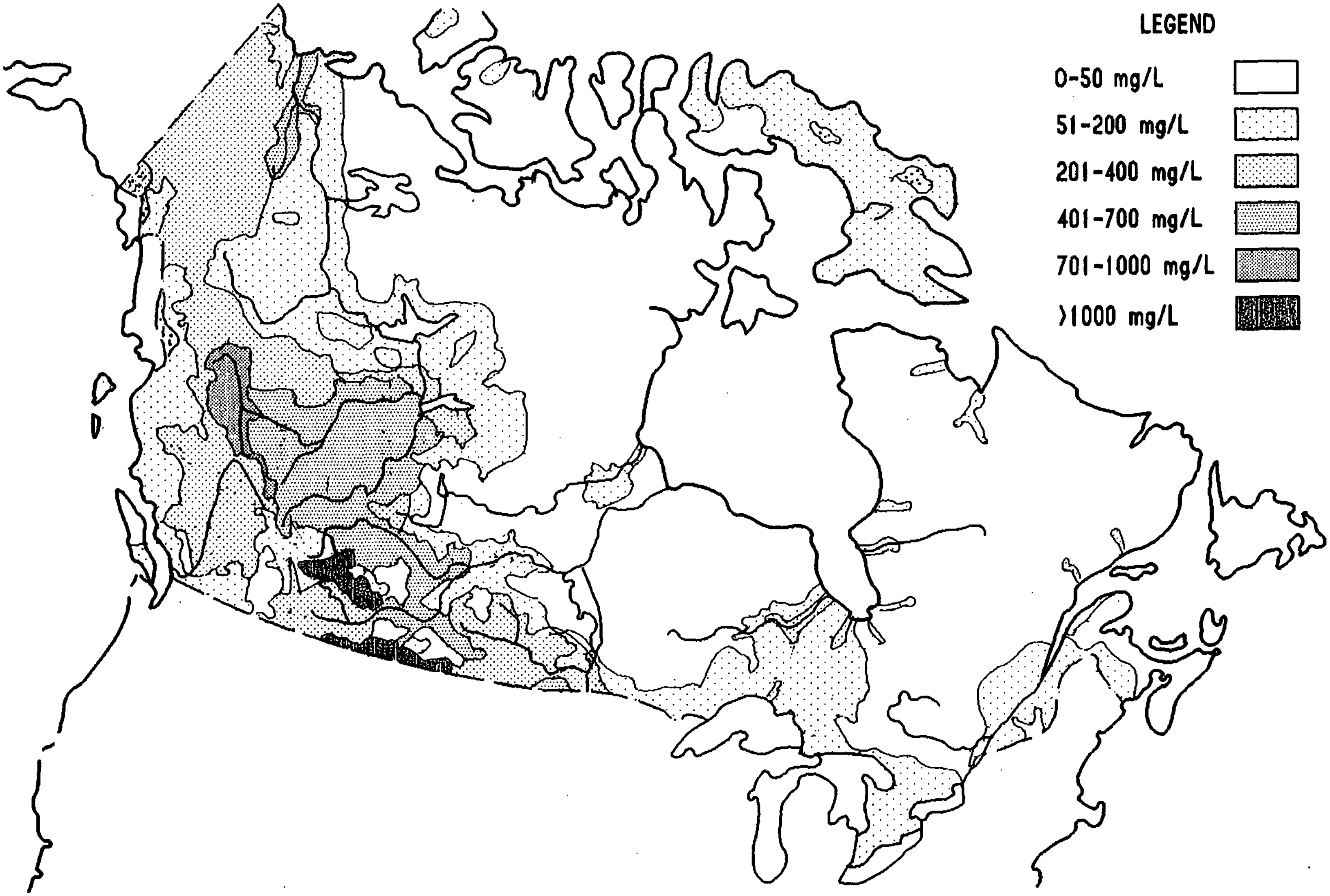


Figure 5.1 Mean load/flow ratio in Canadian Rivers (after Stichling, 1973).

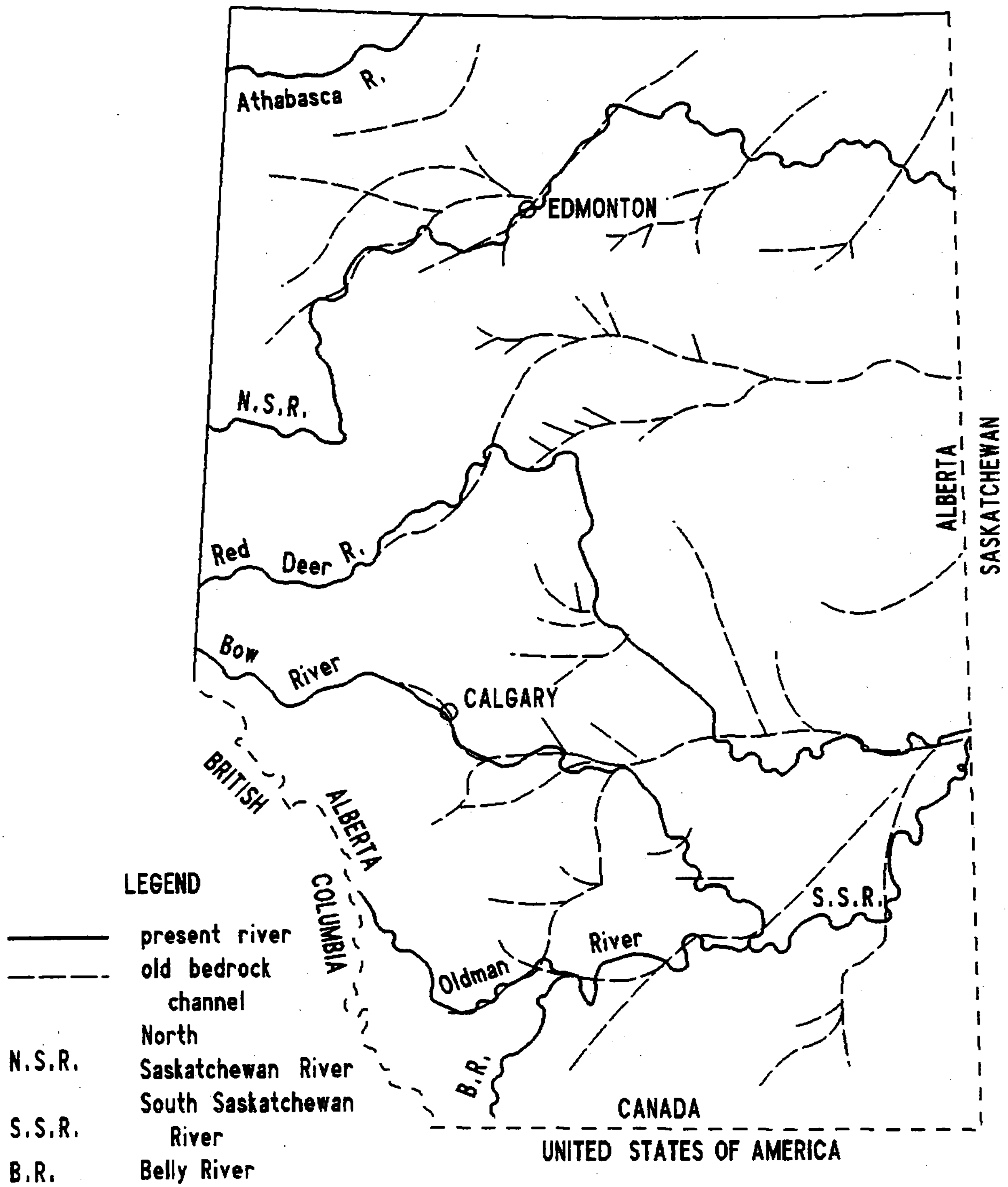


Figure 5.2 Discordance between present and pre-glacial valleys in southern Alberta (after Farvolden, 1963).

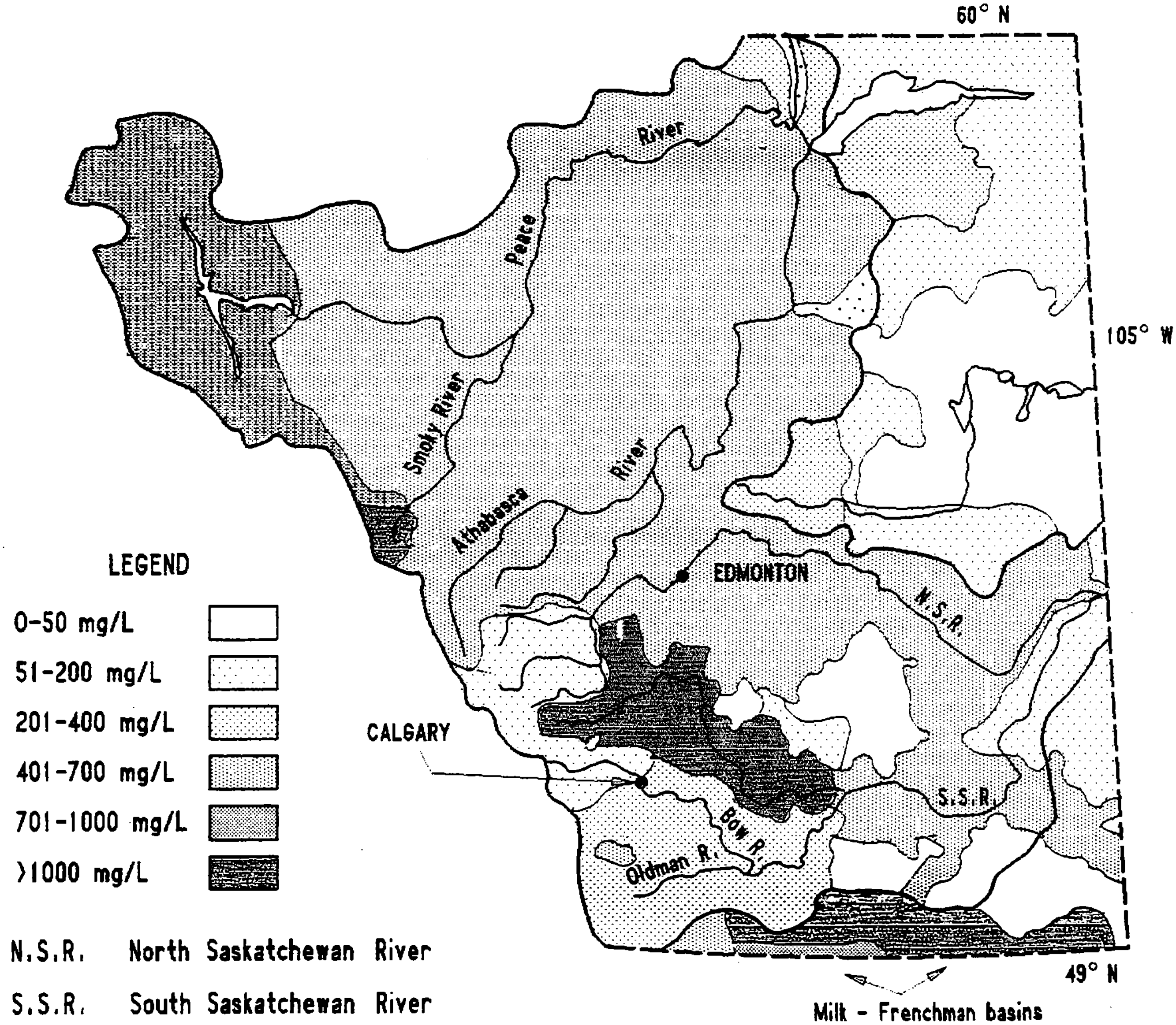


Figure 5.3 Mean load/flow ratio for basins in western prairies (after Stichling, 1973).

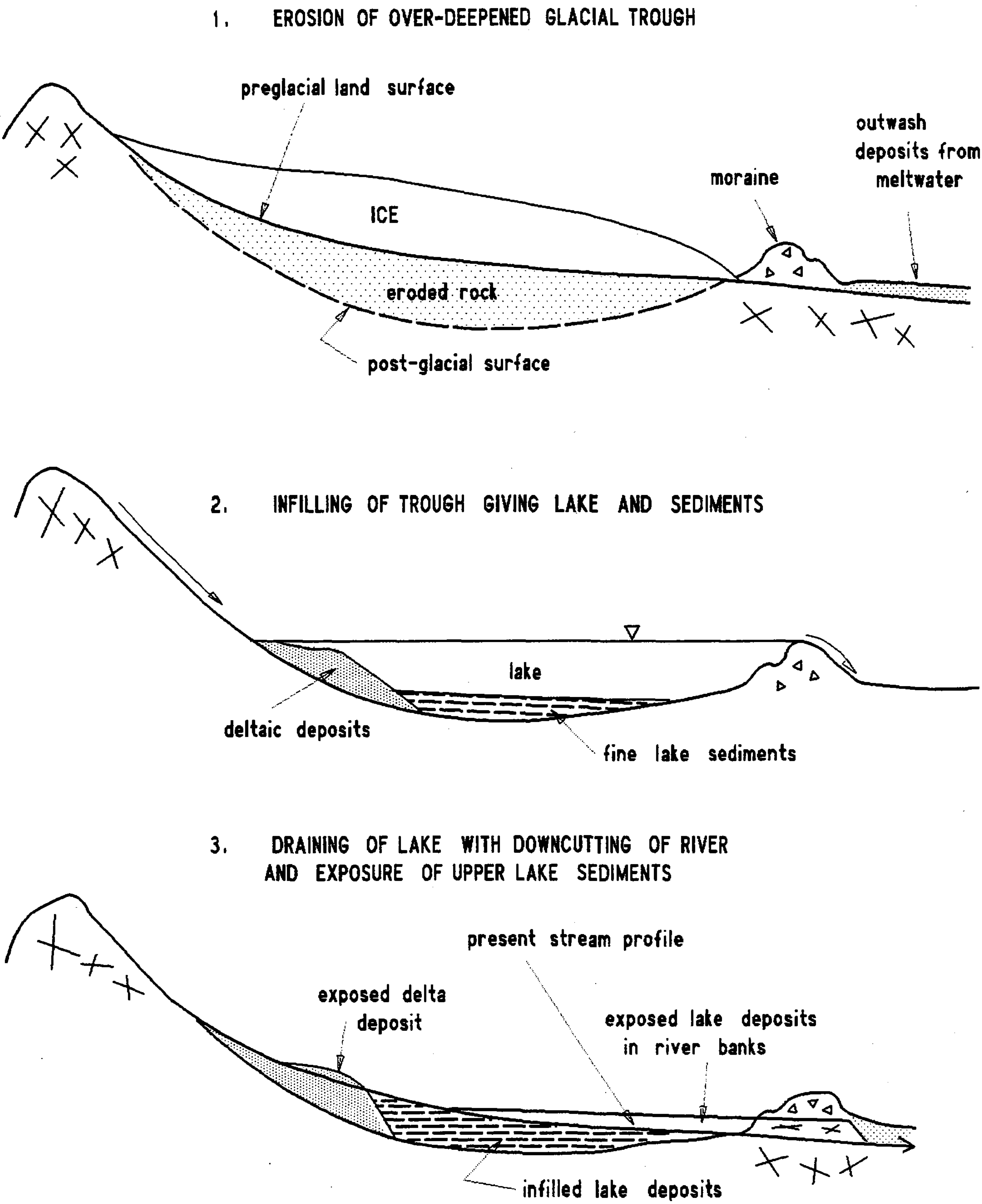


Figure 5.4 Impact of valley glacial erosion on post-glacial drainage.

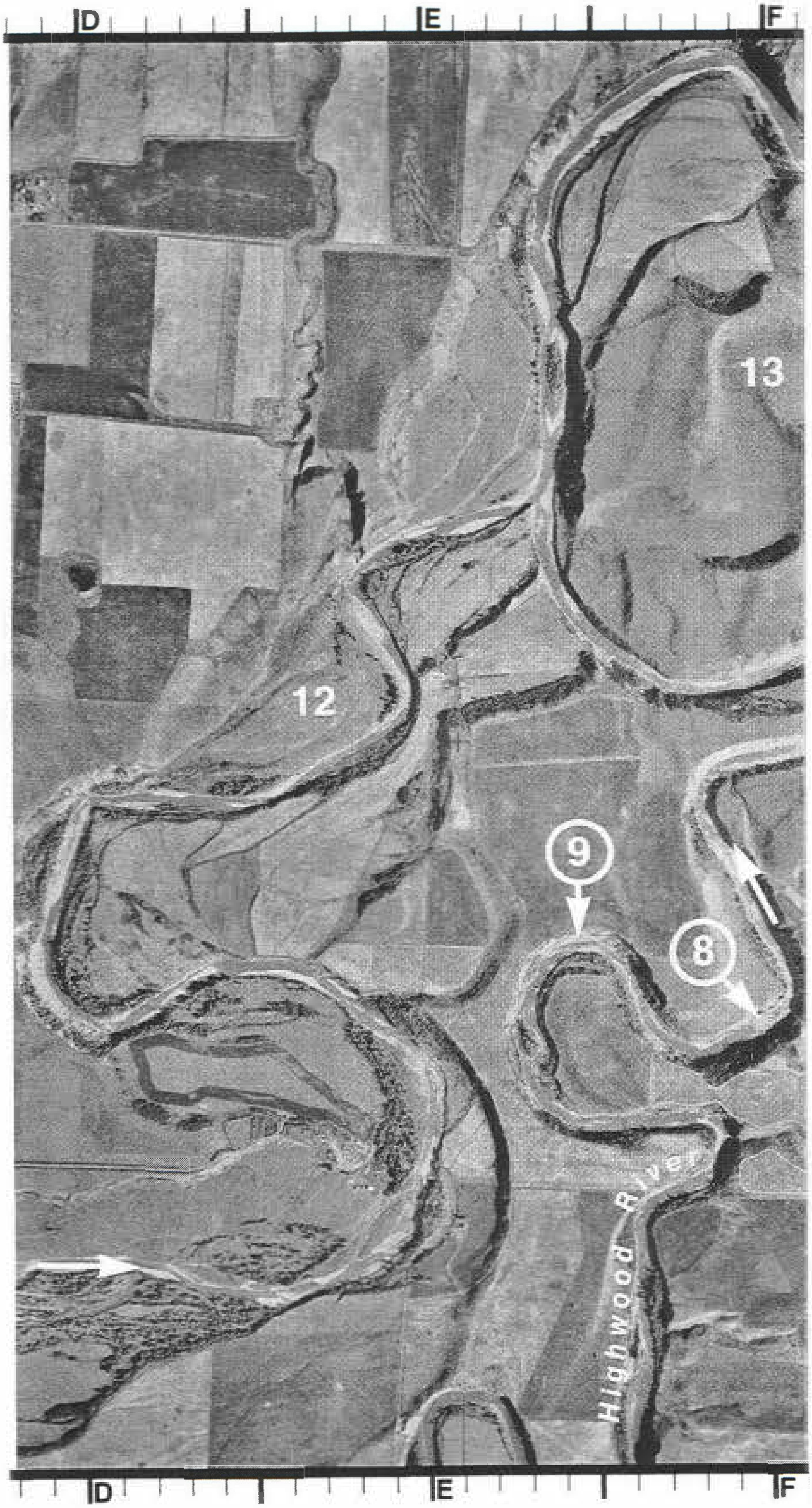


Figure 5.5 Incision of Highwood River, at Sheep R. confluence, Alberta (from Mollard and Janes, 1984).

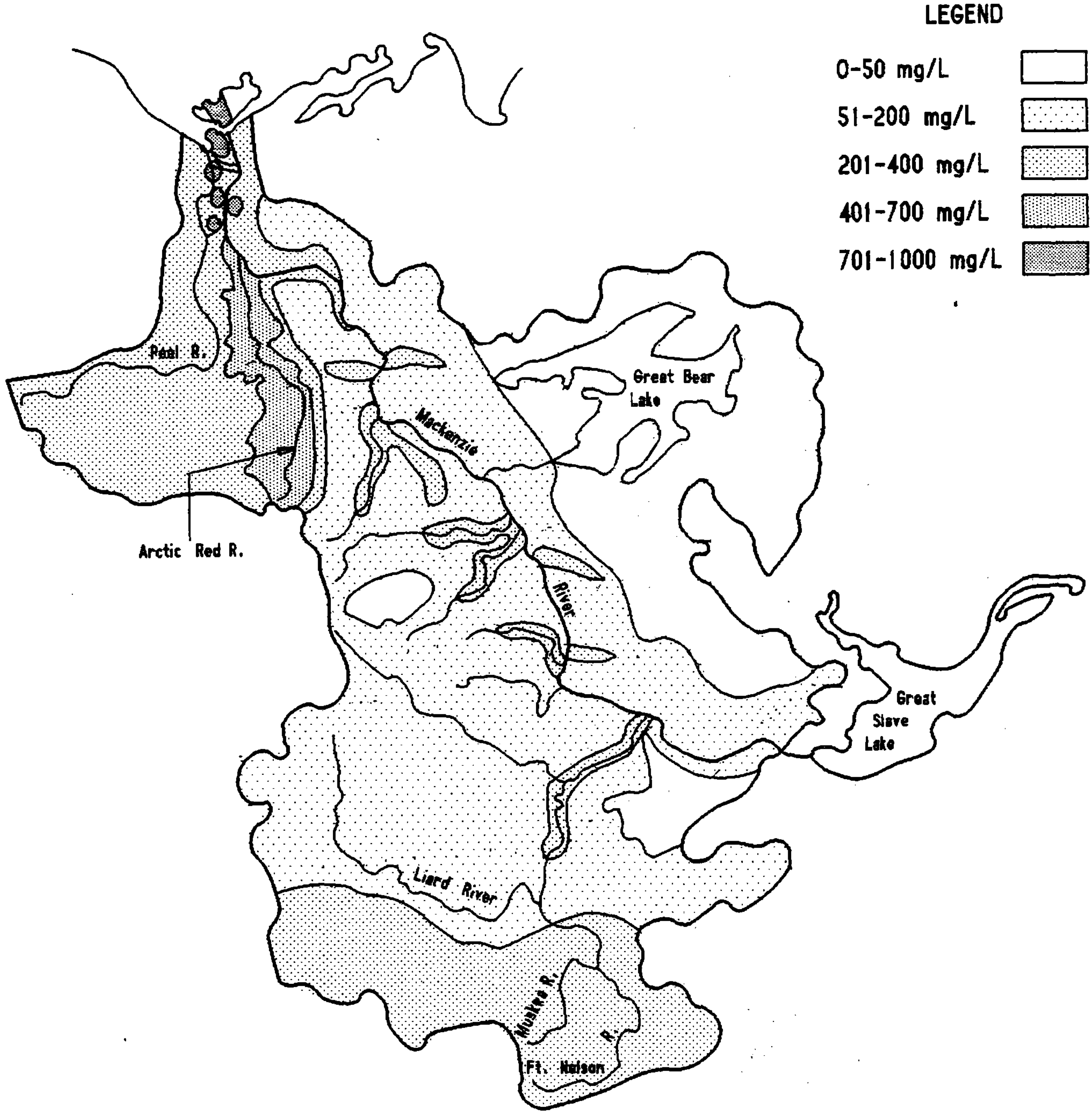


Figure 5.6 Mean load/flow ratio in Mackenzie basin (after Stichling, 1973).

	UM	L	M@AR	LM	TRIB	ARR	PEEL
May	27	810	620	990	1586	2200	1600
June	20	850	720	1070	1239	1600	1270
July	20	840	480	820	662	820	530
August	17	320	320	630	775	1600	570
Sept.	19	160	150	350	395	160	260
Oct.	18	65	100	280	338	40	160
Annual		570	360			1350	950

UM: Upper Mackenzie (above Liard) based on 1972-75 data
L: Liard River near mouth based on 1974-86 sediment rating
M@AR: Mackenzie R. at Arctic Red, 1974-83 sediment rating loads
LM: Liard-Mackenzie, based on M@AR loads but ignoring runoff from upstream of Liard confluence
TRIB: estimated tributary concentrations based on (load at M@AR minus load at Ft. Simpson) divided by (flow at M@AR minus flow at Fort Simpson)
AR: Arctic Red River near mouth, based on 1974-86 sediment rating loads (August load dominated by 1974 flood)
PEEL: Peel River upstream of Fort MacPherson, based on sediment rating loads, 1974-83

Figure 5.7 **Mean load/flow ratios in Mackenzie River system, NWT.**

All data are preliminary

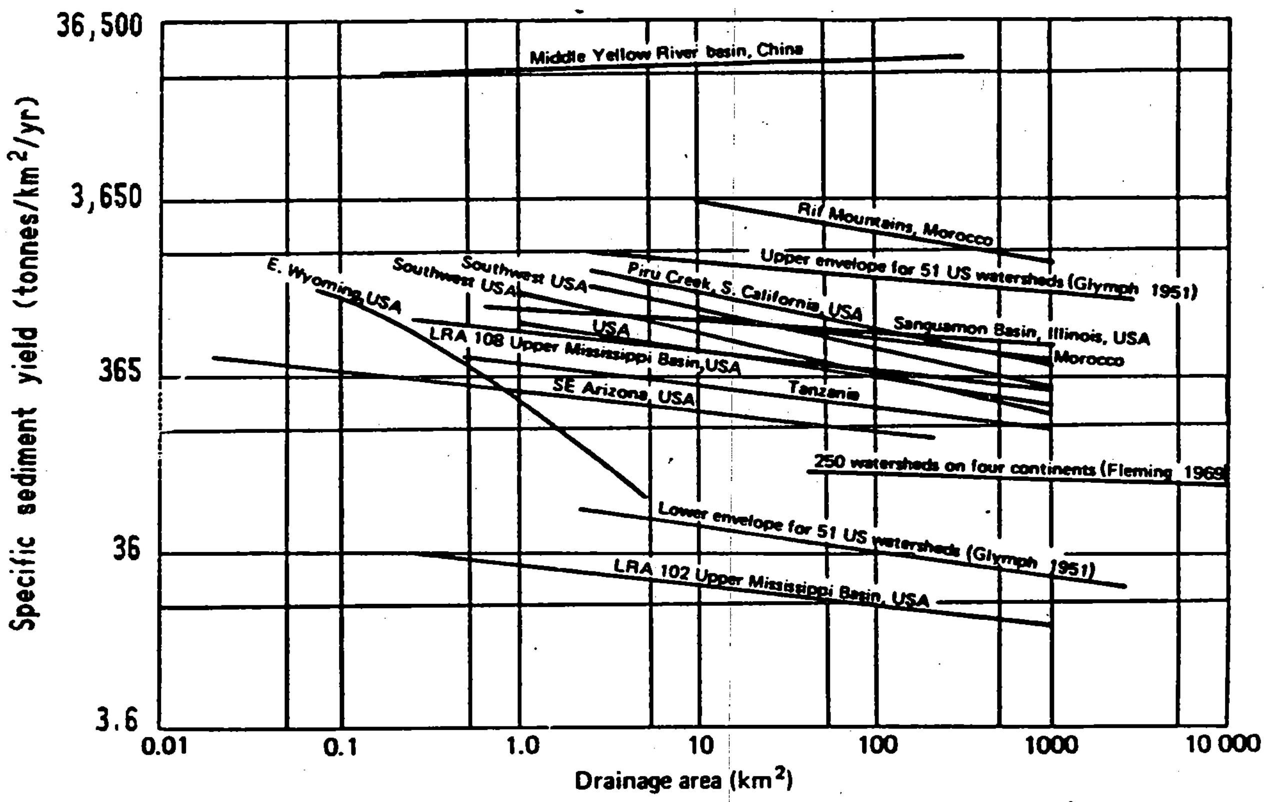


Figure 5.8 Specific sediment yield as a function of basin size (from Church et al., 1989).

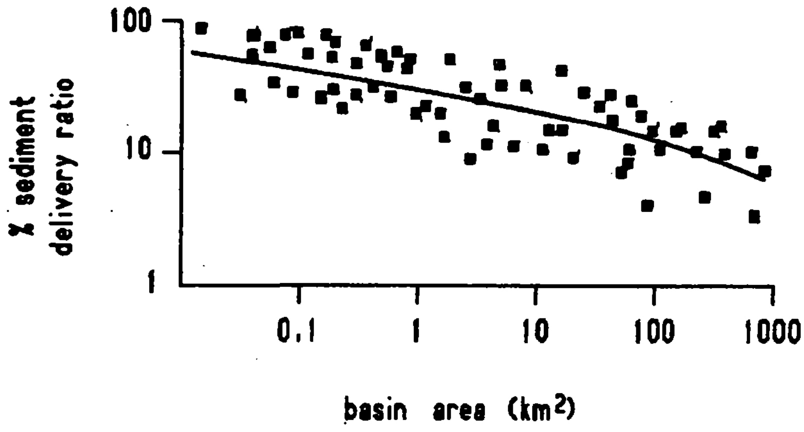


Figure 5.9a Sediment delivery ratio as a function of basin size, south and central United States (after Roehl, 1962).

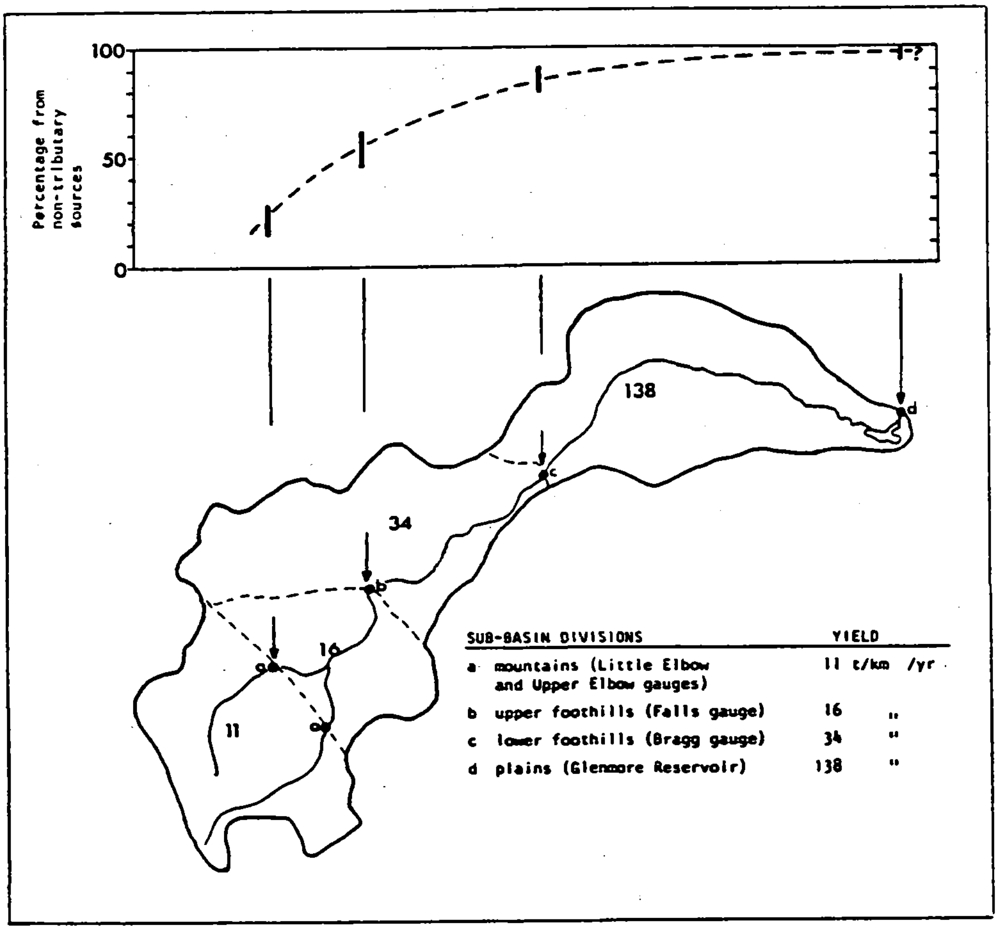


Figure 5.9b Sediment yields and sediment sources in the 1210 km² Elbow River basin, southwest Alberta (Hudson, 1983).

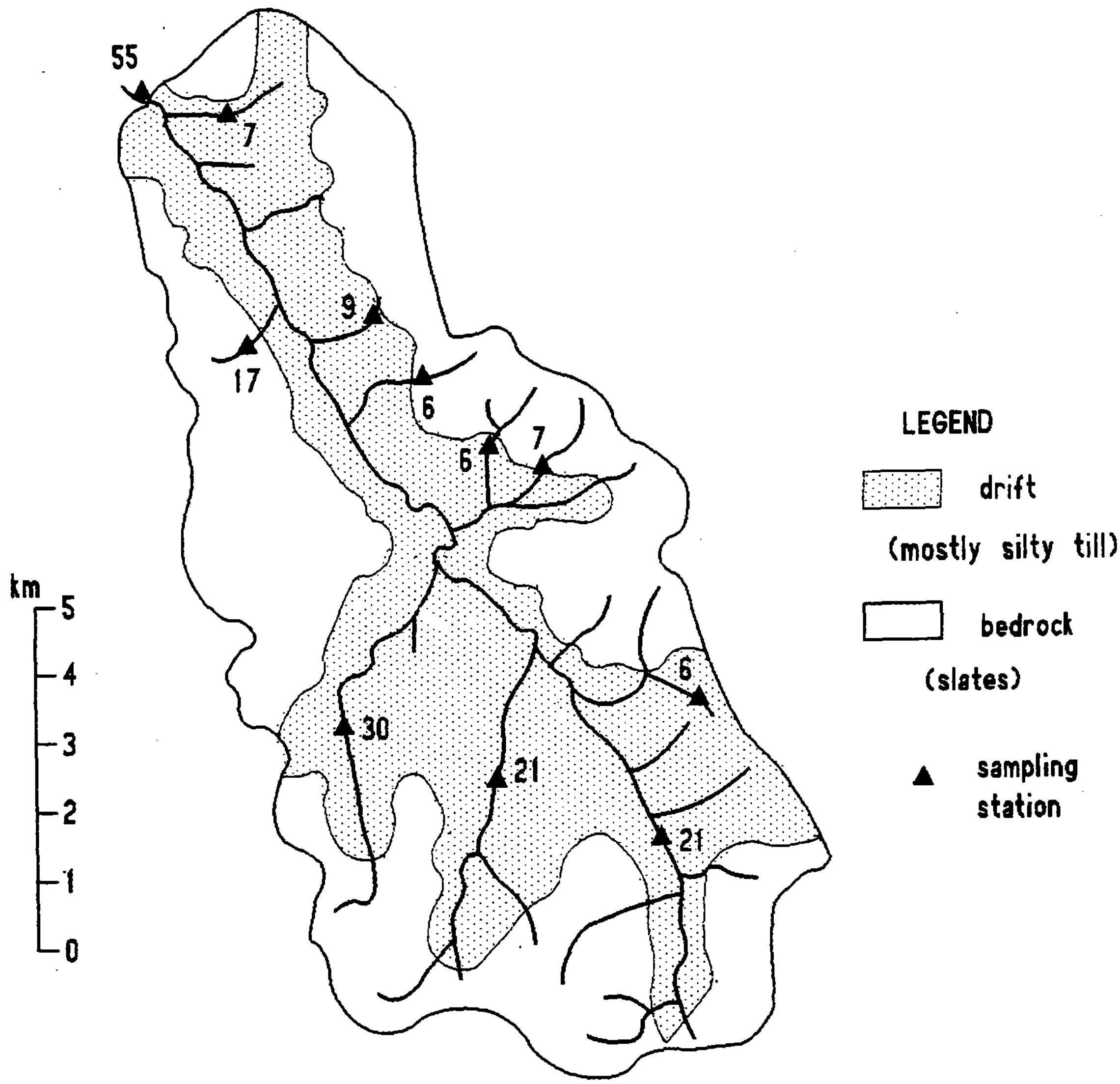


Figure 5.10 Suspended sediment yields (t/km^2) in Eaton River Basin (Quebec) during spring runoff, 1971 (from Carson et al., 1973).

