ASPECTS OF THE QUATERNARY EVOLUTION OF THE PLATEAU REGIONS OF THE NORTHERN RUBY RANGE, SOUTHWEST YUKON TERRITORY

by

Richard Joseph Kodybka

Department of Geography
University of Ottawa
Ottawa, Ontario
K1N 6N5

A thesis
presented to the University of Ottawa
in fulfilment of the
thesis requirement for the degree of
Doctor of Philosophy
in
Geography

ISBN 0-315-95954-1

UNIVERSITÉ D'OTTAWA
UNIVERSITY OF OTTAWA

I hereby declare that I am the sole author of this thesis. I authorize the University of Ottawa to lend this thesis to other institutions or individuals for the purpose of scholarly research.

Richard Joseph Kodybka

I further authorize the University of Ottawa to reproduce this thesis by photocopying or by other means, in total or in part, at the request of other institutions or individuals for the purpose of scholarly research.

Richard Joseph Kodybka

ABSTRACT

The main objective of this thesis was to offer explanations on the evolution of the plateau surfaces in the northern Ruby Range, Yukon Territory. Specifically, to determine whether the geological and geomorphological evidence from the plateau surfaces could support the belief that these regions were subjected to glaciation, or does the evidence indicate other evolutionary processes. The more traditional geological techniques of geochemical and heavy mineral compositional analyses have been used previously in both geological and glacial related studies in the Canadian Shield and other regions of Canada, and have aided in the description of environments. However these types of analyses have had limited application in alpine environments with extensive plateau regions, such as those experienced in the northern Ruby Range. The data also made it possible to delineate, in a regional context, mineralized zones that have not yet been identified. This pursuit constituted an ancillary objective in this thesis.

In pursuit of these objectives, two questions were posed: 1) Can the compositional trends observed in the heavy mineral and geochemical data on the plateaus provide geological evidence (lacking in previous descriptions of these regions) which can contribute to the solution to the question of how the plateau landscape evolved? Specifically, depending upon the genesis of the plateau sediments (ie. local, glacial, etc.), what processes were responsible for these dispersals (glaciation, periglacial, aeolian, tectonic)? and 2) Are similar compositional trends observed in the valleys of the Ruby Range, and what are the implications of these patterns?

It was determined that the average proportion of heavy minerals identified was similar for both plateau and valley sediment samples. The range of the proportions was found to vary considerably in most minerals. The significance of these comparisons were not clear, but probably indicated that the sediments from both plateau and valley sites did not vary greatly in heavy mineral species and proportions, but demonstrated a variable range. As well, there was no appreciable difference in the heavy mineral assemblages between plateau and valley sediment samples, except for the relative lack of allanite from plateaus, and goethite from valleys. The data did not support a clear association between heavy mineral assemblages from plateau and valley sediments, and those derived from bedrock within or outside the study area. However, the sediment assemblages may not be exclusively related to local bedrock sources.

The provenance of orthopyroxene, goethite, and allanite may be attributed, in part, to bedrock from outside the study area, volcanic ash, and meteoritic impacts, or a combination of these. The occurrence of these minerals may also be associated with mineral alterations. Both the heavy mineral and geochemical data indicated there were no discernable dispersal patterns in either an up-valley or down-valley direction. The random dispersals of minerals and elements in the valleys can be characterized as secondary, epigenetic dispersals, and can be further classified as resembling clastic, hydromorphic, or biogenic patterns.

Heavy mineral and element analyses performed on surficial sediments from the plateau surfaces indicated that some of the material may not be of local origin, and discernable dispersal trends were evident in certain heavy minerals and elements which were traced to known sources of mineralizations. The most striking characteristics of these trends were their continuous form, spanning many kilometres, and their patterns which extending over extensive plateau surfaces that were dissected by numerous valleys. The following conclusions were made with respect to dispersal patterns observed on plateau surfaces:

1) The north to northwest alignment of the Tracey Arm Terrane, and the northwest mineral and element dispersal patterns are likely coincidental.

2) Some mineral and element dispersal patterns may be in part, attributed to local terrane conditions.

3) Element dispersal trends cannot be exclusively related to element mobility on the plateau surfaces.

4) Mass-wasting processes probably served to subsequently modify pre-existing dispersal patterns on local scales.

5) The general northwest dispersal patterns observed on the plateau surfaces of the northern Ruby Range were attributed to glaciation.

It is proposed in this study that the heavy mineral and element dispersal patterns observed on the plateau regions of the northern Ruby Range may be the evidence lacking in previous studies which can confirm that the plateau surfaces were glaciated. The glaciation (or glaciations) would, in accordance to the reported glacial history of the area, represent pre-Nisling Ice (pre-Reid) glaciation(s) (late Tertiary or early Pleistocene). Glaciation(s) in late Tertiary and/or early Pleistocene could have preceded major stream incision, hence valleys may be post-glacial (i.e., formed after glaciation in late Tertiary and/or early Pleistocene). Significant uplift of plateau regions may have occurred after late Tertiary and/or early Pleistocene glaciation(s) (i.e., during middle - late Pleistocene). Whether uniform uplift rates were experienced throughout the entire Tracey Arm Terrane is uncertain.

In summary, it is proposed that the northern Ruby Range (and surrounding regions) may have been glaciated (perhaps several times) in late Tertiary and/or early Pleistocene, undergone uplift with stream incision intensifying as uplift progressed, followed by middle - late Pleistocene glaciation(s) which modified the valleys. The plateau surfaces would have experienced periglacial processes (perhaps severe) continually throughout the Pleistocene rendering any glacigenic material residual.

ACKNOWLEDGEMENTS

I wish to thank the many individuals who over the years have contributed in various ways to my PhD program, and to the completion of this thesis. Foremost, I wish to express my sincere appreciation and gratitude to my program supervisor and friend, Dr. Peter G. Johnson, for affording me the unique opportunity to undertake research in the Yukon Territory and his steadfast guidance, insights, encouragement and support during all phases of my PhD program (and afterwards). **Thank you very much Peter.**

I also acknowledge the contributions of Drs. M. Smith, F. Koerner, W.W. Shilts, and D.A. St-Onge in way of comments and advice, and for agreeing to serve as members of my research committee at various stages of the program.

Other individuals and institutions are acknowledged for their professional and technical advice, and assistance. Included are; Dr. Christine Kaszycki, Dr. P. Henderson, Dominique Paré, Patty Higgins-Lindsay, John Lope, Bondar-Clegg Co. Ltd. Ottawa, Arctic Institute of North America, Terrain Sciences Division (Geological Survey of Canada), Earth Physics Division (Geological Survey of Canada), Department of Geography (University of Ottawa).

Throughout the ordeals of thesis preparation, the friendship and good humour of Fredérique Lester, Caroline (the Little One) Lester, Jean-Pierre (the Dude) Lester, Dorothy Dalton, and Mario Paradis proved invaluable. To these very special friends, I offer my sincere thanks.

Financial support for this research was provided, in part, by grants from the University of Ottawa, Northern Research Group, and from the Natural Sciences and Engineering Research Council grants awarded to Dr. Peter G. Johnson.

I wish to dedicate this thesis to my parents, Frank and Romana Kodybka, who have always put their own adversities and personal ordeals aside, in order to provided for me, the encouragement and support necessary to pursue and succeed in my endeavours.

TABLE OF CONTENT

Chapter 1
INTRODUCTION

Chapter 2
GEOLOGY OF SOUTHERN YUKON

Chapter 3
THE QUATERNARY IN SOUTHERN YUKON

Chapter 4
METHODS AND TECHNIQUES

Chapter 5
ANALYSES AND INTERPRETATION

Chapter 6
A DISCUSSION ON THE EVOLUTION OF THE PLATEAU SURFACES OF THE NORTHERN RUBY RANGE

Chapter 7
MINERALIZATION IN THE NORTHERN RUBY RANGE

Chapter 8
SYNOPSES AND CONCLUSIONS

LIST OF FIGURES

LIST OF TABLES

Chapter 1
INTRODUCTION

1.1 The Ruby Range

The Ruby Range, which exhibits a subdued relief with rounded to flat mountain tops and plateaus at elevations between 1220 to 2200 meters above sea level (masl) and major valley systems generally oriented east-west, has been the focus of various environment related studies (e.g.,Muller, 1967; Johnson, 1983; Johnson et al., 1984; Lemaitre, 1985; Smith, 1985; Schwartzburg, 1986; Lagarec, 1986; Lacasse, 1987). With few exceptions (Bostock, 1966; Muller, 1967; Hughes et al., 1969; Dewez, 1988), relatively little data are available on the glaciology of the Ruby Range.

Initial interest in the Ruby Range was stimulated by the complexity of glaciological conditions reported in the literature. Most valleys take on a classical glaciated U-shape (Figure 1) with evidence of glaciation of local (ice cap, cirque, and valley glaciers), and regional (ice sheet) origin in the form of end and lateral moraines. It is reported that ice was not only channelled out from cirques down existing valley systems, but most of the major valleys were affected by ice from an external source, St. Elias and Coast Mountains (Muller, 1967; Hughes et al., 1969).

The plateau regions of the Ruby Range provoke even more curiosity. Initial analysis of areal photographs, and subsequent field observations, demonstrated that the relatively smooth, undulating terrain of the plateaus was void of any conspicuous landforms except for a few tor-like structures (Figure 1). This is in sharp contrast with the valleys which, apart from being occupied by swamps, lakes and streams, contain various glacial landforms and in some locations, rock glaciers. Although the evolution of the plateau regions have not been described in any great detail in the literature, except for being referred to as an old, uplifted, erosional surface, accounts of glaciation in the Ruby Range are accompanied by descriptions that identify the plateau regions as generally being an unglaciated, weathered landscape of unknown age (Muller, 1967).

Specifically, the presence of tor-like structures and other periglacial features found on plateau surfaces indicate that these regions are relatively 'old'. It is also reported that there is an apparent absence of any glacially related features or landforms on these surfaces. However, the presence of tor-like structures and other periglacial features do not necessarily rule out the possibility of glaciation (Sugden and Watt, 1977). In view of this, it is not surprising that any glacigenic sediment or landforms (if in fact they did exist) would have long since been obliterated or altered in this environment.

Figure 1
TOPOGRAPHIC SETTING OF THE NORTHERN RUBY RANGE

Most valleys exhibit a typical glaciated U-shape.

Plateau regions exhibit a relatively smooth, undulating terrain, virtually void of any conspicuous landforms.

1.2 Objectives

The primary objective of this thesis is to offer explanations on the evolution of the plateau surfaces in the northern Ruby Range, Yukon Territory. The techniques employed are an examination of surficial deposits, with the use of heavy mineral and geochemical composition analysis, and other provenance related studies (erratics) as well as geomorphological mapping and interpretation. Specifically, this thesis will attempt to determine whether the geological and geomorphological evidence from the plateau surfaces can support the belief that these regions were subjected to glaciation, or does the evidence indicate other evolutionary processes.

The following research questions are posed:

i) Can the compositional trends observed in the heavy mineral and geochemical data on the plateaus provide geological evidence (lacking in previous descriptions of these regions) which can contribute to the solution to the question of how the plateau landscape evolved? Specifically, depending upon the genesis of the plateau sediments (ie. local, glacial, etc.), what processes were responsible for these dispersals (glaciation, periglacial, aeolian, tectonic)?

ii) Are similar compositional trends observed in the valleys of the Ruby Range, and what are the implications of these patterns?

Geochemical and heavy mineral compositional analyses has been used to help determine whether vast regions of the Canadian Shield and other regions of Canada were glaciated, and have aided in the description of environments (Shilts 1973, 1976, 1980, 1982; Shilts et al., 1979; Henderson 1983, 1985, 1989; Kaszycki et al., 1986; Kaszycki, 1987; Henderson et al., 1987). However these types of analyses have yet to be used in alpine environments with extensive plateaus that show evidence of converging Cordilleran and/or Continental ice.

Research on the Quaternary history of the Canadian Cordillera has produced many descriptive accounts of glaciation (e.g., Beach, 1943; Bostock, 1948; Belyea, 1960; Roed, 1975; Luckman and Osborne, 1979; Holloway et al., 1981; Kearney and Luckman, 1981; Kostaschuk and Smith, 1983). Interpretations ranging from a few distinct glaciations to differing numbers of glacial advances within similar time periods have been made. This implies that, based on our current understanding, some events cannot be correlated between adjacent valleys or ranges, or the evidence is lacking and/or equivocal (Shaw, 1972; Luckman and Osborne, 1979). Further complications arise in convergence zones between Cordilleran and/or Continental ice because the flow dynamics are said to have varied spatially and temporally (Shilts 1982, 1984; Dyke et al., 1982; Prest, 1983; Clague, 1987).

Gaps exist in the literature concerning detailed glacial history of regions where there is evidence of converging Cordilleran and/or Continental ice. Southern Yukon is one of these regions and provides a unique environment in Quaternary studies with its rapid transition from currently glaciated regions in the St. Elias Mountains to areas of central Yukon which were apparently not glaciated throughout the Pleistocene and possibly the Tertiary.

Although the main focus of this thesis is to determine whether the geological and geomorphological evidence from the plateau surfaces can support the belief that these regions were subjected to glaciation, or other evolutionary processes, the data may also make it possible to delineate, in a regional context, mineralized zones that have not yet been identified. This pursuit will constitute an ancillary objective in this thesis. It is anticipated that this study will lead to a better understanding of the evolution of the plateau regions of the northern Ruby Range. Specifically, can the data support the possibility that the plateaus were glaciated at one time, or does the evidence indicate other evolutionary alternatives?

1.3 Regional Climate

The climate of Yukon is classified as sub-Arctic Continental due to large annual, daily, and day to day ranges of temperature, low relative humidity and irregular, low moderate precipitation (Wahl and Goos, 1987). Frequent intrusion of mild air from the Pacific Ocean results in a sub-Arctic rather than Arctic climate. The main controls which influence the climate of southern Yukon include latitudinal effects, proximity to the Pacific Ocean, and mountainous terrain.

Long term meteorological data are not available for the actual study area although short term studies with reference to possible regional trends have been undertaken (Lemaitre, 1985). An approximation of climatic conditions is provided by data from Beaver Creek, Burwash Landing, Haines Junction and Kluane Lake and are summarized in Table 1. It is noted that the variability of climatic conditions in alpine regions necessitates a very general acceptance of data and subsequent trends from these reporting stations.