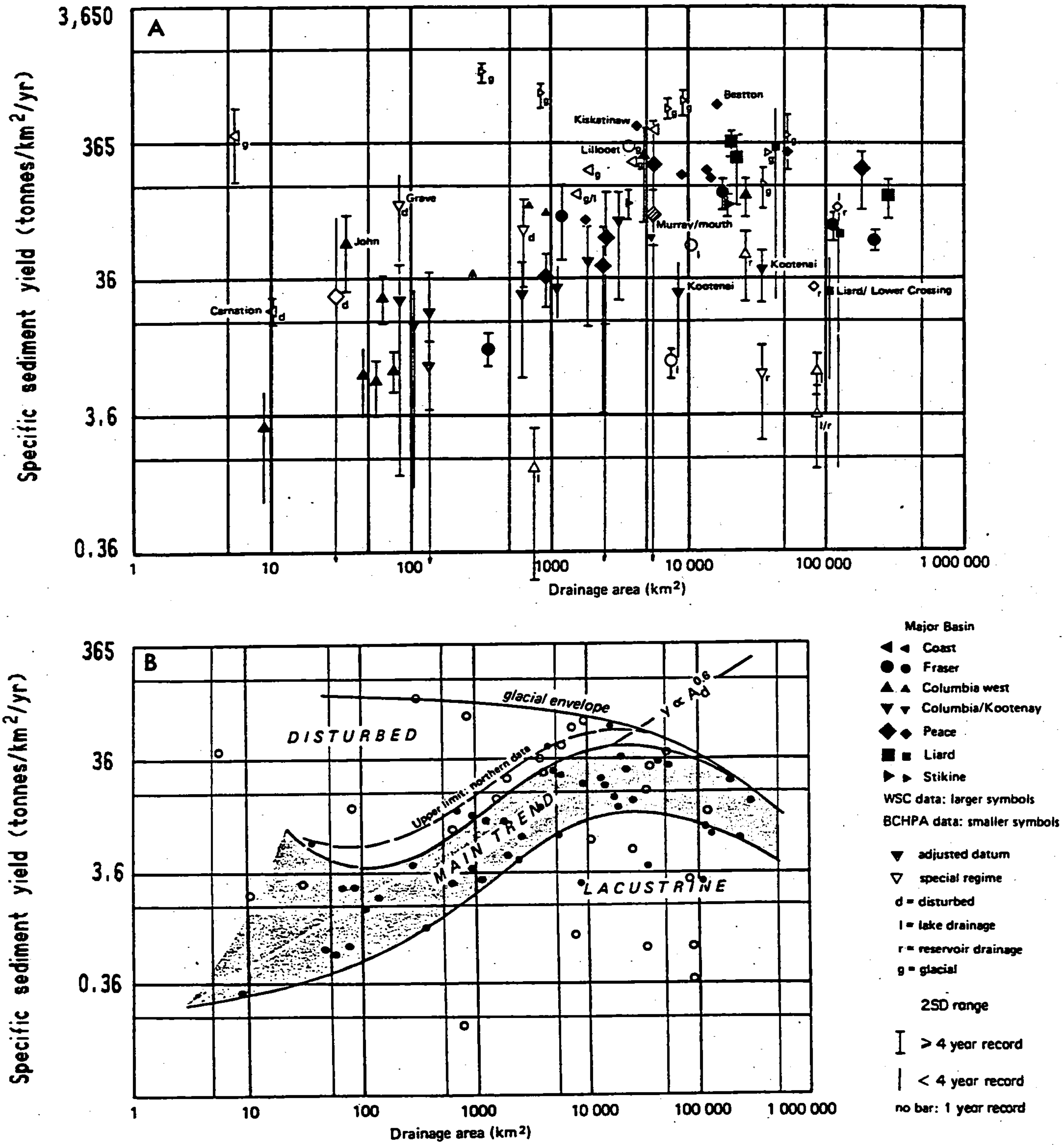


Figure 5.11 Specific sediment yield as a function of basin size in British Columbia (from church et al., 1989).

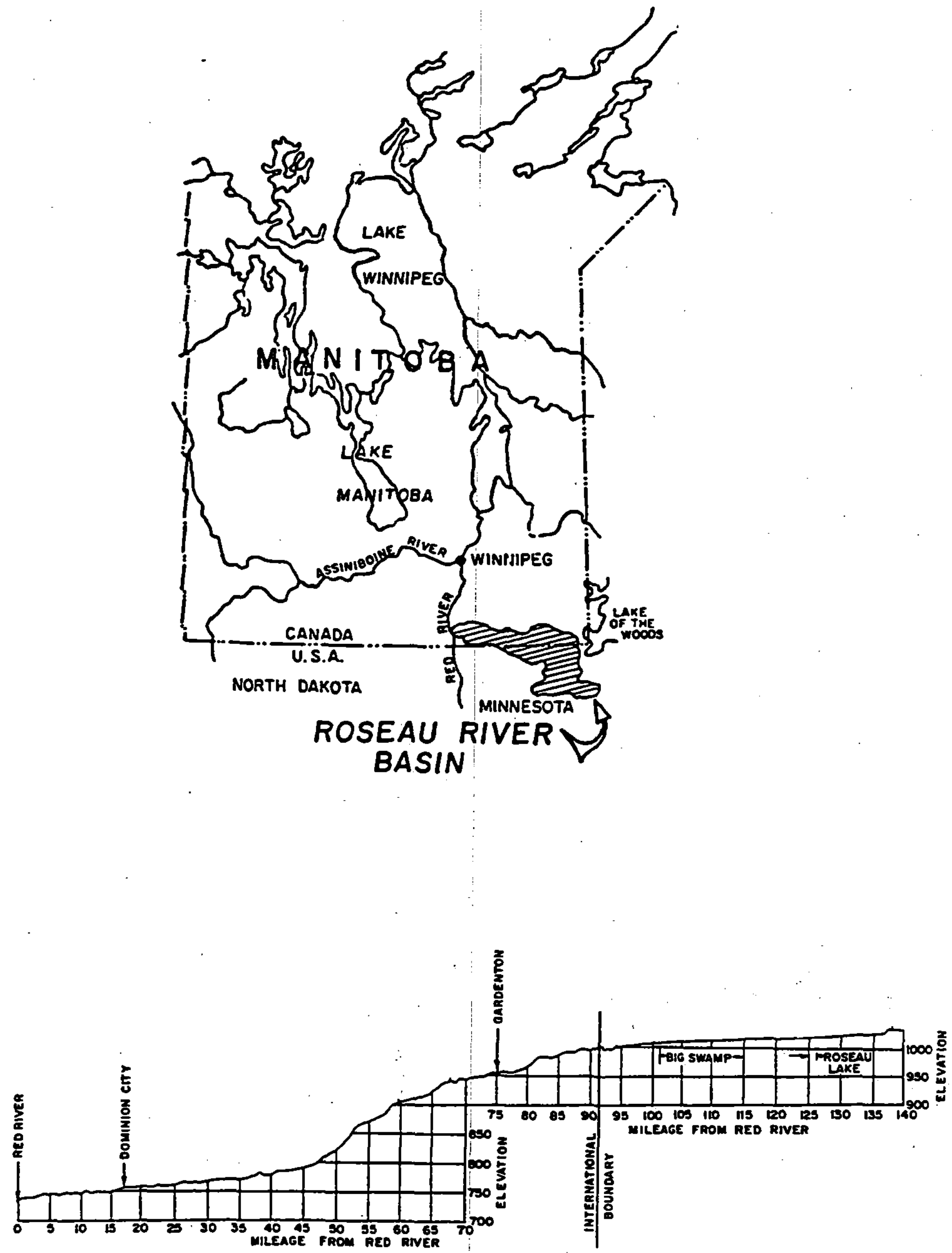


Figure 5.12 Map of the Roseau River basin, Manitoba-Minnesota.

Province	On-farm costs* in $ millions	Farmland area** in million ha	$ per ha
Ontario	68	6.0	11.33
Alberta	200	19.1	10.47
Atlantic	11	1.2	9.17
Saskatchewan	220	25.9	8.49
B.C.	10	2.2	4.55
Quebec	10	3.8	2.63
Manitoba	10	7.6	1.32

* from Rennie (1985)

** from Statistics Canada (1981)

Figure 5.13 Annual on-farm costs of soil erosion in Canada by province.

Figure 5.14 Incision of Smoky River into drift-covered marine shale near Guy, Alberta (from Mollard and Janes, 1984).

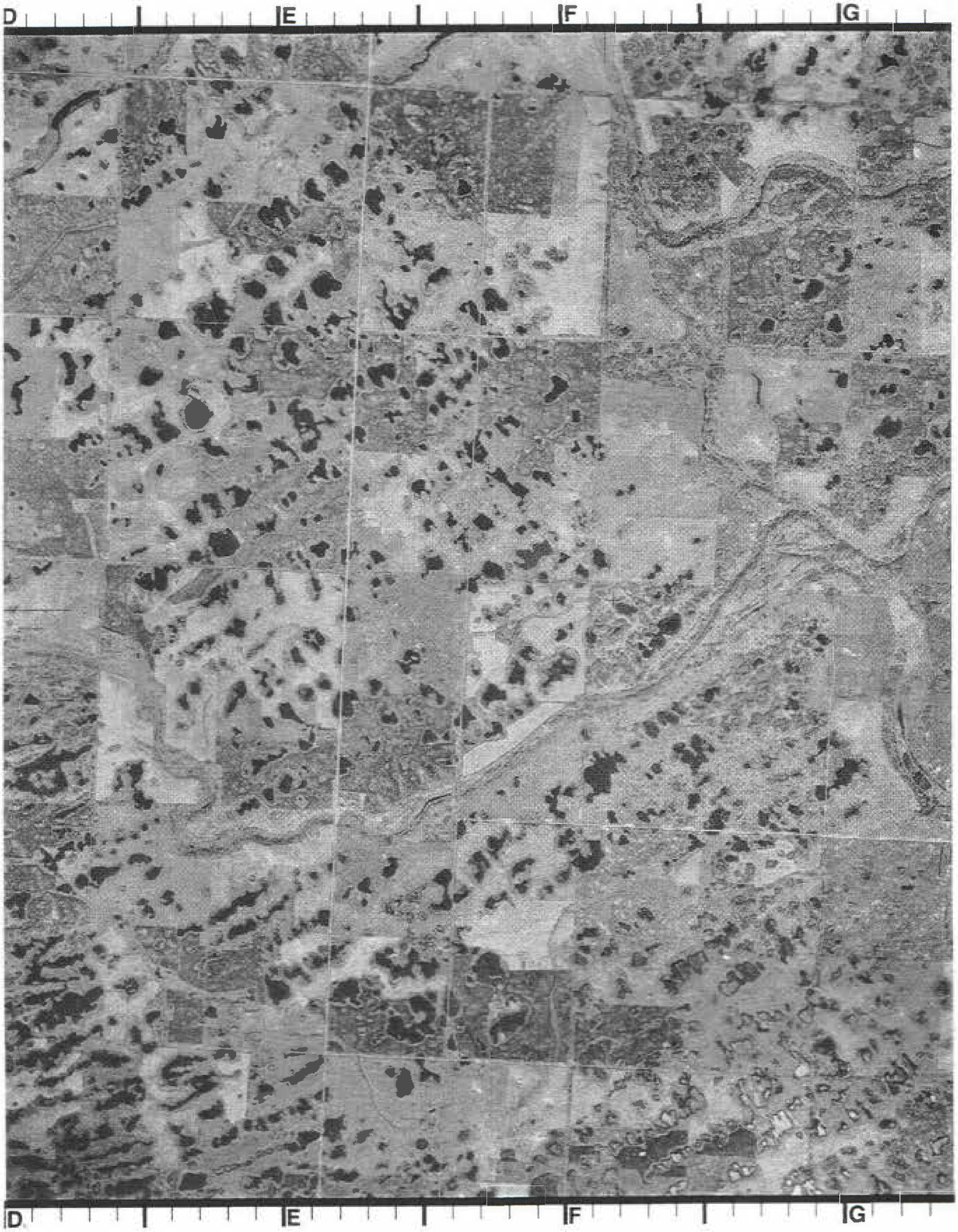


Figure 5.15 Disintegration ground moraine between Pipestem and Assiniboine rivers, Manitoba (from Mollard and Janes, 1984).

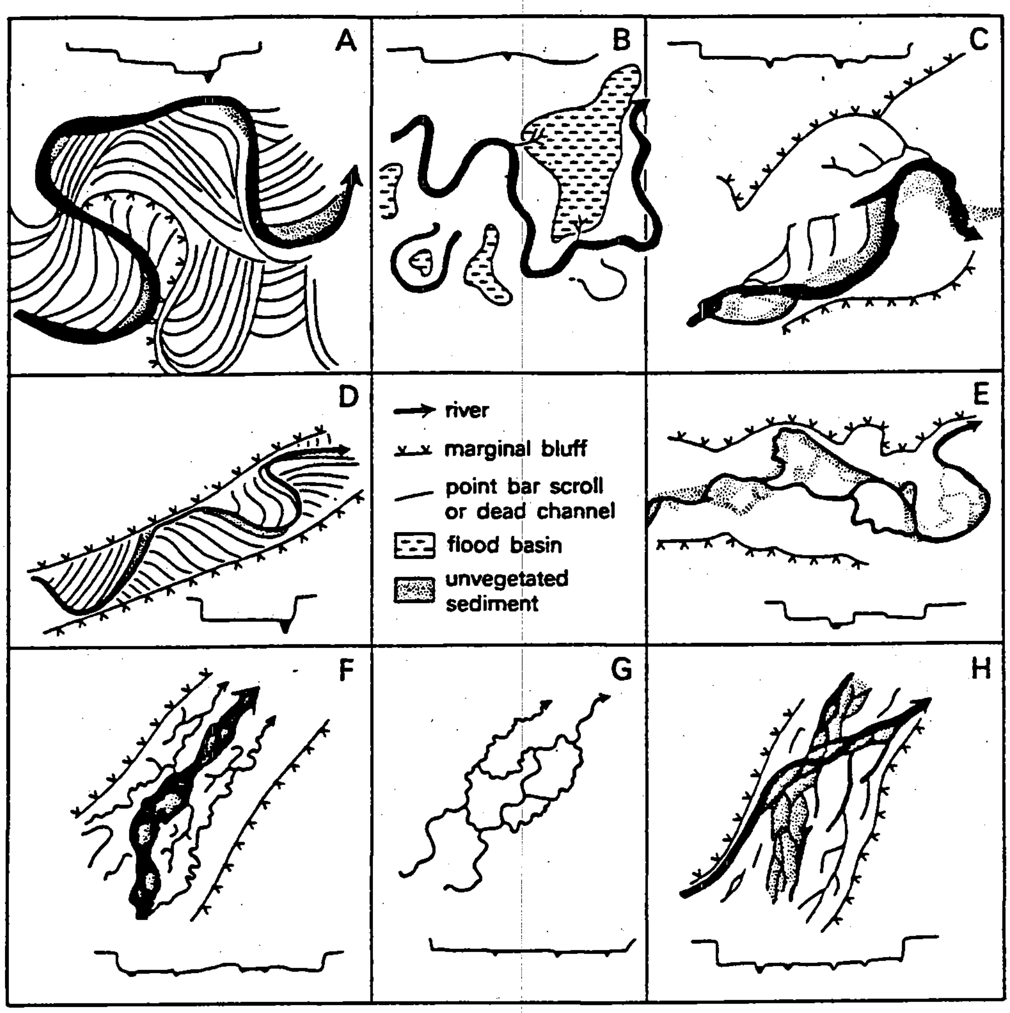


Figure 6.1 Morphology of common types of floodplains (from Lewin, 1978). See text.

Figure 6.2 Braided channel of the Muddy River, Alaska (photograph by bradford Washburn, from Leopold et al., 1964).

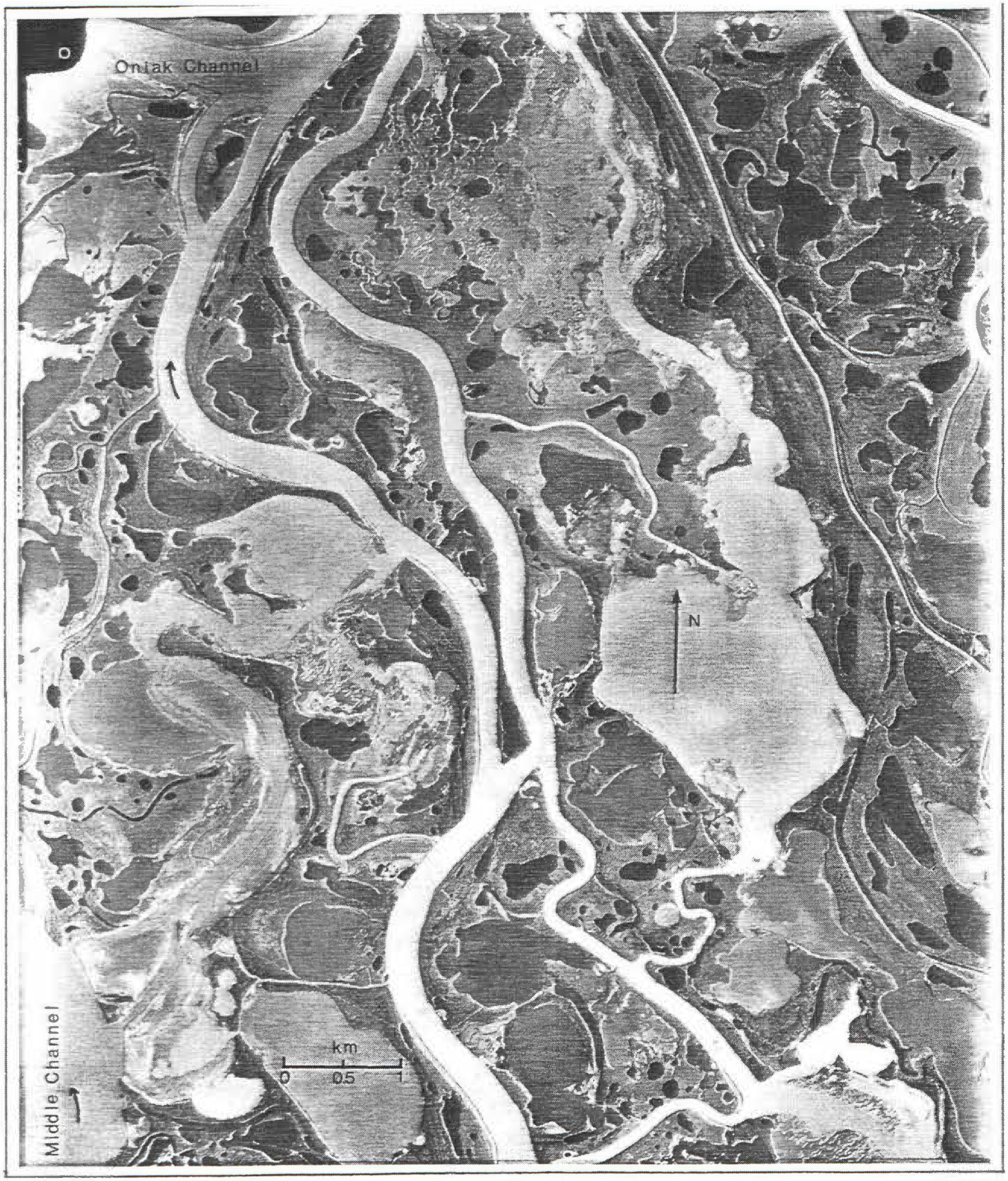


Figure 6.3 Part of the anastomosed floodplain of the Mackenzie Delta (EMR airphoto A2576-133, 1981).

Figure 6.4 Aerial photograph of lower Mountain River, NWT (from Mollard and Janes, 1984).

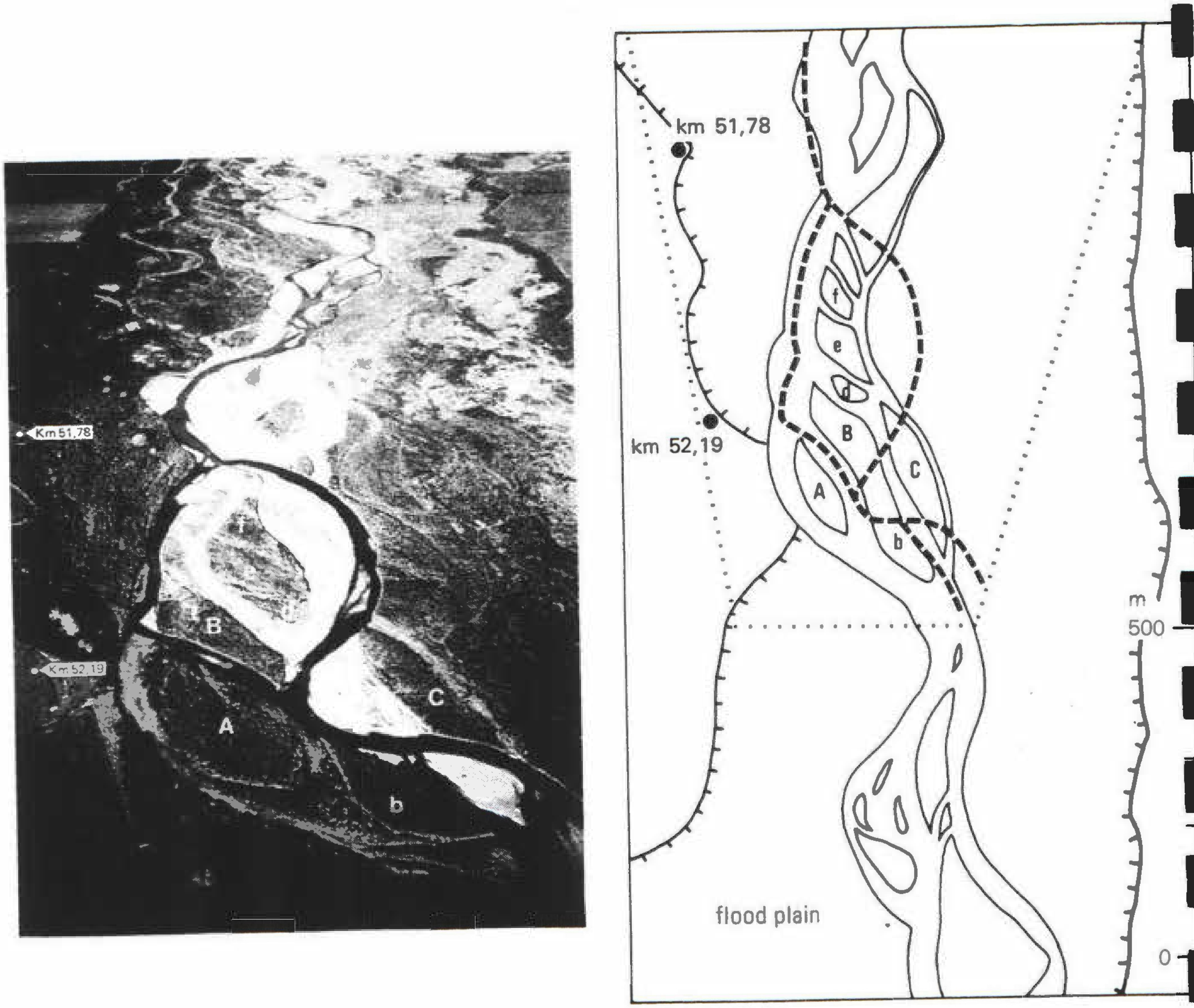


Figure 6.5 Reach of the wandering Selwyn River, New Zealand as seen in June, 1982 (left) and January, 1974 (right) (from carson, 1984). Many of the foothills streams in Alberta show the same floodplain pattern.

Figure 6.6 Aerial photograph of the Beaver River, Saskatchewan (from Mollard and Janes, 1984).

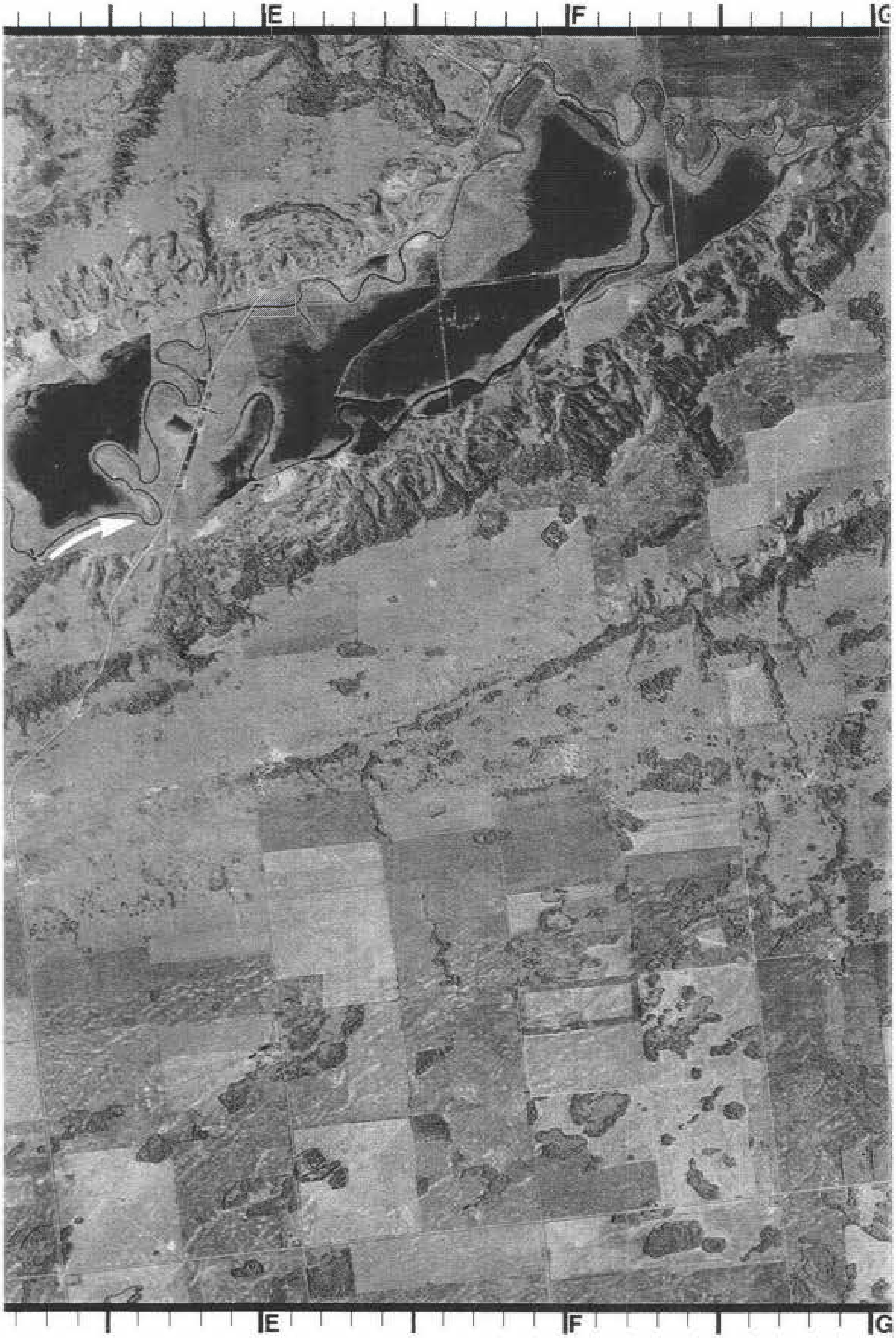


Figure 6.7 Aerial photograph of the Qu'Appelle River floodplain (from Mollard and Janes, 1984).

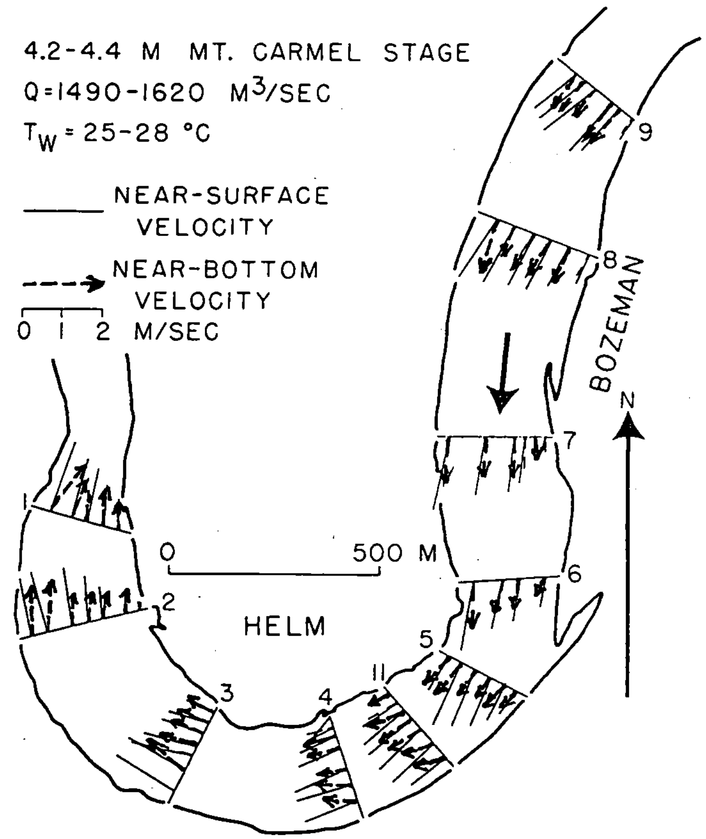


Figure 6.8 **Contrast in near-surface and near-bottom velocity pattern in a bend on the lower Wabush River, Indiana (from Jackson, 1975).**

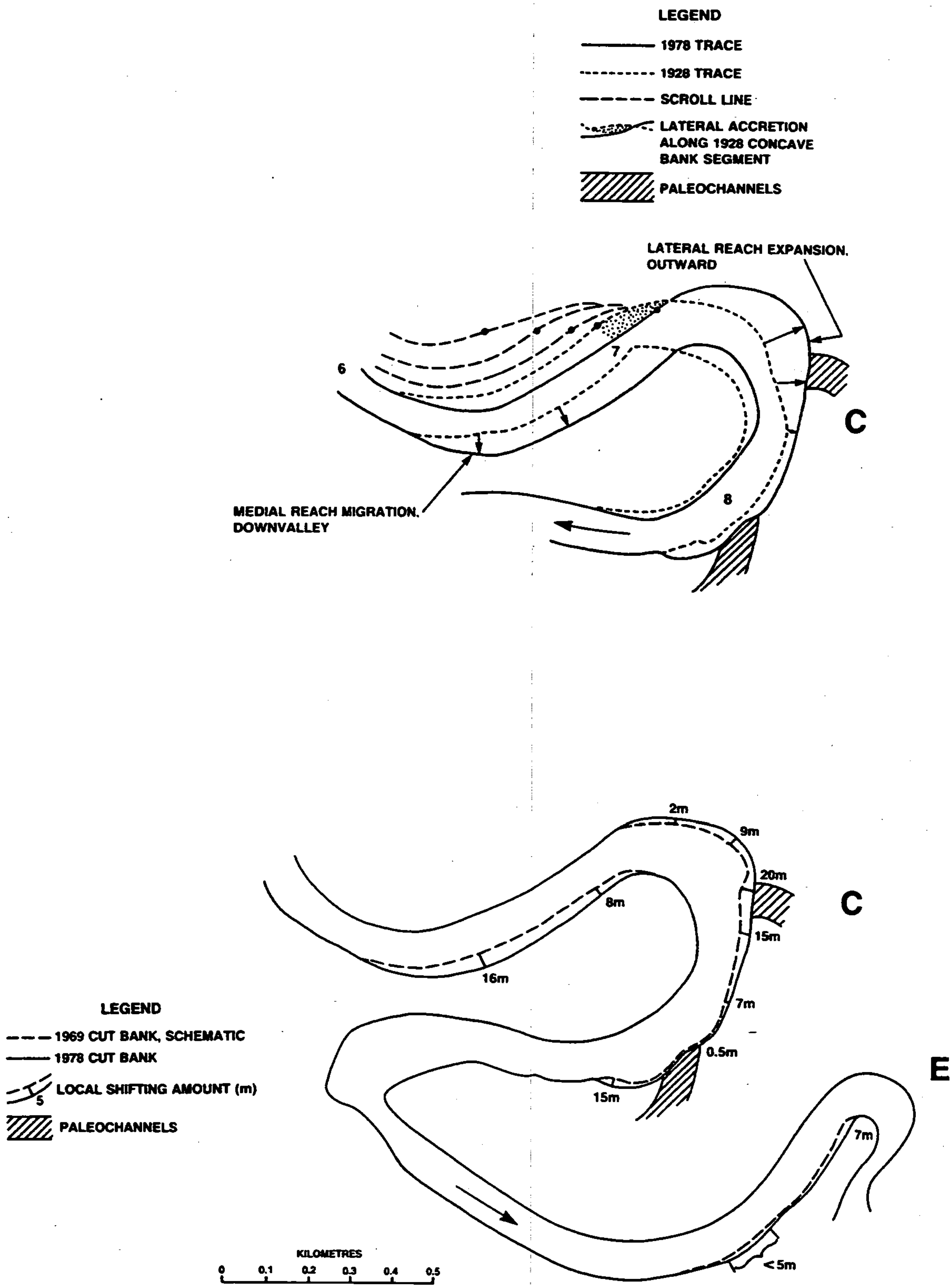


Figure 6.9 Channel bend shifting on the Rouge River, Quebec (fr… ROUGE RIVER, QUEBEC
… See setting of
of Bend C on Figure 6.10.

Figure 6.10 1978 aerial photograph: Rouge River, Quebec (from Lapointe and carson, 1986). 1928 channel boundary also marked on photograph. Flow is to bottom of photograph. Note asymmetry in meander geometry, typical of confined meanders.

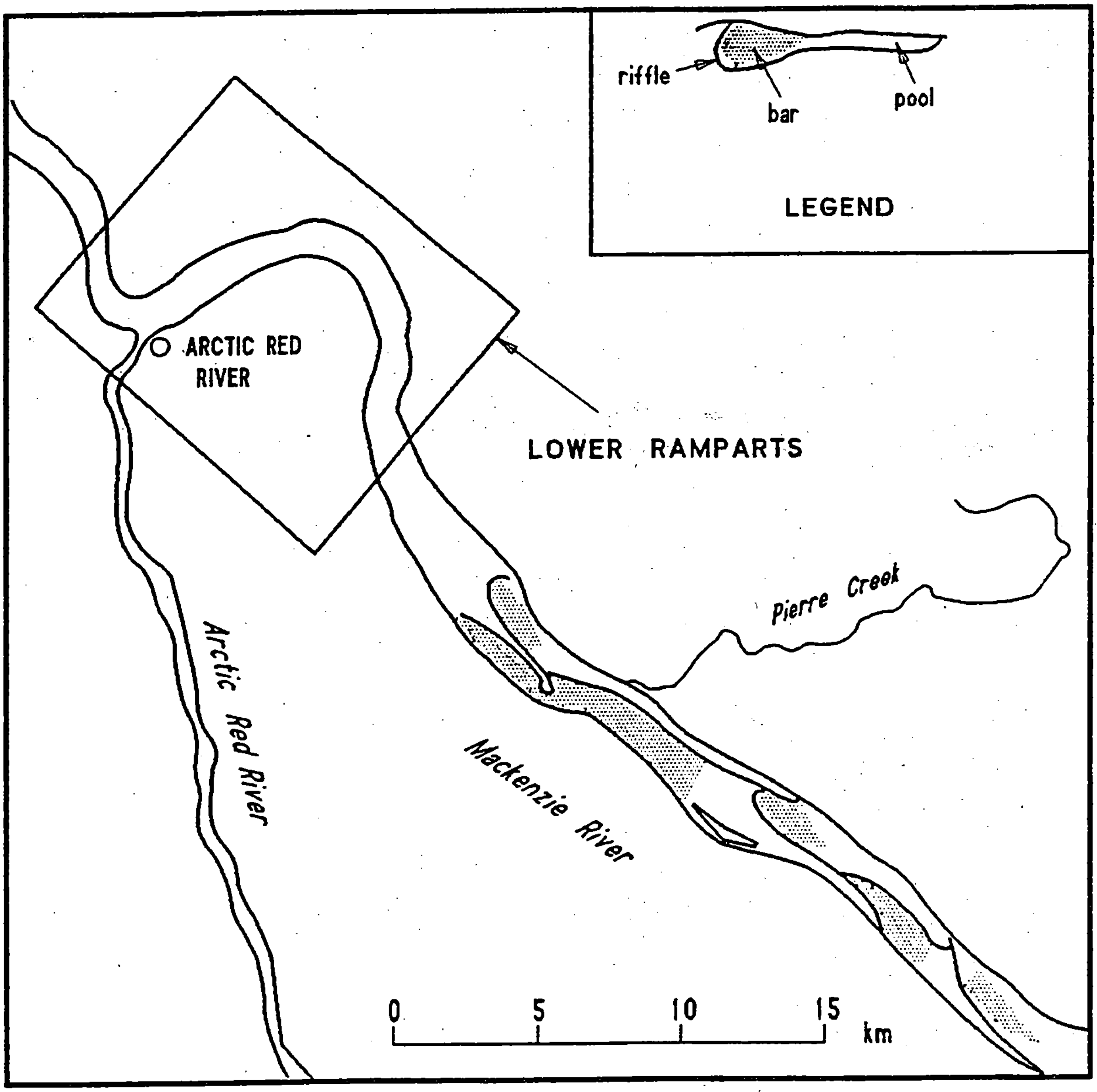


Figure 6.11 Pool-riffle reach of lower Mackenzie River, NWT.

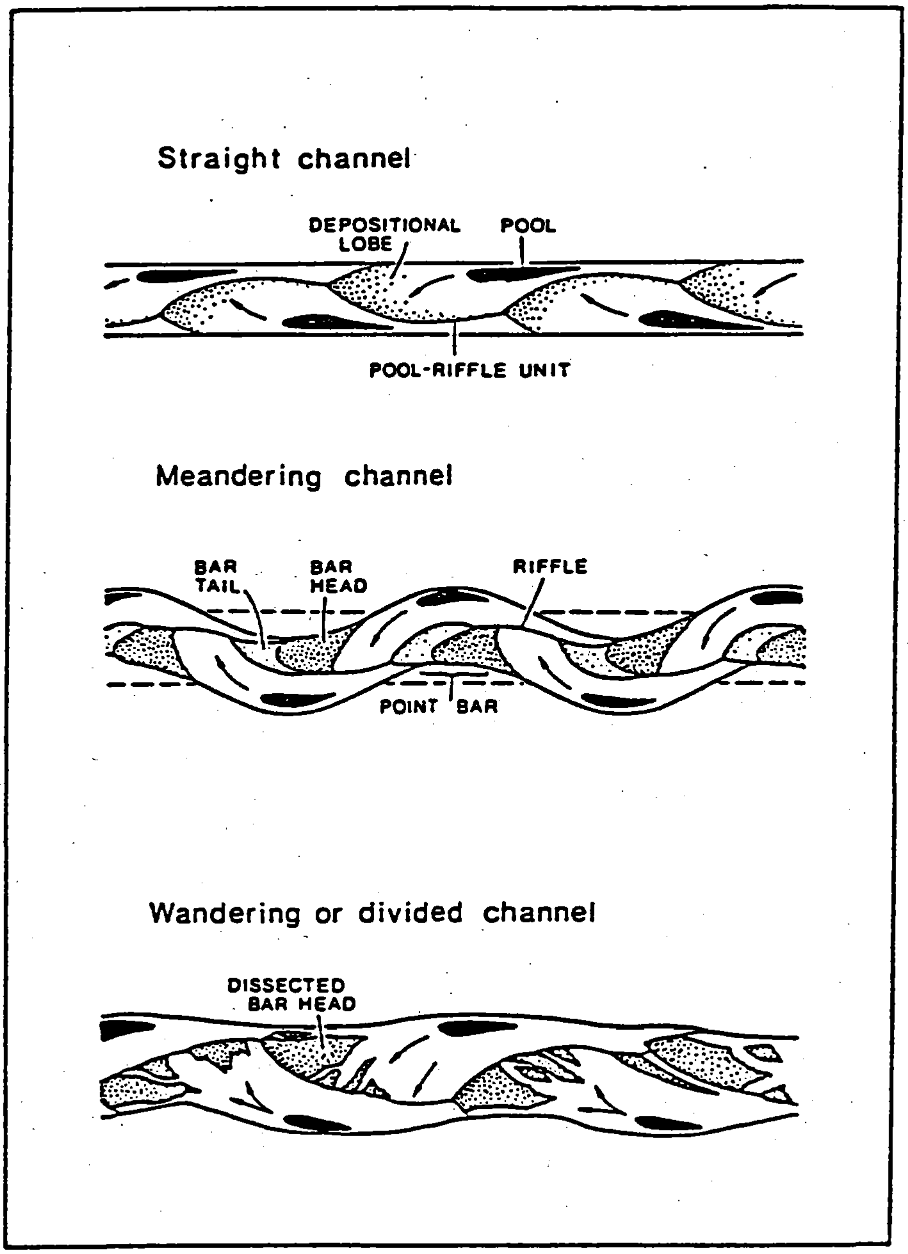


Figure 6.12 Idealized morphology of pool-riffle units in alluvial channels (from Thomson, 1986).

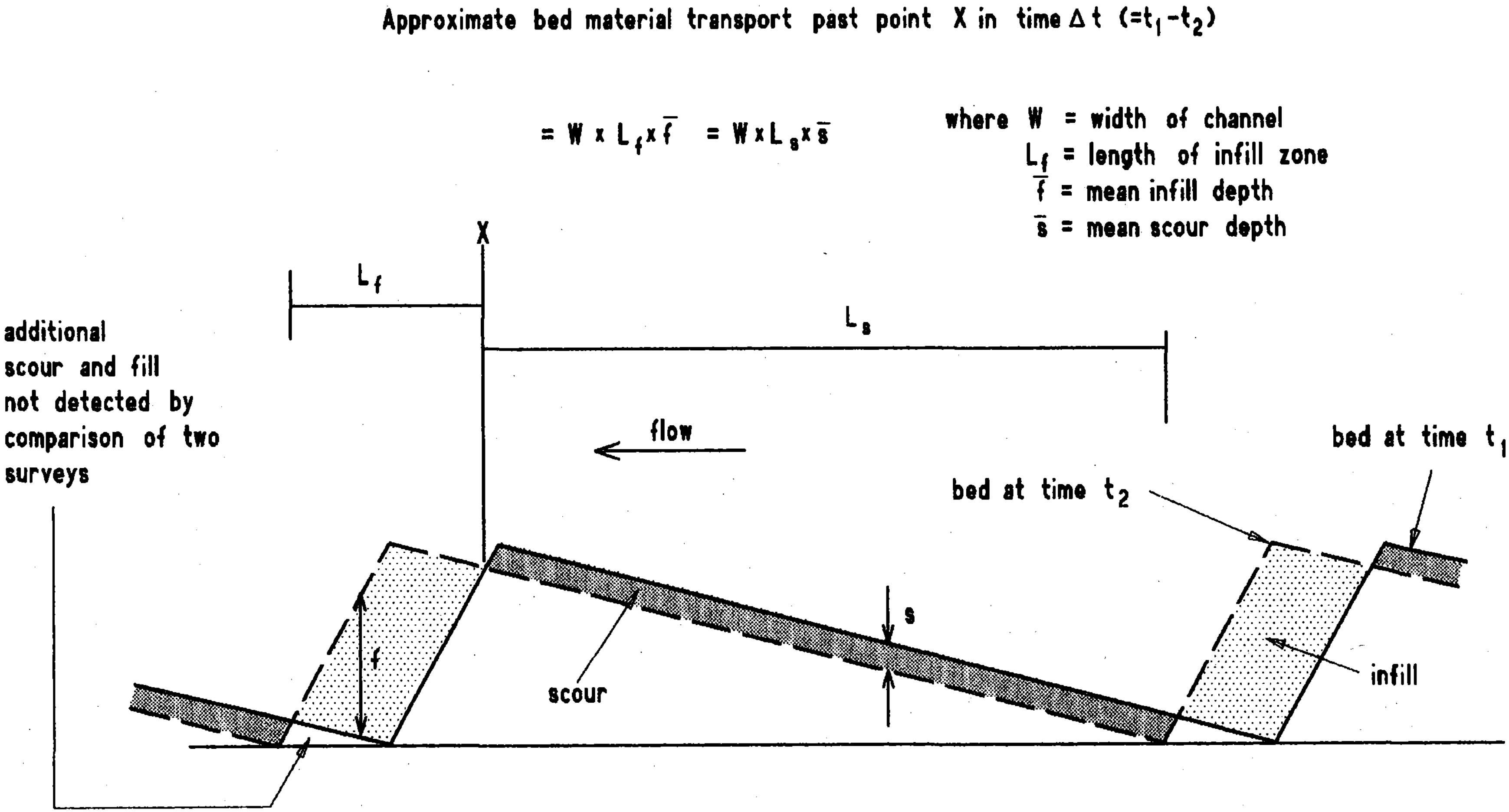


Figure 6.13 Schematic calculation of bed material transport in a channel with two-dimensional bars.

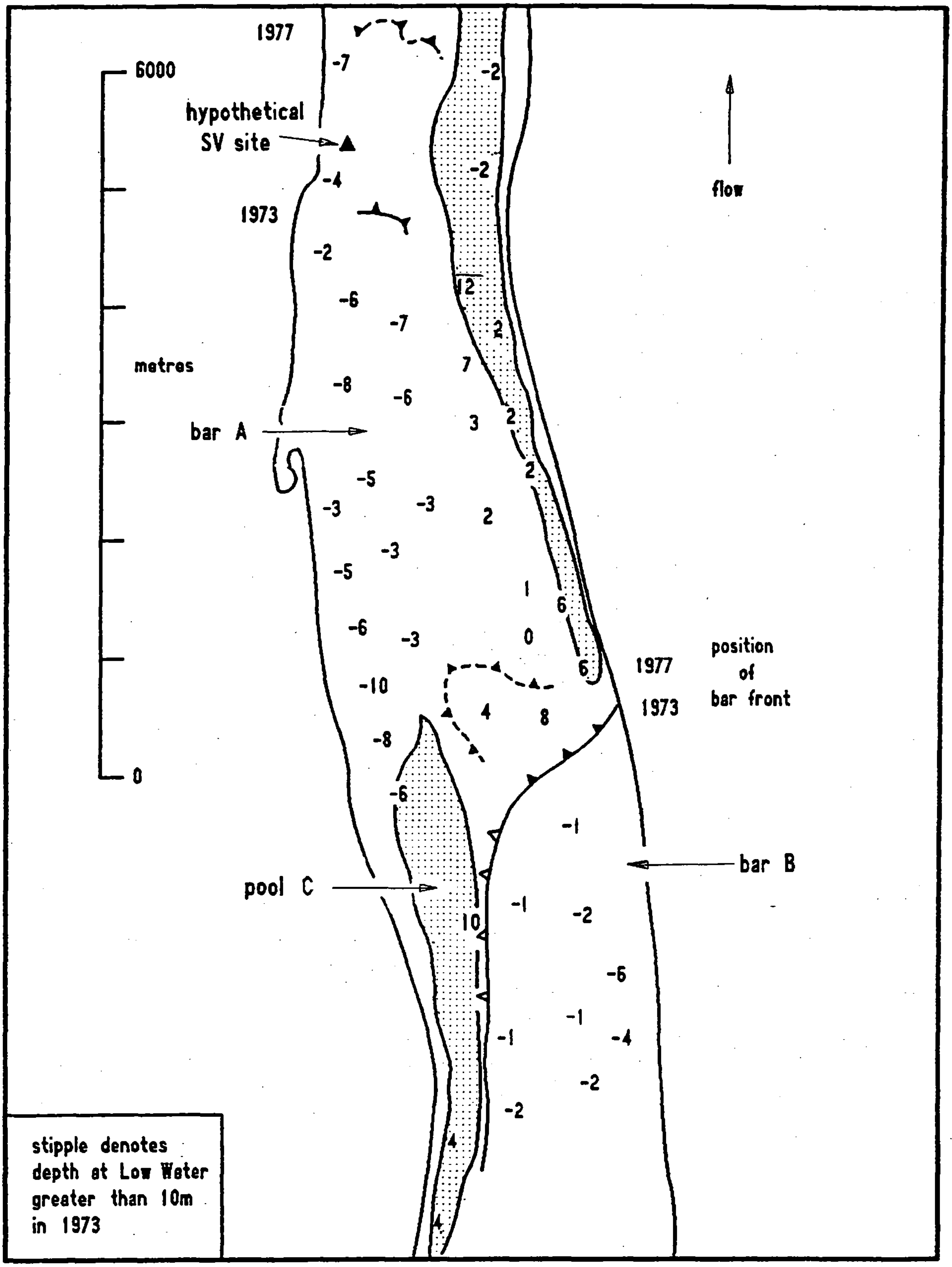


Figure 6.14 Downstream migration of pool-riffle units in Mackenzie River. Spot values of scour (negative) and deposition (positive) between August 1973 and August 1977 in metres.

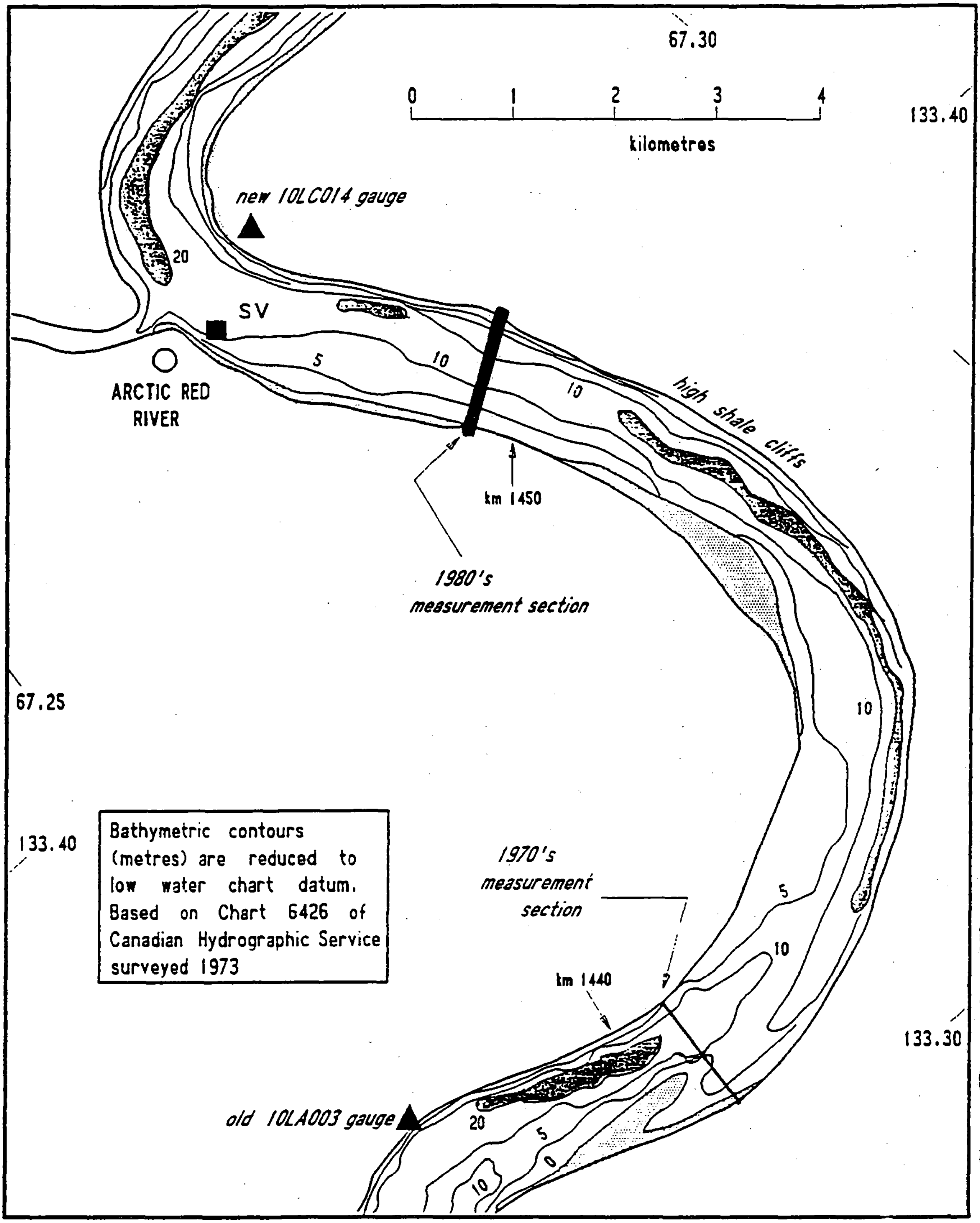


Figure 6.15 Bathymetry of Mackenzie River at the Lower Ramparts, NWT. Flow is to top of page.

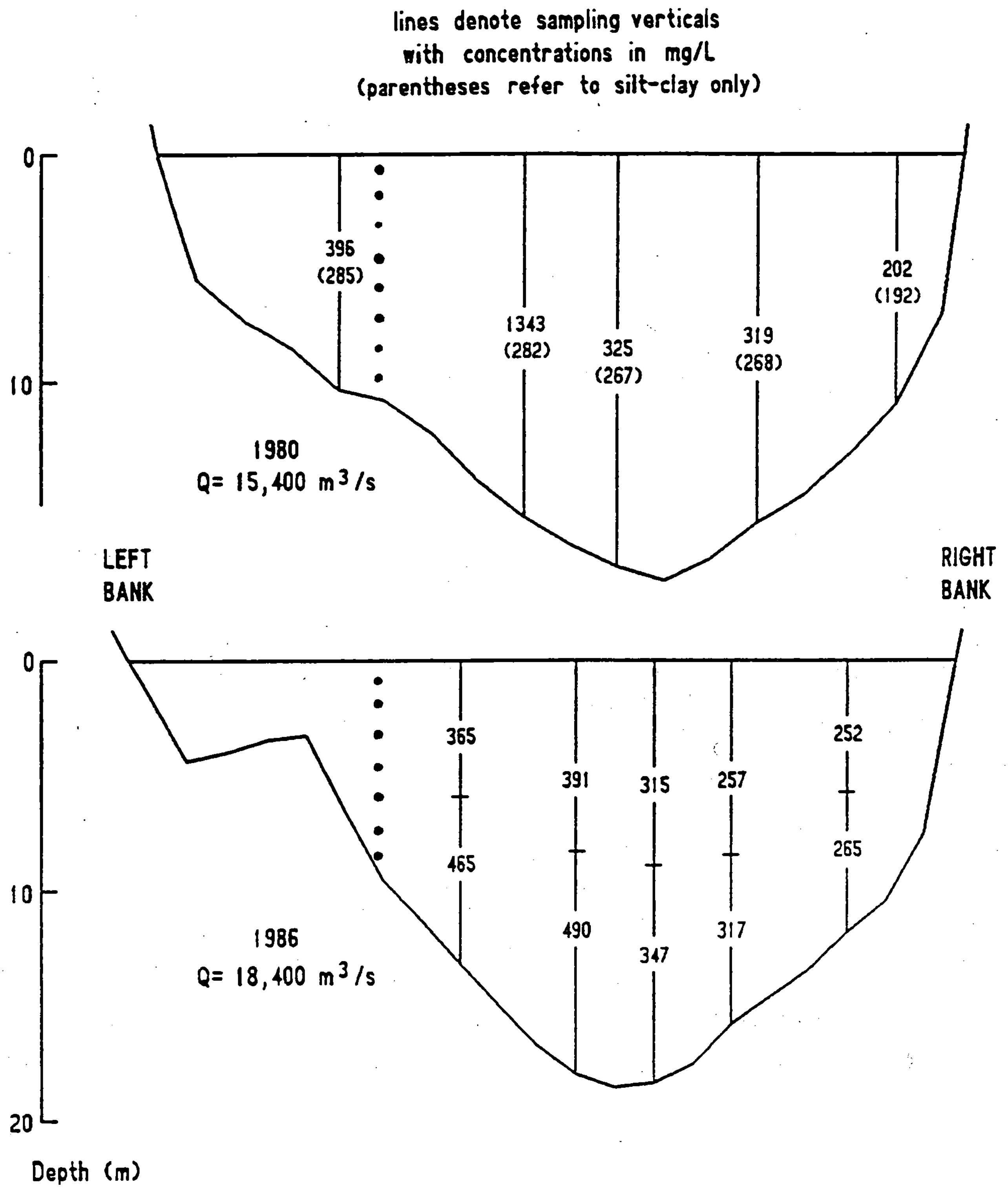


Figure 6.16 Cross-sectional distribution of sediment, Mackenzie River, upstream of Arctic Red River, NWT. Dotted vertical denotes SV site downstream of measurement section.

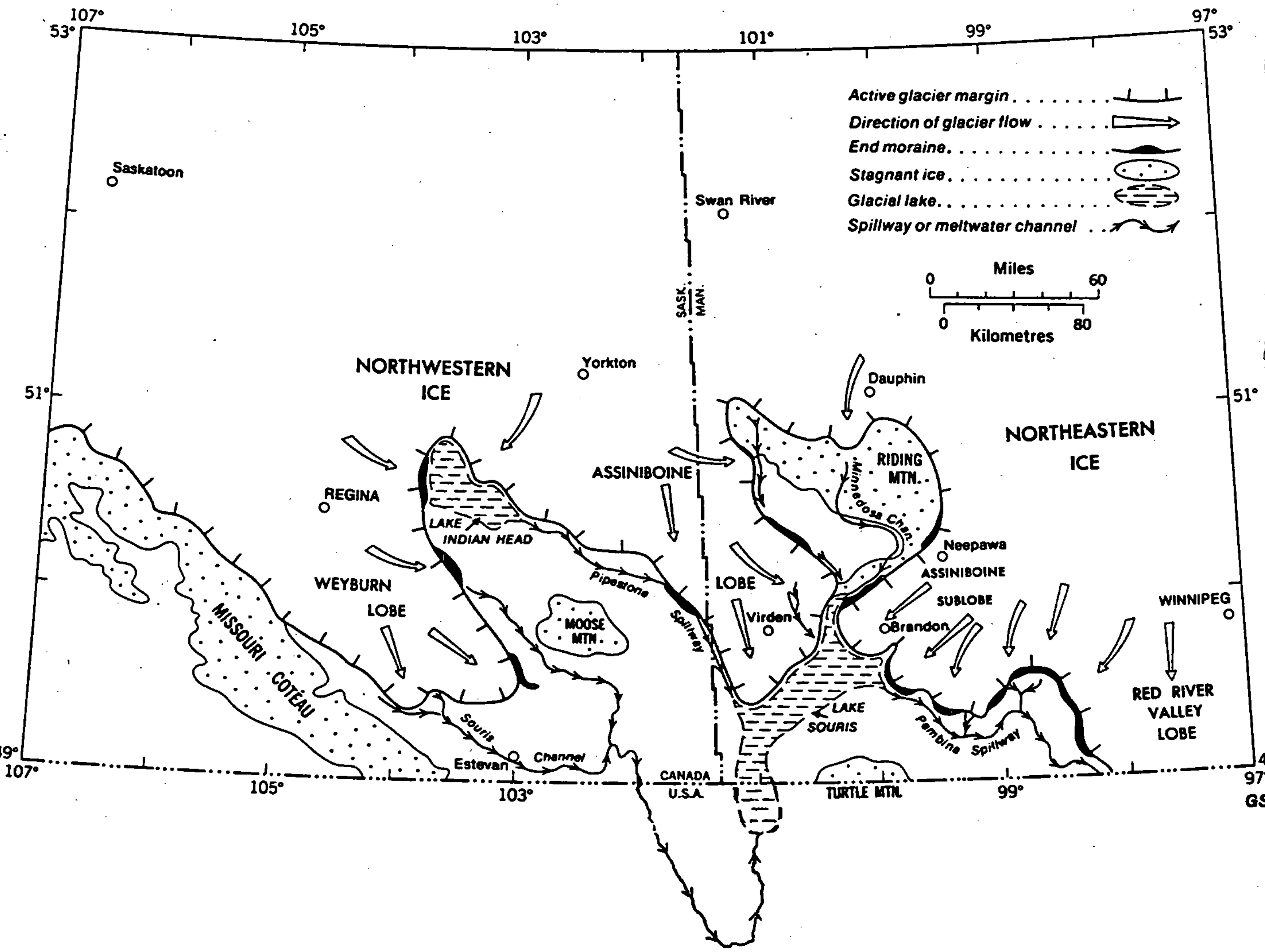


Figure 7.1 Pattern of ice-marginal drainage in southern Saskatchewan and Manitoba at an early stage of deglaciation (about 14,500 years ago) (from Klassen, 1975).

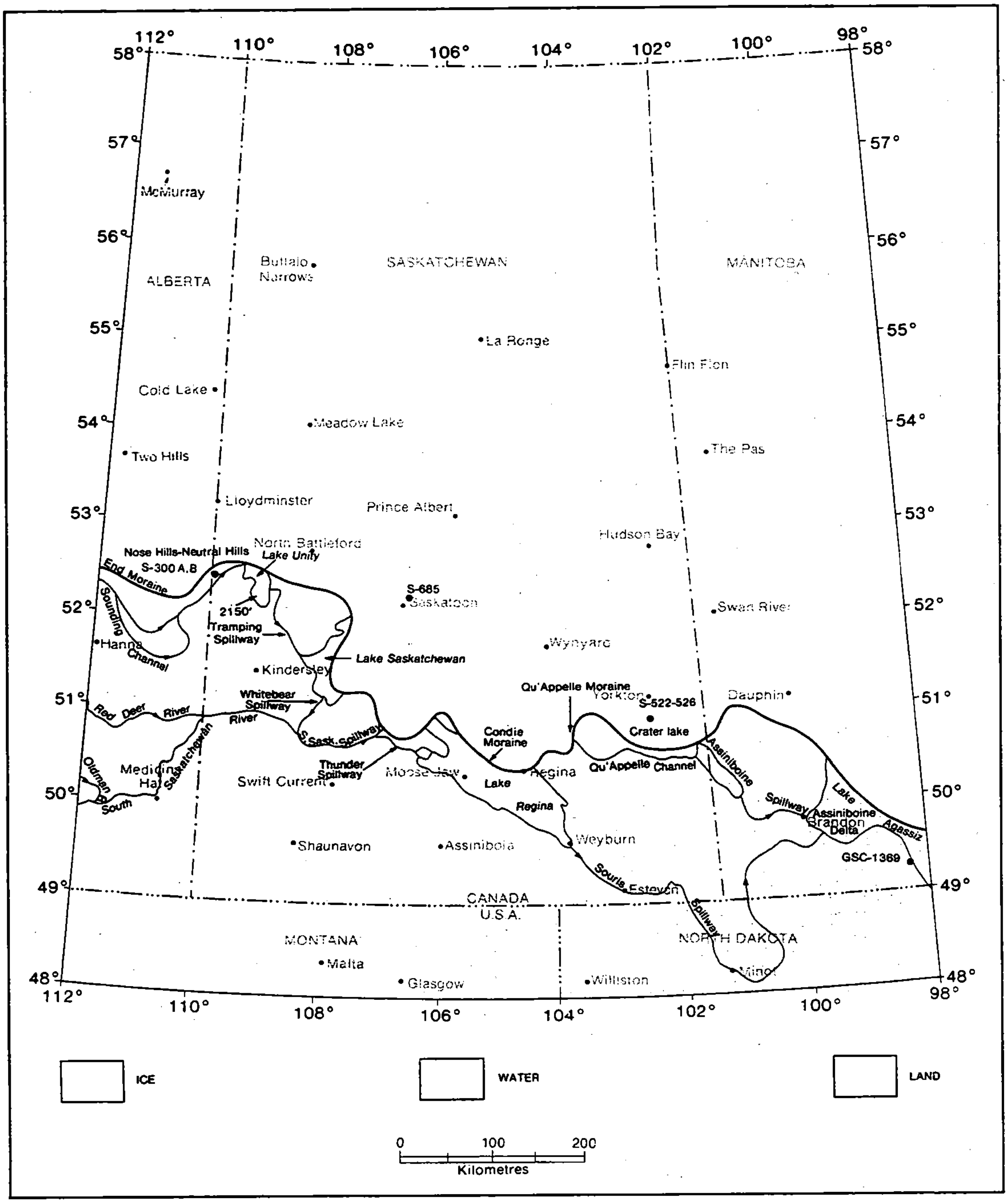


Figure 7.2 **Ice-marginal drainage pattern in south-central prairies about 14,000 years (from Christiansen, 1979).**

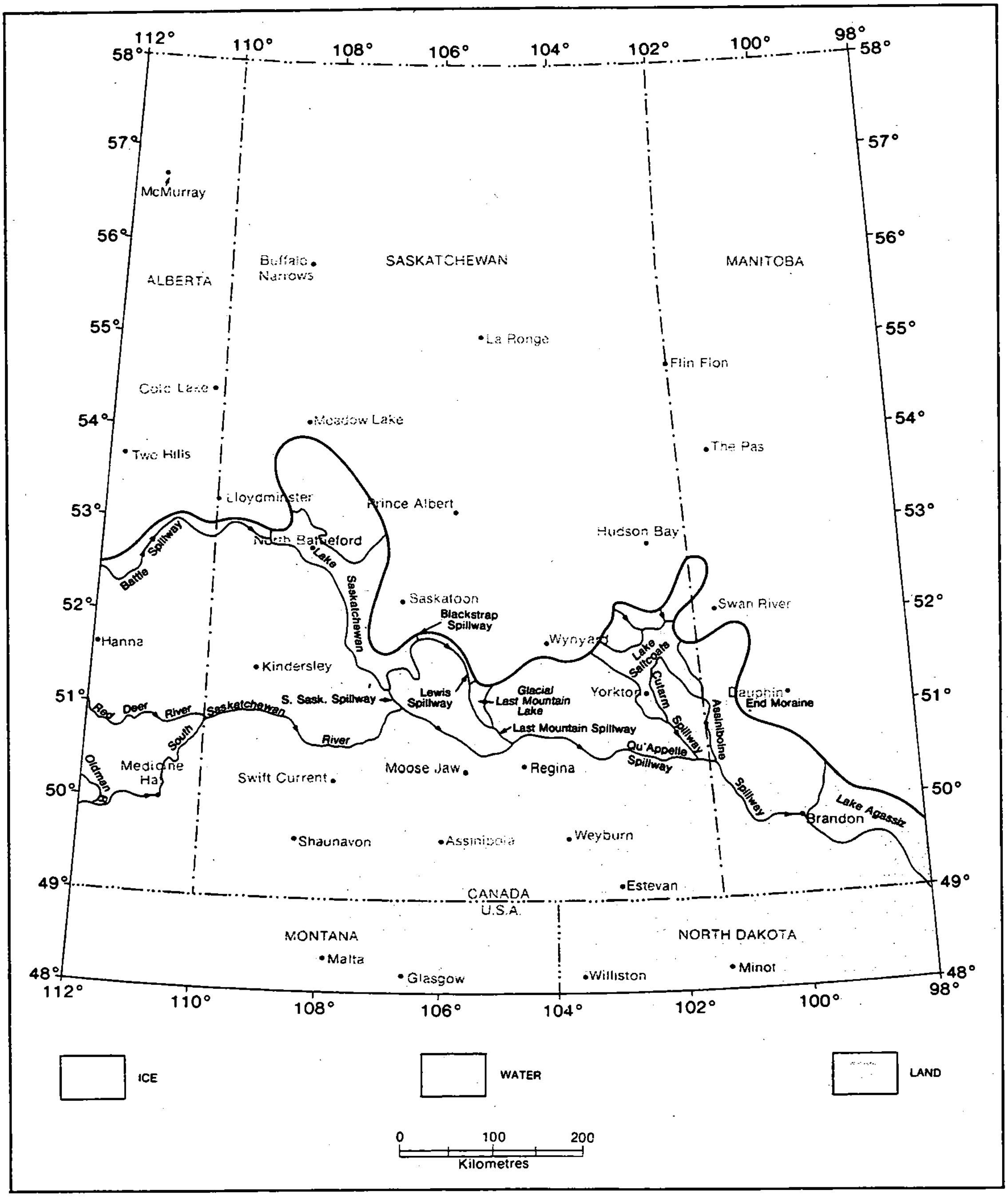


Figure 7.3 Ice-marginal drainage pattern in central prairies about 12,500 years ago (from Christiansen, 1979).