Environment Canada reports a mean winter temperature for the region of -20°C. Mean spring temperatures are given as 10°C. The mean summer temperature is reported as 17°C. Deviations from these data are expected, for example at Burwash in August 1981, a maximum temperature of 29°C was recorded. On the other extreme, a record low temperature of -62°C was recorded at Snag, north of the study area (Muller, 1967). August is generally cool with frost common at higher elevations. Snow has been recorded in the Ruby Range in late July and August. In some localities, snow patches persist throughout the summer months. Table 1 reports precipitation for various stations around the study area. When Environment Canada data are compared with

Table 1
GENERAL REGIONAL CLIMATIC CONDITIONS

	June				July				August			
	Temperature (°C)			Precipitation	Temperature (°C)			Precipitation	Temperature (°C)			Precipitation
	max.	min.	mean	(mm)	max.	min.	mean	(mm)	max.	min.	mean	(mm)
Beaver Creek 62°23' 140°52'	17.3	1.8	9.6	51.4	20.3	3.8	11.9	71.8	19.3	1.1	10.2	64.5
Burwash Landing 61°21' 139°00'	16.0	1.8	8.9	77.5	13.8	2.8	8.3	33.6	17.8	5.2	11.5	27.6
Haines Junction 60°45' 137°30'	16.7	1.7	9.2	26.6	20.3	6.0	13.2	8.5	17.4	6.2	10.3	39.3
Kluane Lake 61°02' 138°23'	15.4	2.8	9.1	37.2	19.9	6.5	13.3	33.2	-	-	-	-
Headwater Lake 61°25' 138°21'	16.3	2.0	8.8	239.9	20.1	6.1	13.1	31.5	16.5	5.6	10.6	Trace

Sources: Environment Canada meteorological observations, 1981. Headwater Lake data coutesy of Dr. D. Lagarec (Dept. of Geography, University of Ottawa).

precipitation data at Headwater Lake (Schwartzburg, 1986), differences in local and regional conditions become evident. While the precipitation data reflects some regionality in storm systems, it is evident that local variations do exist. For example, Headwater Lake station receives an average of 239.9 mm of precipitation for the month of June. By comparison, Beaver Creek, Burwash Landing, Haines Junction, and Kluane Lake received 51.4 mm, 77.5 mm, 26.6 mm and 37.2 mm respectively. This comparison clearly demonstrates the orographic effect of the Ruby Range.

1.4 Flora

The flora of southwest Yukon is as diverse and distinctive as the physiography of the region. Although vegetational analysis does not constitute a major portion of this thesis, the recognition of dominant or climax species and plant associations provides a useful tool for the analysis of geomorphic environments. Vegetation is dependent on various environmental factors such as climate, soils, slope characteristics, moisture content and migration, and parent material.

Forest cover in the study region is relatively sparse with an upper limit of approximately 1200 masl (Muller, 1967; Terasmae, 1967). Because of the permafrost, trees develop a shallow, horizontal matted system of roots which makes them susceptible to uprooting. The main species encountered in the lower valleys are; Picea mariana (black spruce), glauca (white spruce) and Populus balsamifera (balsam poplar). Also indigenous to the area are Populus tremuloides (aspen), Betula papyrifera (white birch), Ablies lasiocarpa (alpine fir), Larix laricina (tamarack) (Muller, 1967; Terasmae, 1967). Some alders and willows are present above 1,200 m but are generally isolated. Plant associations in poorly drained valley bottom areas include basically black spruce/moss associations.

Of significance on some isolated plateau areas is the occurrence of peat. In relatively well drained valleys and particularly on south facing slopes, the forest associations are dominated by aspen, white birch and black spruce. Above tree line and particularly on relatively flat plateau surfaces, shrubs, mosses, sedges and lichens dominate.

1.5 Permafrost

Many of the permafrost observations in the Yukon have been made by geologists involved with mineral exploration operations. Although mapped by Brown (1978) as scattered discontinuous and alpine permafrost, very little is known about permafrost conditions for this region of the Yukon. Brown's (1978) permafrost map incorporates data from only four boreholes for all of the Yukon. The division between continuous and discontinuous permafrost is based on the -5°C mean annual isoline which is derived from a minimum of data in the southwest Yukon. Several landforms diagnostic of permafrost conditions occur within the study area. These include poorly sorted stone circles and stripes, and in some localities beaded drainage patterns. On higher plateau regions, frost or mud boils, and solifluction lobes are observed. Rock glaciers are also present at some cirque and valley bottom locations and palsas in Gladstone Creek.

Although no detailed studies of active layer depths have been made for the study area, some comparative observations from the Ruby Range and other regions are available. For example, Rampton (1977) reports that permafrost may be absent under south-facing slopes at low elevations throughout the Ruby Range. Price (1971) has recorded active layer depths of up to 150 cm for upper elevations in the Ruby Range. North of the study area, in the Eagle Plains area, Pollard (1983) recorded active layer depths of 60 to 100 cm. Elsewhere in the Ruby Range, Johnson (1981) has documented active layer depths of between 20 to 100 cm. In the Aishihik Lake region, Hughes (1990) reports active layer depths to be within 50 cm of the surface. In this study, active layer depths have been observed to range between 20 to 100 cm.

Chapter 2
GEOLOGY OF SOUTHERN YUKON

2.1 Introduction

The following sections emphasize the geology of the southern Yukon, within the framework of the Canadian Cordillera and set the stage for a more detailed description of the regional setting, geology, and Quaternary history of the Ruby Range. The discussions include summaries of, i) the structural setting of Yukon, ii) geological terranes, iii) Quaternary tectonic activity within, and in proximity to the southwest Yukon, and iv) the pre-Quaternary landscape.

2.2 Structural Setting: Diastrophism

The Canadian Cordillera comprises part of the deformed western margin of the North American craton and an assemblage of foreign crustal fragments and terranes which became accreted to the craton due to the convergence of the North American and Pacific plates in Mesozoic and Cenozoic time (Monger at al., 1982; Clague, 1991). The structure of the Canadian Cordillera in Yukon (as throughout western Canada) is divided into the Columbia and Pacific Orogen zones. The southwest Yukon constitutes the northern extension of the Pacific Orogen zone while the southeast Yukon forms the northern extension of the Columbia Orogen zone (Figure 2) (Price et al., 1973). The strong northwest - southeast structural grain of the Canadian Cordillera is a result of accretion and related faulting. The Cordillera comprises five major northwest trending geological belts which exhibit unique stratigraphy, plutonism, metamorphism, volcanism and structure (Monger and Berg, 1984; Clague, 1991).

The Columbian Orogen zone includes the Omineca Crystalline Belt and an eastern Foreland Belt. The Pacific Orogen zone includes the Intermontane Belt, Coast Belt, and Insular Belt. Two smaller structural units, the Northern Yukon Fold Complex and St. Elias Fold Belt have been identified as parts of more extensive units projecting into Yukon from Alaska (Price, et al., 1972). The Northern Yukon Fold Complex (in northern Yukon), forms the structural and stratigraphic link between the structures of the Foreland Belt in central Yukon and those of northern Alaska. The St. Elias Fold Belt (in the extreme southwest Yukon) is the youngest in the Canadian Cordillera. Much of the structural character of this belt was probably developed as a result of mid-Jurassic and mid-Cretaceous tectonism that also affected the regions of southeastern Alaska. Further deformation took place in the Cenozoic related to the northwest underflow of the Pacific plate (Norris, 1971).

Figure 2

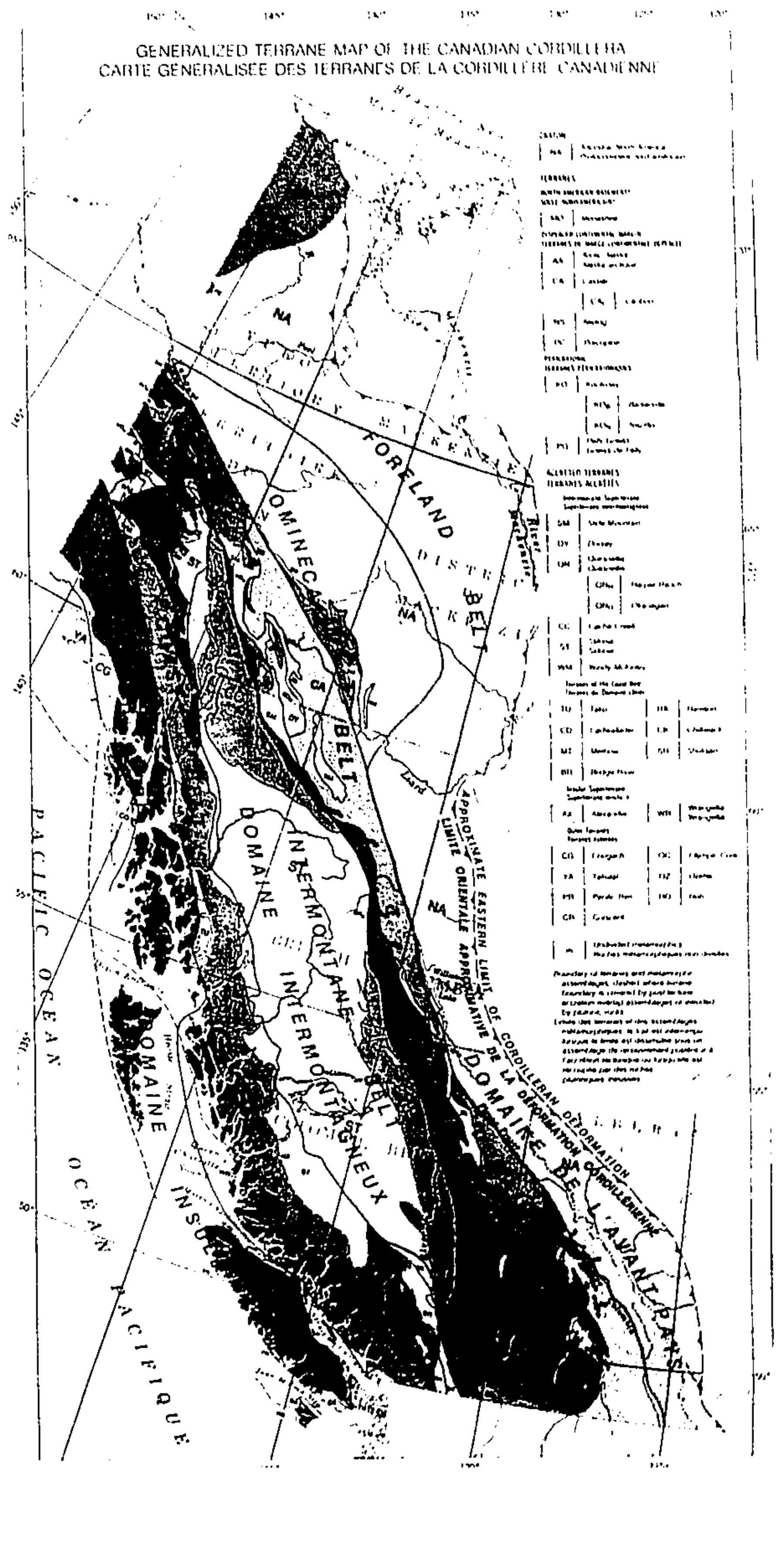

GENERALIZED TERRANE MAP OF THE CANADIAN CORDILLERA
CARTE GENERALISEE DES TERRANES DE LA CORDILLERE CANADIENNE
FORELAND
BELT
OMINECA
BELT
INTERMONTANE
BELT
DOMAINE
INTERMONTAGNEUX
PACIFIC OCEAN
OCEAN PACIFIQUE
APPROXIMATE EASTERN LIMIT OF CORDILLERAN DEFORMATION
LIMITE ORIENTALE APPROXIMATIVE DE LA DEFORMATION CORDILLERIENNE
DOMAINE DE L'AVANT-PAYS
NA

Table 2
GEOLOGICAL BELTS

Omineca Crystalline Belt	Straddles the boundary between ancestral North America and the allochthonous (displaced over a considerable distance from their origin) terranes to the west. The belt is characterised by deformation, regional metamorphism, and granitic plutonism. The eastern portion of the belt comprises of middle Proterozoic to middle Paleozoic miogeoclinal rocks. The western portion consists of accreted Paleozoic and lower Mesozoic volcanic and sedimentary rocks. Granitic plutons are frequent throughout the entire belt.
Foreland Belt	Borders un-deformed rocks of the Interior Platform to the east and consists of middle Proterozoic to Upper Jurrassic miogeoclinal and platform carbonates and some craton derived clastics. The belt exhibits numerous northwest trending folds.
Intermontane Belt	This belt is a composite of three major foreign terranes formed of upper Paleozoic to middle Mesozoic, allochthonous, marine volcanic and sedimentary rocks. These rocks are superposed by autochthonous (in place) Jurassic and Cretaceous clastic wedges and Tertiary volcanic and sedimentary rocks. The belt also contains granitic intrusions which are comagmatic (common source) with the volcanics.
Coast Belt	Metamorphic and plutonic belt consisting of Jurassic to Tertiary granitic rocks and varied metamorphosed sedimentary and volcanic strata ranging in age from Paleozoic to early Tertiary.
Insular Belt	This is a composite allochthonous terrane belt with Upper Cambrian to Tertiary volcanic and sedimentary rocks, and some granitics.

Clague (1991) assigns the Northern Yukon Fold Complex to the Foreland Belt, and the St. Elias Fold Belt to the Insular Belt. A summary description of the five belts is given in Table 2. It is of interest to note that potassium-argon dates of lavas associated with the White River tills and tillites indicate that intermittent glaciation occurred in the St. Elias Fold Belt (Denton and Armstrong, 1969). Evidence suggests that many of the ancient glaciations occurred around 9 to 10 m.y. ago, and even earlier. At least four intervals of major glacier expansion are tentatively dated at less than 2.7 m.y. and thus probably Pleistocene in age.