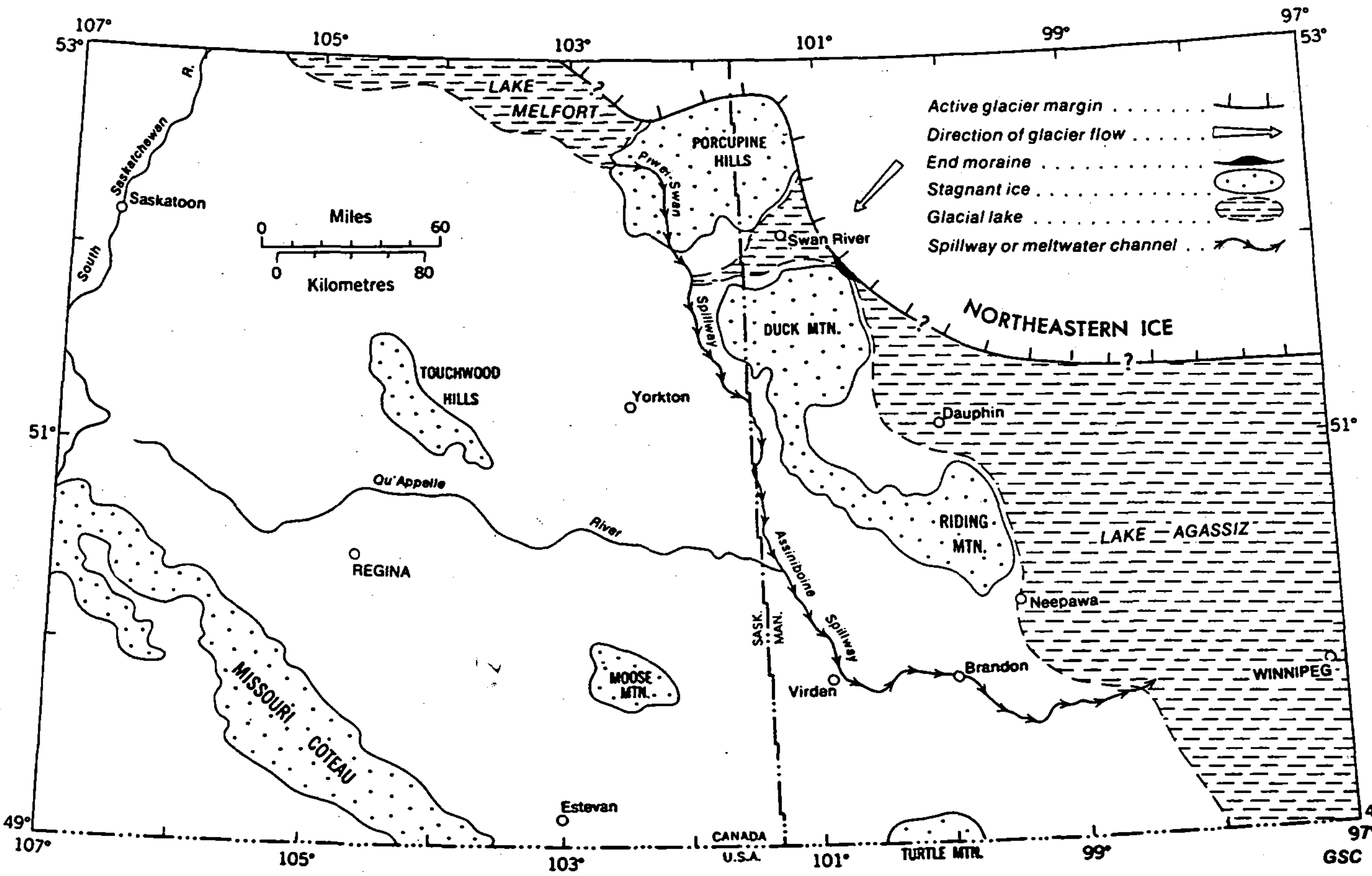


Figure 7.4 Drainage pattern in the Assiniboine basin during last stages of deglaciation (about 11,700 BP) (from Klassen, 1975).

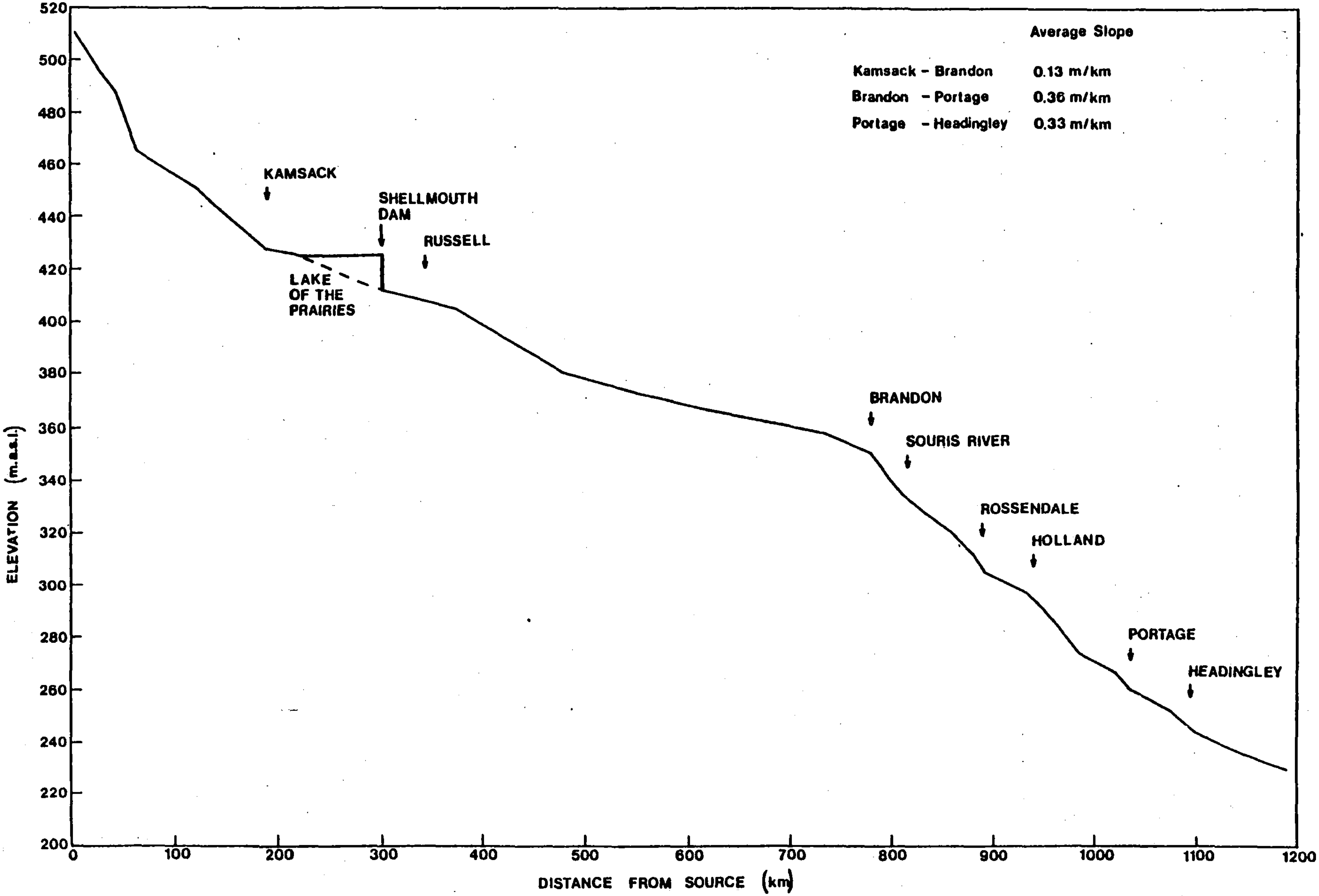


Figure 7.5 Long profile of Assiniboine River(from Ashmore, 1990).

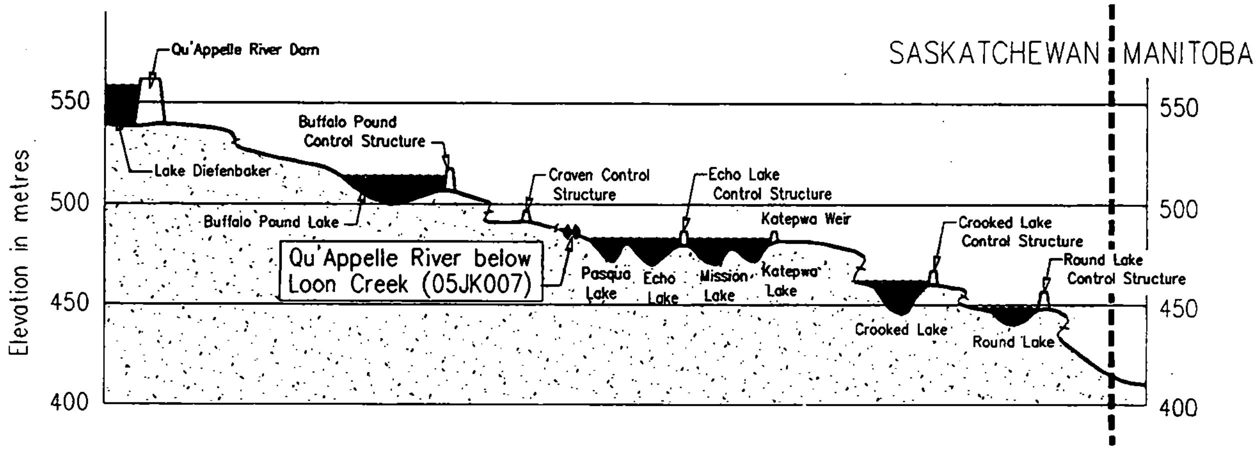
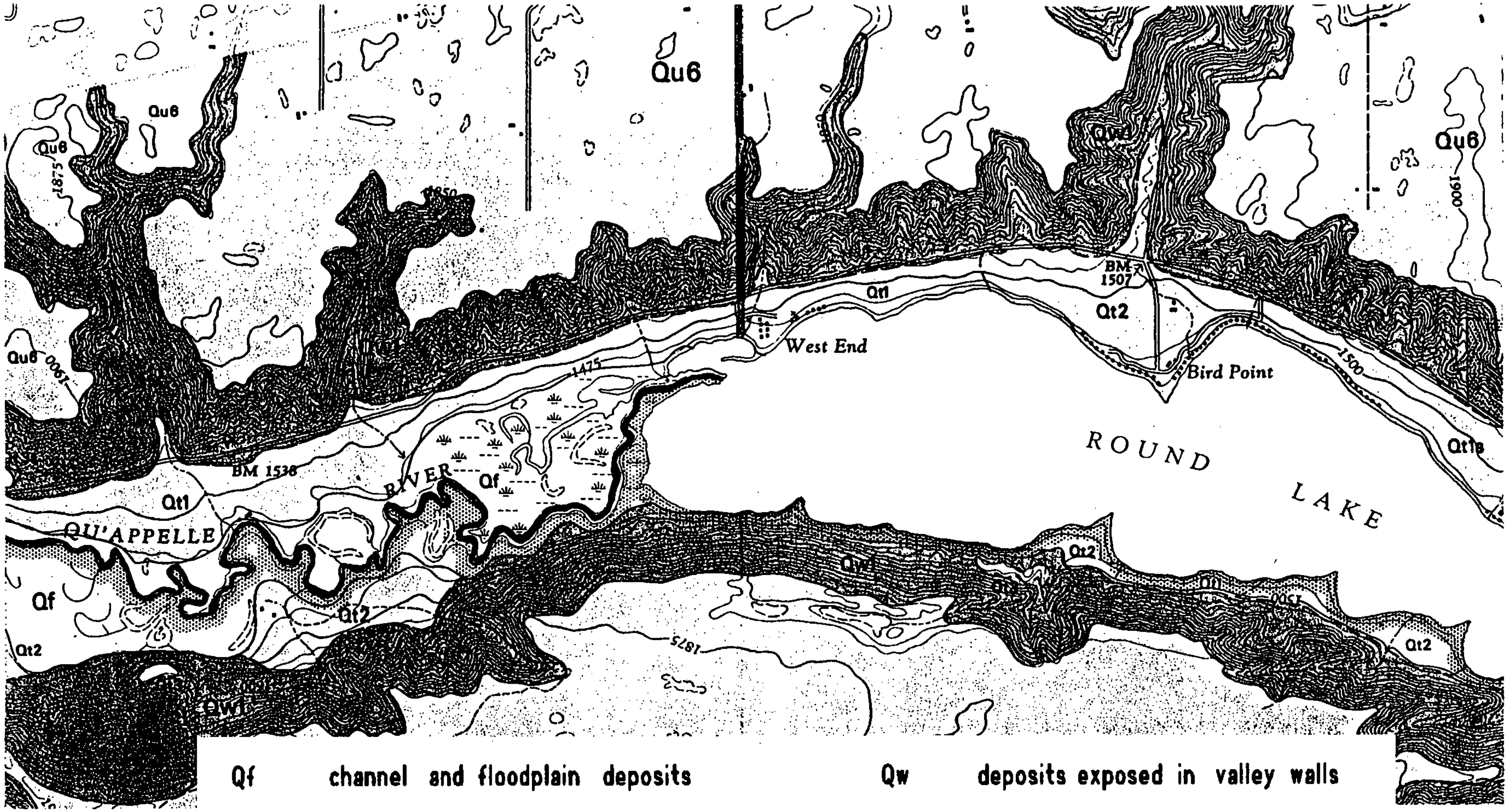


Figure 7.6 **Long profile of Qu'Appelle River (from WRB/IWD Fact Sheet for Qu'Appelle River below Loon Creek).**



Qf	channel and floodplain deposits	Qw	deposits exposed in valley walls 1 mainly till 2 silt, sand, gravel
Qt	terrace and alluvial-colluvial deposits 1 low terrace and valley bottom 2 alluvial fans 3 high terrace	Qu	upland deposits 3 outwash 6 ground moraine till

Figure 7.7 Progradation of Qu'Appelle River into Round Lake (from Klassen, 1975).

October 12, 1987 Scale 1:10 000 CSMA 87349 049 11

Longitudinal Profile of Study Reach

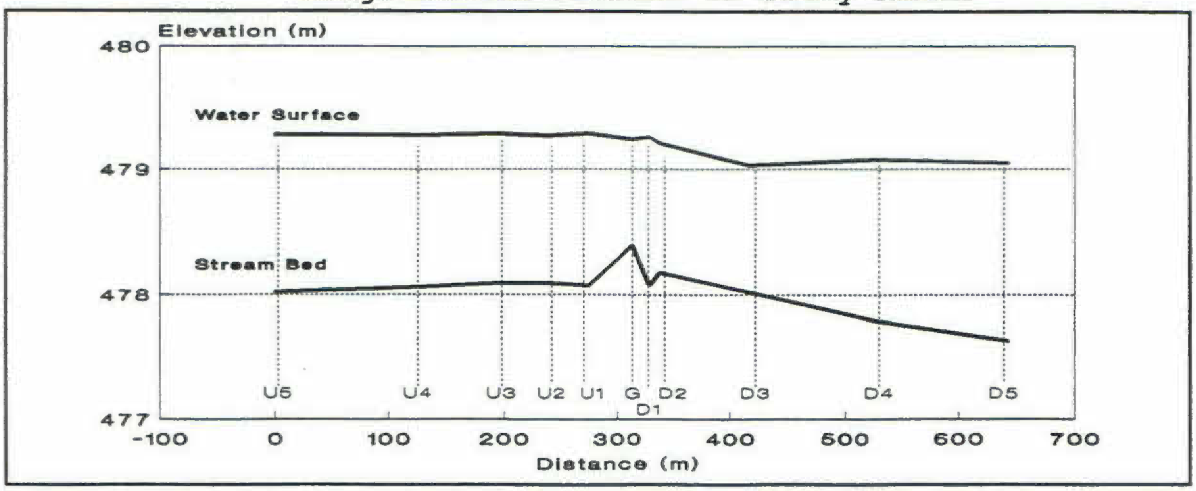


Figure 7.8 Air photograph and long profile of Qu'Appelle River below Loon Creek (from WRB/IWD Fact Sheet).

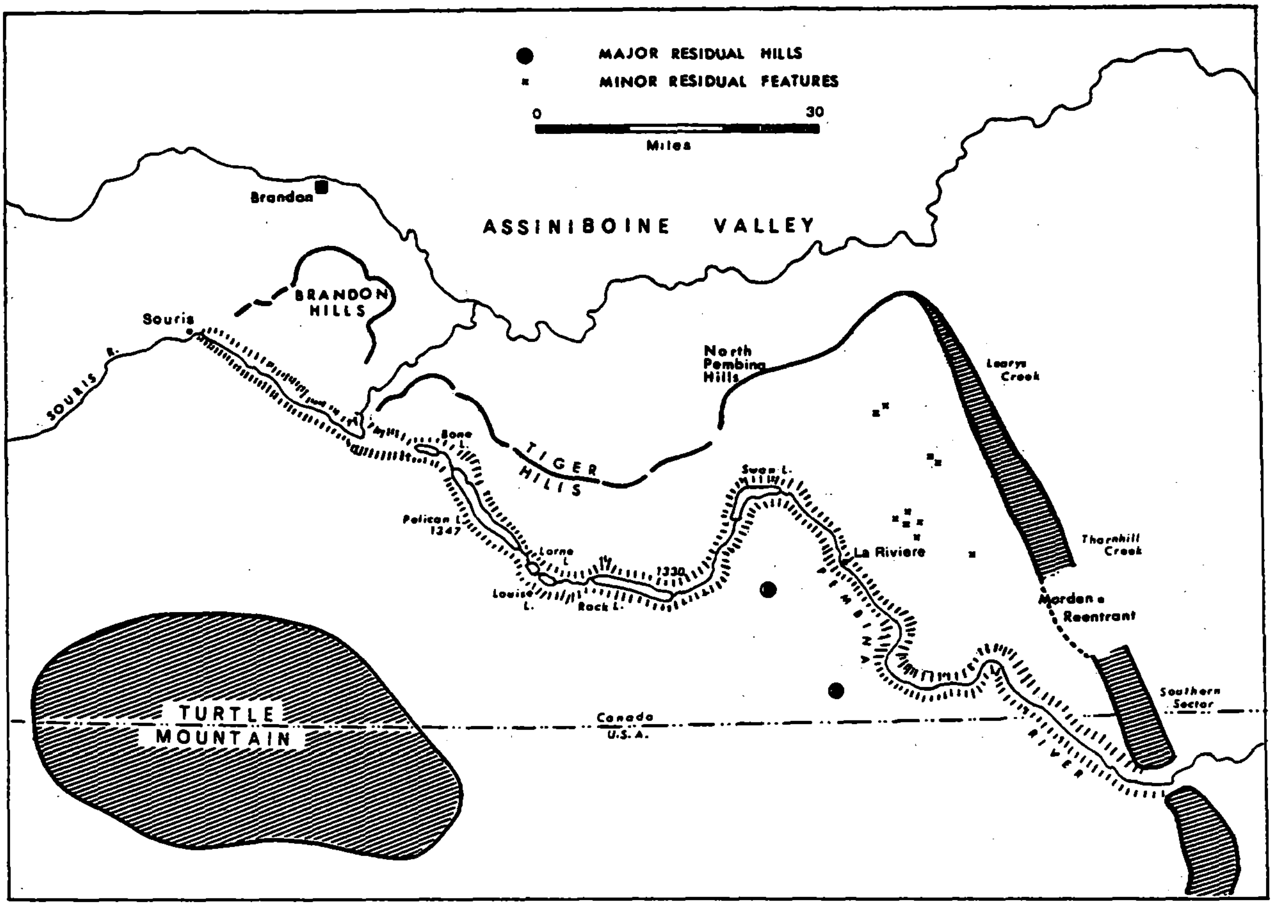


Figure 7.9 The Souris-Pembina spillway, Manitoba (from Bird, 1972).

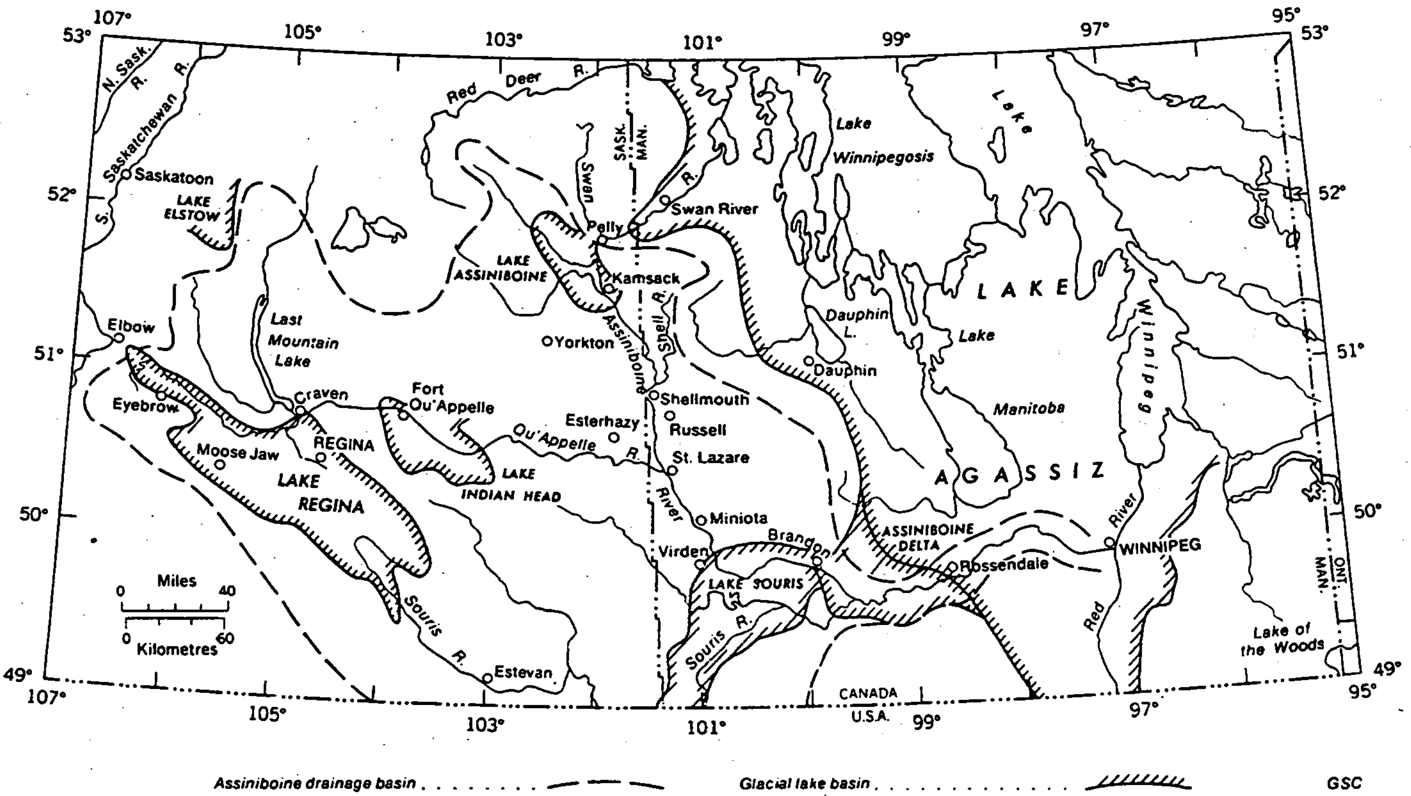


Figure 7.10 Map of Assiniboine drainage basin showing location of major glacial lakes (from Klassen, 1975).

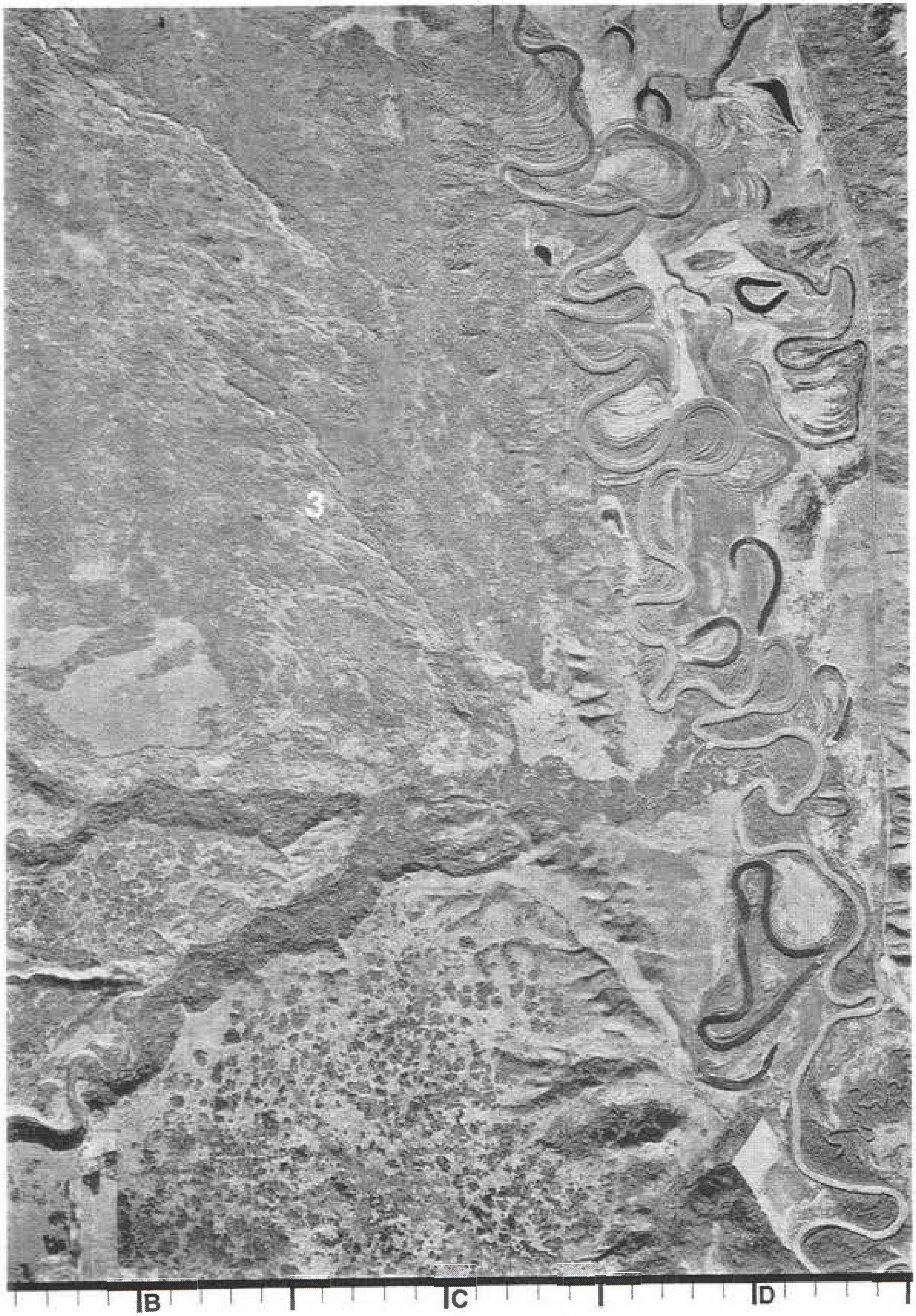


Figure 7.11 Air photograph of Assiniboine valley, near St-Lazare Manitoba, downstream of Qu'Appelle confluence (from Mollard and Janes, 1984).

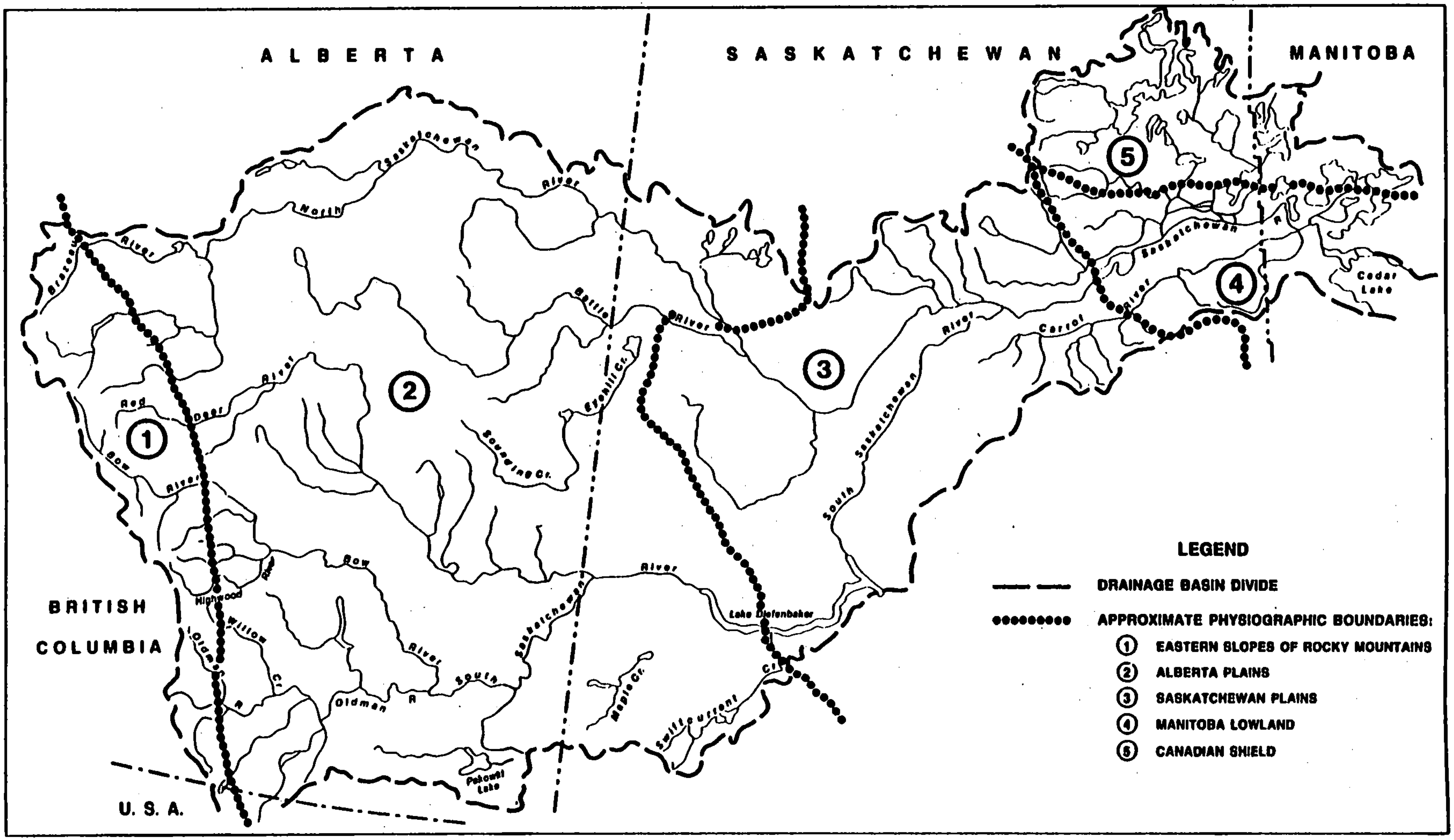


Figure 8.1 Drainage and physiography of the Saskatchewan River basin (from Ashmore, 1986).

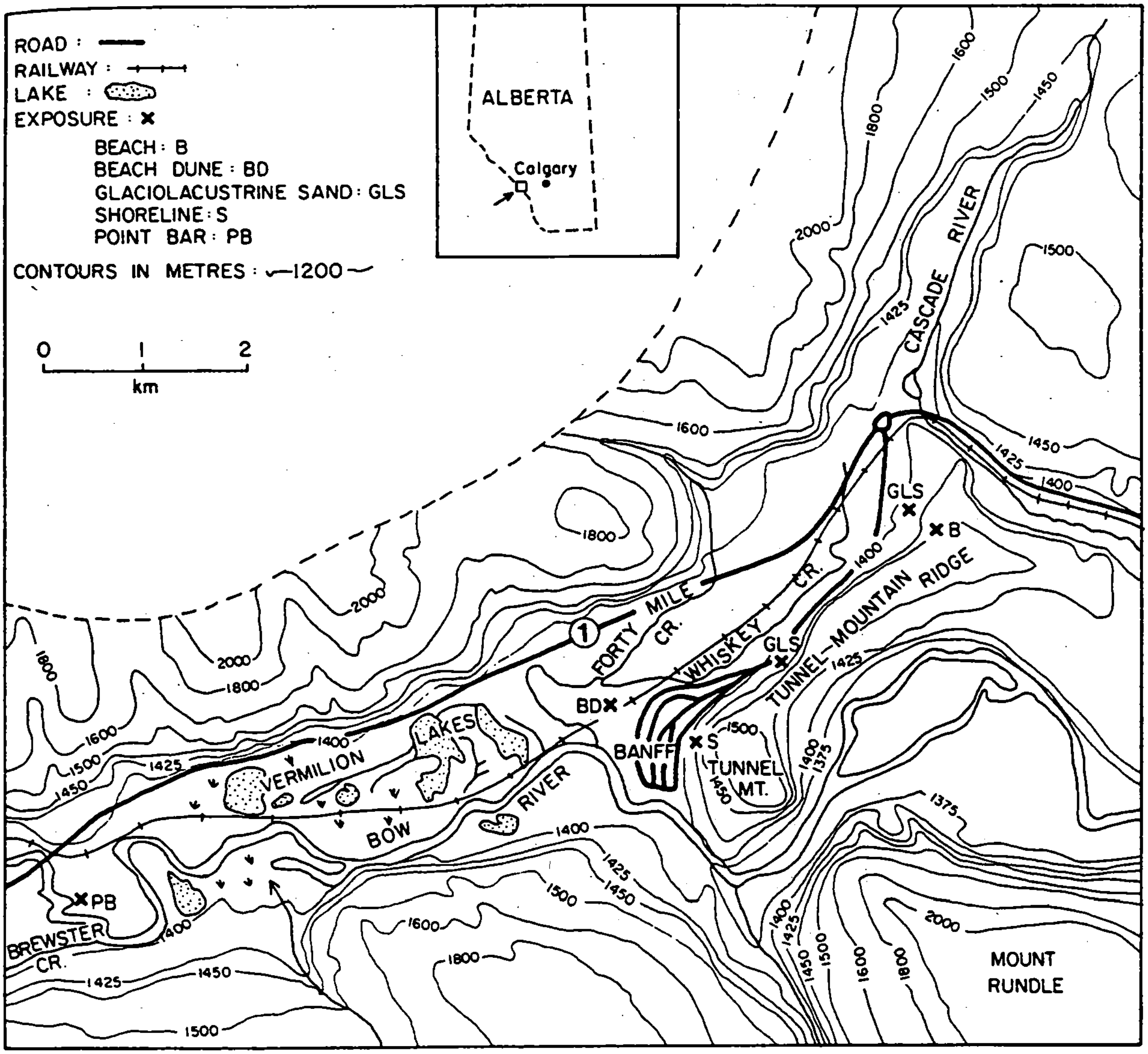


Figure 8.2 Floodplain of Bow River near Banff, Alberta (from Kostaschuk and Smith, 1983).

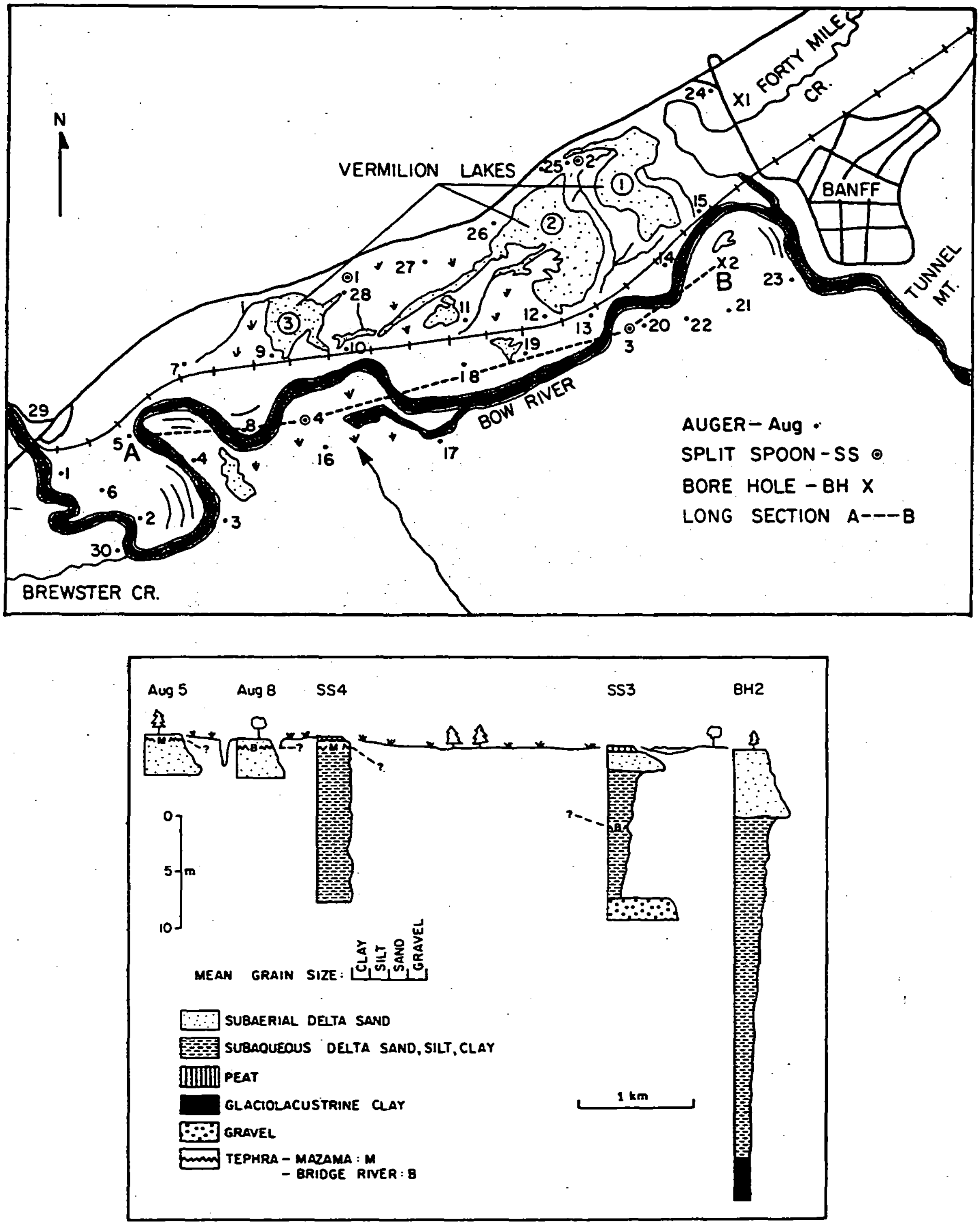


Figure 8.3 Stratigraphy of valley bottom deposits, Bow River, Banff (from Kostaschuk and Smith, 1983).

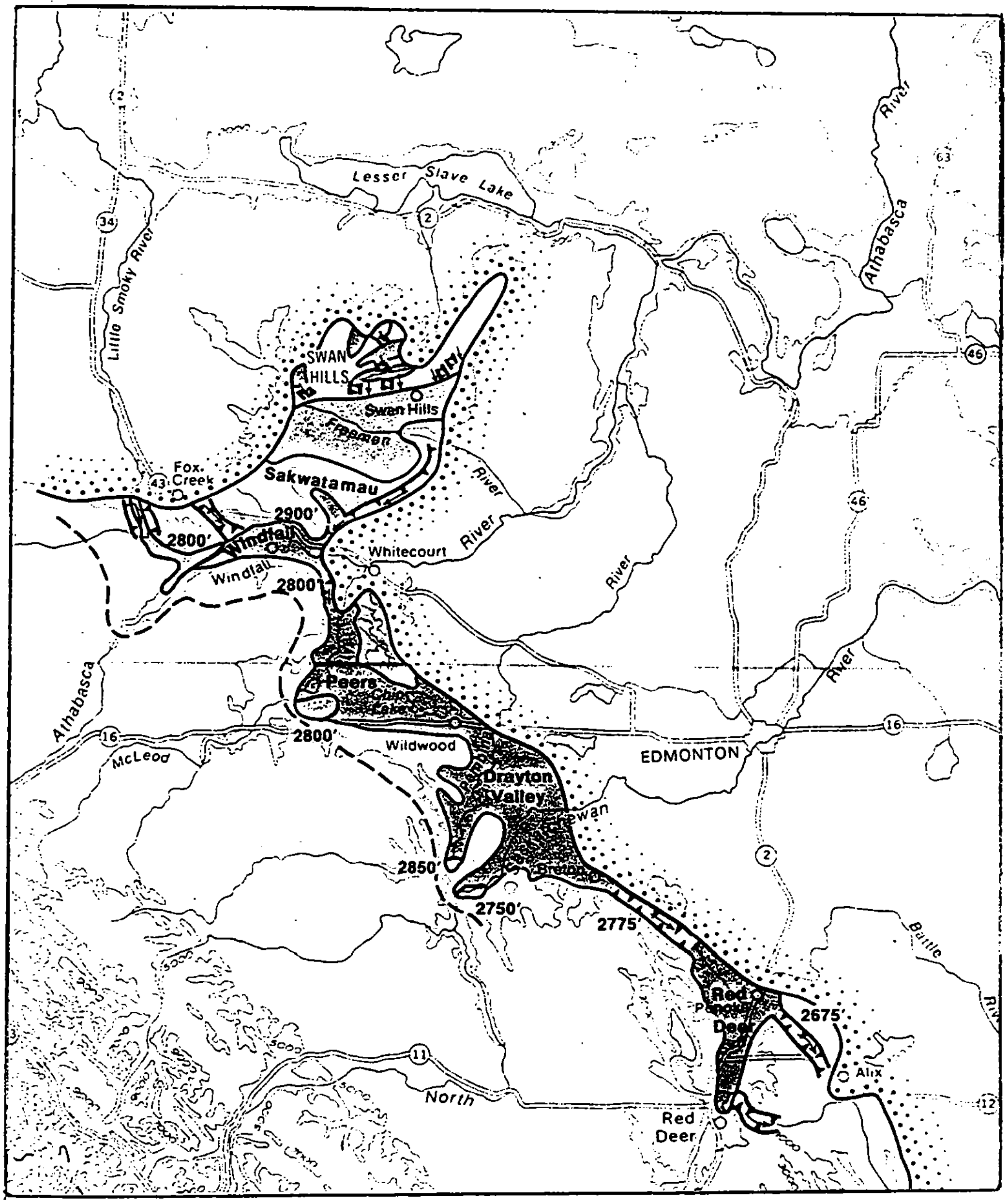


Figure 8.4 Ice-marginal drainage pattern, Edmonton area, about 13,000 BP (from St-Onge, 1972).

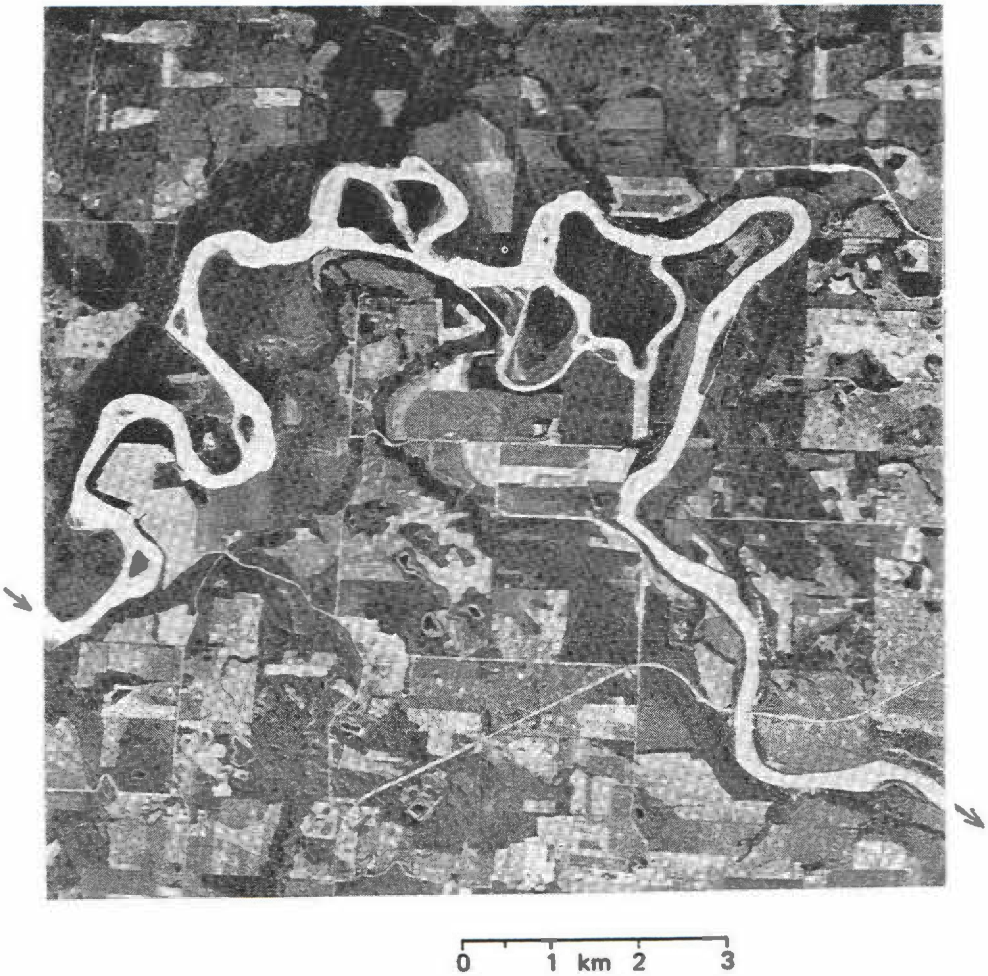
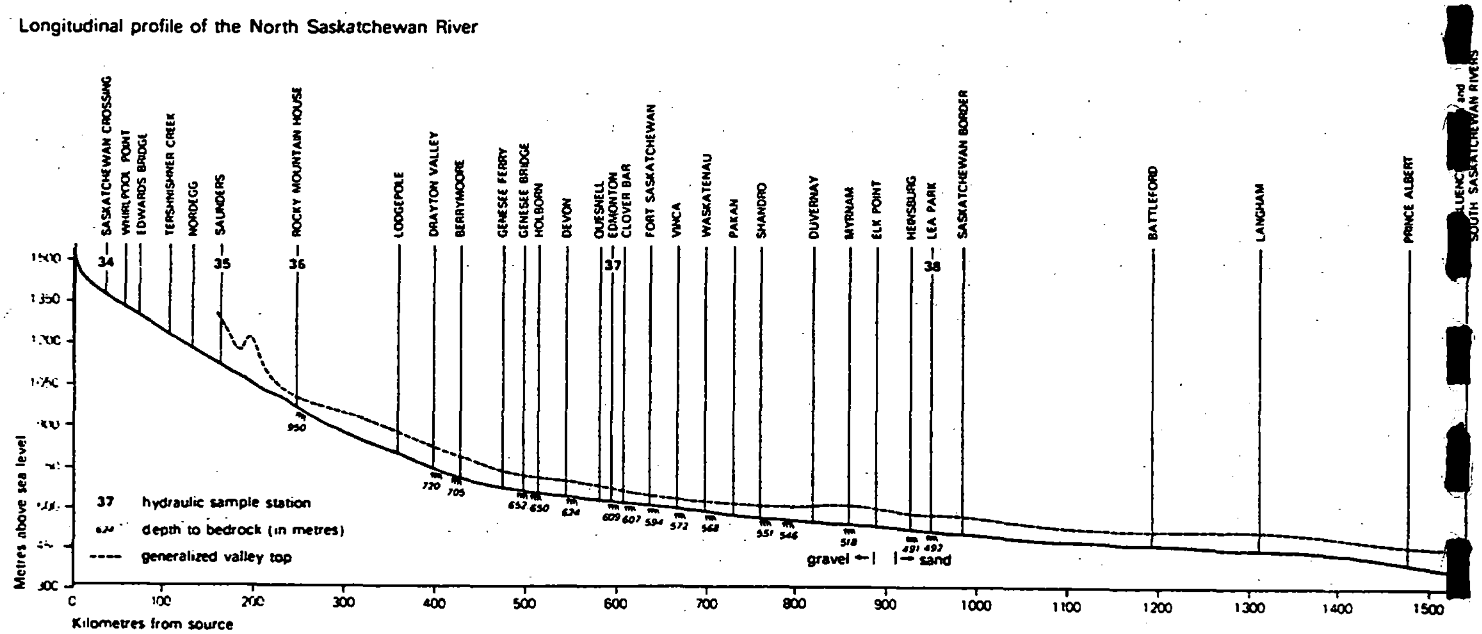


Figure 8.5 Air photograph of North Saskatchewan River south of Wabamun Lake, Alberta (National Air Photo Library photo A21644-13).



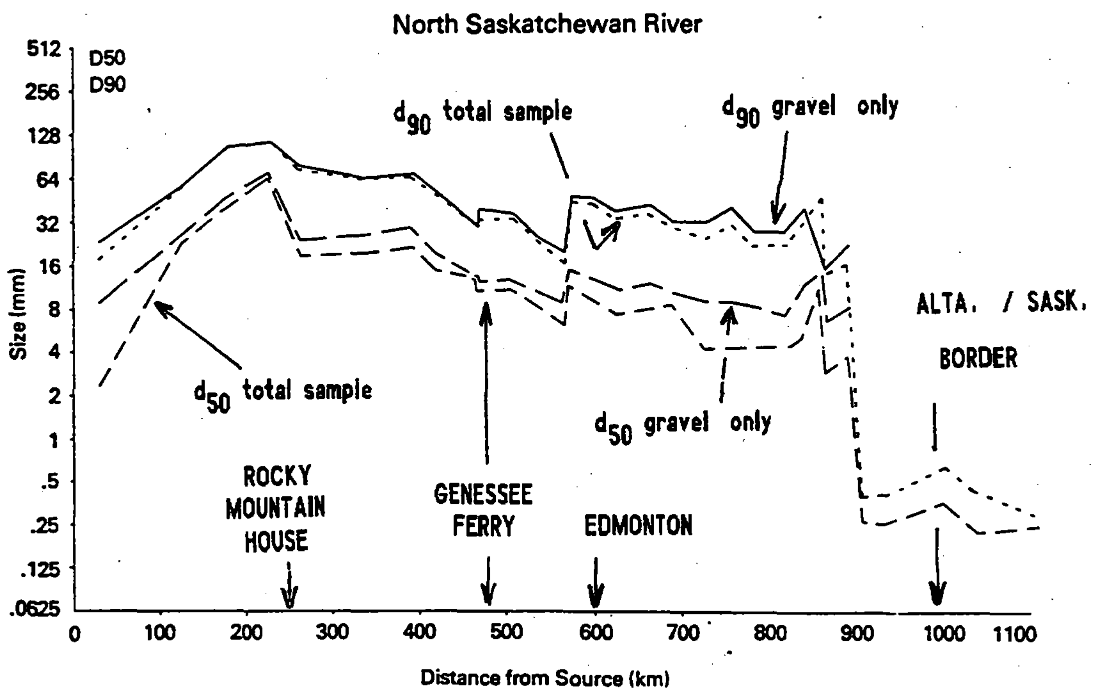
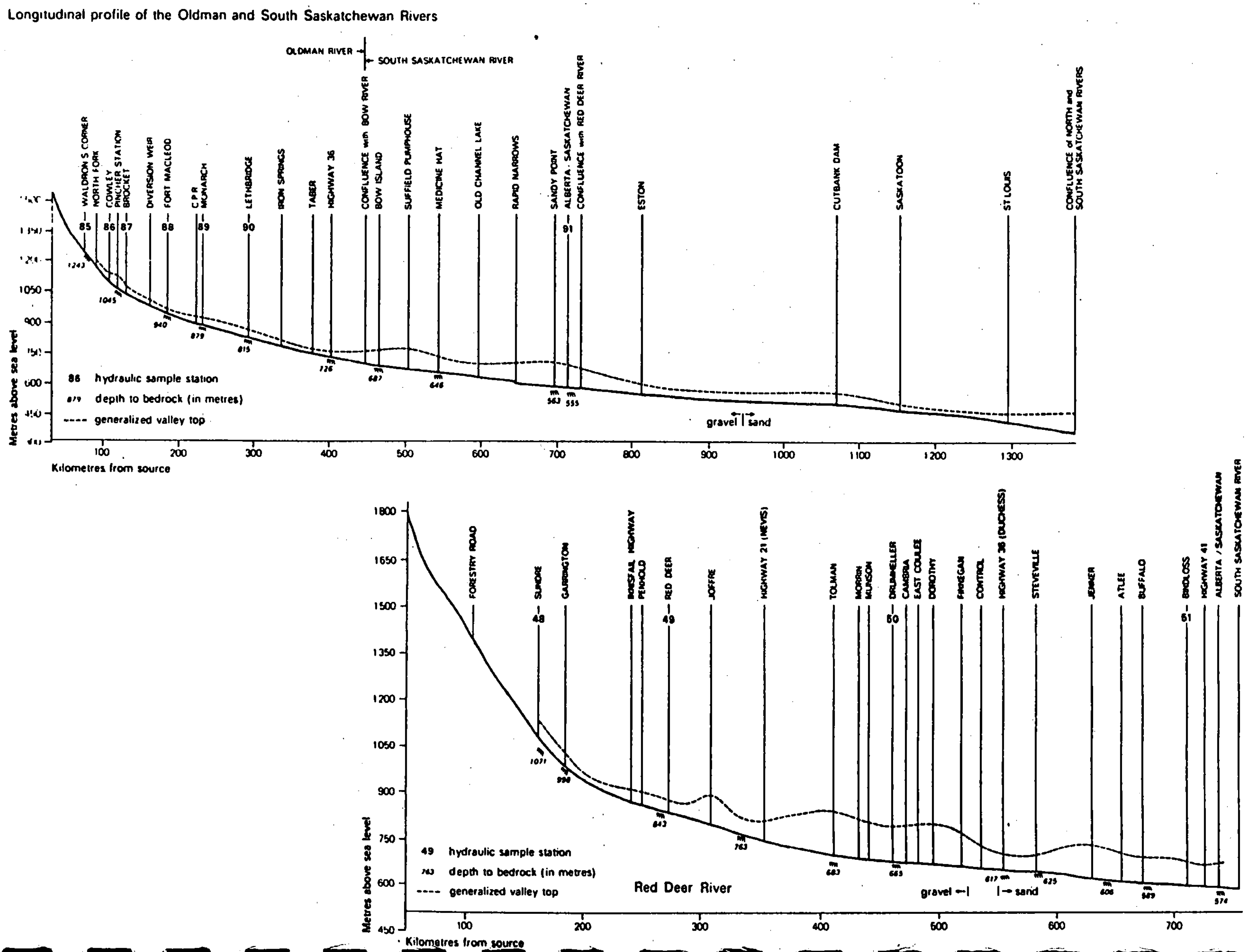


Figure 8.6 North Saskatchewan River long profile and bed material size (from Shaw and Kellerhals, 1982).

Figure 8.7 Long profiles of Oldman - South Saskatchewan River and Red Deer River (from Shaw and Kellerhals, 1982).



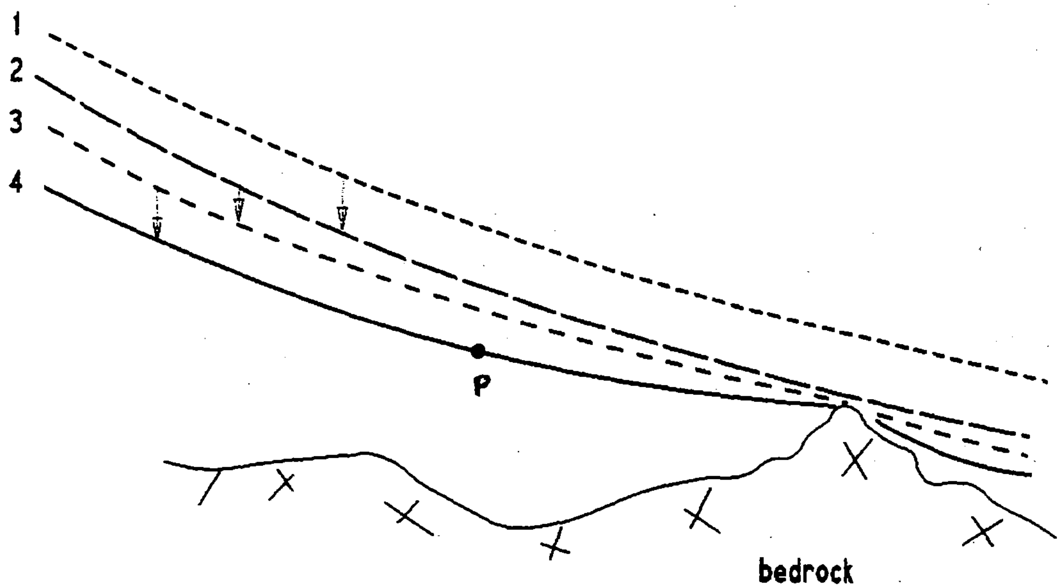


Figure 8.8 Base level effect of resistant outcrop on stream long profile. Four stages in schematic pattern of downcutting.

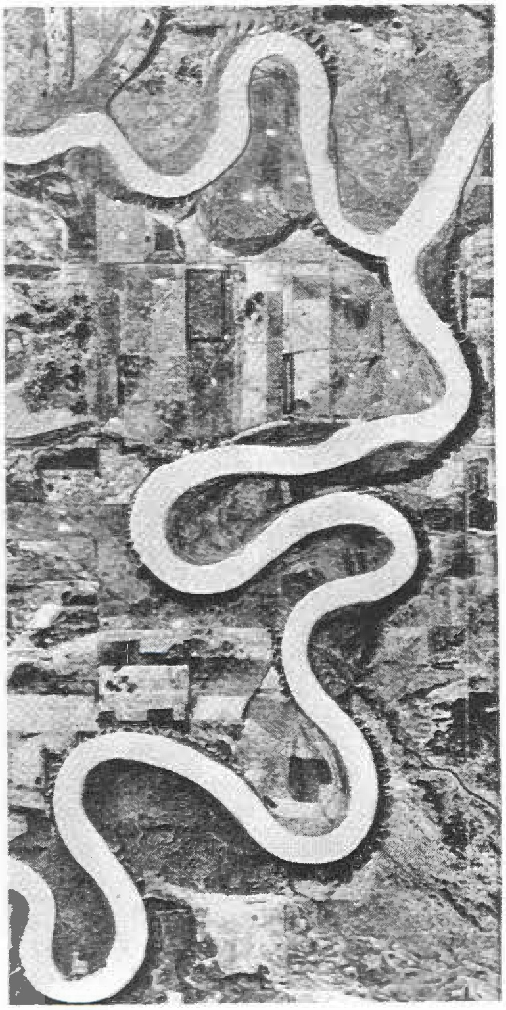


Figure 8.9 Meandering style of North and South Saskatchewan Rivers at their confluence (from Mollard, 1973).

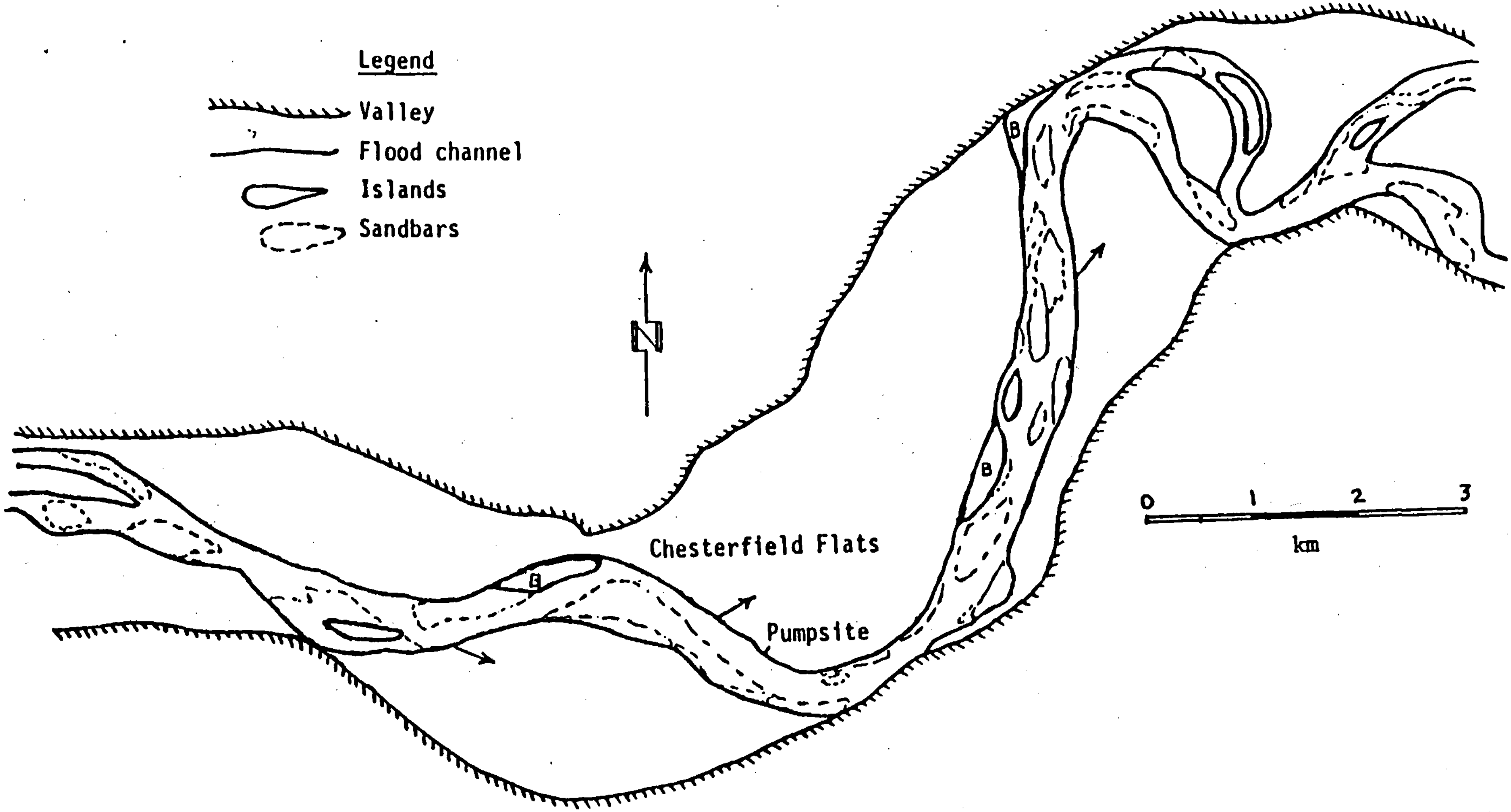


Figure 8.11 Map of south Saskatchewan River and valley at Chesterfield Flats (from Smith and Wigham, 1990).

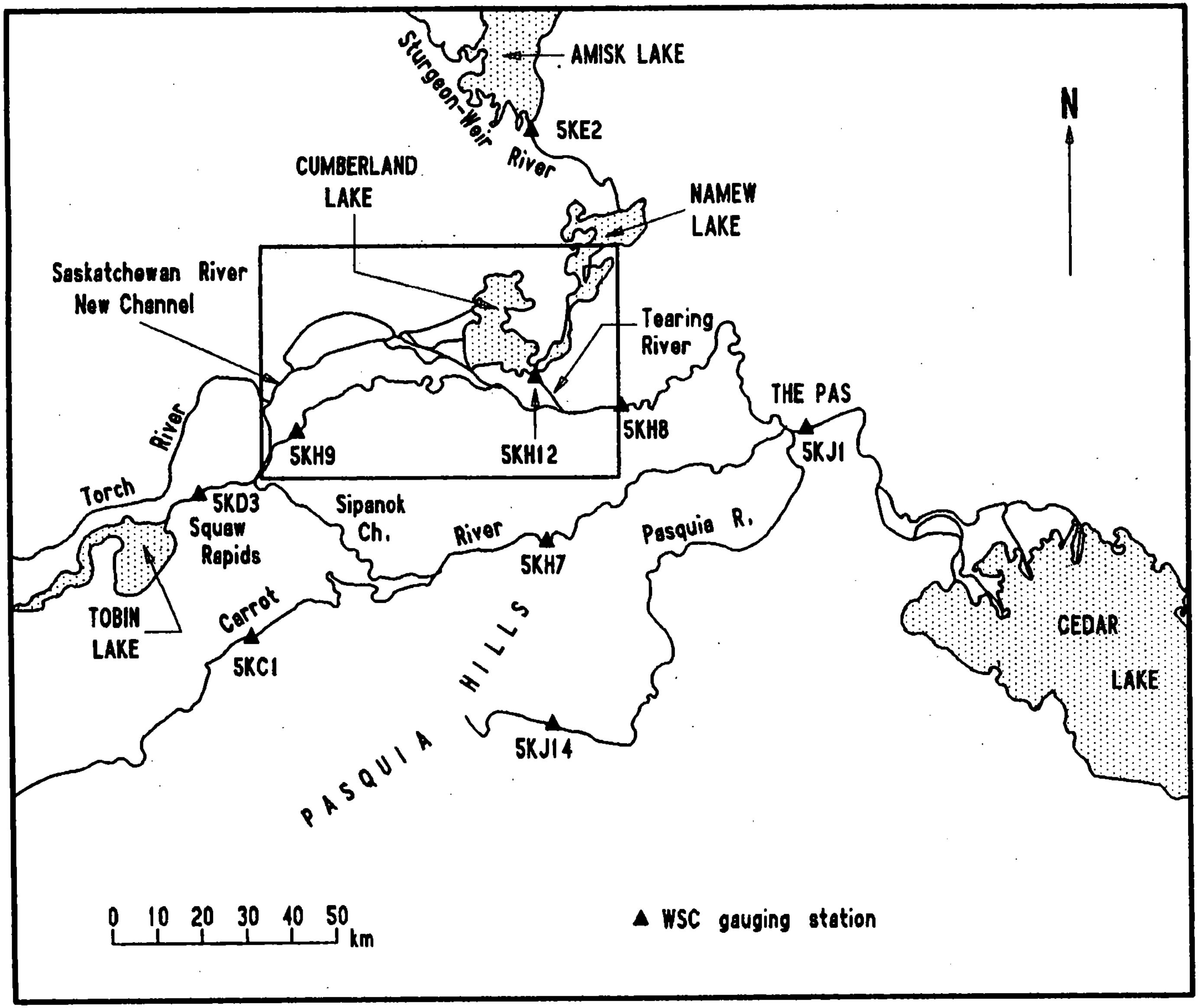


Figure 8.12 Map of Saskatchewan River Delta complex.