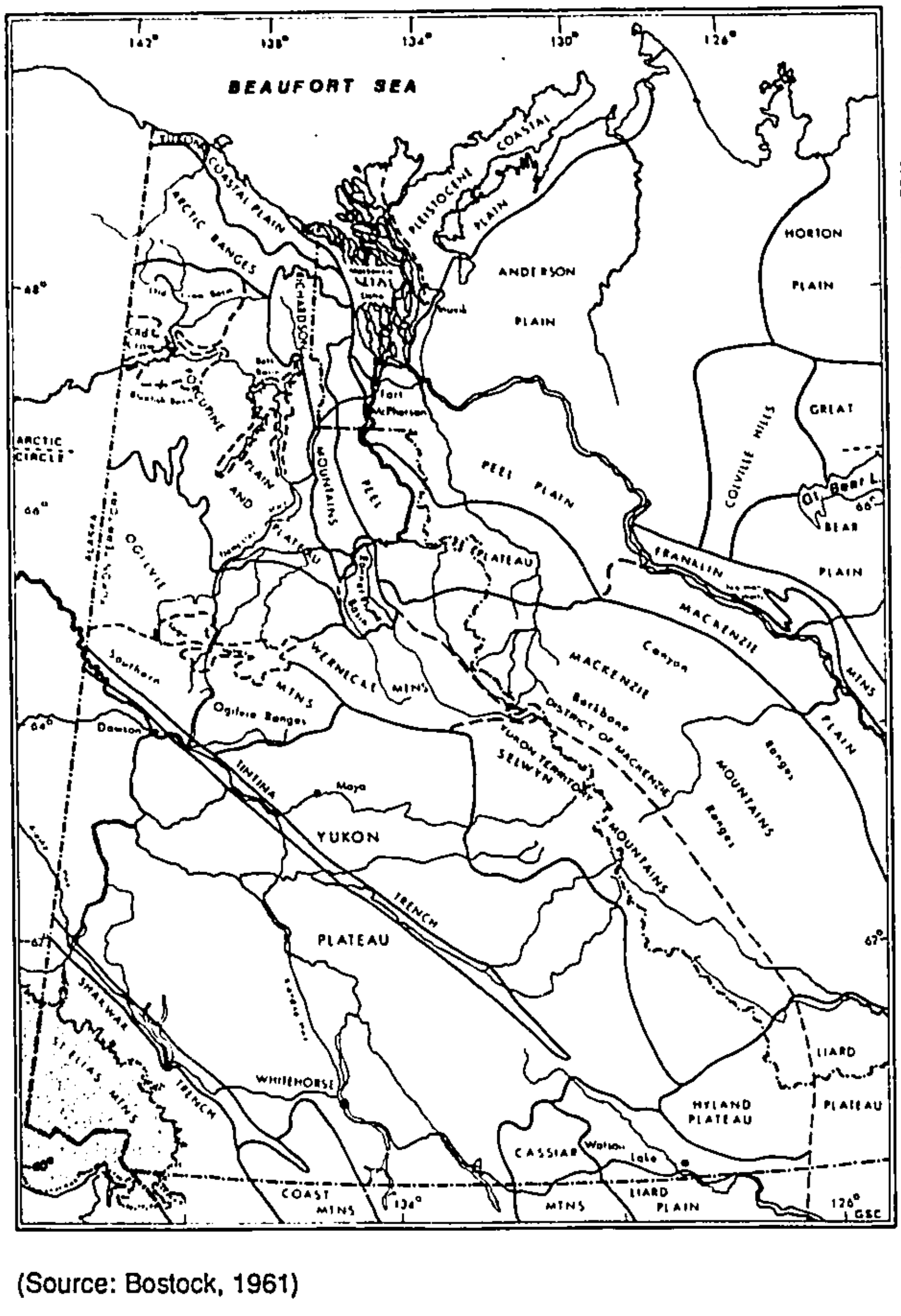
2.3 Geological Terranes

Geological terranes of the Canadian Cordillera are regions characterized by an assemblage (or assemblages) of rocks whose paleogeographical relationships with those in bordering terranes are unknown (Wheeler et al. 1988). Paleontological and paleomagnetic data suggest that some currently juxtaposed terranes were originally separated by distances of thousands of kilometres. Terranes are generally bounded by faults although in some places these may be concealed by cover rock or intrusions. Terranes are categorized according to their relationship to ancestral North America. Accreted terranes represent oceanic or island arc lithologies, generally of unknown paleogeographic origin which are allochthonous with respect to miogeoclinal strata. Accreated terranes are grouped into two superterranes, the Intermontane and Insular, based on their time of assembly and collision with North America. Between these two lie the terranes of the Coast Belt (Figure 2).

Southern Yukon is further divided into three parts by two northwest trending strike-slip faults, the Tintina and Denali faults (Figure 3). Terranes northeast of the Tintina Fault are also present southwest of it, but the Denali Fault separates discrete terranes and probably follows a fundamental suture in most of Yukon. Both faults localize narrow northwest oriented grabens which are Pliocene physiographic trenches; the Tintina Trench along the Tintina Fault and the Shakwak Trench along the Denali Fault.

The identification and delineation of geological terranes in southern Yukon have followed an evolutionary process, combining previous interpretations and new information (e.g., Tempelman-Kluit, 1979, 1980; Campbell and Dodds, 1982; Monger and Berg, 1987; Wheeler and McFeely, 1987; Wheeler et al., 1988; Currie, 1992; Gareau, 1992). The characterization of terranes in southern Yukon used in this thesis (Table 3) emulates the divisions identified by Tempelman-Kluit (1979), Monger and Berg (1987), Wheeler et al., (1988), and Gareau (1992). Terranes northeast of the Tintina Fault include the Mackenzie Platform, Selwyn Basin, Cassiar Platform and Slide Mountain. Between the Tintina Fault and the Denali Fault (to the southwest) the terranes consists of Yukon-Tanana, Tracey Arm, Stikinia, Slide Mountain, Cassiar Platform, Quesnellia, and Cache Creek. Southwest of the Denali Fault, the terrane includes Wrangellia, Alexander and Chugach.

Figure 3
PHYSIOGRAPHY OF YUKON



(Source: Bostock, 1961)

Table 3
YUKON TERRANES

Mackenzie Platform Terrane	Consisting of North American miogeosynclinal strata and younger intrusions; Precambrian clastic strata, Cambrian to Devonian carbonate succession overlies these. Mississippian and younger clastic rocks cover the carbonate strata.
Selwyn Basin Terrane	Consisting of North American miogeosynclinal strata and younger intrusions; shale, chert and basaltic volcanics overlie calcareous shale with intercalated volcanics.
Cassiar Platform Terrane	Includes sedimentary rocks ranging in age from late Precambrian to Devonian, condensed sequence of carbonate and sandstone.
Slide Mountain Terrane	Includes a lower sedimentary sequence comprising chert, argillite, sandstone, conglomerate, and minor carbonate rocks. The age of the terrne ranges from early Mississippian to late Permian. It is believed that the Slide Mountain Terrane has undergone a northward displacement but the timing is not certain.
Yukon-Tanana Terrane	A heterogeneous metamorphic terrane. Contains suits of ductile deformed metamorphic rock that rest on a basement of older gneiss. The gneiss may be a Precambrian basement which underwent displacement in the Devonian. Above the gneiss are Mesozoic sedimentary and volcanic rocks. Above this is Paleozoic greenstone and ultramafic rocks. This is capped with Devonian granodiorite.
Tracey Arm Terrane	A metamorphic terrane consisting of schist, paragneiss, amphibolite, marble, serpentine, and other metamorphosed sedimentary and igneous rocks. Age is not certain but probably predate early Triassic.
Stikinia Terrane	A coherent terrane comprising of statigraphically stacked Mississippian, Permian, Triassic, and Jurassic marine and non-marine volcanic and sedimentary strata. The Stikinia Terrane underwent a northward displacement during the Triassic or Jurassic. Also includes the Nisling sub-Terrane (Currie, 1992).
Quesnellia Terrane	Predominantly Upper Triassic and Lower Jurassic strata of marine, calc-alkaline to alkaline volcanic rocks.
Cache Creek Terrane	Mississippian to Upper Triassic oceanic volcanics and sediments, with Upper Triassic island arc volcanics including chert, argillite, basalt, carbonates. Terrane experienced northward displacement during (mid?) Cretaceous.
Wrangellia Terrane	Upper Paleozoic to Jurassic felsic to mafic volcanic rocks and intrusions, limestone, pelite; Middle and Upper Triassic basalt overlain by carbonate rocks. It is believed that the Wrangellia Terrane was displaced northward but the timing is uncertain.
Alexander Terrane	Comprised of late Precambrian to Triassic strata including schist and gneiss derived from feldspathic sediments, felsic to mafic volcanic rocks, pelite and carbonate rocks.
Chugach Terrane	Upper Cretaceous graywacke, argillite, and slate with a matrix of Upper Jurassic to Lower Cretaceous cherty argillite and blocks of mafic volcanics, chert, limestone, and ultramafic rocks.

Detailed terrane descriptions for the southern Yukon and the entire Canadian Cordillera are available from numerous sources (e.g., Price and Douglas, 1972; Campbell and Dodds, 1982; Monger and Berg, 1987; Wheeler and McFeely, 1987; Wheeler et al., 1988). The terrane descriptions provided in Table 3 are succinct syntheses of the terranes in southern Yukon. Hansen (1989) describes the terranes northeast of the Tintina Fault (including the Cassiar Platform Terrane, between the Tintina and Dinali Faults) as autochthonous (still in place of formation, have not been displaced). These terranes constitute continental margin strata of ancestral North America and include the Mackenzie Platform Terrane, Selwyn Basin Terrane, and Cassiar Platform Terrane.

Terranes southwest of the Tintina Fault (excluding the Cassiar Platform Terrane) are described by Hansen (1989) as allochthonous (displaced over a considerable distance from their place of formation). These include; Slide Mountain Terrane, Yukon-Tanana Terrane, Tracey Arm Terrane, Stikinia Terrane, Quesnellia Terrane, Cache Creek Terrane, Wrangellia Terrane, Alexander Terrane, and Chugach Terrane. Generally, terranes of southern Yukon are described (or implied) as being aligned and trending predominately north to northwest (Cockfield, 1922, 1926, 1927; Bostock, 1948; Muller, 1967; Price et al., 1973; Green, 1972; Tempelman-Kluit, 1974, 1980; Monger and Berg, 1987; Wheeler et al., 1988; Hughes, 1990; Gareau, 1992). The implications of this trend with respect to element and heavy mineral dispersal patterns will be considered later in this thesis.

2.4 Quaternary Tectonic Activity

The following summary of Quaternary tectonic phenomenon affecting southwest Yukon will focus on earthquakes and related fault movements, terrane uplift, and volcanic activity. The summary will emphasize the affects of tectonic phenomena on terrane movements, overburden disturbances, and sediment compositional changes brought about by volcanic ash-fall.

2.4.1 Earthquake Activity

Most Quaternary tectonic activity has been restricted to the Pacific - North American Plate boundaries and along parts of the Denali Fault system, particularly in Alaska. Earthquakes up to magnitude 8.6 have occurred at the plate boundary and Denali Fault respectively (Power, 1988). Only parts of the Denali Fault in south-central Alaska are active, there has been little significant movement along the fault in Yukon (Clague, 1979). However, Horner (1983) notes that the activity along the Denali Fault in Alaska indicates some motion at a depth of approximately 15 km in either an east-west or north-south direction, oblique to the trend of the fault. Similar motion along at least some portions of the Denali Fault in Yukon is not ruled out.

The Denali Fault is a major crustal break extending for more than 2000 km across south-central and southwest Alaska, southwest Yukon, and northern British Columbia. The fault system, which in part probably formed in the late Paleozoic or early Mesozoic, consists of many individual faults which have different displacement histories (Clague, 1979). The fault system as a whole is characterized by large Cenozoic displacements (dextral strike slip). In reviewing the displacement history of the Denali Fault System, Lanphere (1978) concluded that about 350 km of dextral displacement occurred on the McKinley, Shakwak and Dalton segments between 38 - 55 Ma BP.

In southern Alaska, deformation of glacial deposits due to earthquake activity are well documented (e.g., Hamilton et al., 1989; Kline and Bundtzen, 1989; Schmoll and Yehle, 1989; Thorson, 1989). In southwest Yukon reported evidence of overburden displacement due to earthquakes has been minimal, but some observations suggest that both vertical and lateral displacement has occurred. For example, Power (1988), Morison (1987), and Campbell and Dodds (1982) report that till and alluvial deposits dated as Pliocene and Pleistocene have undergone localized displacement. The extent of the displacements appears to be of the order of a few hundred meters. Clague (1981) has noted the association between the location of large rockslides of unknown age and historic epicentres in the St. Elias Mountains. However, Hansen (1989) maintains that major Quaternary displacements attributed to earthquakes have not been proven for the Denali Fault system in Yukon.

2.4.2 Tectonic Uplift

Also associated with tectonic activity in southwestern Yukon are regional vertical tectonic crustal movements. Data on crustal movements in Canada are limited. Vanicek and Nagy (1981) present a partial map of contemporary vertical crustal movements in Canada which include both uplift and subsidence. They note that it is not currently possible to distinguish vertical crustal movements due to tectonic processes and those due to glacial unloading (Hicks and Shofnos, 1965; Evans, 1991). Generally, the Cordillera is an area of uplift but the patterns of movement are spatially variable within geological terranes. Field observation of uplift in Yukon has been reported in a number of studies.

Evidence in central Yukon indicates that the aggradation of Pliocene and early Pleistocene gravels along the Yukon River are due, in part, to tectonic tilting (Hughes et al., 1972; Naeser et al., 1982). Tipper (1963), and Souther and Stanciu (1975) have observed Miocene lavas in the St. Elias Mountains that have been tilted, faulted, and uplifted. Parrish (1981) concluded that there has been sustained uplift of Mount Logan (St. Elias Mountains) averaging about 0.3 m/ka for the past 15 Ma. Other studies indicate that parts of the St. Elias Mountains and northern Coast Mountains are presently rising at the rate of a few centimetres per year (Clague, 1989). Vanicek and Nagy (1981) conclude that the maximum rate of uplift for southwest Yukon is currently 2.4 cm per year. Since contemporary uplift is evident in other parts of the Cordillera, these movements are mainly attributed to tectonic activity (Clague et al., 1982; Riddihough, 1982; Clague, 1991).

It may be of interest to note the relationship between uplift and denudation rates with reference to the Cordillera. Slaymaker and McPherson (1977) have divided the Canadian Cordillera into three major relief regions which, based on distinctive landform assemblages, exhibit different rates of denudation. These are; i) high mountain areas, ii) plateau and foothills areas, and iii) trenches, plains and low-lands. Denudation rates are measured in Bubnov units, where one Bubnov (B) is equivalent to 1 mm denudation per 1000 yrs (Fischer, 1969).

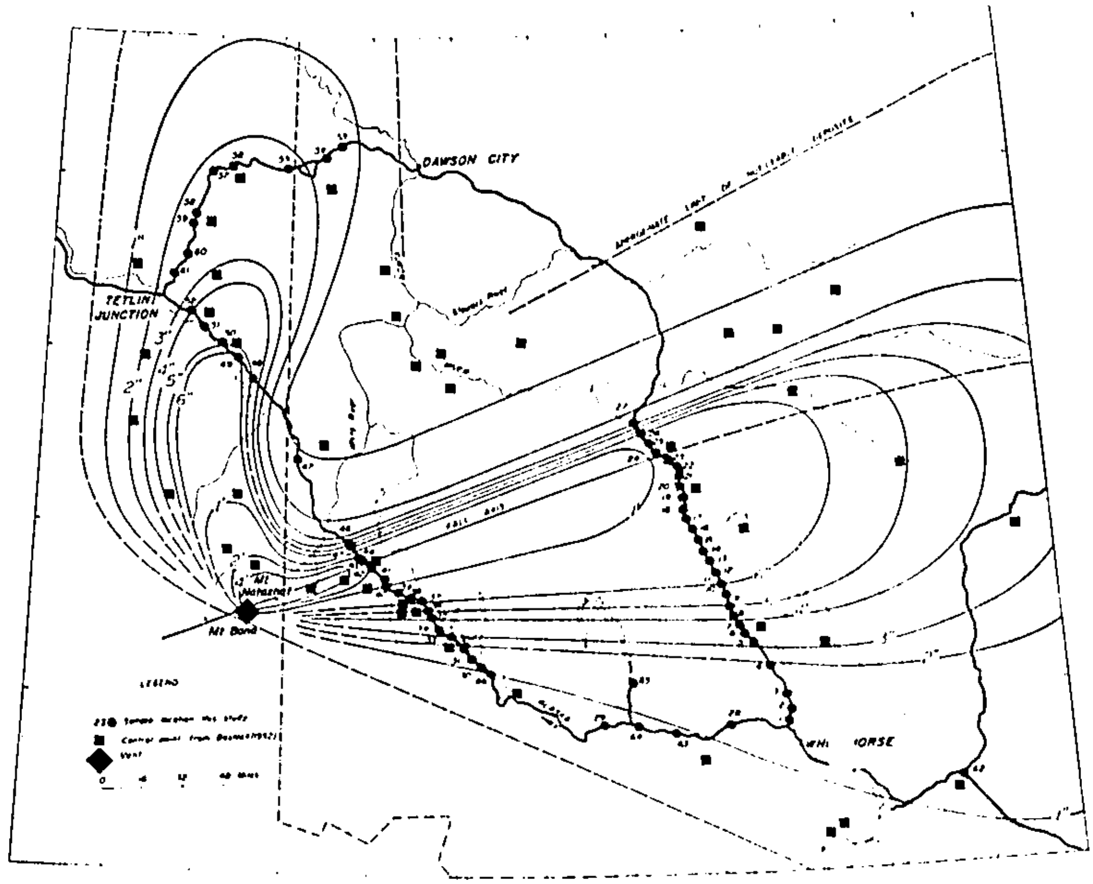
Denudation rates for the three major relief regions are; i) 92 - 750 B in high mountain areas, ii) 41 - 80 B in plateau and foothills, iii) 11 - 20 B in trenches, plains and low-lands (Slaymaker and McPherson, 1977). Based on these figures, if we consider the denudation rates for the Ruby Range study area of 41 - 80 B (which equates to 0.0041 - 0.0080 cm/yr, or an average of approximately 0.0060 cm/yr), and compare it to the tectonic uplift rate of 2.4 cm/yr for southwestern Yukon, it's clear that the areas of maximum denudation do not correspond to the areas of maximum uplift. Hence, it can be concluded that there a considerable disparity between rates of regional denudation and uplift in southwestern Yukon. Specifically, with reference to the Ruby Range study area, the region has (likely throughout the Quaternary) experienced uplift at a faster rate than it has weathered and eroded.