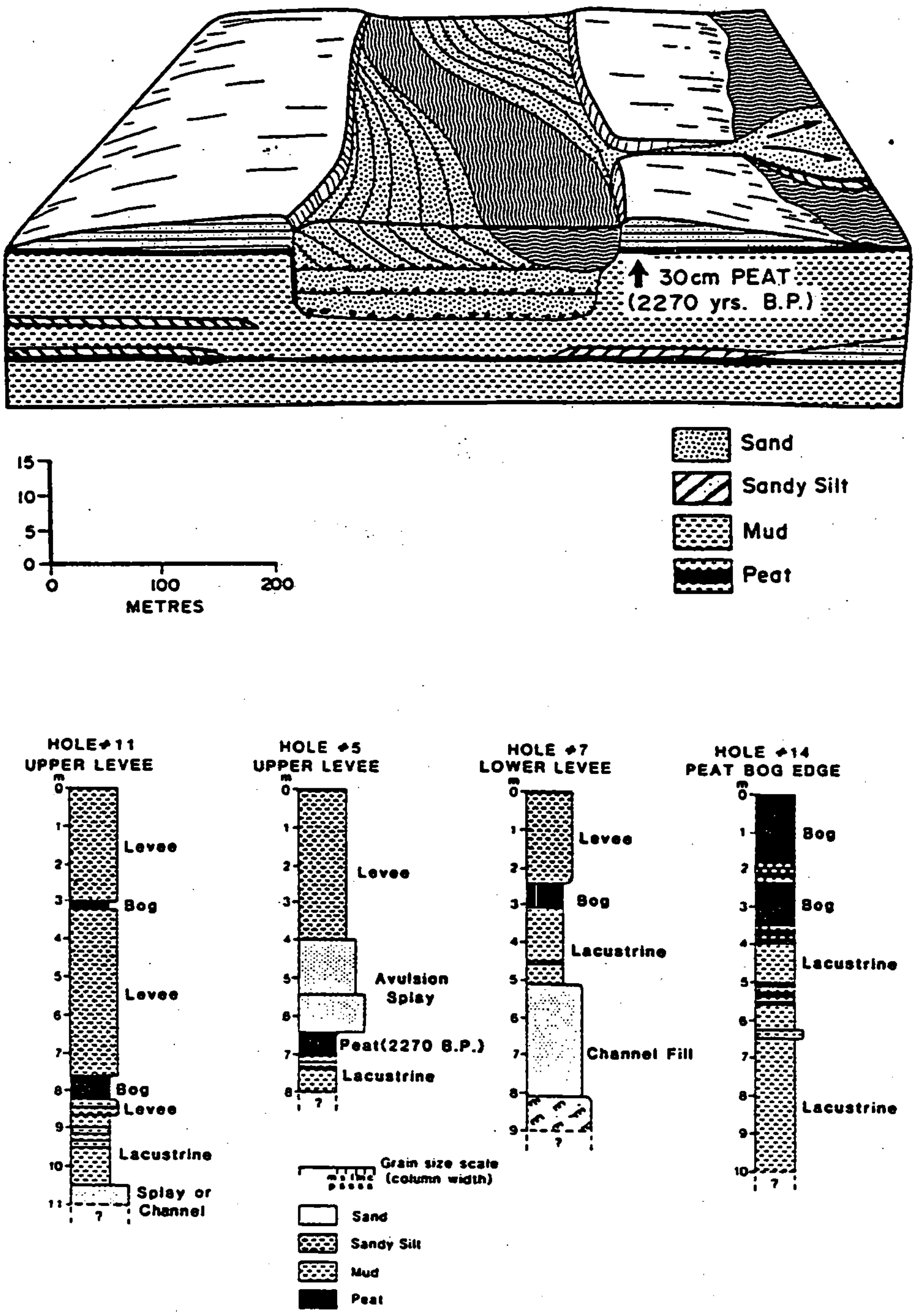


Figure 8.13 Stratigraphy of sediments beneath floodplain of Saskatchewan River delta complex (from Smith, 1983). Top figure is schematic representation of post-1875 infilling of old channel; arrow denotes site of core hole #5 below.

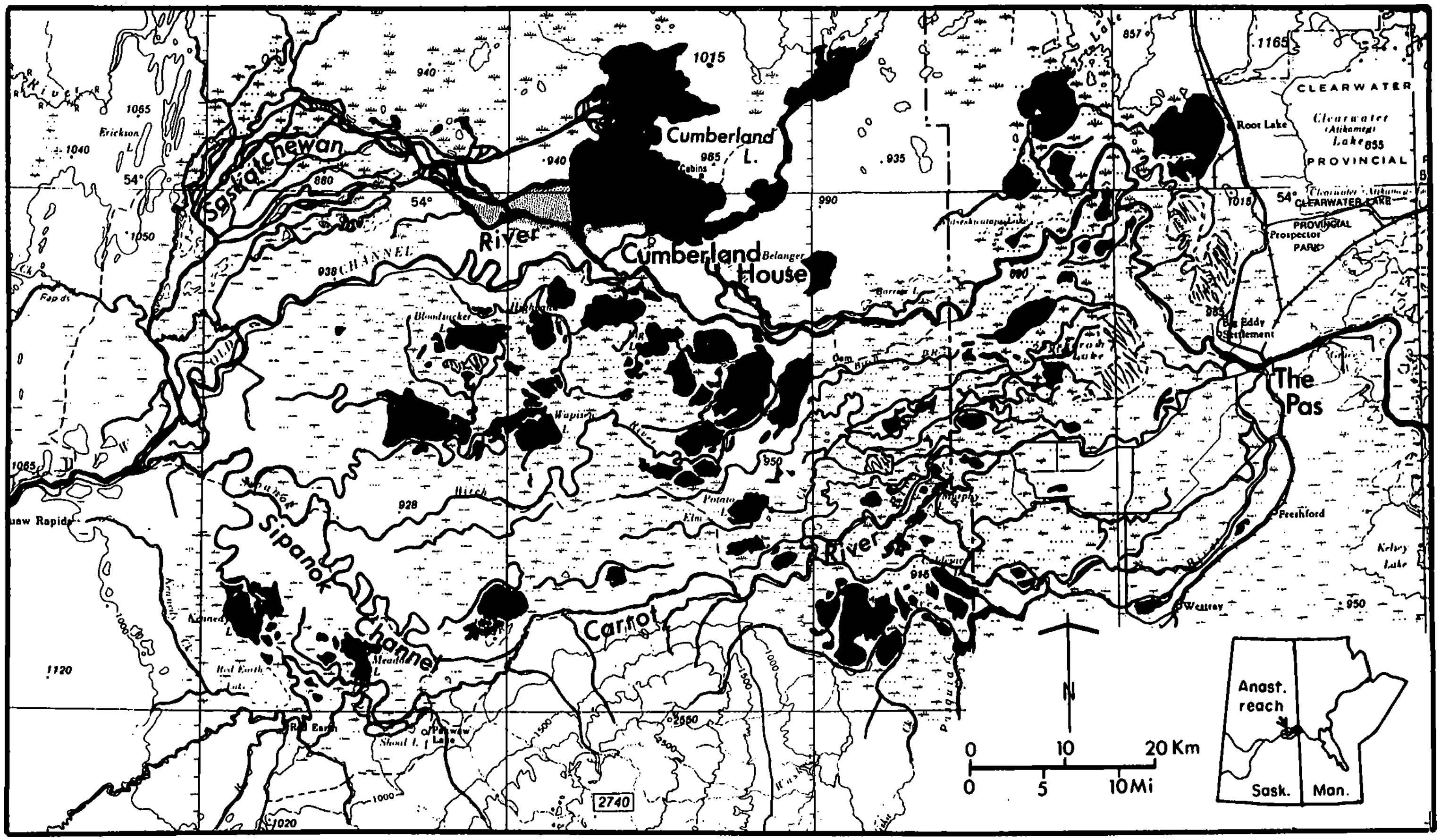


Figure 8.14 Anastomosed channel complex of lower Saskatchewan River (from Smith, 1983).

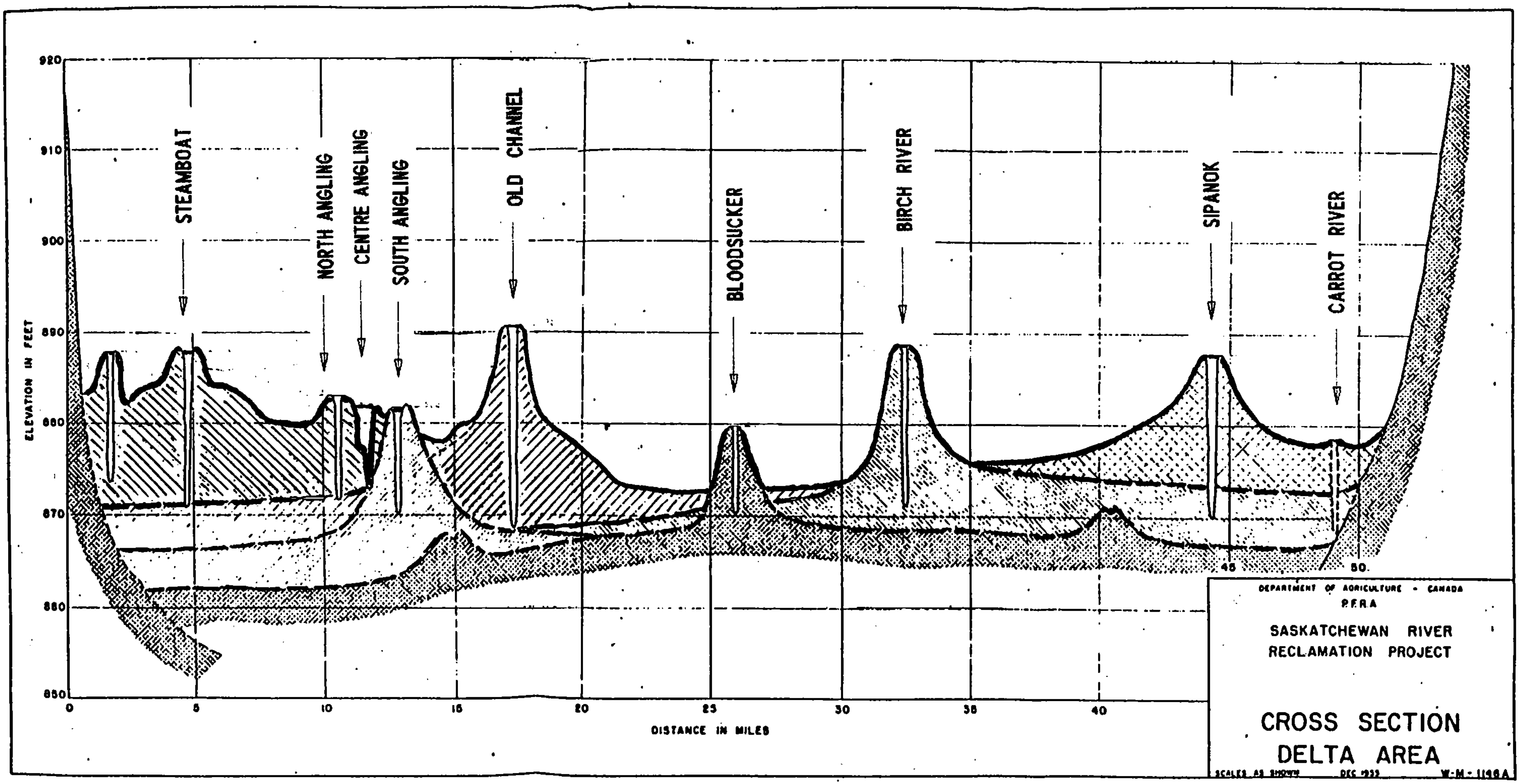


Figure 8.15 Saskatchewan River delta complex: floodplain cross-section (from Kuiper, 1960).

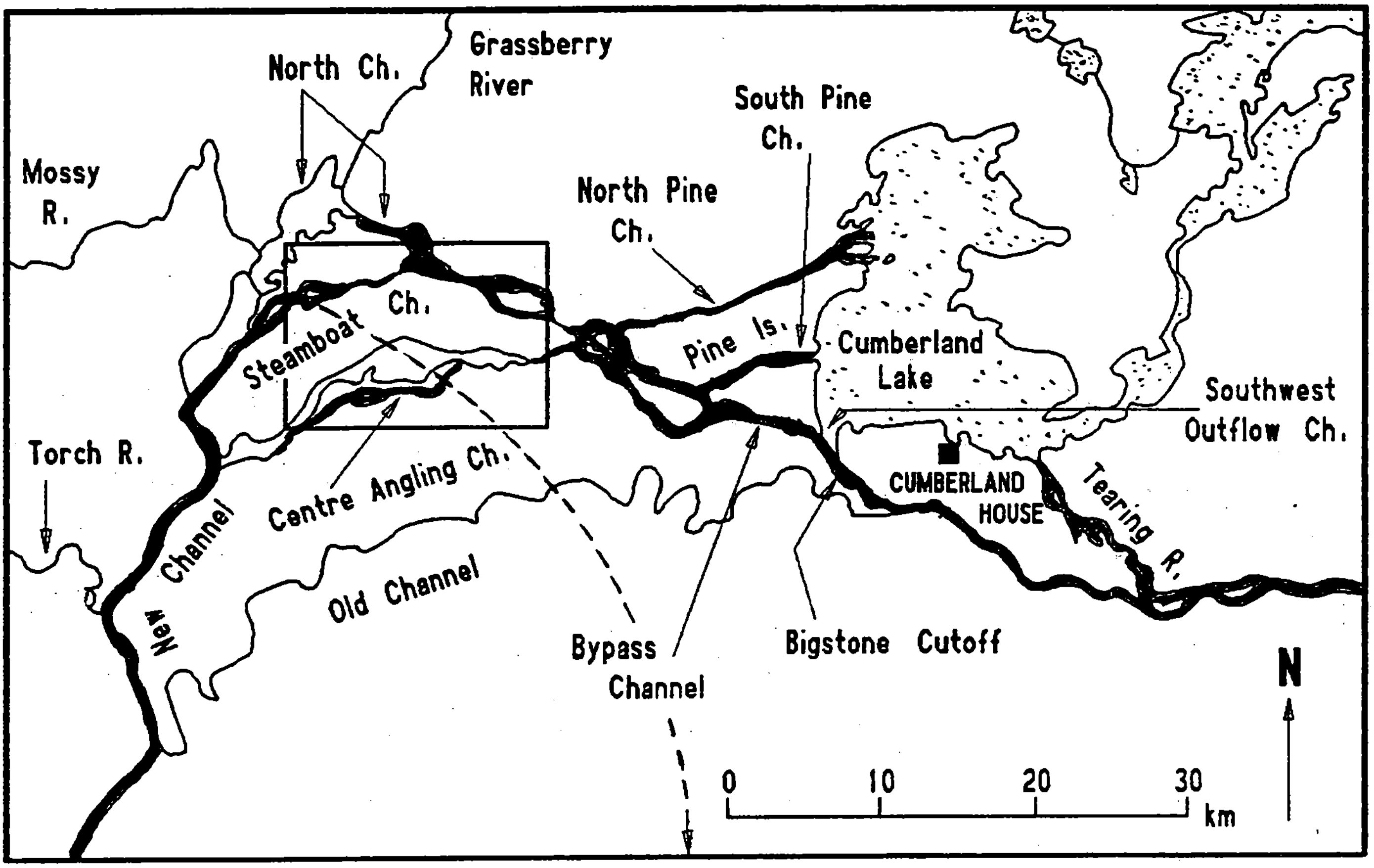
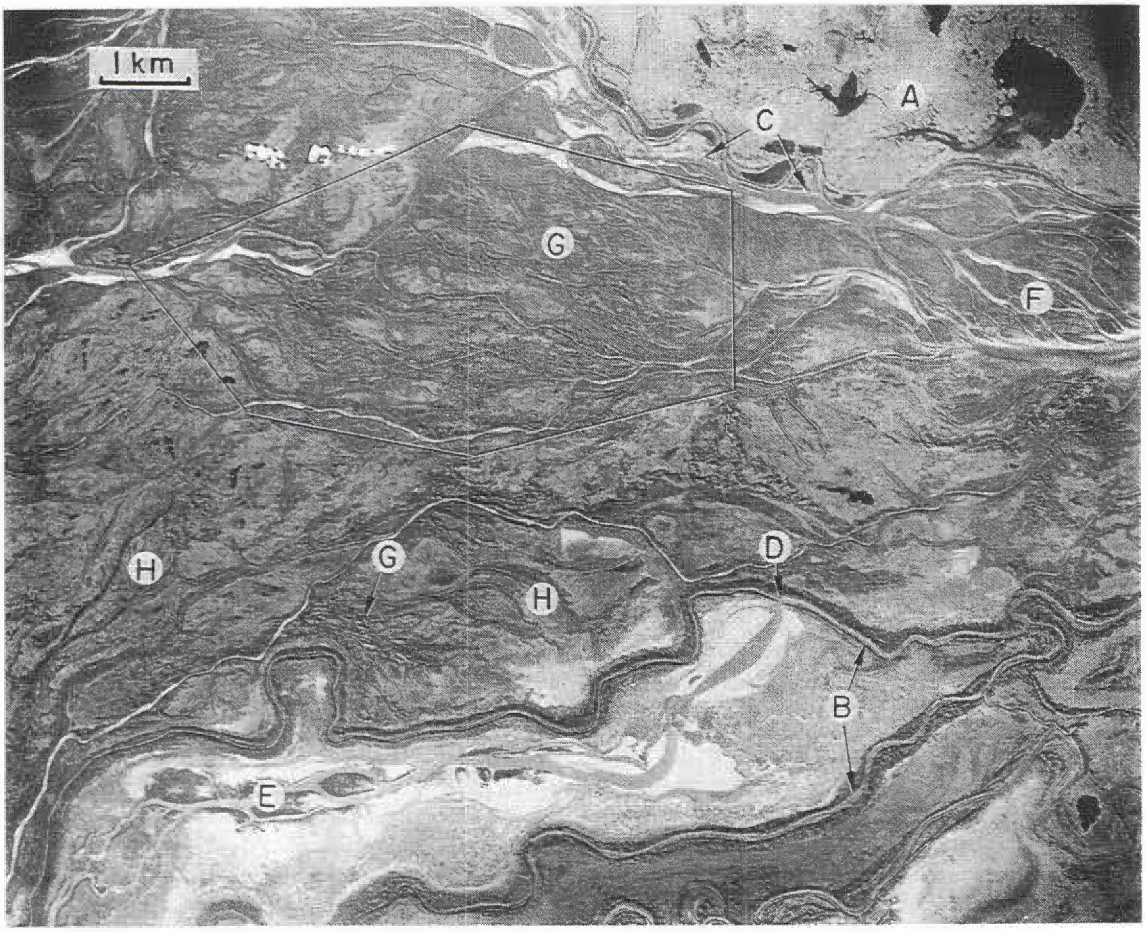


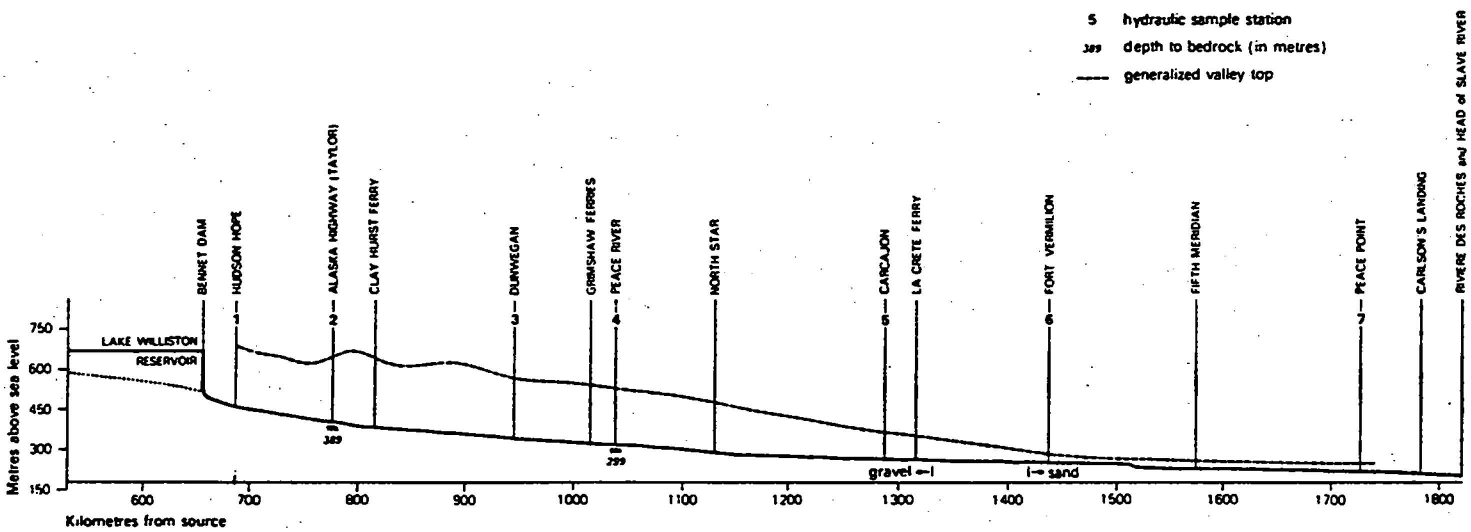
Figure 8.16 Saskatchewan River distributary network: Cumberland Lake area. Inset denotes location of photograph shown in Figure 8.17.



1953 aerial photograph showing portion of breakout area . Flow is generally from left to right. (A) Wetland terrain still unaffected by avulsion, including fen meadows and shallow lakes. (B) Channels active before the 1883 avulsion; note prominent forested levees. (C) Cutoffs in pre-avulsion channel (Mossy River) caused by increased discharge supplied by avulsion. (D) Widening of pre-avulsion channel (to right of arrow) due to capture of Centre Angling Channel . (E) Centre Angling Channel, a relatively large channel formed recently in wetland basin between levees of two older (pre-avulsion) channels. Note active crevasse splays and the irregular and rather indistinct channel boundaries, indications that the channel is still actively evolving. (F) Active Stage II splay complex. (G) Abandoned Stage II splay complexes; note multiple crevasse channels leading from trunk channels. (H) Abandoned Stage III splay complexes.

Figure 8.17 Aerial photograph (1953) of portion of anastomosed floodplain of Saskatchewan River delta showing newly formed Centre Angling Channel (from Smith et al., 1989).

Longitudinal profile of the Peace River



Longitudinal profile of the Athabasca River

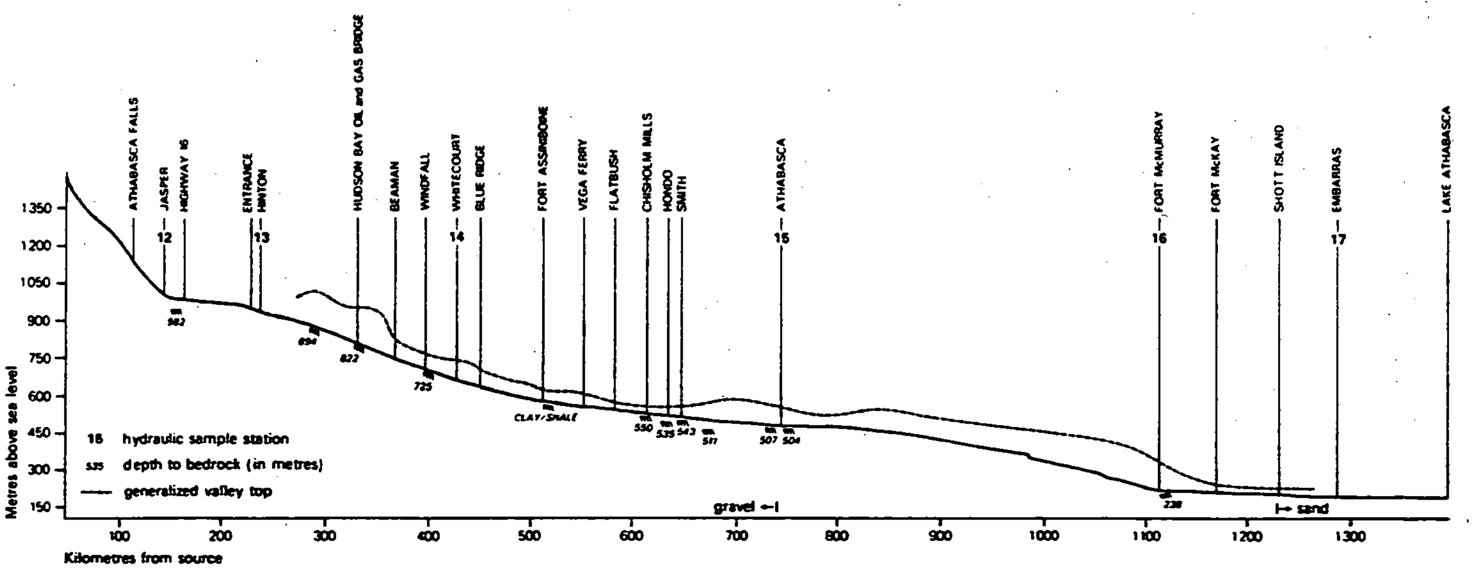


Figure 9.1 Long profiles of Athabasca and Peace Rivers (from Shaw and Kellerhals, 1982).

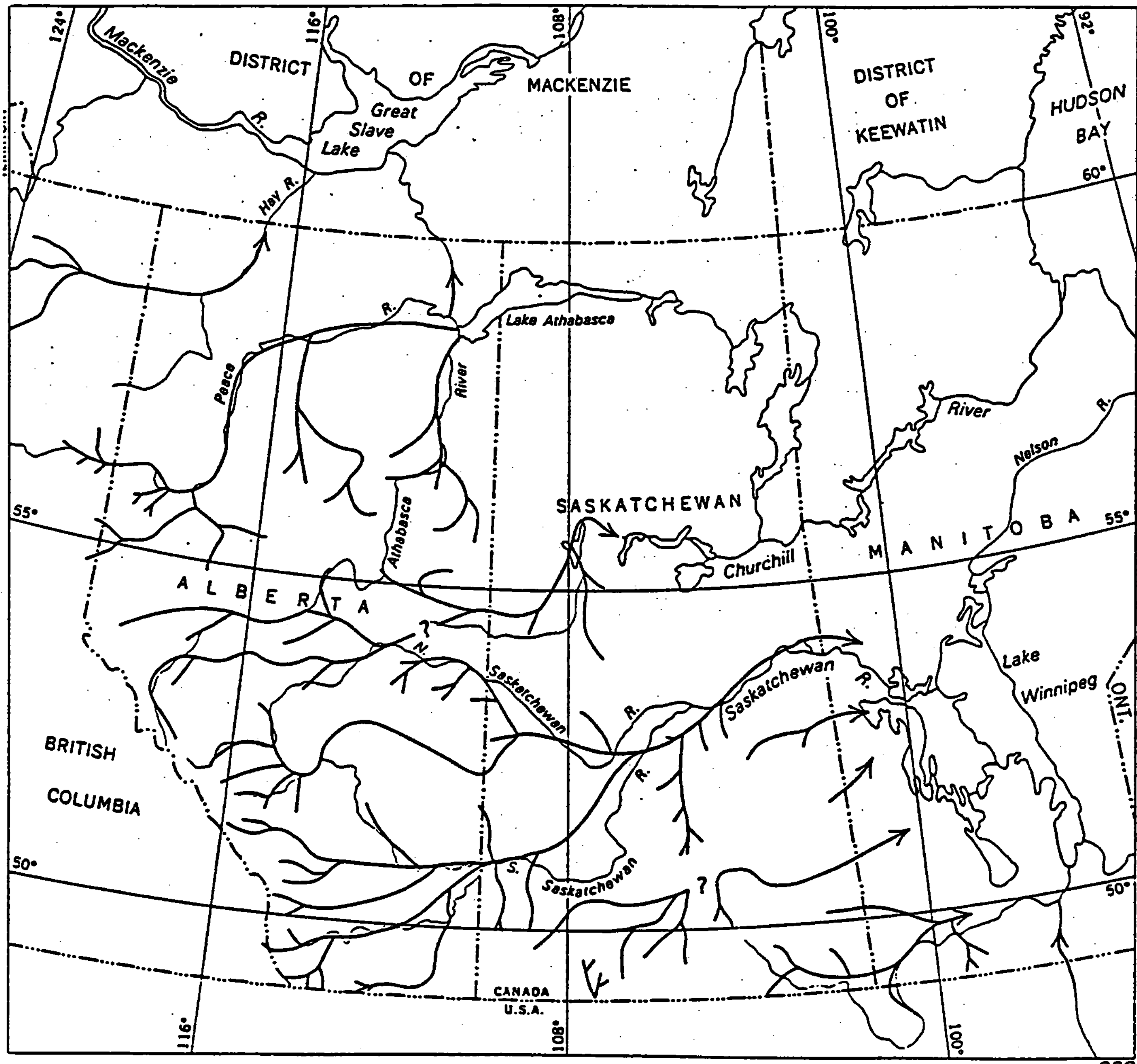
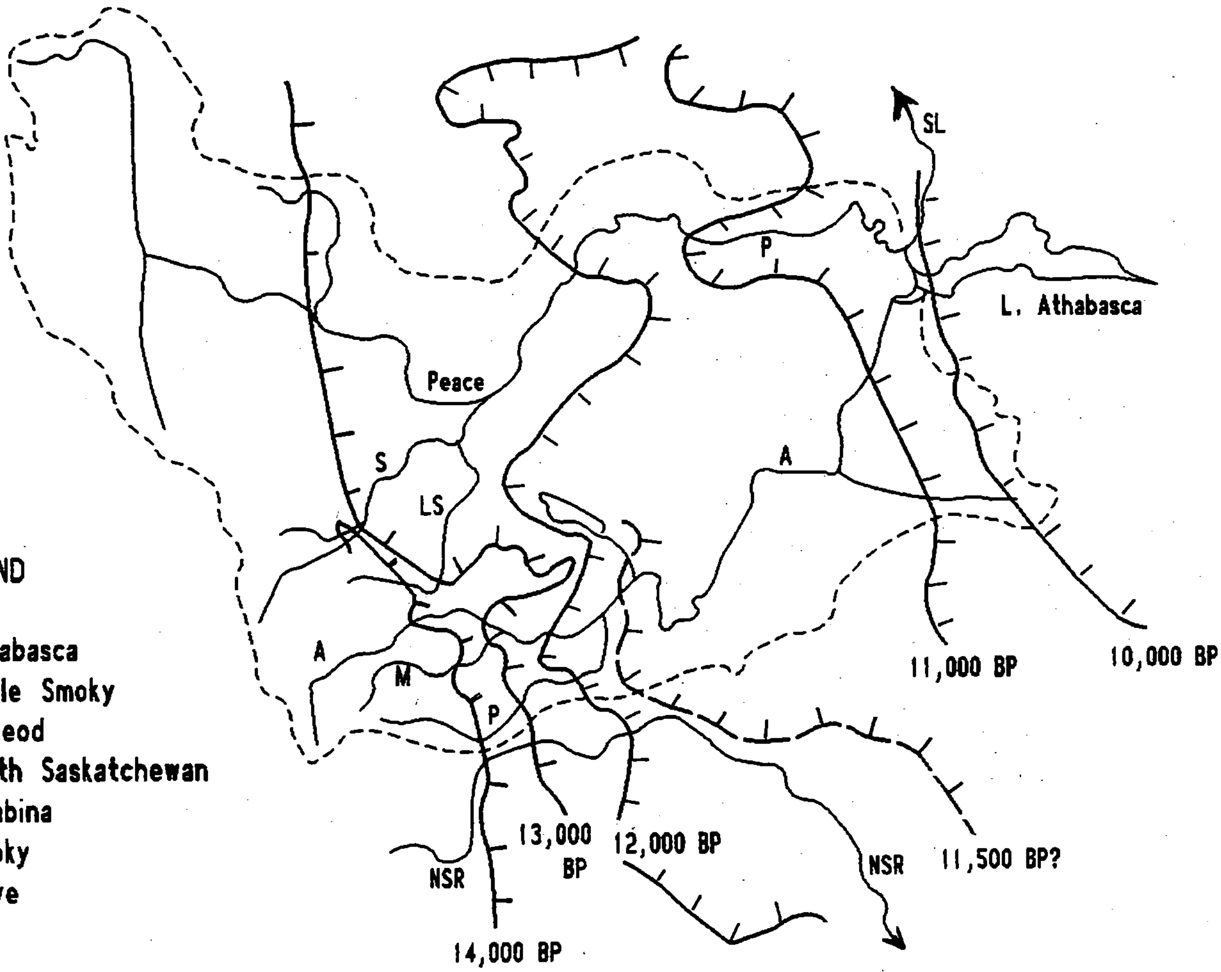


Figure 9.2 Inferred preglacial drainage system of Alberta-Saskatchewan (from Prest, 1971).



LEGEND

A : Athabasca
LS : Little Smoky
M : McLeod
NSR : North Saskatchewan
P : Pembina
S : Smoky
SL : Slave

Figure 9.3 Approximate ice front positions in deglaciation of Athabasca-Peace drainage basin 14,000-10,000 BP (after Dyke and Prest, 1987b).

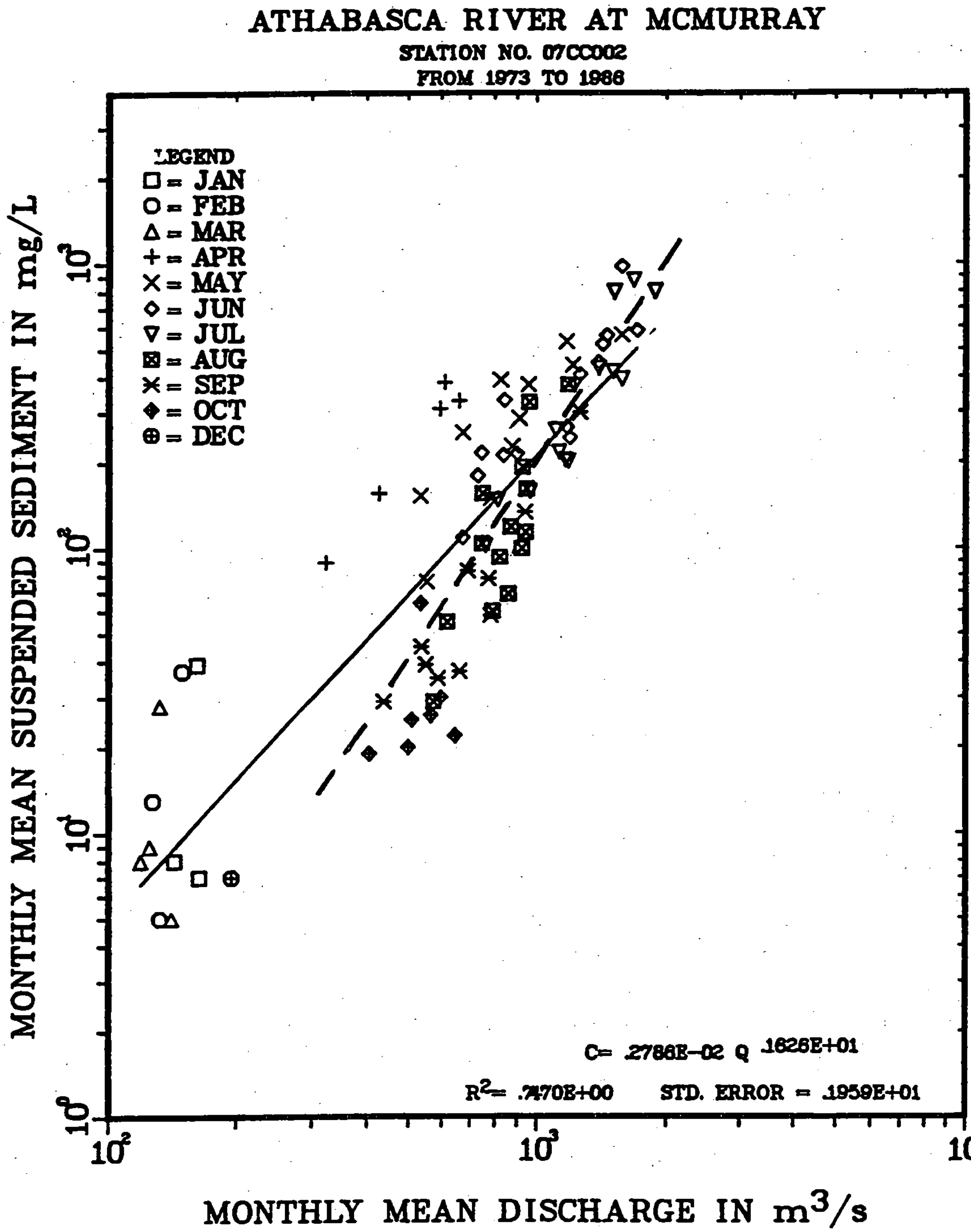
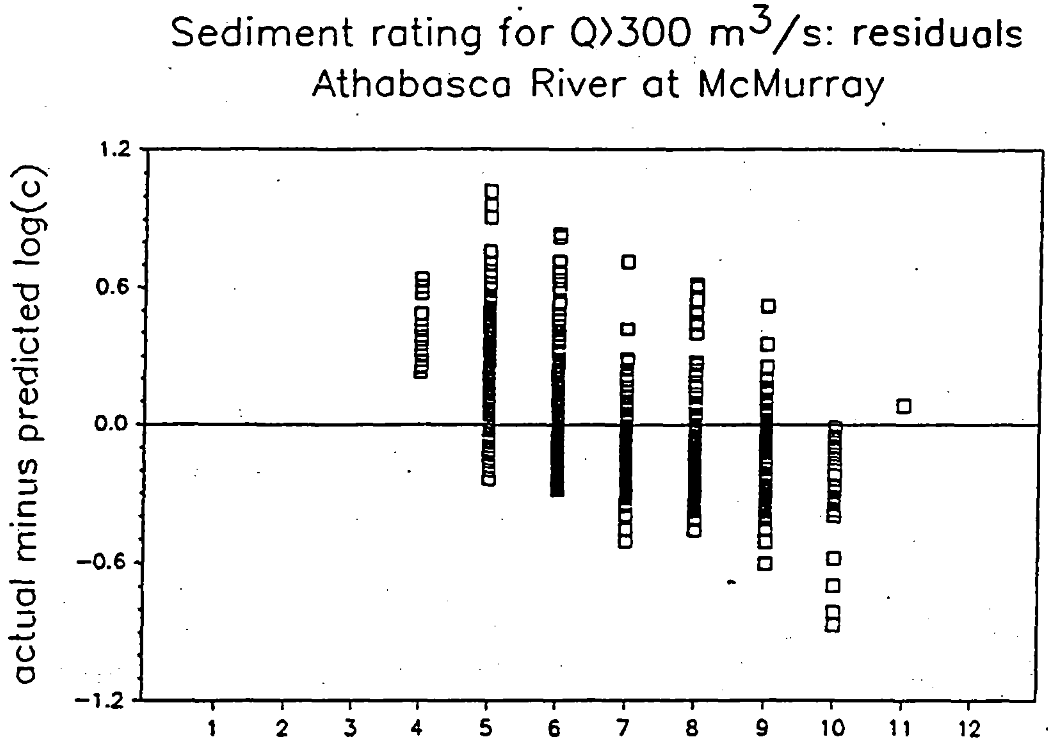


Figure 9.4 Sediment rating diagram for Athabasca River at McMurray.



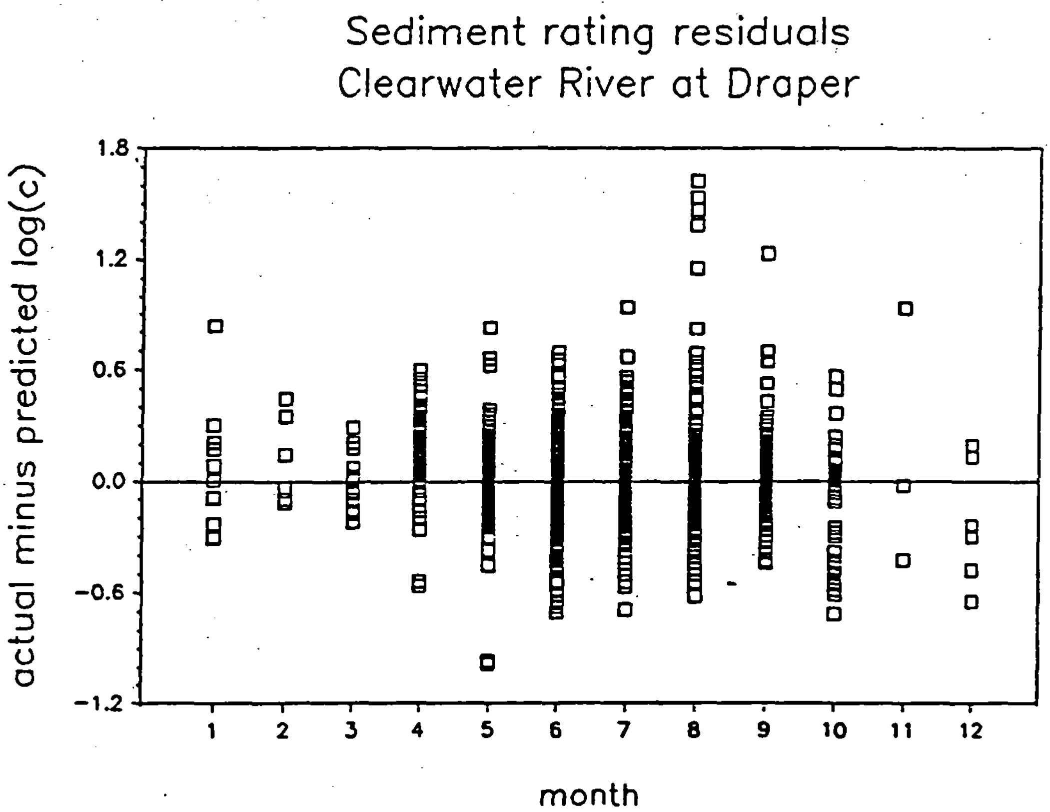


Figure 9.5 Sediment rating residuals plotted by month for (a) Athabasca River at McMurray and (b) Clearwater River at Draper.

Figure 9.6 Flood conditions on the Pembina River floodplain, downstream of Paddle River confluence, Alberta (from Mollard and Janes, 1984).

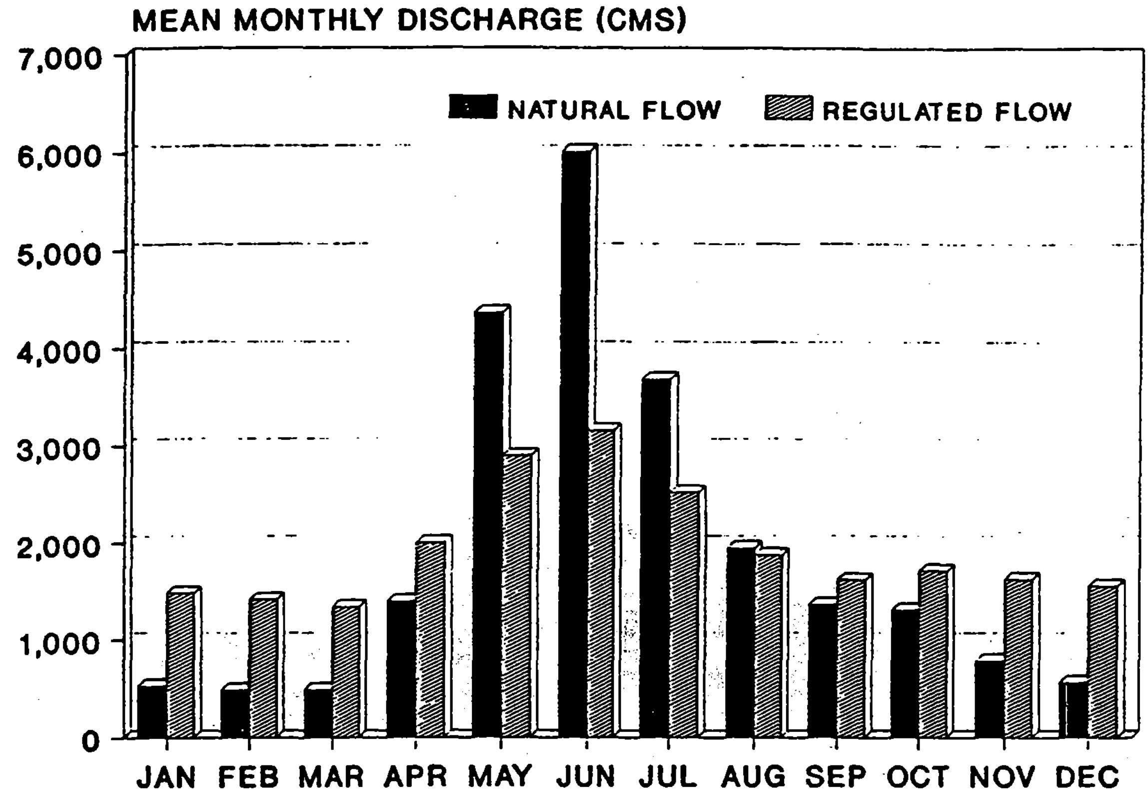


Figure 9.7 Natural and regulated flow regime of the Peace River at the town of Peace River, Alberta (from Shaw et al., 1990).

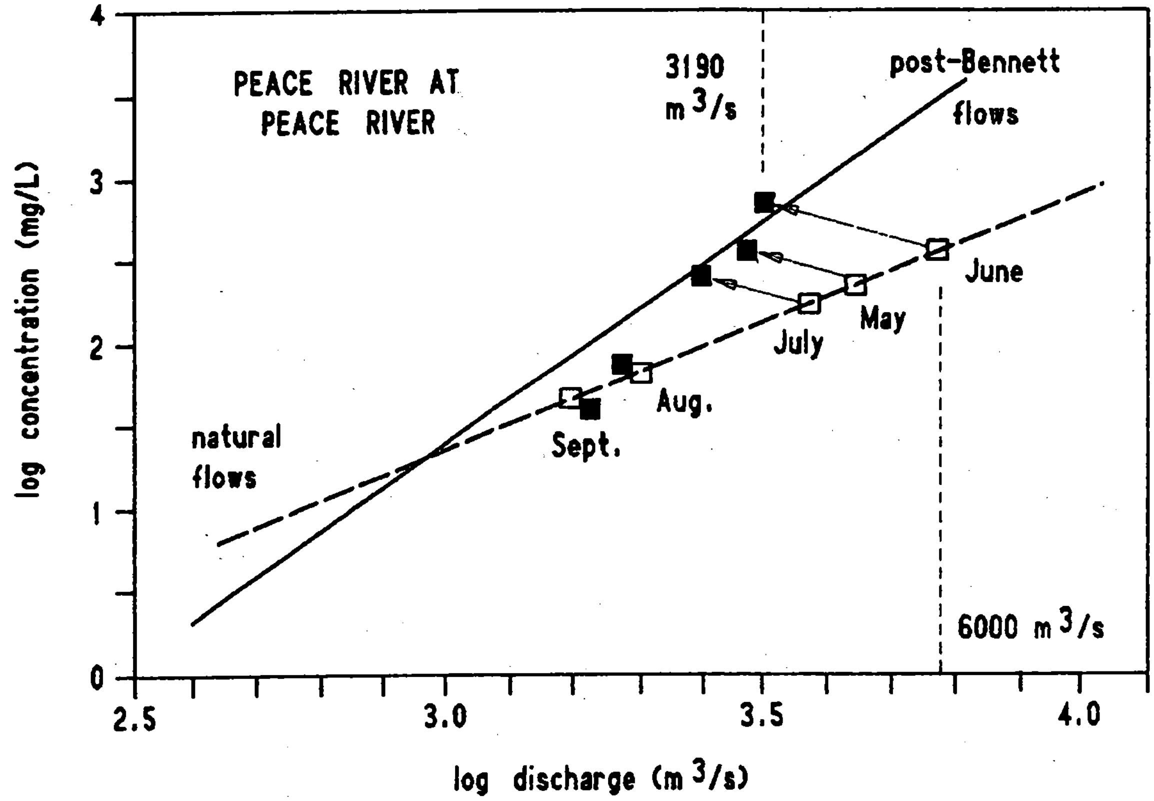


Figure 9.8 Sediment rating lines for Peace River at Peace River before and after flow regulation.

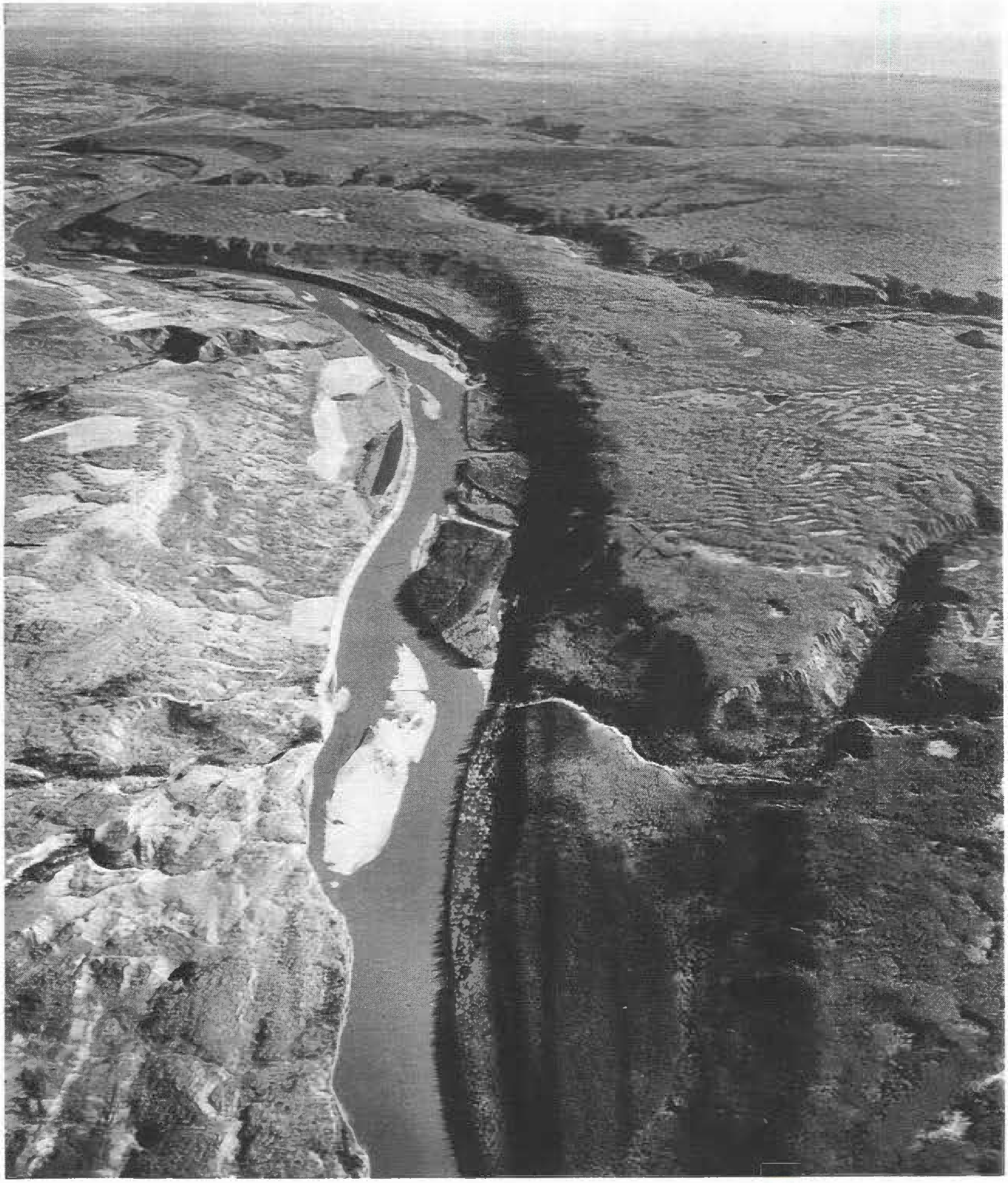


Figure 9.9 Oblique aerial photograph of Peace River looking downstream towards Moberly River confluence (B.C. air photo 1951:45).

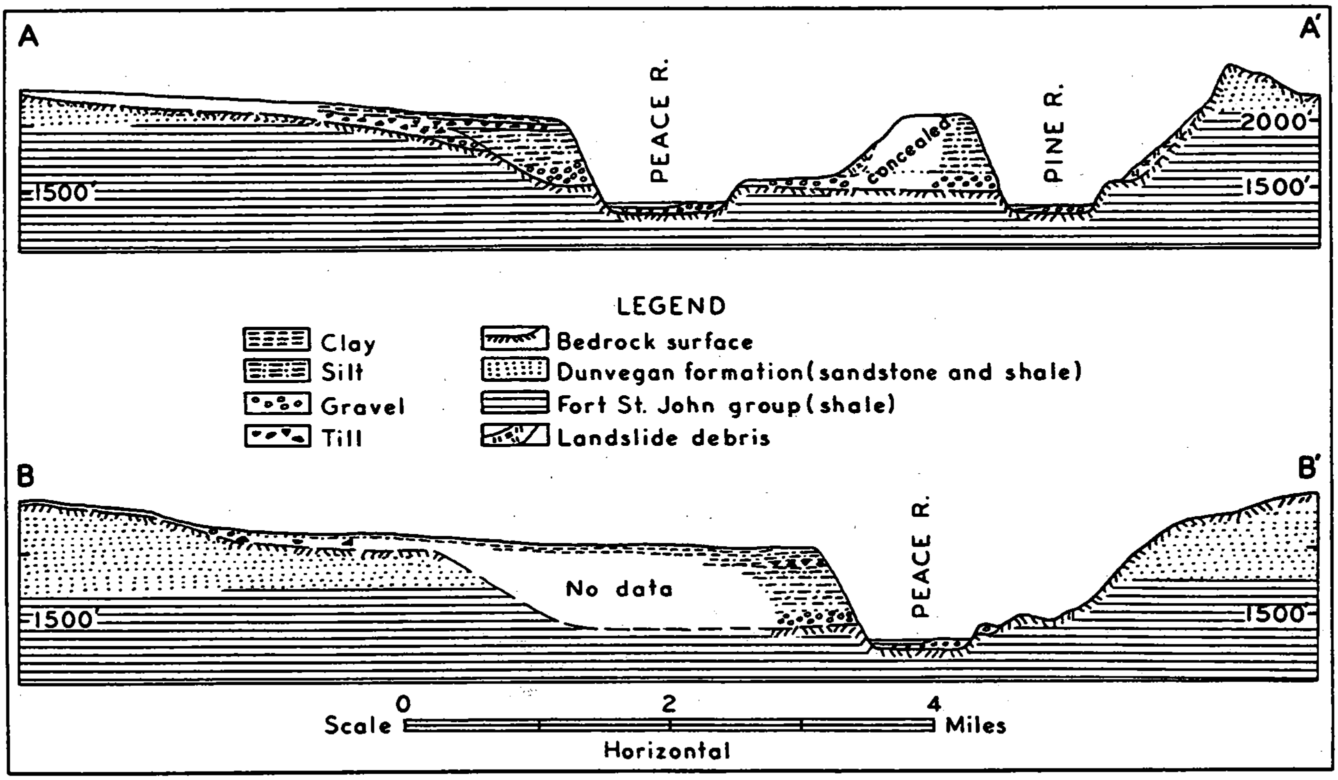


Figure 9.10 Stratigraphy of valley walls of Peace River near Taylor, B.C. (from Mathews, 1978).

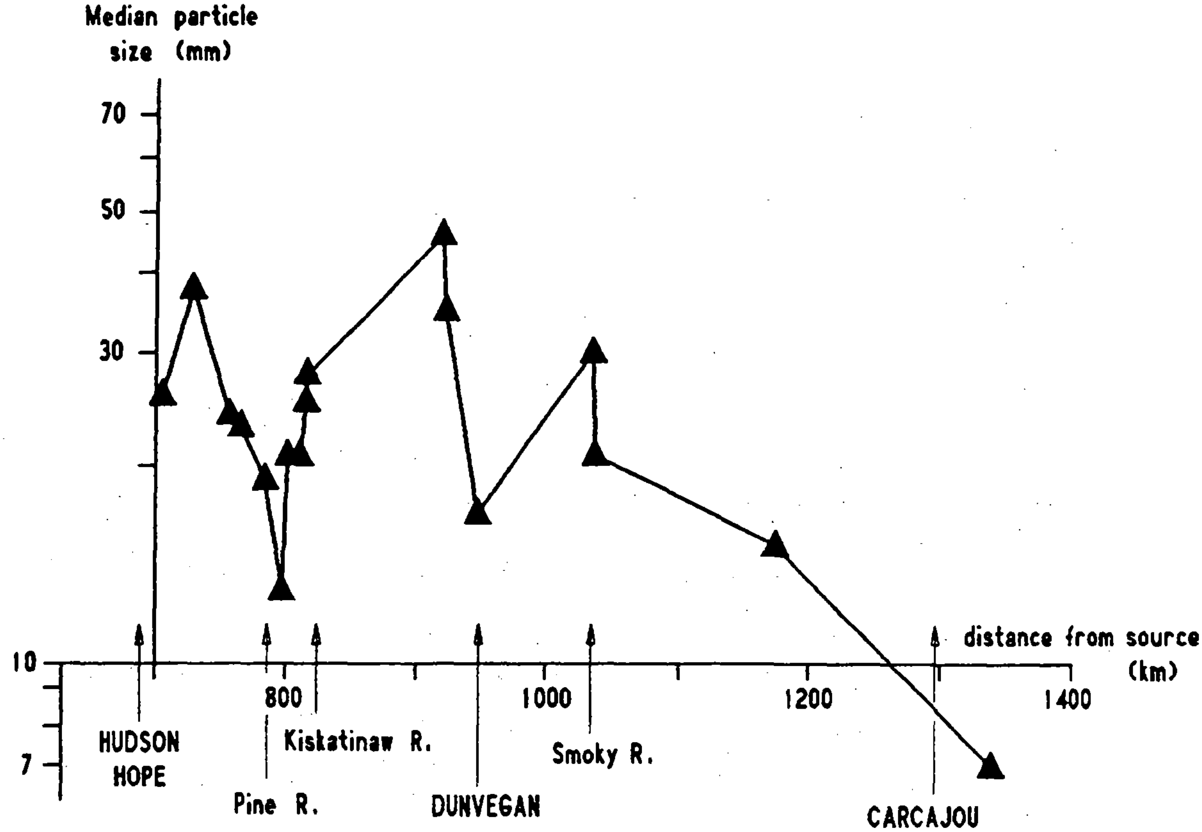


Figure 9.11 Changes in bed material size along Peace River between Hudson Hope (B.C.) and Carcajou (after Shaw and Kellerhals, 1982).

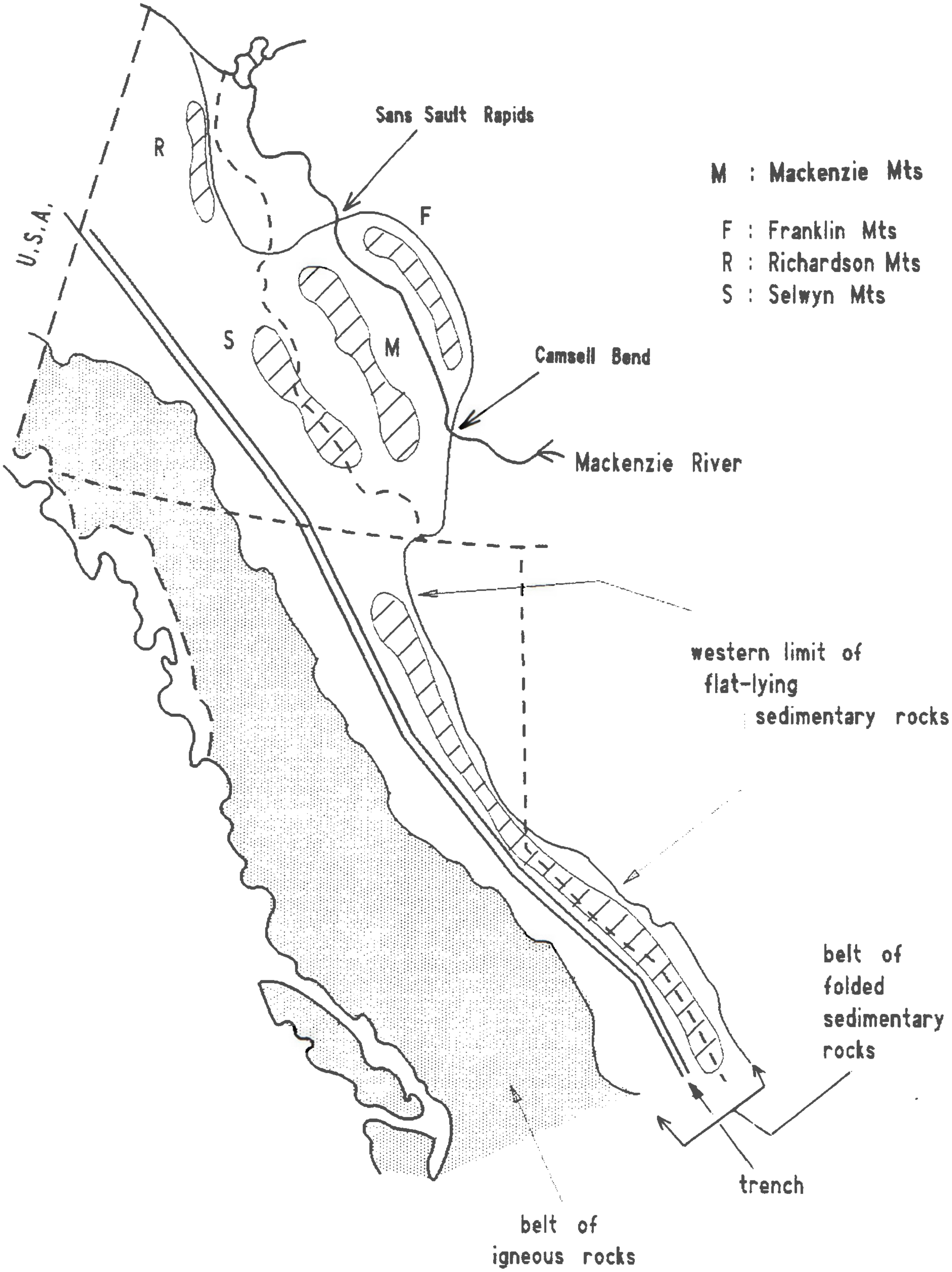


Figure 10.1 Schematic bedrock geology of Cordillera.

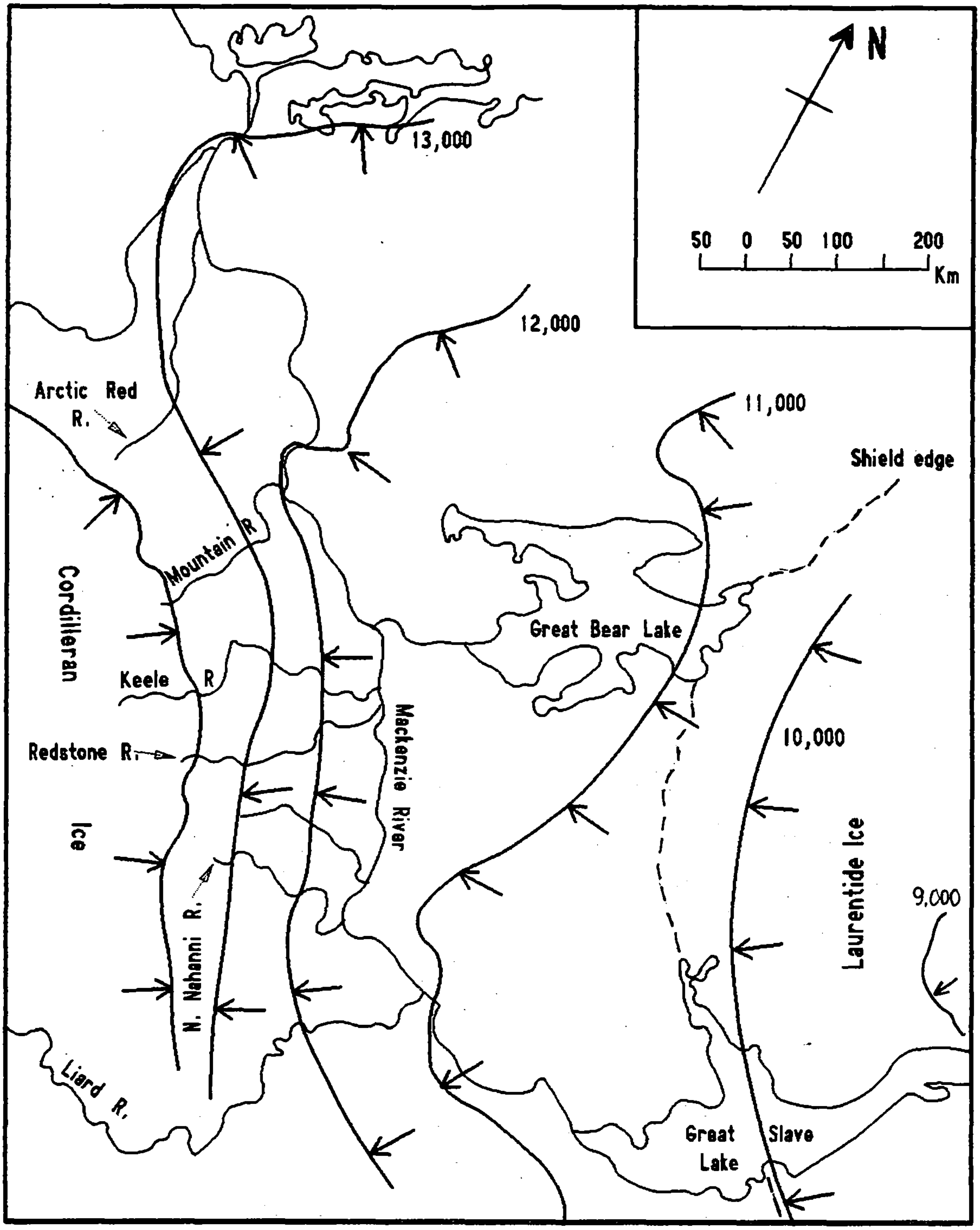


Figure 10.2 Approximate ice front positions in Mackenzie Basin, 13,000 to 9,000 years ago (after Dyke and Prest 1987a,b).