2.4.3 Volcanic Activity: White River Ash

Volcanism, which generated widespread basaltic lavas in the British Columbia interior (including northwestern British Columbia), in the late Tertiary continued on a reduced scale during the Quaternary. Volcanic activity within and in proximity to the study region is restricted to two major volcanic belts. The Stikine volcanic belt occupies parts of northwestern British Columbia and southeastern Alaska (Panhandle). This belt is located to the southeast of the study region and has experienced periods of activity throughout the Quaternary. A lava flow was reported as occurring in the Stikine volcanic belt at Aiyansh as early as 230 years BP (Sutherland Brown, 1969; Evans, 1991).

The second major volcanic belt, the Wrangell, which extends through the St. Elias Mountains and into Alaska, has also been active. Two recent eruptions in the Wrangell Mountains (in southeast Alaska) gave rise to two distinct ash-fall deposits that are collectively known as the White River Ash (Figure 4). A northern lobe was deposited approximately 1880 years BP, and a more extensive

Figure 4
WHITE RIVER VOLCANIC ASH PLUME



From Bostock (1952)

eastern lobe, which covers much of southern Yukon, was deposited about 1230 years BP (Downes, 1985). Combined, the ash deposit covers an area of approximately 325,000 km^2 of southern Yukon and eastern Alaska. In the Kluane Lake region, and specifically in the Ruby Range study area, the thickness of ash has been measured from 2 - 7.5 cm (Downes, 1985; Lerbekmo and Campbell, 1968).

The White River Ash had been recognized in a number of early excursions in Yukon (e.g., Schwatka, 1883; Dawson, 1887), and has been described in various related environmental studies (e.g., Lerbekmo, et al., 1973, 1975; Brazier, et al., 1983; Dewez, et al., 1984; Campbell, 1987). Detailed mineralogical and chemical analyses of the ash have been reported in a number of studies (e.g., Knopf, 1910; Lerbekmo and Campbell, 1969; Brazier et al., 1983; Downes, 1985; Westgate, 1987; McGimsey et al., 1992). The ash is composed of quartz, feldspar, hornblende, hypersthene, and magnetite. Lerbekmo and Campbell (1969) report the mineral composition (Wt %) of the ash as; quartz (23.2%), orthoclase (14.8%), albite (34.7%), anorthite (13.9%), hornblende (9.6%), hypersthene (1.0%), ilmenite (0.9%), and magnetite (1.8%).

Since this thesis is attempting to address the question of how heavy mineral and geochemical compositional trends in the Ruby Range were derived, it is important to determine to what extent volcanic ash-fall minerals may have complicated (or altered) these patterns by incorporation into the existing sediment. Because the heavy mineral separation procedures used in this thesis involves the use of Methylene Iodide (specific gravity 3.3) as the separation between heavy and light minerals, any volcanic ash-fall minerals with a specific gravity of 3.3 or greater may have been incorporated into the heavy fraction (depending on sample depth and comminution). The minerals in question include, hornblende (sg 2.9 - 3.3), hypersthene (sg 3.1 - 3.5), ilmenite (sg 4.5 - 5.0), and magnetite (sg 5.5 - 6.5). Dewez (1988) determined that ash-fall derived hornblende, hypersthene, ilmenite, and magnetite could be differentiated on the basis of a pitted and glassy, perlitic texture. This thesis uses the reported mineral composition of the ash as a basis for discussion. These concerns will be addressed in subsequent discussions.

2.5 Pre-Quaternary Landscape: An Overview

Although it is difficult to ascertain the pre-Quaternary landscape in southern Yukon, some generalizations can be made regarding the physiography of the Cordillera. Implications for the southern Yukon can be made from these generalizations. Mathews (1989) states that the Cordilleran landscape at the beginning of the Quaternary was dominated by mountainous terrain, dissected plateaus, and alluvial plains, probably not unlike the present day landscape of northern Yukon. The general distribution of mountains, plateaus and valleys was probably as it is currently, but almost certainly the relief variations were much less. Mountainous regions were likely lower than their modern counterparts and their glacial modifications and features were probably much more restricted than currently observed.

Valleys were not incised to the extent they are today, nor is it believed that glaciation extensively modified them, although Pliocene glaciation was a certainty. Subaerial erosion accompanying and following late Cenozoic uplift played an important role in developing major valley systems. Fluvial incision could be expected to form V-shape valleys. Subsequent modification by glaciers could then result in their widening and deepening. Mathews (1989) states that most Cordilleran valleys were probably initiated by streams prior to glaciation (Quaternary) although some exceptions have been noted (Mackay and Mathews, 1973). The streams that occupied these valleys may have differed in size and direction of flow from those now existing. Stream diversion and reversals are reported to have frequently taken place during the Quaternary (Mathews, 1989; Clague, 1989).

In the unglaciated central and northern regions of Yukon, Mathews (1989) states that streams alone have been responsible for valley formation. However, Hughes et al., (1972), and Mathews (1989) note that periodic glacial activity in eastern and southern Yukon could have influenced stream behaviour in these unglaciated regions by contributing to the sediment load of some rivers. Extensive plateau surfaces dominated (as they currently do) much of southern Yukon.

In particular, the Yukon Plateau has long been identified as an uplifted and dissected surface. It has been described by Cockfield (1926) as;

> *"...a broad geosyncline, which is higher at the margins then at the centre, sloping from both the east and west towards an axis which coincides very closely with the position of the Yukon river and its tributary, Lewes river. The summits of the plateau back from the main waterways form a gently undulating plain which is broken only here and there by isolated mountain masses that rise above the general level. Into this upland surface the streams have cut broad valleys to varying depths."*

The Ruby Range and areas to the south in the Dezadeash region have been described by McConnell (1905) as follows;

> *"They probably represent erosion remnants of an old low level plain, since elevated some thousands of feet and partly destroyed."*

Bostock (1948) states that affects of Quaternary glaciations serve to separate the Yukon Plateau into two general areas of differing topography; glaciated and unglaciated. The implication being that the general landscape and characteristics of the pre-glaciated areas would be similar to the currently observed unglaciated areas. On a regional scale, the Yukon Plateau and associated geological terranes have, likely since their formation, trended in a general northwest direction. This is exemplified by the dominant northwest flow of the Yukon River which has a drainage area of about 275,000 km^2 in Yukon, most of which is comprised of the Yukon Plateau (Environment Canada, 1972). Although diversions of the Yukon River likely did occur during pre-Quaternary as well as Quaternary, there is no evidence to indicate that the historic northwest flow direction was altered in a significant way (Hughes et al., 1972).

2.6 Regional Setting: Study Area

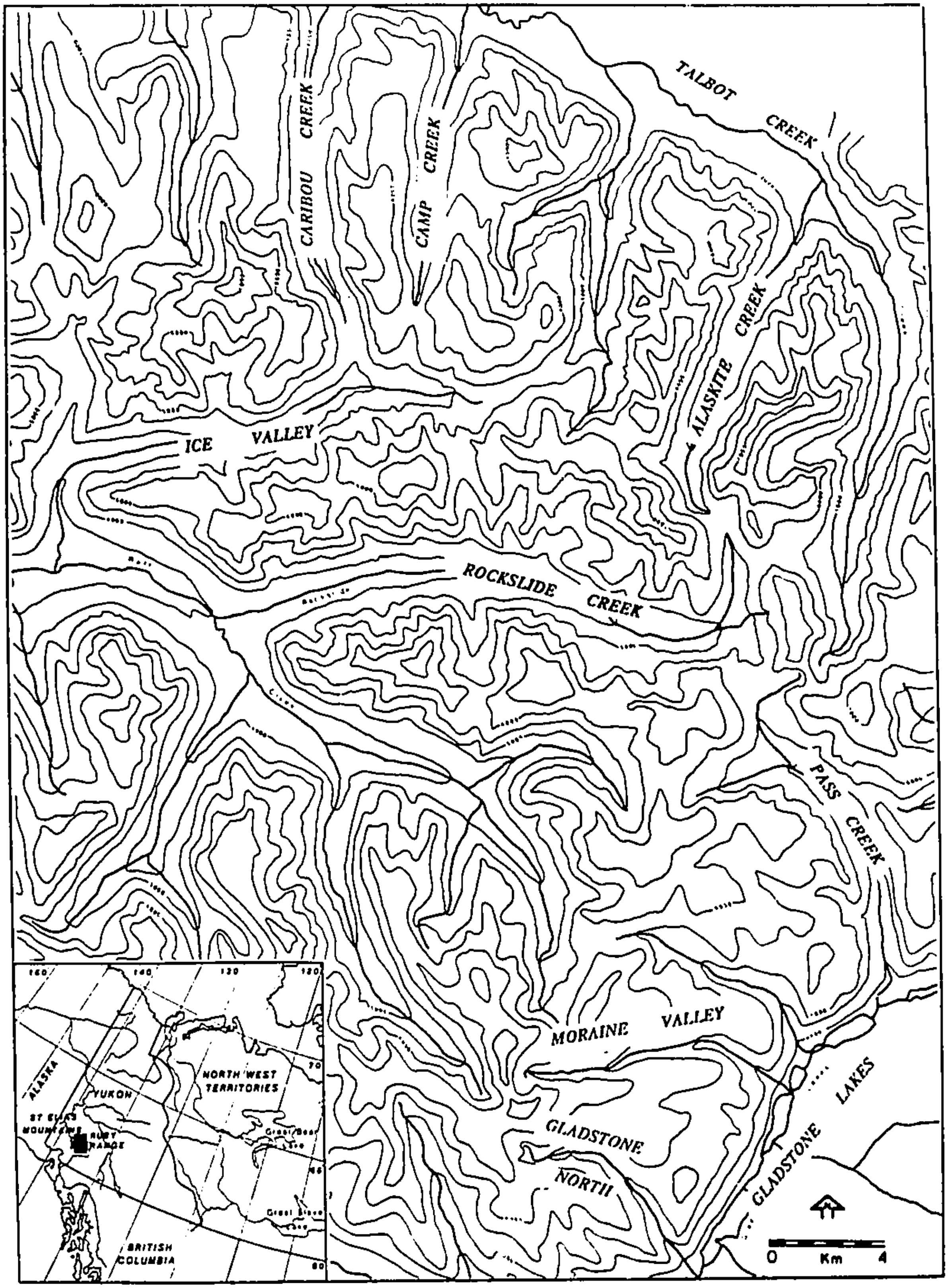
The study area falls within the Kluane Lake map area (sheets 115G, 11F E1/2), Yukon Territory and is bounded by Talbot Creek to the north (approximately 61°35'N) and Gladstone Creek to the south (approximately 61°20'N). The Talbot Arm serves as the approximate western boundary (approximately 138°38'W) and the heads of Albert and Talbot Creeks serve as the eastern boundary (approximately 138°5'W) (Figure 5). The physiography of the region, along with the identification and description of various subdivisions have been fully described by Bostock (1948, 1952) and summarized by Muller (1967), and Dewez (1988). The following synthesis draws upon these descriptions.

The two major physiographic subdivisions that dominate this region are the Yukon Plateau and St. Elias Mountains. They are separated by the Shakwak Trench which follows the Denali Fault. The northwest aligned Shakwak Trench is 8 to 16 kilometres wide and is occupied at its deepest part by Kluane Lake at an elevation of approximately 772 meters above sea level (masl).

The Kluane Ranges, which constitute the front ranges of the St. Elias Mountains, rise abruptly from the western edge of the Shakwak Trench to elevations of approximately 2550 masl. The Kluane Ranges are deeply incised by V-shaped transverse valleys. The courses of the Slims, Duke, Donjek, Koidern and White Rivers have created major breaches in the Ranges. Bordering west on the Kluane Ranges is the Duke Depression which is made up of a series of valleys and plateau surfaces at elevations of about 1515 masl.

The Duke Depression is bound to the west by the Icefield Ranges which form the backbone of the St. Elias Mountains. Vast icefields, valley glaciers and rugged prominent peaks (e.g. Mount Lucania 5196 masl; Mount Steele 5044 masl) dominate the landscape. The region northeast of the Shakwak Trench, known as the Yukon Plateau, is in sharp contrast to the St. Elias Mountains to the southwest. The divisions of the Yukon Plateau that are of concern include the Ruby Range and Nisling Range, with the valleys of Talbot Creek and Tincup Creek considered as the dividing line between the two Ranges. Various authors have described the general physiographic character of the

Figure 5
NORTHERN RUBY RANGE STUDY AREA



Yukon Plateau or regions of it. For example, Cockfield (1922) described the plateau as;

> *"...an upland surface standing at an elevation of 5000 to 5500 feet. Into this surface the streams have cut to depths ranging from 4000 feet, giving a very irregular topography. The summits of the unreduced ridges lying between waterways are the remains of a gently rolling plain, which is broken only here and there by isolated residuary masses that rise above the general level. The surface is apparently an uplifted and dissected peneplain."*

Bostock (1948), and Tempelman-Kluit (1974) characterize the Ruby and Nisling Ranges as remnants of an old uplifted erosion surface. Hughes (1990) emulates Bostock's (1948) portrayal of the Klondike Plateau (subdivision of the Yukon Plateau) in describing the Nisling Range and applies it to the entire Yukon Plateau:

> *"The topography is a maze of deep, narrow valleys separated by long, smooth-topped ridges whose elevations are very uniform and which are remnants of an old uplift erosion surface. This surface shows gentle undulations rising here and there along converging ridges to culminate in monadnocks that consist of dome-like eminences or groups of relatively smooth-sloped mountains".*

The northern Ruby Range study area exhibits a relatively subdued relief with undulating plateau surfaces and rounded mountain tops at elevations of 2120 masl. A few peaks of just over 2120 masl are evident, flanked by cirques in some locations. Tors or castellated outcrops are sparsely distributed between elevations of approximately 1675 and 2040 masl. The geomorphological significance of tors will be reviewed in subsequent discussions. The region is cut by major east-west oriented valleys (Gladstone, Rockslide, Raft, and Talbot Creeks) which are connected through a system of low passes. Some valleys contain small lakes, swamps and slow flowing, meandering streams where apparently little erosion or aggradation takes place. This is in sharp contrast with the St. Elias Mountains where rejuvenation has been considerable in late and post Pleistocene time (Muller 1967).