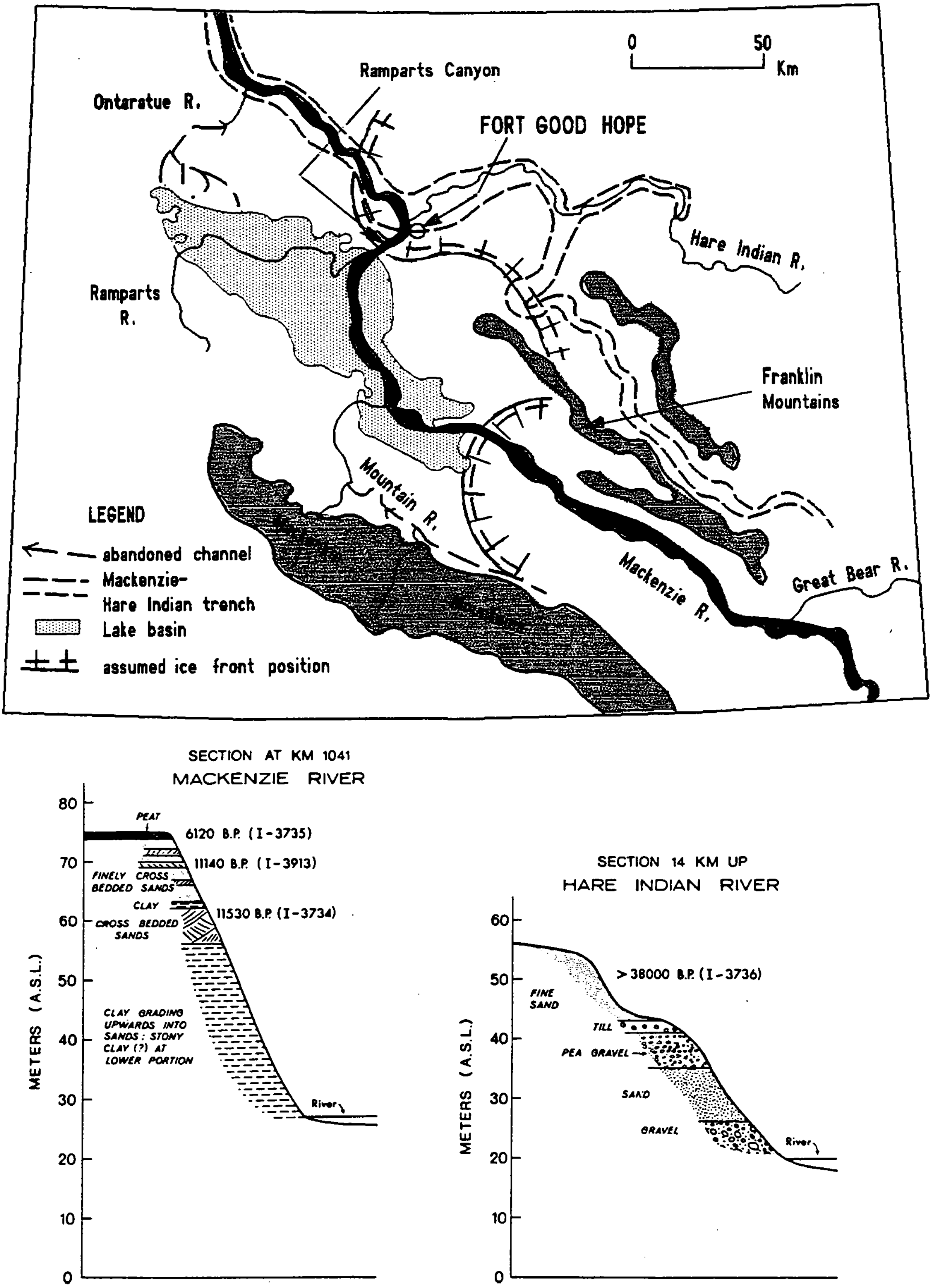


Figure 10.3 Late glacial drainage conditions and valley fill stratigraphy, Mackenzie River near Norman Wells (from MacKay and Mathews, 1973).

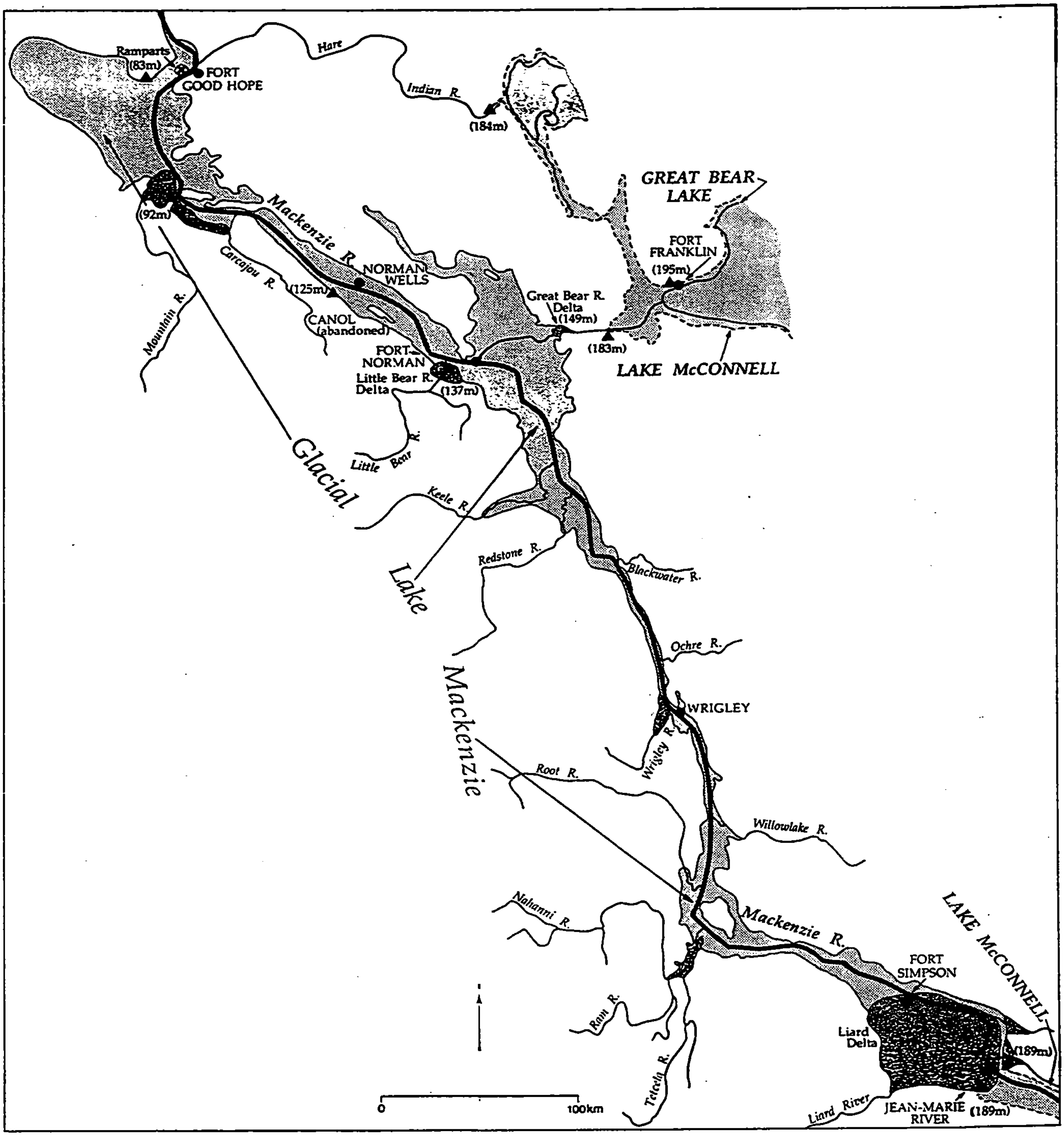


Figure 10.4 Extent of Glacial Lake Mackenzie (from Smith, 1989).

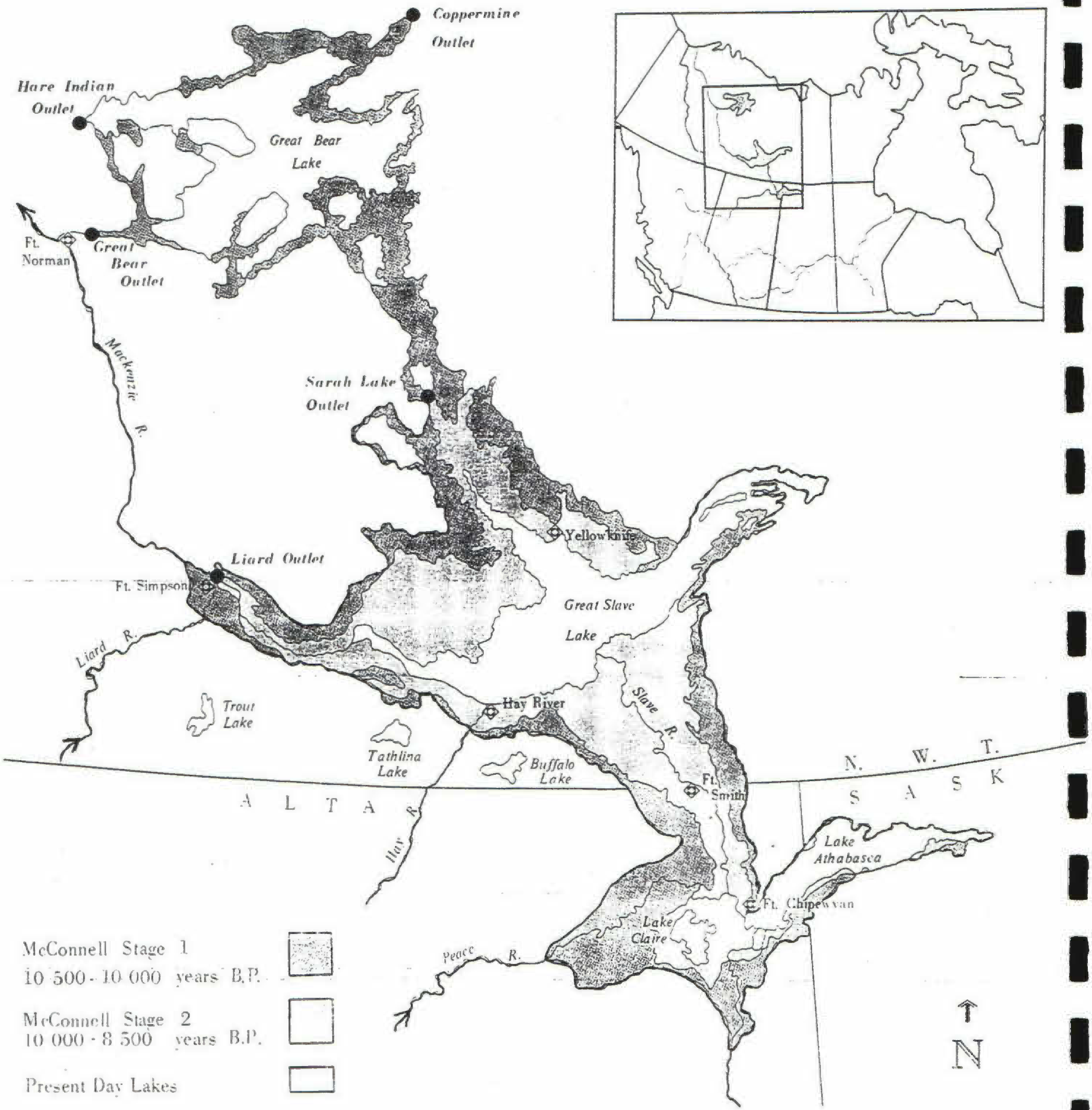
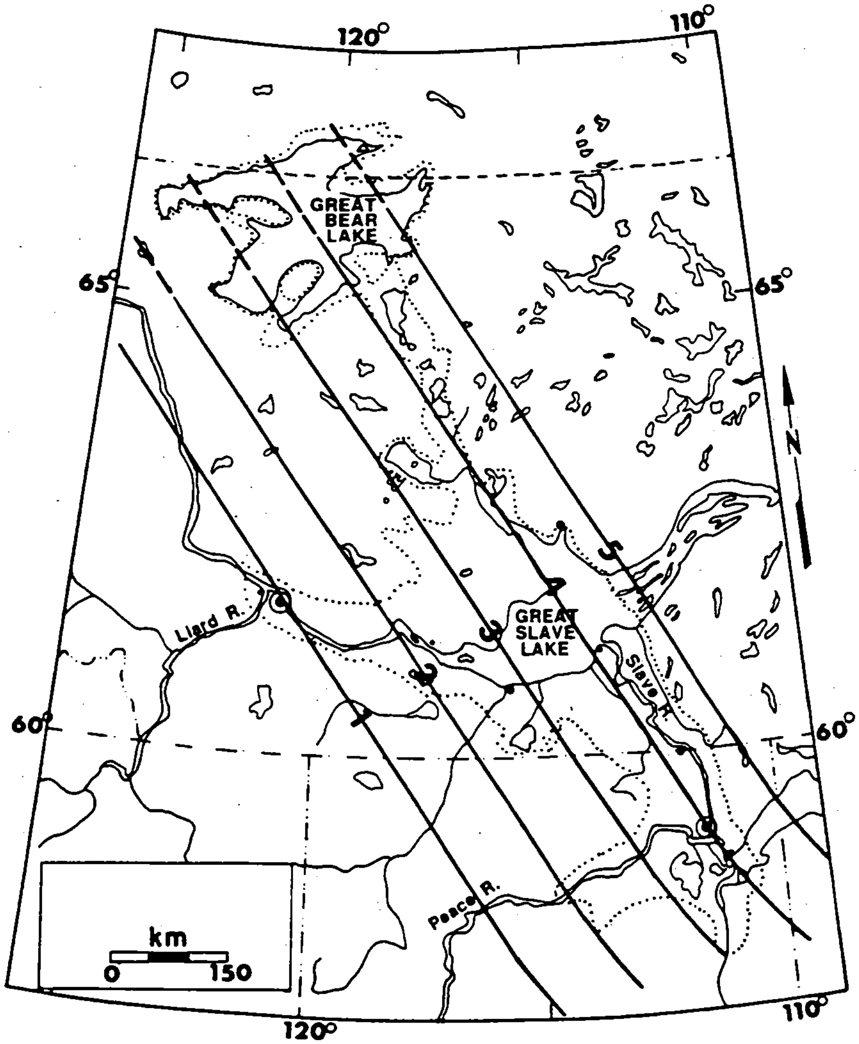


Figure 10.5 Extent of Glacial Lake McConnell (from Smith, 1991, pers. comm).



6 Lines of Equal Postglacial Uplift With Numbered Isobases At 100 km Intervals

Limit of Lake McConnell (Craig 1965)

Locations of Western Outlet and Peace-Athabasca Delta Surface

Extent of Glacial Lake McConnell (after Craig 1965) and rebound isobases between the Liard and Slave deltas. Rebound estimates are relative to the Liard Delta at Fort Simpson between 8780 BP and the present. North of Great Slave Lake the isobases are speculative.

Figure 10.6 Pattern of isostatic rebound in Glacial Lake McConnell area (from Vanderburgh and Smith, 1988).

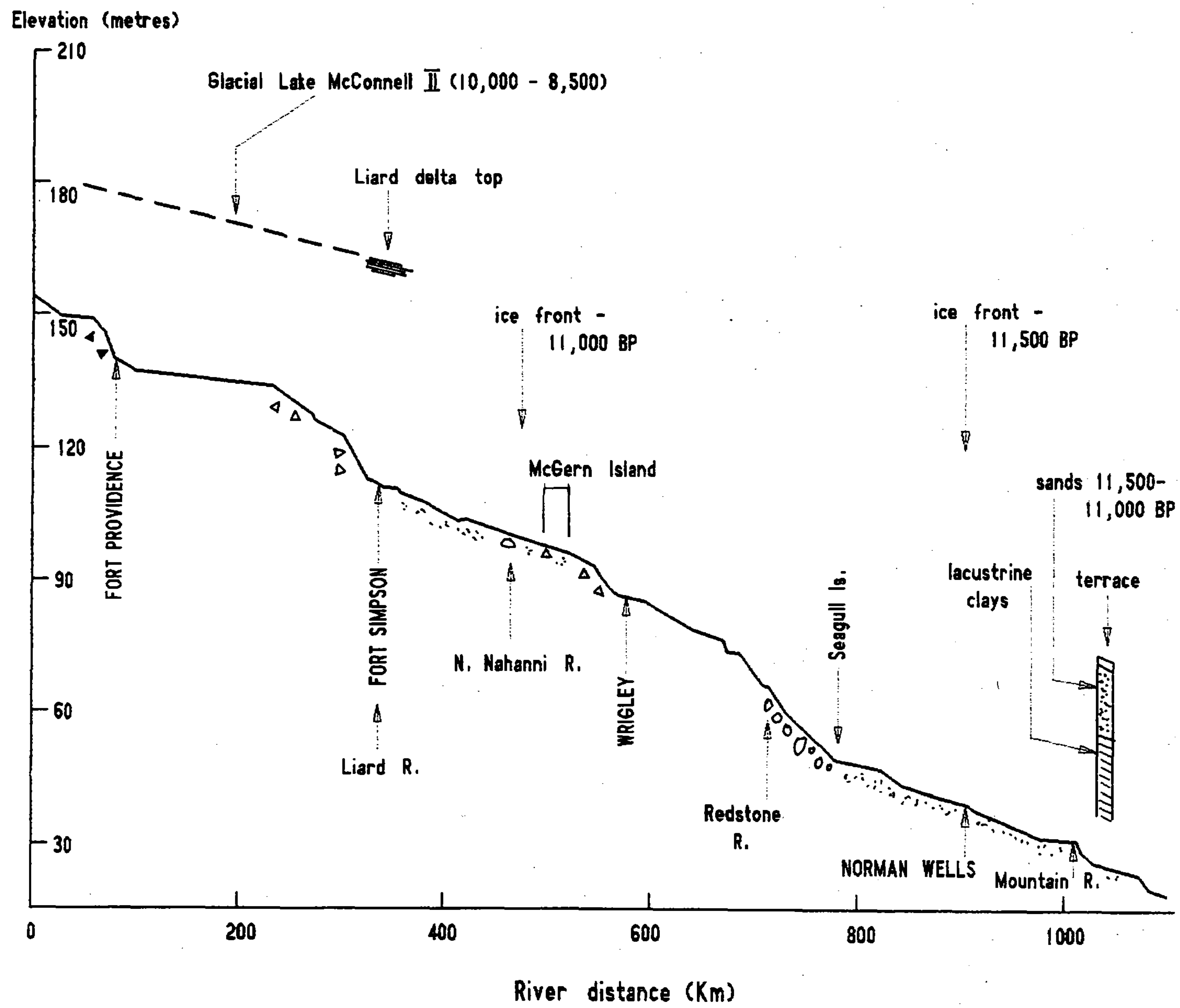


Figure 10.7 Long profile of Mackenzie River.

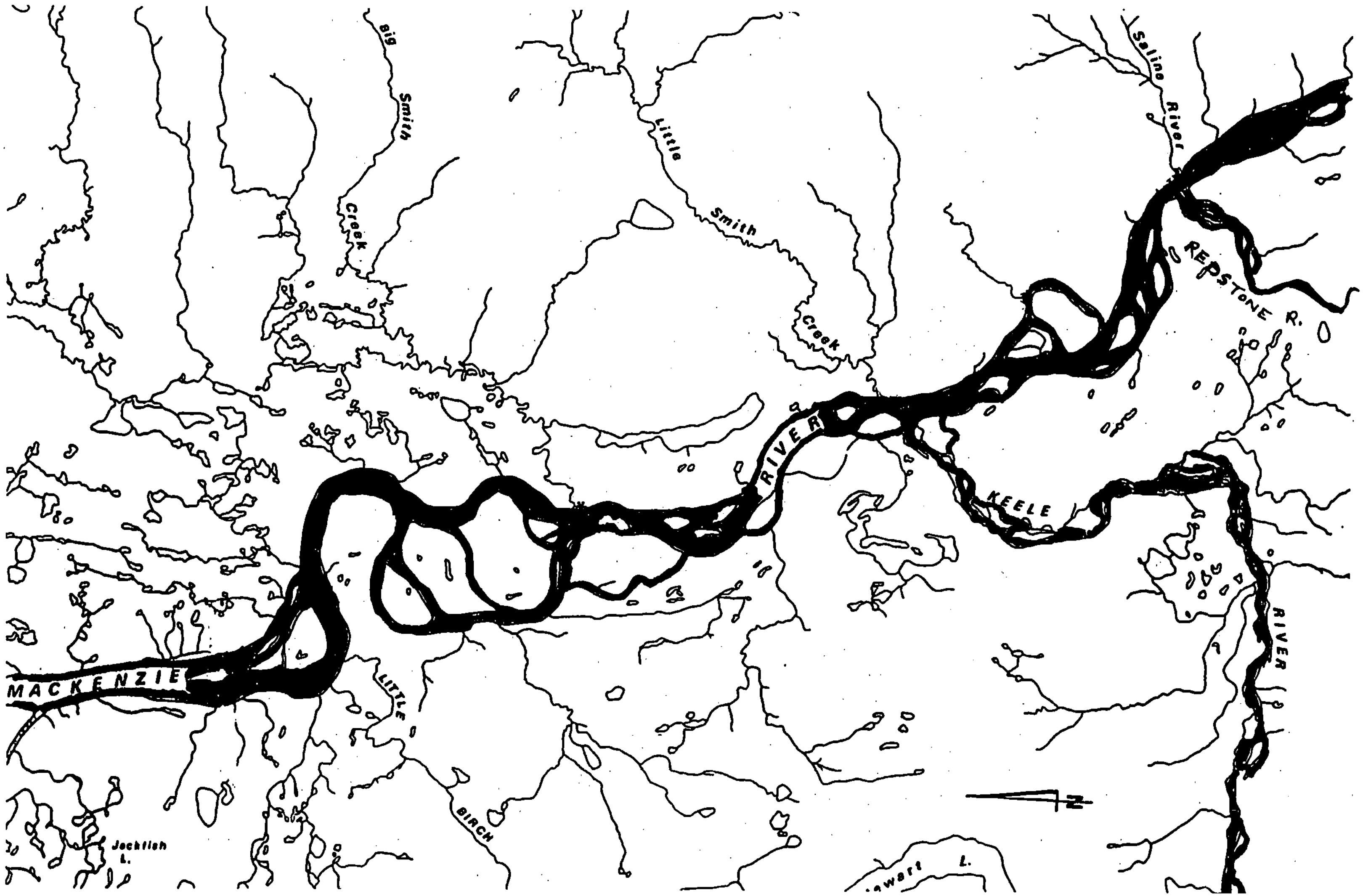


Figure 10.8 Morphology of Mackenzie River downstream of Redstone River. Flow is right to left.

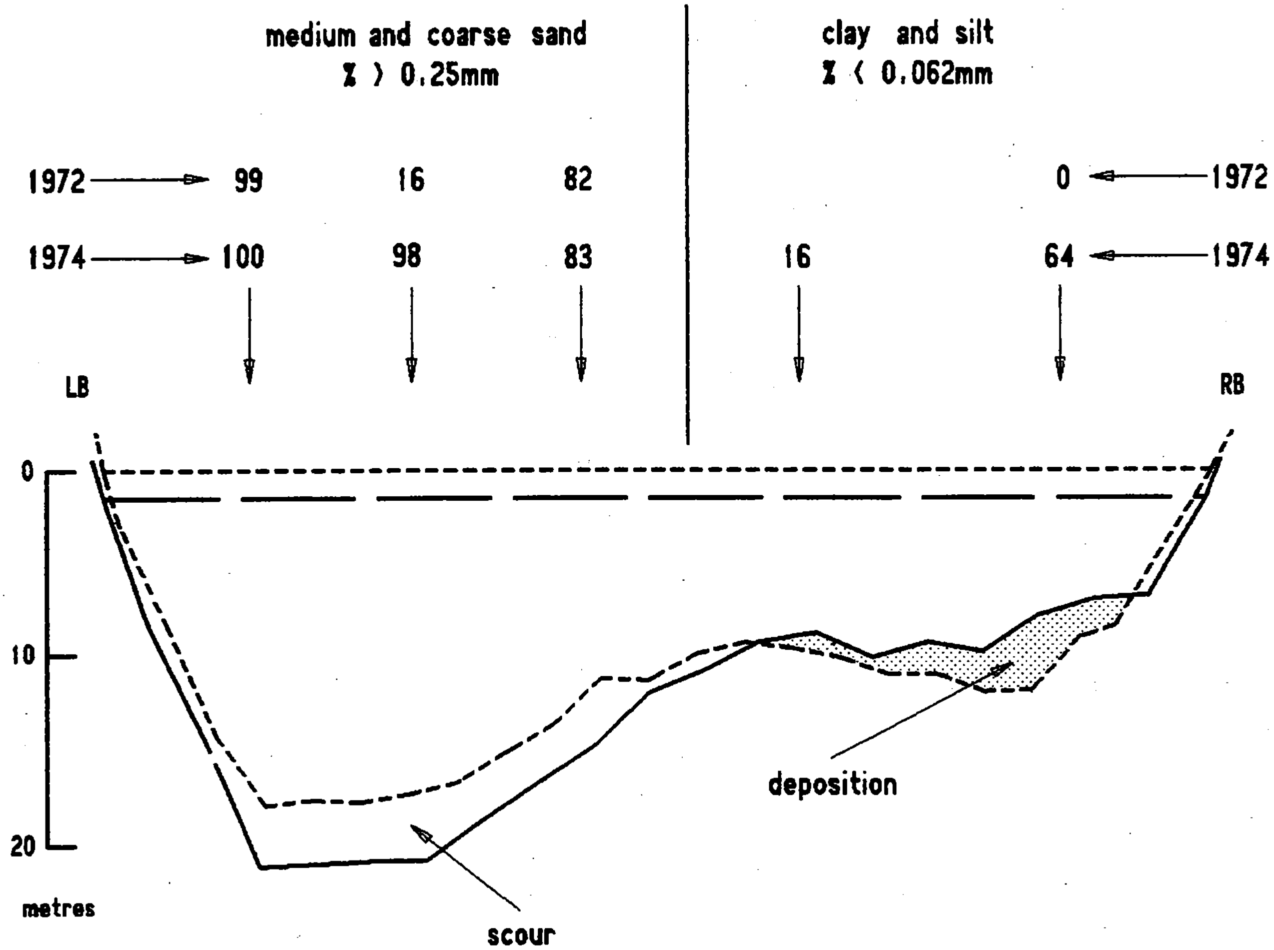


Figure 11.1 Mackenzie River upstream of Arctic Red River: channel geometry at 1970s measurement section
short dash: 1972 July 13 Discharge = 17,260 m ^{3}/s
long dash : 1974 Sept 24 Discharge = 11,600 m ^{3}/s

Data taken from Hydrometric Survey Notes

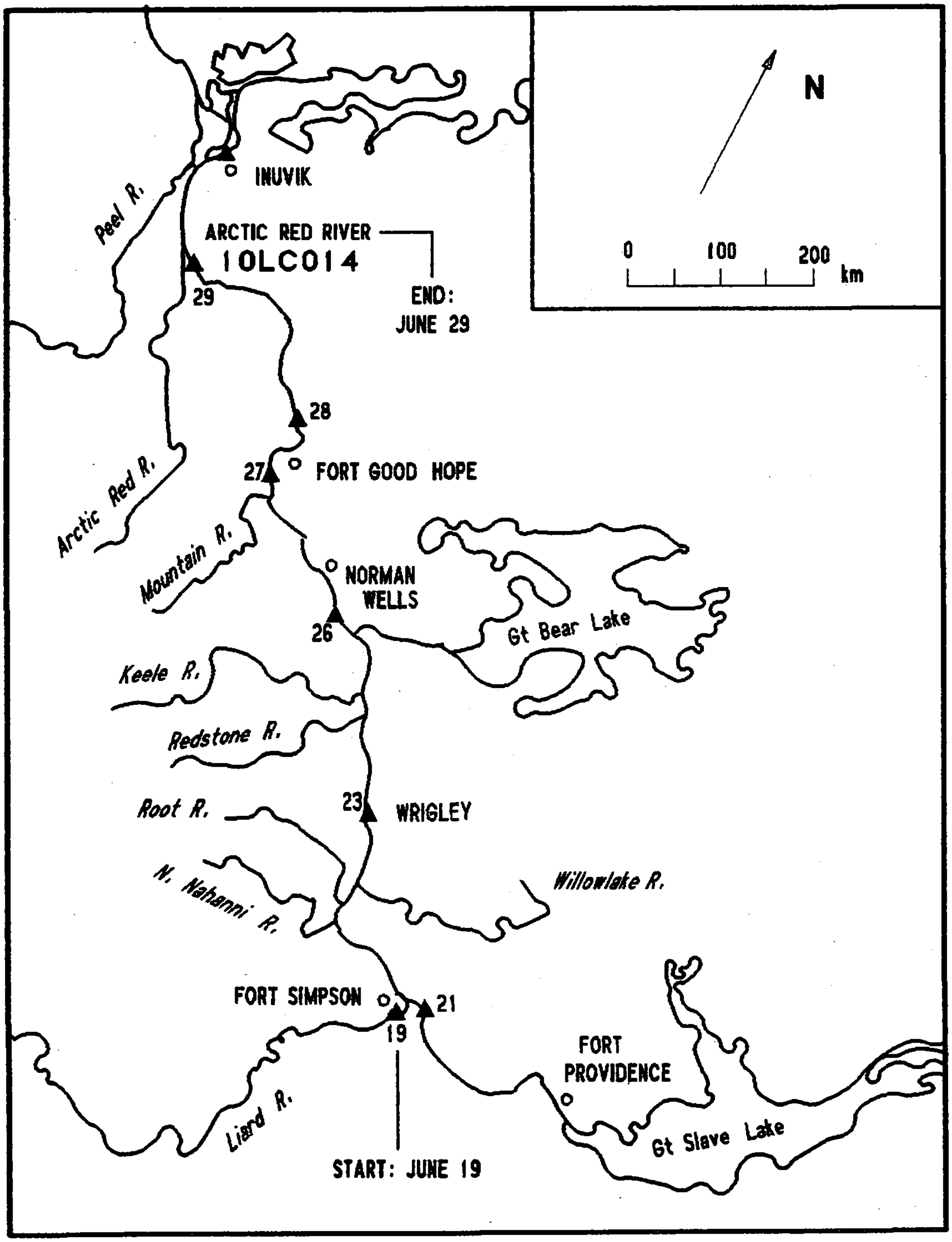


Figure 11.2 The NHRI Mackenzie River hydrocarbon study, June 1986.

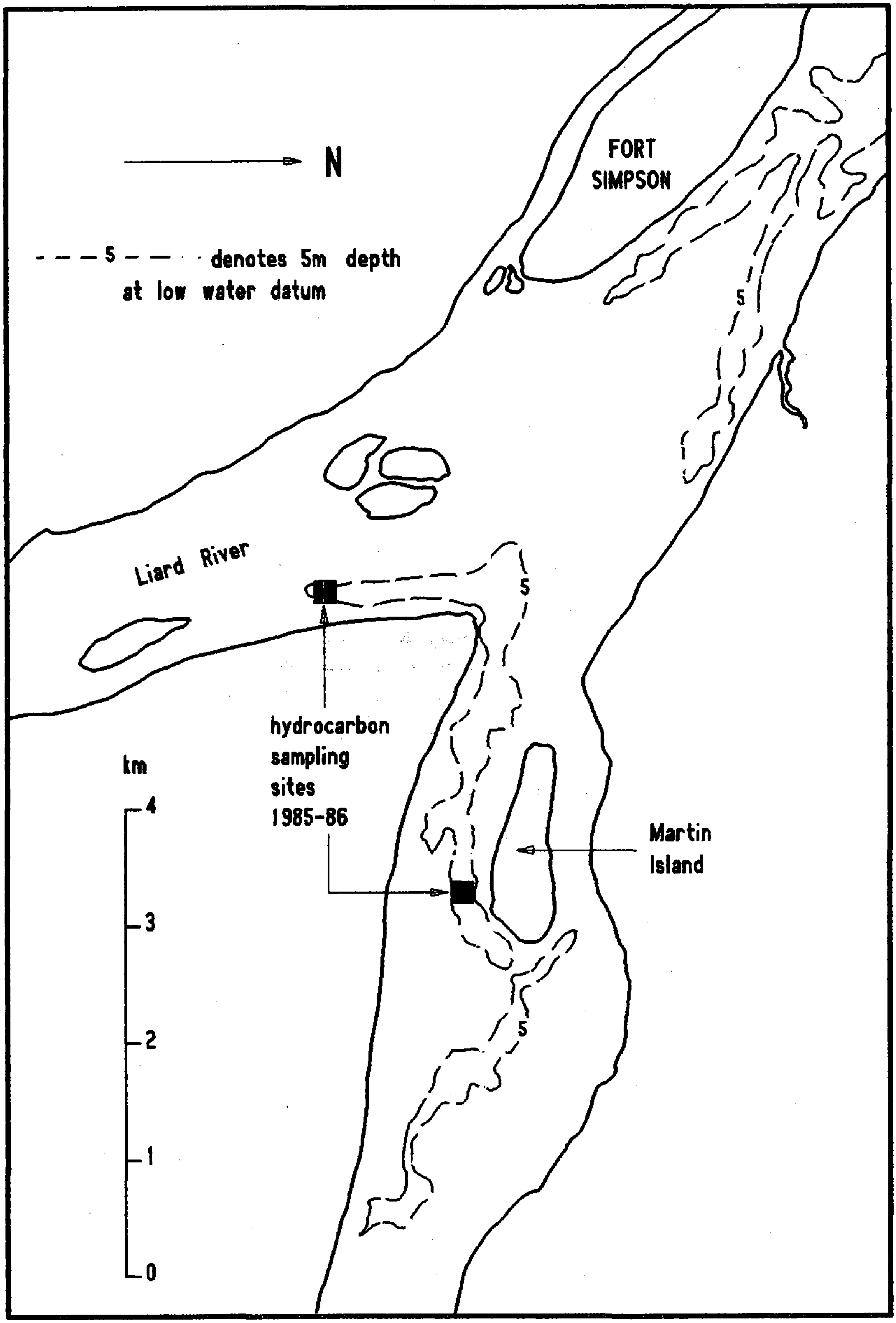


Figure 11.3 Mackenzie River at Liard River confluence: NHRI hydrocarbon sampling sites.

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Applications of geomorphology for Inland Waters Directorate staff : Western and Northern region.