Cirques are evident in the area which attest to the former presence of local glaciation. Rock glaciers are located in some cirques (notably in Alaskite Creek) and on some upper valley floors (Rockslide and Gladstone Creeks). The presence of rock glaciers (which are well vegetated) suggest that during the Holocene, the region did not experience local glaciation (Johnson per comm.). Muller (1967) states that except for the disappearance of ice, the morphology of the Ruby Range has changed relatively little since the Pleistocene. With few exceptions, slopes are generally well vegetated with only isolated signs of mass movements (primarily at the heads of Rockslide, Talbot and Alaskite Creeks where the bedrock consists of Alaskite). Slope disturbances have been active at the heads of these same valleys where a mining road was forced through to locate and service core drilling operations.

The Nisling Range exhibits an even more subdued relief with undulating plateau surfaces and flat to rounded mountain tops at elevations ranging between 910 to 1820 masl. Major valleys such as Tyrrell Creek, Dwarf Birch Creek and Onion Creek are generally oriented north-south with many east-west tributary valleys. Oxbow lakes in alluvial floodplains along with smaller lakes and ponds within regions of thermokarst alluvial sediments are common (Hughes, 1990).

2.7 General Geology: Introduction

The description of the general geology of the region, emphasizing lithology and mineralogy (where available), is important to this thesis. Specifically, a review of the general geology of the Kluane Lake, Aishihik Lake and Dezadeash map areas will set the stage for a discussion on assemblages of heavy minerals in various terranes with reference to provenance implications. Most of the geological descriptions of the Kluane Lake, Aishihik Lake and Dezadeash map areas are comparable (Muller, 1967; Tempelman-Kluit, 1974; Kindle, 1953; respectively). However, Dewez (1988) pointed out that lithological divisions vary and hence the delineation of lithologies may not correspond from region to region. The descriptions which follow are obtained from Muller (1967), Tempelman-Kluit (1974), and Kindle (1953), and are complemented by summaries from Dodds (1982), Campbell and Dodds (1978), Dewez (1988), and Hughes (1990).

2.7.1 Kluane Lake Region

The Kluane Lake region (of which the Ruby Range study area is part) is divided into two distinct geological regions along the same lines as the physiographic subdivision described above; the St. Elias Mountains, and Yukon Plateau. The St. Elias Mountains consist of a eugeosynclinal assemblage of sedimentary, volcanic and intrusive rocks where fossils establish ages ranging from Devonian to early Tertiary (Muller, 1967). The Yukon Plateau is generally underlain by granitic rocks and associated migmatites and gneisses. Sedimentary rocks, where present, are metamorphosed in the low to medium grade. The St. Elias Mountains generally form part of the Alexander, Wrangellia, and Chugach Terranes. Muller (1967) has divided the St. Elias Mountains into eleven generalized formations or groups (Table 4). Subsequent work under the Geological Survey of Canada's research project *'Operation Saint Elias'* has furthered the description and delineation of lithology and structure in this region (Campbell and Dodds, 1978; Dodds, 1982).

Table 4
FORMATIONS OF THE ST. ELIAS MOUNTAINS
(KLUANE LAKE REGION)

Kaskawulsh Group	Recrystalized limestone or marble interbedded with argillite; some phyllite and greenstone.
Greenschist Complex	Slates and schists with minor massive greenstone, low grade metamorphism corresponding to greenschist facies; angular quartz and albite fragments with sericite, chlorite, calcite and minor epidote; volcanic greenstones show finely layered assemblages of epidote, zoisite, actinolite, quartz, calcite and albite.
Cache Creek Group	Subdivided into volcanic and sedimentary divisions. Volcanic division: Primarily altered tuffs and breccias; plagioclase, chlorite, epidote, some magnetite, carbonate and quartz. Sedimentary division: Contain fossils; argillite, greywacke, chert, limestone.
Basic & Ultrabasic Intrusions	Include gabbro and peridotite. Gabbro: Light green feldspar and dark green augite with some serpentine; plagioclase and diopsidic augite with diabasic texture; feldspars partly sericitized and contain epidote and chlorite; instances of hornblende being replaced by diopside; chlorite, epidote, zoisite, serpentine occur as secondary minerals. Peridotite: Ultrabasic rocks with olivine and pyroxene; minor biotite and chlorite.
Mush Lake Group	Thick assemblage of volcanic and sedimentary divisions. Volcanic division: Include basaltic and andesitic lavas and associated pyroclastic rocks; contain amphibole and pyroxene; lava flows contain plagioclase and diopside; some epidote, chlorite and sericite; diopside is present in varying amounts as are carbonate, chalcedony, magnetite, hematite, chlorite and epidote fillings. Sedimentary division: Massive limestone, thin bedded shaly limestone and shale.
Dezadeash Group	Interbedded light to dark buff grey lithic greywacke, sandstone, siltstone, thin dark grey shale, argillite and conglomerate; conglomerate consist of limestone and volcanic boulders; feldspar and minor pyroxene.
Kluane Ranges Intrusion	Granodiorite, quartz diorite and diorite; some quartz monzonite with quartz, feldspar, hornblende and biotite.
Donjek Range Intrusion	Primarily granite, quartz monzonite, alaskite (?); hornblende, quartz, feldspar, albite and biotite.
Amphitheatre Formation	Sandstone, pebbly sandstone, polymictic conglomerate, siltstone, mudstone; minor shale and coal.
St. Clare Group	Consists of a lower volcanic formation of mainly basalt with matrix of labradorite, augite and magnetite; minor apatite, hematite and epidote. Upper conglomerate formation of pebbles and boulders of lavas, volcanic breccia and tuff; fragments of labradorite, andesine and augite.
Tertiary Intrusions	Porphyritic latite, trachyte, rhyolite, gabbro; with feldspar, hematite, epidote, calcite, aegirine-augite, tourmaline, magnetite and labradorite.

Table 5
FORMATIONS OF THE YUKON PLATEAU
(KLUANE LAKE REGION)

Unit 1	Two belts of quartz-biotite schist and gneiss, some minor limestone. Containing some garnet; also feldspar, graphite, biotite and chlorite; magnetite, apatite and zircon are accessory constituents.
Unit 2	Quartz-sericite schist; banded quartz mosaic with albite and some fine-grained graphite interlayed with chlorite-sericite laminae; tourmaline is considered an accessory constituent.
Unit 3	Recrystallized limestone, mica schist and amphibolite; quartz, muscovite, chlorite, albite, hornblende, epidote; pyrite, magnetite and other opaques; minor rutile and garnet;
Unit 4	Andesitic and basaltic tuffs and breccias; plagioclase, hornblende, magnetite; some albite.
Unit 5	Ruby Range batholith: Hornblende-biotite granordiorite and quartz diorite are main components; diorite and gabbro also occur; quartz, feldspar, hornblende, amphibole, biotite, main constituents; some epidote, augite; apatite, pyrite, sphene, zircon classified as accessory minerals, constituting less than one per cent of rock.
Unit 6	Nisling Range granodiorite: Granodiorite, quartz monzonite, biotite-hornblende granodiorite; quartz, plagioclase, orthoclase, hornblende, biotite; some amphibole, pyroxene, chlorite, magnetite, pyrite and apatite.
Unit 7	Nisling Range alaskite: Alaskite, leucogranite or leuco-quartz monzonite; quartz, orthoclase, plagioclase main constituents; some biotite, pyrite and fluorite; may contain wolframite; general crumbly nature, less firmly bonded compared to granodiorite.

Units 1, 2 and 3 constitute part of the Yukon Complex (Muller, 1967).

The Yukon Plateau constitutes in part, the northern extension of the Coast Plutonic Complex. Specifically, the Ruby Range is considered part of the Tracey Arm Terrane while the Nisling Range is associated with both the Tracey Arm Terrane and Stikinia Terrane (Nisling sub-terrane). Muller (1967) divides the formations of the Yukon Plateau in the Kluane Lake region into seven units or sub-divisions. These are described in Table 5. Tempelman-Kluit (1974) stated that most of the northwest trending structural grain of the Yukon Plateau has been concealed by metamorphism and modified by intrusions of various plutonic masses. Superimposed on these structural elements are the northward trends related to intrusions of the granitic rocks. Because of the preponderance of the structureless granitic and volcanic rocks, the Yukon Plateau appears as a somewhat homogeneous structural entity.

2.7.2 Aishihik Lake Region

The Aishihik Lake region lies predominately within the Stikina and Tracy Arm Terranes. The region consists primarily of northwest trending metamorphic rocks which are assigned to the Yukon Complex. They include, for the most part, schists, granites and gneisses, along with some volcanic and sedimentary rocks (Tempelman-Kluit, 1974; Erdmer, 1991). The metamorphic rocks have been divided into a number of lithologic units, some of which may be metamorphic equivalents (Tempelman-Kluit, 1974). The descriptions provided in Table 6 are a compilation of Tempelman-Kluit's (1974) original lithological narrative of the Aishihik Lake region and summaries of these depictions by Dewez (1988), Johnson (1988), and Hughes (1990). Hughes (1990) assigns Tempelman-Kluit's (1974) Unit 2 (Marble), Unit 4 (Amphibolite), and Unit 11 (Nisling Range alaskite) as part of Unit 1 (Biotite schist), and Unit 18 (Undifferentiated volcanics) as part of Units 16 and 17 (Laberge Group and Tantalus Formation, respectively).

2.7.3 Dezadeash Region

The Dezadeash region lies entirely within the Alexander, Tracy Arm, and Stikinia Terranes. A description of the general geology of the Dezadeash region by Kindle (1952) was enhanced by Eisbacher (1975), and Erdmer (1990, 1991). Kindle (1952) states that formations of the St. Elias Mountains, the Coast Mountains, and the Yukon Plateau are all found in this region. Kindle has divided the geological formations of the Dezadeash region into a number of units (Table 7). Although the names of some units (or groups) may differ from those reported in the Kluane Lake region (Muller, 1967) and Aishihik Lake region (Tempelman-Kluit, 1974), many lithological descriptions of these units correspond. More recent studies (e.g., Erdmer, 1989), have served to augmented Kindle's (1952) description and delineation of formations.

Table 6
FORMATIONS OF THE AISHIHIK LAKE REGION

1. **Biotite schist:**	Coarsely crystalline muscovite-biotite schists and micaceous quartzites containing garnet; some garnetiferous amphibolite and marble; equivalent to Muller's Yukon Complex Unit #1.
2. **Marble:**	Occurs as small lenses; interfoliated with biotite schist; contains grossularite, idocrase, epidote and diopside in some places; minor calcite.
3. **Hornfelsed schist:**	Homogeneous unit containing cordierite, staurolite, quartz, biotite, muscovite, plagioclase, chlorite, graphite, tourmaline, andalusite; garnet is rare.
4. **Amphibolite:**	Black-green rock composed almost entirely of actinolite; some garnet.
5. **Ruby Range granodiorite:**	Equivalent to Muller's (1967) Ruby Range batholith unit (see #5 above); composition approx. 25% quartz, 15% potash feldspar and 50% plagioclase; hornblende and biotite also occur in roughly equal amounts.
6. **Hornblende granodiorite:**	Composition ranges from quartz monzonite to quartz diorite; composition approx. 40% andesine, 25% quartz, 15% potash feldspar and 12% hornblende; sphene is common; minor biotite.
7. **Pink quartz monzonite:**	Contains quartz, potash feldspar and albite; some biotite; hornblende may be as much as 10%; minor pyrite, pyrrhotite and hematite.
8. **Porphyritic quartz monzonite:**	Quartz, biotite, hornblende with feldspar.
9. **Hornblende diorite:**	Fine textured; consisting of plagioclase, hornblende and biotite; apatite and magnetite are accessory minerals; minor epidote and chlorite.
10. **Coffee Creek granite:**	Biotite granite to quartz monzonite; contain equal amounts of quartz, potash feldspar and albite; some biotite; Muller's (1967) Nisling Range alaskite, unit #7 includes equivalents of the Coffee Creek granite.
11. **Nisling Range alaskite:**	Miarolitic leucocratic granite; quartz and potash feldspar are main constituents; some biotite and hornblende; minor fluorite; the Coffee Creek granite and Nisling Range alaskite correspond to Muller's (1967) Nisling Range alaskite (see unit #7).
12. **Feldspar porphyry:**	Occur as dyke and flow rocks of intermediate to acid composition; may include Nisling Range alaskite; feldspar phenocrysts are common; hornblende also common; minor biotite.
13. **Mount Hansen Group:**	Refers to acid volcanic rocks; uniform appearance and show no texture or structure on fresh surfaces; tuffs and tuff-breccias fragments.
14. **Varicoloured acid tuff:**	Tuffs and minor flow rocks; texture varies; quartz crystals; some biotite and hornblende.
15. **Massive green vocanics:**	Aphanitic, massive and structureless flow rocks with some tuffs; dark green epidotized basalt.
16. **Laberge Group:**	Folded sandstone and conglomerate; interbedded brown shale and thin pebble conglomerate beds; sandstones composed of rock fragments, feldspar and quartz.
17. **Tantalus Formation:**	Chert pebble conglomerate with minor interbedded sandstone and shale.
18. **Undifferentiated volcanics:**	Quartz feldspar porphyry dyke and flow rocks; altered andesite and basalt; contain feldspar and augite; some labradorite.
19. **Carmacks Group:**	Brown augite olivine basalt and flow breccia; includes altered anddesite and basalt; contains feldspar and augite.
20. **Little Ridge volcanics:**	Basalt and flow breccia; some labradorite, minor augite and zeolite.

Units 15 - 20 are described as non-metamorphed volcanic and sedimentary rocks occuring in the Aishihik Lake region.

Table 7
FORMATIONS OF THE DEZADEASH REGION

1. Yukon Group:	Consists primarily of quartz-mica schists, gneiss, slate, quartzite, crystalline limestone; minor quartz-sericite-andalusite schist, hornblende schist, sericite-cordierite schist, garnetiferous schist and gneiss.
2. Kaskawulsh Group:	Crystalline limestone with intercalated zones of slates, quartzite and argillites; some chert, andesite and schist.
3. Mush Lake Group:	Thick assemblage of volcanic and sedimentary rocks; andesite, basal, tuff, dacite, rhyolite, volcanic breccia; crystalline limestone, slate, greywacke, argillite.
4. Dezadeash Group:	Sedimentary rocks comprise slate, greywacke, argillite, quartzite, chert, limestone, grit, conglomerate, tuffaceous sandstone and bedded volcanic tuffs; minor intrusive bodies of peridotite, serpentine, and granodiorite.
5. Peridotite and serpentine:	Finely crystalline with gradations between dark peridotite and green serpentine; some crystals of chrome diopside with olivine; constituents of magnetite, iron, chromium.
6. Gabbro:	Hornblende gabbro with magnetite and pyrite.
7. Coast (granitic) intrusions:	Granodiorite, granite, diorite, porphyritic granite, augen-gneiss, gabbro; quartz, andesine feldspar, orthoclase, biotite, hornblende, copper, lead, zinc, magnetite, gold.
8. Granite porphyry, quartz porphyry:	Appear as dykes and stocks; contain pyrite, biotite, gold.
9. Conglomerate, sandstone, shale:	Thick massive beds of conglomerate separated by layered sandstone and shale; conglomerates contain granite, quartz, greywacke, slate, quartzite, limestone, greenstone, red anesite, and schist; some coal.
10. Soda Syenite:	Fine grained massive crystalline rock; quartz, orthoclase, oligoclase.
11. Volcanics:	Mountain cap rocks; volcanic breccia, tuff, rhyolite, dacite, andesite, basalt and some sandstone; contain iron, hematite, limonite.

2.8 Regional Geological Synopsis

There are two major physiographic and corresponding geologic subdivisions in southwestern Yukon which are separated by the Shakwak Trench; the St. Elias Mountains and the Yukon Plateau. The Kluane Lake, Aishihik Lake and Dezadeash regions constitute part of these subdivisions. The St. Elias Mountains form part of the Alexander, Wrangellia, and Chugach Terranes, and consist of a eugeosynclinal assemblage of sedimentary, volcanic and intrusive rocks which are subdivided into eleven general formations or groups.

The Yukon Plateau constitutes in part, the northern extension of the Coast Plutonic Complex. Specifically, in southwestern Yukon, the Ruby Range is considered part of the Tracey Arm Terrane while the Nisling Range is associated with both the Tracey Arm and Stikinia Terranes (Nisling sub-terrane). The Yukon Plateau is underlain primarily by granitic rocks and associated migmatites and gneisses. Sedimentary rocks, where present, are metamorphosed in the low to medium grade. Granitic rocks occupy about half the area of the Yukon Plateau. The Ruby Range batholith is the largest granitic body in the region and underlies most of the study area. It consists primarily of hornblende-biotite granodiorite and is considered the backbone of the plateau. Also found in some locations in the Ruby Range are stocks of alaskite, and recrystallized limestone, mica schist and amphibolite which occur as roof pendants.

The Nisling Range is characterized by smaller scale intrusions of granodiorite and alaskite. The Nisling Range granodiorites differ from their Ruby Range counterparts in terms of grain size and colouring. Nisling Range granodiorites are medium to coarse grained and darker in colour than the Ruby Range equivalents, primarily due to a darker tone of feldspars and large biotite flakes. Nisling Range alaskites are somewhat light coloured than the Ruby Range equivalents, and are described as fine to coarse mineral grains which are less firmly bonded than the granodiorites. The low percentage of biotite and hornblende attribute to the light colouring.

Chapter 3
THE QUATERNARY IN SOUTHERN YUKON

3.1 Introduction

Since the effects of glaciation(s) (specifically, glacial erosion and deposition being possible agents or processes responsible for the heavy mineral and geochemical compositional trends), are a central theme to this thesis, current themes on the glacial history of the region need to be reviewed. In this regard, it is important to establish the Quaternary framework from which observations and assertions regarding the likelihood that dispersal patterns of elements and heavy minerals were glacially derived can be made.

This discussion consists of a review of; i) the Quaternary framework of North America, ii) the Cordilleran Ice Sheet (glacier) concept, iii) the general glacial history of southern Yukon, iv) glacial sequences in southwestern Yukon, and southeastern and central Alaska, v) Quaternary correlations between regions in southern Yukon and southeastern/central Alaska, and vi) Pleistocene glaciation in the Ruby Range.

3.2 Quaternary Framework

The Quaternary is the most recent major period of the geological record and extends up to, and includes present day. The Quaternary includes two intervals of epoch status, the Pleistocene and Holocene. It is generally recognized that the Pleistocene spanned a period of approximately 2 million years B.P. to 10,000 years B.P. (van Eysinga 1978; Berggren et al., 1980; Lowe and Walker, 1984). The Holocene, which commenced approximately 10,000 years B.P. is still current (INQUA 1969; van Eysinga 1978; Berggren et al. 1980; Fairbridge and Michel 1980). However, it has been argued that the Holocene can be seen as part of the Pleistocene since it is a current warm interval that is part of a long-term climatic cycle and is comparable to previous warm episodes of the Quaternary (West, 1977). To most researchers, Quaternary and Pleistocene are used interchangeably and have long been considered synonymous with widespread glaciation (Forbes 1846; Cooke 1972, 1984; Fullerton 1980; Clayton and Moran 1982; Lowe and Walker, 1984). This practice and interpretation is adopted in this thesis.

The conventional division of the Quaternary is into glacials and interglacials, with further subdivision into stadial and interstadial episodes. Glacials are generally considered to be lengthy cold phases during which the major expansions of ice sheets and glaciers took place. Interglacials are usually considered to be warm intervals during which temperatures at the thermal maximum were as high or higher than those experienced during the Holocene. Stadials are regarded as shorter cold episodes in which local ice advances occurred. Interstadials are considered as relatively short-lived periods of thermal improvement during a glacial phase (Lowe and Walker, 1984).

The classical scheme of glacials and interglacials was formulated on evidence found in central North America. Chamberlain (1894), and Leverett (1896) have been credited with formulating the original interpretation. The classification is based primarily on till sheets, the morphostratigraphy of landforms, and soil stratigraphic units. It was greatly influenced by the Alpine Model devised by Penck (1882) for central Europe. Based on these criteria, the Quaternary (Pleistocene) is subdivided into a series of glacial and interglacial periods (Flint, 1971). At least four major glaciations are proposed for the Pleistocene with three intervening interglacial periods.

3.3 Concept of the Cordilleran Ice Sheet (Glacier)

Pleistocene studies in the Canadian Cordillera have produced many descriptive accounts of glaciation (for example, Dawson, 1891; Kerr, 1936; Beach, 1943; Davis and Mathews, 1944; Bostock, 1948; Mathews, 1955; Wilson et al., 1958; Belyea, 1960; Crandell, 1965; Lemke et al., 1965; Fulton, 1967; Prest et al., 1968; Hughes et al., 1969; Flint, 1971; Roed, 1975; Luckman and Osborne, 1979; Holloway et al., 1981; Kearney and Luckman, 1981; Kostaschuk and Smith, 1983; Porter et al., 1983; Clague, 1989). It is not the purpose of this discussion to detail the Pleistocene throughout the entire Cordillera, but rather to highlight some benchmark Quaternary contributions and focus primarily on the southern Yukon region of the Canadian Cordillera.

Repeatedly in late Tertiary time, during the Quaternary glaciers spawned and enveloped vast regions of the Canadian Cordillera. Only certain regions of central and northern Yukon, the eastern Mackenzie Mountains, and some peaks escaped glaciation (Clague, 1989). The first significant studies of the Quaternary of the Canadian Cordillera were made by George M. Dawson (Geological Survey of Canada) in the later part of the Nineteenth Century. In his many reports, Dawson (1877, 1881, 1889, 1891, among others) recognized that much of British Columbia and adjacent areas (particularly Yukon) had at one time been glaciated on a very large scale. In his most comprehensive paper on glacial geology, Dawson (1891) detailed his proposal of a Cordilleran glacier of continental magnitude:

> *"Though at first in doubt as to the probable origin of the traces met with of this first and most general epoch of the glaciation of the Cordillera, much additional information gained in later years convinced me that it clearly indicated the former existence of a great glacier-mass such as that here described. Still more recent observations have proved the north-western movement of the northern part of the great glacier, and after having thus ascertained its area, I venture to designate it as the Cordilleran glacier."*
> *(Dawson, 1891; page 28)*

This glacier he argued, originated in the Coast Mountains of northwest British Columbia, expanded and coalesced with other valley glaciers and local ice caps to form a continuous cover over plateaus and coastal lowlands. Once fully established, the ice sheet had a dome-like central area in northern British Columbia from which ice flowed primarily northwest and southeast:

> *"...when the Cordillera became completely buried, a general movement was initiated from a region situated between the fifty-fifth and fifty-ninth parallels of north latitude, in a south-easterly and north-westerly bearing."*
> *(Dawson, 1891; page 27)*

Dawson was also first to suggest that the entire Cordillera from the 48th parallel to regions near the 63rd parallel (in Yukon) were completely engulfed in a dome-like glacier, analogous to that of Greenland. It extended off the west coast to Vancouver Island, approached the Queen Charlotte Islands, and reached into Alaska. Subsequent investigations (e.g., Fulton, 1984) showed that Vancouver Island and the Queen Charlotte Islands apparently developed their own ice caps early during the last glaciation. However, Vancouver Island was later overrun by flow from the mainland Cordilleran ice sheet whereas the Queen Charlotte Islands remained essentially independent of mainland Cordilleran ice (Clague, 1981).

Sutherland Brown (1968) stated that ice originating from the Queen Charlotte Islands flowed eastward where it coalesced with the mainland Cordilleran ice and was deflected north towards Alaska, and south down the coast of British Columbia. Dawson (1891) maintained that the northwest movement of the Cordilleran glacier into southern Yukon from the main source area in the Coast Mountains in northwestern British Columbia was far less extensive than the corresponding southeast movement. With reference to observations made in southern Yukon, Dawson states:

> *"...evidence was there obtained of its northward or north-westward direction of movement, and this has been confirmed and added to by observations in surrounding regions... On the Lewes and Pelly Rivers, branches of the great Yukon River, striated rock-surfaces, evidently due to the general Cordilleran glacier, were noted; in the case of the first mentioned river as far north as latitude 61° 40' on the Pelly to latitude 62° 30', longitude 135° 45'... Granting that the north-western extremity of the Cordilleran Glacier reached the furthest point above assigned to it, we find that its extension from the central gathering ground (or from the approximate margin of this gathering ground already given) was much shorter than that attained by the south-easterly flowing part, the approximate lengths being 350 and 600 miles respectively."*
> *(Dawson, 1891; pages 29-30)*

He attributed the variation of the northwest and southeast movements to greater relative topographic controls in the northwestern part of the continent which would inhibit glacial movement, and/or a reduced supply of precipitation in these regions.

The remarkable insights of Dawson (e.g., 1877, 1881, 1889, 1891) lead the way for more detailed study of Cordilleran Ice Sheet growth and decay. In a comprehensive summary of the Quaternary Geology of the Canadian Cordillera, Clague (1989) advances a successional four-phase model for the growth of the Cordilleran Ice Sheet:

Phase 1: Alpine Phase of Glaciation: At the end of each major Pleistocene nonglacial period, glaciers were restricted to high mountain locations of the Cordillera. As climate deteriorated during the early part of each glaciation, small mountain ice fields grew and alpine glaciers advanced.

Phase 2: Intense Alpine Phase: With continued cooling and increased precipitation, glaciers expanded and coalesced to form a more extensive cover of ice in mountain areas.

Phase 3: Mountain Ice Sheet Phase: During sustained cold periods, glaciers advanced across plateaus and lowlands and eventually grew into large confluent masses covering much of British Columbia and adjacent regions.

Phase 4: Continental Ice Sheet Phase: Ice thickened to such an extent that one or more domes with surface flow radially away from their centres became established over the interior of British Columbia.

Although the model provides a useful framework for conceptualizing the growth of the Cordilleran Ice Sheet, the actual glacial development in the Canadian Cordillera during the Pleistocene was likely more complex because ice did not build-up in a uniform or synchronous manner. Periods of growth were certain to be interrupted by intervals during which glaciers stabilized and/or receded. These fluctuations were controlled mainly by global climatic changes and, to a lesser degree, by local and regional factors induced by glaciation such as eustatic sea level lowering, ocean cooling, and changes in local atmospheric circulation brought about by glacier expansion (Clague, 1989).

Wilson et al., (1958) estimated that when the Cordilleran Ice Sheet was fully formed, it was upwards of 900 kilometres wide and 2000 to 3000 meters thick over much of the Stikine and Interior plateaus in British Columbia. In southern Alaska, the ice sheet consisted of ice fields and large valley and piedmont glaciers that flowed from their mountain source areas across the continental shelf to the south and into the low regions drained by the Yukon River to the north (Hamilton and Thorson, 1983; Clague, 1989).

Flint (1971) stated that ice was thicker and more extensive on the Pacific side as opposed to the interior because of a greater supply of snowfall. Most of interior Alaska and central-northern

Yukon were dry throughout the Quaternary and consequently remained unglaciated (Clague, 1989). The main source areas of the Cordilleran Ice Sheet in southern Yukon was the Selwyn and Cassiar Mountains with some supply from the St. Elias Mountains (Hughes et al., 1969). Each period of deglaciation during the Quaternary was characterized by both complex frontal retreat in peripheral glaciated areas and downwasting accompanied by widespread stagnation throughout much of the interior regions (Rice, 1936; Fulton, 1967; Tipper, 1971; Clague, 1981; 1989). A four stage process of deglaciation has been proposed for areas of low and moderate relief in the more interior regions of the ice sheet in northwestern British Columbia (Fulton, 1967).

Stage 1. Active Ice Phase: Regional flow continued but diminished as ice thinned.

Stage 2. Transitional Upland Phase: Highest uplands appeared through the ice sheet but regional flow continued in major valleys.

Stage 3. Stagnant Ice Phase: Ice was confined to valleys but was still thick and extensive enough to flow.

Stage 4. Dead Ice Phase: Valley tongues thinned to the point where plasticity was lost.

At the end of each glaciation the western periphery of the Cordilleran Ice Sheet became unstable, likely because of eustatic rise in sea level (Clague, 1989). As a result, the British Columbia continental shelf was rapidly deglaciated. Southern Yukon also experienced relatively rapid frontal retreat. Active glaciers probably persisted in some mountain valleys. These glaciers may also have coexisted with large remnant dead ice masses on some plateau regions. The decay of the Cordilleran Ice at the end of each glacial cycle was interrupted repeatedly by stagnation and readvances. Most of these fluctuations affected relatively small areas and were not necessarily synchronous from one region to another, and hence are attributed to local factors as opposed to global climatic changes (Shaw, 1972; Luckman and Osborne, 1979; Clague, 1989).

3.4 Glaciation in Southern Yukon: An Overview

Southern Yukon provides a unique environment for the study of Quaternary history. The glacierized St. Elias Mountains are in sharp contrast to the south-central and northern Yukon which were not glaciated during the Wisconsin and likely throughout the Pleistocene (Bostock, 1966; Hughes et al., 1969; Fulton, 1984; Clague, 1989). Eastern Yukon shows signs of glaciation from both Cordilleran and Continental ice sources. Evidence of pre-Quaternary Cordilleran glaciation has been reported in the St. Elias Mountains, parts of the Yukon Plateau, and bordering regions of Alaska.

Table 8

REFERENCES TO PLEISTOCENE GLACIATION IN YUKON

Sharp (1943), on the geology of the Wolf Creek area, St. Elias Mountains:

Deposits of two Wisconsin substages are mapped and designated for convenience as earlier and later Wisconsin... The younger substage is latest or possibly post-Wisconsin and includes moraines of two separate but closely related phases.

Kindle (1945) along the Canol Road:

Bedrock on the mountain top (elevation 4,600 feet) on the east side of Quiet Lake is rounded and polished by glacial ice, and rounded granite boulders lie upon greenstone and schists. The slopes are mantled with boulder clay, and the pass at mile 65.6 is underlain by a thick mass of glacial till. four mountain peaks on the west side of Quiet Lake range from 6,066 to 6,642 feet high. These mountains show a rounded, glaciated aspect below an elevation of 6,000 feet.

Abandoned glacial cirques that bottom in U-shaped valleys at an elevation of 4,800 feet were seen on the north side of 6,000-foot peaks, 3 miles west and southwest of mile 82. These and a high, strongly glaciated, U-shaped valley that drains north and west to Ross River from 4 miles east of mile 78 are evidence of ice accumulation there during the glacial period. Other cirques, also free of ice, were noted on the northwest sides of high granite peaks west from miles 96 to 98.

Alpine glaciers were active at one time on the north slope of the mountain west of mile 116, but the mountain peaks nearer the Pelly, between miles 117 and 130, are in a drier belt and appear to have escaped glaciation. Lapie Lakes and Rose Lake are giant kettle-holes that were occupied by large masses of stagnant ice during the late stages of valley glaciation. These lakes are roughly 3,500 feet above sea-level. An esker that formed along the west side of Rose Lake and upper Lapie Lake rises to 125 feet above the water. Sand, gravel and boulder terraces several hundred feet wide occur in a number of places along the valley sides near Rose Lake and Lapie Lakes, and rise about 250 feet above the level of the lakes. They are believed to have been formed by running water when the valley was filled to an equal or higher level with melting ice and glacial debris, with melting most pronounced along the flanks of the disintegrating ice mass. A few other terraces, noted farther down the valley as far as mile 121, are located at the mouths of tributary streams and represent an accumulation due to damming by ice or till deposits in the main valley.

Throughout most of its length Pelly River occupies a trench-like depression that trends north 55 degrees west for several hundred miles... During the glacial period the trench was filled by ice that moved northwesterly down grade. The low mountains to the northeast were ice covered at that time, as proved by the occurrence of glacial striae crossing the top of the ridges north of mile 146. Striae and deep grooves seen there at an elevation of 3,500 feet strike north 60 degrees west. Glacial ice brought much debris into the trench and filled the lower hollows and ravines in the low mountainous sectors to the northeast, whereas the tops of these mountain ridges were rounded and scraped bare of most of their soil. Terraces of gravel that occur at various places along the sides of Pelly Valley probably formed in streams flowing along the border of the melting glacial ice during stages in its recession, and ponding of some of the streams resulted in deposition of silt beds in many places along the valley walls. The country adjoining the road between Pelly River (mile 141) and Sheldon Lake (mile 220) is much like some rugged areas of the Canadian Shield, with rock-ribbed, glaciated hills rising from a few hundred to 1,000 feet above a valley bottom that is heavily mantled with glacial deposits.

Two cirques seen on the north side of Sheldon Mountain are now free of ice. On the north and northeastern slope of Sheldon Mountain perched boulders of large size were observed to elevations of 4,600 feet, and glacial striae seen below this elevation and at many places in the valley below have common strike of about 10 degrees north of west.

The highest peak of Itsi Mountains is judged to be more than 7,500 feet above sea-level. Large masses of glacial ice occur in cirques along the north sides of these mountains... This was evidently a great gathering ground for ice during the Pleistocene epoch, and the ice moved down the valley of the South Branch of Macmillan River. Glacial striae on the rounded mountain top at elevation 4,600 feet, north of mile 251, trend south 25 degrees west.

All of the valleys traversed by the Canol Road southwest of Macmillan Pass are carpeted by deposits of boulder clay left by great valley glaciers of the last glacial epoch.

Bostock (1948), on the physiography of southern Yukon:

Liard Plateau:	*Most off the plateau appears to have been covered by Pleistocene ice, and lower parts seem to have been well scoured.*
Stikine Ranges:	*In Pleistocene time the ice covered them to an elevation of 6,500 feet. The upper levels show only light glacial erosion, but the main valleys are deeply scoured.*
Nisutlin Plateau:	*Pleistocene ice probably covered all of it.*
Dease Plateau:	*Its upland surface appears to be one with that of Nisutlin Plateau to the north.*
Stikine Plateau:	*Nearly all of the Stikine Plateau was covered by Pleistocene ice to a height of about 6,500 feet, only a few isolated mountains in the plateau areas being high enough to project above it.*
Liard Plain:	*The greater part of its lower areas is mantled by thick deposits of glacial drift for which it formed a general area of deposition during the close of Pleistocene time. The drift is more than 200 feet thick in many places. Glacial and post-Glacial stream courses, marked by eskers, abandoned canyons, and broad gravel beds extend across it southward towards Liard River and eastward on both sides of the river, but notably on the north side.*
Hyland Plateau:	*All parts of the Hyland Plateau were overridden by Pleistocene ice, but in contrast with the Liard Plateau, which appears to be relatively lacking in drift deposits, Hyland Plateau, like Liard Plain, was an area of deposition. A mantle of glacial drift is spread over much of it, and great trains of glacial outwash clog the floors of the main valleys.*
Logan Mountains:	*The river valleys are deeply entrenched between the ranges, and their floors are a mile or so wide in the central part of the group and widen southward. In Pleistocene time these valleys were scoured in their northern parts, which are relatively bare of drift, but further south are widely strewed with glacial deposits. The higher parts of this southern group have been much carved by alpine glaciers, and today they contain many small glaciers and ice-fields, two or three more than 3 miles wide.*
Hess Mountains:	*In Pleistocene time the ice reached everywhere to at least an elevation of 6,000 feet on the west flank of Hess Mountains, but most of their peaks appear to have stood above it. It gathered in alpine ice-fields and glaciers on the peaks, coalesced on the lower levels into a sheet studded with nunataks, and flowed outward, mainly westward, in valley glaciers.*
Wernecke Mountains:	*In Pleistocene time the ice moved down the valleys to the southwest and north from the Wernecke Mountains. About 50 miles to the southwest, beyond the mountains, the surface of the ice lay as high as 5,000 feet above sea-level. To the north, where it left the mountains in Wind River Valley, the surface reached an elevation of about 3,000 feet, and was at least 1,000 feet higher in Braine Pass.*
Ogilvie Mountains:	*In their interior parts, Pleistocene ice formed local ice-fields and valley glaciers, which did not extend beyond the south borders of the mountains, and decreased in extent and number to the west... It is believed that the effects of Pleistocene glaciation dwindled northward and westward, though to the northeast some valley glaciers pushed to near the borders of the mountains.*
Yukon Plateau:	*Another feature not noted further south, and which separates Yukon Plateau into two general areas of differing topography, is the limit of glaciation. Both in detail and in its larger features the topography is strikingly different in the glaciated and unglaciated areas... the Western Yukon Plateau includes glaciated areas in its southern part, but its main distinction is that it contains the unglaciated country southwest of Tintina Valley.*
Tintina Valley:	*Differential uplift and the advance of Pleistocene ice across the valley from the east have both played important parts in forcing its drainage to escape to the west.*

Table 8 (continued)

Bostock (1948), cont.:	
Pelly Plateau:	*The entire plateau, except, perhaps, for some isolated mountains in its southeast part, was covered by Pleistocene ice. In general the ice moved westerly, deflected locally by topography, but some ice north of Frances Lake may have escaped southward.*
Macmillan Plateau:	*Pleistocene ice filled the valleys, but many of the mountainous areas between them, particulary those in the northwest, stood above the glaciers as nunataks.*
Stewart Plateau:	*All of Stewart Plateau is broken into tablelands by a network of deeply cut, broad valleys. Some of these tablelands are remarkably level and little dissected, the streams on them flowing in open valleys with relatively gentle gradients. Others, notably in the western part of the plateau, north of McQuesten River, which is mainly unglaciated, are, as a whole, deeply and intricately dissected by their streams, and tablelands are almost non-existent... This type of dissection is typical of the unglaciated regions, but is lacking in the glaciated areas of Yukon Plateau. In the east part of Stewart Plateau, and in other, similar glaciated areas, glaciation seems to have been a major factor in developing the present topography.... Drift deposits, attributed to glaciation that antedated the last advance of the ice, have been found by the writer in two places in this plateau. These are remnants of glacial till in the upper part of Dublin Gulch, and large patches of drift on the north side of McArthur Range. Both places are beyond the limit of the last glacial advance and at high level.*
Teslin Plateau:	*All but perhaps the highest peaks of Teslin Plateau, those more than 6,500 feet high, were covered by Pleistocene ice, which moved northward, except along the southern fringe of the plateau where it may have moved to Taku River.*
Nisutlin Plateau:	*Pleistocene ice probably covered all of it.*
Lewes Plateau:	*Several small areas north of Little Salmon River and, perhaps, in Miners Range appear to have escaped glaciation in Pleistocene time, but elsewhere the plateau was completely covered by ice.*
Pelly Mountains:	*In Pleistocene time the main ice surface of the last glaciation stood at an elevation of more than 6,000 feet in most of Pelly Mountains and, perhaps, considerably higher on the southwest side of St. Cyr Range. Its surface was lower westward along that side of Big Salmon Range where ice was moving west and northwest. The northeast side of St. Cyr Range is an area of relatively light precipitation compared with the southwest side, and some features suggest that the ice was lower, perhaps as much as 1,00 feet lower, on that side, or that it was an area where the ice had little movement. A large part of Glenlyon Range stood above the main surface of the ice as nunataks. Alpine glaciers extended northeastward from the higher parts of these nunataks to coalesce with the main ice moving northwest down Tintina Valley. Similar conditions prevailed in parts of Big Salmon and St. Cyr Ranges, but the general level of the ice rose to the south and southeastward and less of these ranges stood above it.*
Klondike Plateau:	*Pleistocene ice, on its last advance, pushed against the east side of the southeast end of the range, where it reached a maximum elevation of nearly 4,000 feet. Cirques in the valley heads around Apex Mountain are evidence of a group of small alpine glaciers in Pleistocene time... At the southeast end of Dawson Range, in the valley of Victoria Creek beyond the limit of the last glaciation, there is evidence of earlier and somewhat more extensive glaciation in the form of remnants of glacial till beneath pay-streaks of placer deposits.*
Kluane Plateau:	*Pleistocene ice reached to a height of about 5,000 feet above sea-level in the southern part of the basin, and moved northwestward along Aishihik Lake to near Nisling River. The ice also overflowed northeastward through the gaps in the divide between Aishihik Lake and Nordenskiold River. A great part of this basin is drift covered... The high upland surface, notable here for its broad, rolling character, changes to a much dissected surface of the Klondike Plateau type where it slopes to Nisling River. During the last glaciation the ice overrode the lower parts of the southwest side of the upland and spread through the entrenched valleys, splitting and reuniting its tongues in the valleys of the trough, but it appears to have advanced through only three valleys in Nisling Range sufficiently far to reach that of Nisling River.*
Shakwak Valley:	*In Pleistocene time the valley formed a large trough where the ice from the great glaciers of St. Elias Mountains coalesced and spread before it pushed on through the gaps in the ranges to the north. As a consequence, it has been heavily scoured in its narrow parts and elsewhere mantled by widespread drift deposits.*
Boundary Ranges:	*In Pleistocene time, glaciation was very active in Boundary Ranges, and ice probably covered all but the highest parts. At its maximum this ice moved westward and eastward from near the summits of the ranges, much of that moving eastward subsequently turning to escape west through the great valleys such as that of Stikine River. Some ice that pushed east into the Central Plateau and Mountain area escaped northward from the neighbourhood of Atlin Lake in the north and, judging by the configuration of the valleys as they appear from air photographs, southward in the south.*
St. Elias Mountains:	*In Pleistocene time the ice-fields of St. Elias Mountains were higher, and pushed their glaciers out on all sides. On the interior side they extended far out on Kluane Plateau. In all the higher parts of Kluane, Alsek, and Donjek Ranges there appears to have been an upper limit of the ice in some part of later Pleistocene time at an elevation of about 6,000 feet, but with considerable variation in level from place to place due to the influence of local topography on the movement of the ice.*

Bostock (1952), on the geology of northwest Shakwak Valley:

Evidence of Pleistocene Glaciation is widespread in most of the region mapped. It is lacking in the extreme northwest where the ice did not reach, and on high summits that stood above the ice. Only one major advance has been recognized, and the glacial phenomena referred to here are attributed to this last well-marked major glaciation, which is tentatively considered late Wisconsin. It is probable, however, that earlier and more extensive advances took place in this region, as in areas farther north, and that in a few places features attributed to the last advance may represent an earlier one...

The extent and maximum elevation of the ice surface during the last well-marked glaciation have been recorded in many localities. They were greatest in the St. Elias Mountains...

In the Kluane Hills and Ruby Range east of Kluane Lake, the ice surface reached elevations of 6,150 and 6,200 feet, but many large summit areas here were not covered...

The main courses of the ice are indicated by a few striae and by drumlin-like features in the main valleys. Most of the ice came from the Icefield Ranges, entering Shakwak Valley by the valleys of Slimms, Donjek, and White Rivers. It spread into Shakwak Valley, moving northwesterly in most places to where it could escape northward along channels through the northeast side of the valley, such as those of the north arms of Kluane Lake, Kluane, Donjek, and White Rivers, and other intervening gaps. On its passage northward from Shakwak Valley the ice divided into many valley glaciers, spreading around mountains and hills too high to override, and finally lost movement in the mountain valleys northeast of the region mapped.

The earliest reports of glacial phenomena in southern Yukon were by geologists in the late 19th and early 20th Centuries, inspired in part by the discovery of the rich Klondike placer gold fields in 1896 (e.g., Dawson, 1889; Hayes, 1892; Tyrrell, 1898; Brooks, 1900, 1906; McConnell, 1905, 1905; Camsell, 1906; Cairnes, 1907, 1915; Capps, 1910; Tarr and Martin, 1914; Cockfield, 1926). Although the recording of glacial phenomena was not a priori in the field, it was often mentioned in the objectives of field work.

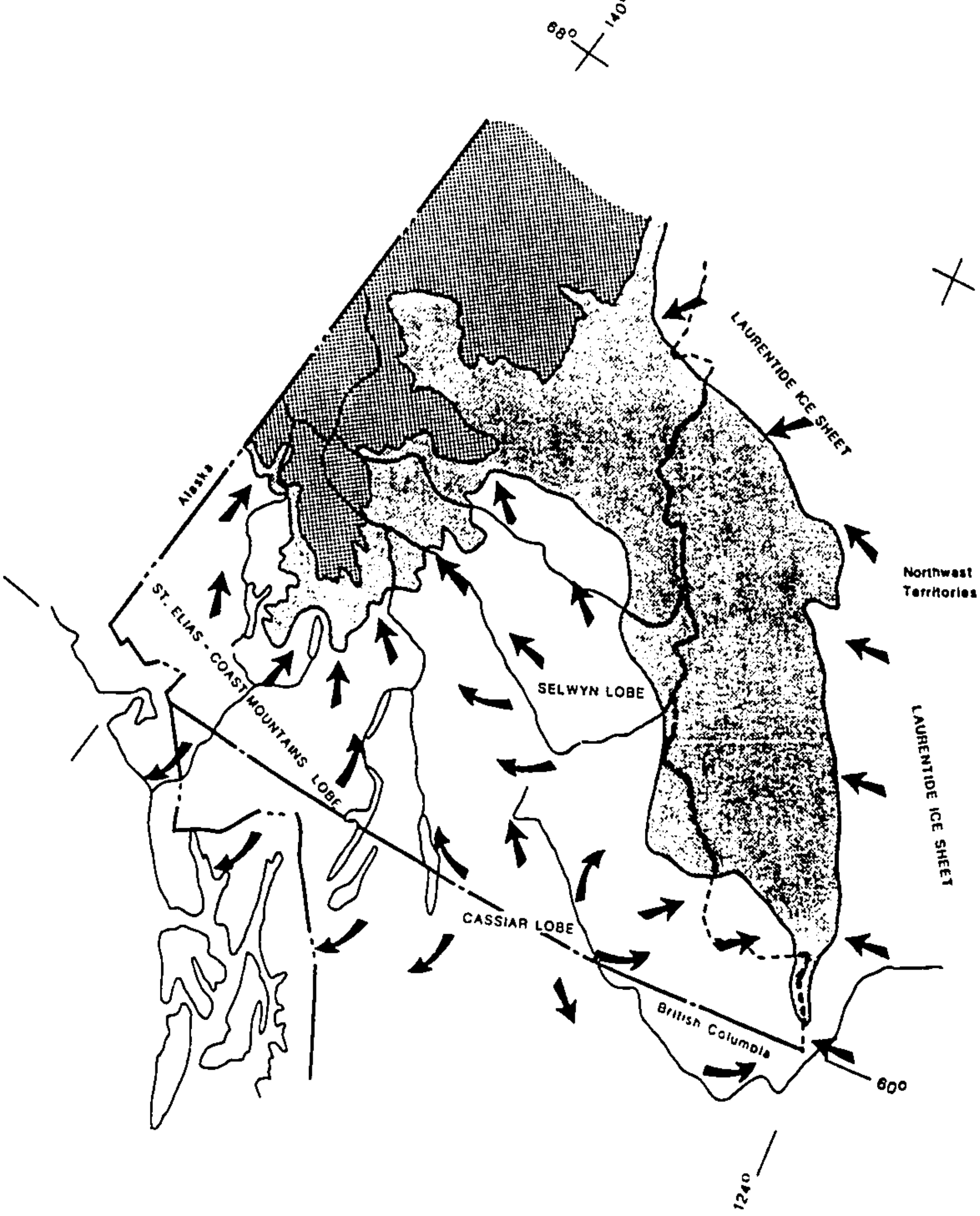
> *"The principal object of your exploration will be to obtain as much geological and general information as possible... The glacial and other superficial deposits will not escape your attention as these are likely to have intimate relations to the occurrence of placer gold."*
> *(Tyrrell, 1889)*

The impact of operations in far northern areas during and after World War II intensified interest in the Yukon. Roads were built to ensure access to strategic northern regions, and oil supplies. This resulted in increased geological reconnaissance for construction purposes, and opened-up remote regions to geological exploration. However, observations relevant to the Quaternary geology and history of most of southern Yukon were still made by geologists whose main interests were related to bedrock geology and physiography (e.g., Sharp, 1943; Bostock, 1952; Kindle, 1945, 1953; Wheeler, 1963). As a result, accounts of Pleistocene glaciation were somewhat brief and fragmented.

For example, Sharp (1943) described and mapped the geology of the Wolf Creek area, St. Elias Mountains, and comments briefly on Quaternary deposits. Kindle (1945) described various glacial phenomena while mapping the bedrock geology and describing the physiography along the Canol Road in south-central Yukon. Similarly, Bostock (1948) in describing and delineating the physiography of the Canadian Cordillera north of the fifty-fifth parallel, made summary comments concerning glaciation in southern Yukon. In a report on the geology of the northwest Shakwak Valley, Bostock (1952) commented on Pleistocene glaciation in the region and remarked on the probability of multiple glaciations. The reconnaissance nature of Sharp's (1943), Kindle's (1945), and Bostock's (1948), (1952) observations are demonstrated in Table 8 where their comments regarding Pleistocene glaciation are cited.

Bostock (1966) assembled an account of Pleistocene glaciation in the central Yukon from previous reports and airphoto interpretation. This account was one of the first Geological Survey of Canada publications that dealt exclusively with Pleistocene glaciation in southern and central Yukon and is regarded a benchmark contribution to Yukon Quaternary studies (Morison and Smith, 1987). Only during the last few decades have studies been undertaken to synthesize and correlate the fragmented, glaciation related information of past decades, and to deal predominately with the

Figure 6
GENERALIZED GLACIER FLOW



Quaternary geology of southern Yukon (e.g., Rampton, 1971; Fulton et al., 1984; Fulton, 1984; Hughes et al., 1969; Hughes et al., 1972; Hughes et al., 1989; Hughes, 1987; 1989; 1990; Rutter, 1984; Vincent, 1989; Ryder and Clague, 1989; Clague, 1989).

Our understanding, and the basis for the formulation of the Quaternary history in southern Yukon has come primarily from the study of glacial sediments and landforms. Hughes et al., (1969) summarized the extent of glacial limits and flow patterns south of the 65th parallel in Yukon by stating:

> *"The great range in age of the glacial features in the area is unique in Canada. In southwestern Yukon, ice fields and cirques of St. Elias Mountains support numerous large valley glaciers and uncounted smaller ones. The snouts of most of these are bordered by fresh, in part ice-cored, moraines of Neoglacial age. Well-preserved moraines found northeast of St. Elias Mountains mark the limits of an extensive late Wisconsin ice-advance; beyond these are subdued moraines produced by a more extensive (probably) early Wisconsin advance. In central Yukon comparable moraines mark the limits of presumed early and late Wisconsin advances of the main Cordillera ice sheet, but beyond the subdued early Wisconsin moraines are remnants of the deposits of at least two older glaciations (Bostock, 1966). The oldest of such deposits are probably of early Pleistocene age. Similarly, moraines indicate (probably) early and late Wisconsin advances of valley glaciers in the Ogilvie Mountains, with evidence of still older glaciation(s) beyond the early Wisconsin moraines. Finally, in west-central Yukon a large unglaciated area lies beyond the limits of the most extensive advances of the Cordilleran ice sheet, and of ice emanating from St. Elias and Ogilvie Mountains."*

During the Pleistocene, virtually all of southern Yukon was glaciated, and in some regions glaciated several times. Glaciation in southwest Yukon was dominated by ice from the St. Elias Mountains. Ice from the Coast Mountains in northwestern British Columbia, combined with St. Elias ice to produce a large centre of Cordilleran flow in southwest Yukon (Figure 6). The general direction of flow of this ice mass was north and northwest. Other minor regional flows were west and east.

The main Cordilleran ice sheet in south-central and southeast Yukon had two major source areas; one in the Selwyn and northern Logan Mountains, and the other in the Cassiar Mountains. Ice from the Selwyn and Logan Mountains (Selwyn lobe), flowed mainly in a westerly direction (Campbell, 1967; Jackson Jr., 1989). Ice from this source also flowed south and southeast across the Liard Plain where confluence with the Continental ice sheet occurred in extreme southeast Yukon. The main ice flow direction from the Cassiar Mountains (Cassiar lobe), was northwest (Wheeler 1961). A zone of confluence between the St. Elias/Coast Mountains lobe and the Cassiar lobe occurred east of Kluane Lake in the Aishihik - Sekulmun Lakes and valley regions (Hughes et al., 1969; Tempelman-Kluit 1974). Minor flows from the Cassiar lobe in a north and northeast direction were presumably a result of topographic barriers.

The Pelly Mountains mark the zone of confluence between the Cassiar and Selwyn lobes. Changes in flow (apparently non-synchronous), between these two lobes produced a zone of confluence which fluctuated, creating complex patterns of flow directions and deglaciation related landforms. Hughes et al., (1969) suggested that during deglaciation, the ice front of the Cassiar lobe retreated more rapidly than that of Selwyn lobe in some regions, notably in northwest Pelly Mountains. Ensuing advances of Selwyn ice may have completely inundated areas previously glaciated by Cassiar ice.

The more central and northern regions of Yukon are described as experiencing limited glaciation or escaping glaciation altogether. Non-glaciated central and northern Yukon include diverse terrains ranging from lowlands, undulating plateaus to rugged mountains. The non-glaciated regions of Yukon include parts of the Yukon Plateau, northern Ogilvie Mountains, the Porcupine Plain and Basin, southern Richardson Mountains, and the Arctic Ranges. The most conspicuous attribute of these regions is the apparent lack of glacial deposits (Hughes et al., 1989). These regions, along with parts of Alaska, eastern Siberia and the intervening shelf areas, comprise the Beringia refugia (Hopkins, 1967).

Refugia are important for a number of reasons; not only do they offer the possibility of a relatively long and continuous geological record of non-glaciated environmental changes, but they also provide centres for dispersal of flora and fauna following deglaciation (Youngman, 1975). The Yukon portion of the Beringia refugia displays a unique periglacial landscape which is the result of prolonged cold and aridity.

Many of the landforms (tors, cryoplanation surfaces, and felsenmeer) developed in response to a climate more severe than the present (Pollard, 1983). Although the limits of the non-glaciated regions of Yukon may be adjusted as more research is conducted, it is generally agreed that these regions were not glaciated during the Wisconsin and probably throughout the entire Pleistocene (Rutter, 1984; Hughes et al., 1989; Vincent, 1989).

3.5 Glacial Sequences in Southern Yukon and Southeast/Central Alaska

Pleistocene investigations in southern Yukon have produced many descriptive accounts of glaciation (e.g., Dawson, 1889; McConnell, 1907; Bostock, 1948, 1952, 1966; Hughes et al., 1969, 1972; Muller, 1967; Tempelman-Kluit, 1974; Klassen, 1978; Fulton, 1984; Clague, 1989; Hughes et al., 1989). Succinct summaries of these descriptions and interpretations are available in the literature (most notably; Dewez, 1988; Hamilton, 1989; Hughes et al., 1989; Hughes, 1990). Similarly, Pleistocene and earlier glaciations in southeastern and central regions of Alaska have been documented (e.g., Capps, 1915; Taliaferro, 1932; Miller, 1953, 1964; Péwé et al., 1953; Wahrhaftig, 1958; Pewe, 1975; Denton, 1974; Fernald, 1965; Denton and Armstrong, 1969; Thorson, 1986; Mann, 1986; Molnia, 1986; Weber, 1986; Nichols, 1986; Hughes, 1989).

Table 9 presents various interpretations of glacial events for regions of southern Yukon. Selective interpretations of glacial sequences in bordering regions of Alaska are shown in Table 10. The interpretations in Tables 9 and 10 are by no means the only studies conducted in these regions, but are some of the most often referred to in the literature. The discussion which follows summarizes these interpretations, and reviews the regional correlations proposed to provide a framework in which the Pleistocene of the Ruby Range can be addressed. The classification of events (glacial and nonglacial) in Tables 9 and 10 into a Quaternary (and Tertiary) chronology is based on reported or implied sequence(s) and is not intended for correlative purposes. The locations of the areas under review, as well as the central positioning of the Ruby Range, are identified on Figure 7.

3.5.1 Southern Yukon

Central Yukon

Bostock (1966) proposed a fourfold glacial sequence of the Cordilleran ice sheet in southern and central Yukon, with each successive advance being less extensive than its predecessor. Bostock did not attempt to assign the glacial advances to a Pleistocene chronology, but rather, to a sequential order from oldest to youngest; Nansen, Klaza, Reid, and McConnell. However, Hughes (1989) placed the Nansen and Klaza Glaciations as early Pleistocene, Reid Glaciation as middle Pleistocene, and McConnell glaciation as middle - late Wisconsin. The maximum extent of the McConnell advance is well marked by relatively fresh moraines and ice contact features. The Reid advance can also be identified by similar features. The maximum extent of the Klaza and Nansen advances, though more extensive, are not well defined. Scattered glacier erratics, outwash gravels, and local till deposits have been used to imply their limits. The development of soil on till, described as well drained, truncated Brunisols with a thick B horizon, is considered diagnostic and is used as an indicator of Klaza and Nansen limits in valleys (Foscolos et al., 1977; Hughes, 1989).

Table 9
GLACIAL SEQUENCES IN SOUTHERN YUKON

Age (Ka)	Era / Period / Epoch	Liard Plain (1) Klassen (1978, 1987)	Central Yukon (2) Vernon & Hughes (1966)	East-Central Yukon (3) Jackson (1989)	Yukon Plateau (4) Bostock (1966) Hughes (1989)	Aishihik Lake (5) Hughes (1990)	Kluane Lake (6) Muller (1967)	Kluane National Park (7) Rampton (1981)	St. Elias Mts (8) Denton and Stuiver (1967)	Snag-Klutlan (9) Rampton (1971)	Southwest Yukon (10) Krinsley (1965)
5–10	Cenozoic – Quaternary – Holocene				Postglacial			Postglacial	Neoglacial Nonglacial		
15–30	Cenozoic – Quaternary – Pleistocene – Late – Wisconsin – Late	Glacial (Till-D)	Last Glaciation	McConnell Glaciation	McConnell Glaciation	McConnell Glaciation	St. Elias Advance	Kluane Glaciation	Kluane Glaciation Boutellier Nonglacial	Macauley Glaciation	Tchawsahmon
35–60	Cenozoic – Quaternary – Pleistocene – Late – Wisconsin – Middle	Nonglacial		Reid Glaciation	Nonglacial	Reid Glaciation		Boutellier Nonglacial	Icefield Glaciation		
65–100	Cenozoic – Quaternary – Pleistocene – Late – Wisconsin – Early	Glacial (Till-C)			Reid Glaciation		Ruby Ice-Sheet	Icefield Glaciation	Silver Nonglacial	Mirror Creek Glaciation	Snag
128	Cenozoic – Quaternary – Pleistocene – Middle – Pre-Wisconsin	Nonglacial	Intermediate Glaciation	Pre-Reid Glaciation	Klaza Glaciation			Silver Nonglacial			Donjek
	Cenozoic – Quaternary – Pleistocene – Middle – Pre-Wisconsin	Glacial (Till-B)					Nisling Ice-Sheet	Shakwak Glaciation	Shakwak Glaciation		Nisling
340	Cenozoic – Quaternary – Pleistocene – Early – Pre-Wisconsin	Nonglacial	Old Glaciations								
1000	Cenozoic – Quaternary – Pleistocene – Early – Pre-Wisconsin	Glacial (Till-A)			Nansen Glaciation			Glaciation			
2000	Cenozoic – Tertiary – Pliocene										
7000	Cenozoic – Tertiary – Miocene										
27000	Cenozoic – Tertiary – Oligocene										
38000	Cenozoic – Tertiary – Eocene										
55000–65000	Cenozoic – Tertiary – Palaeocene										

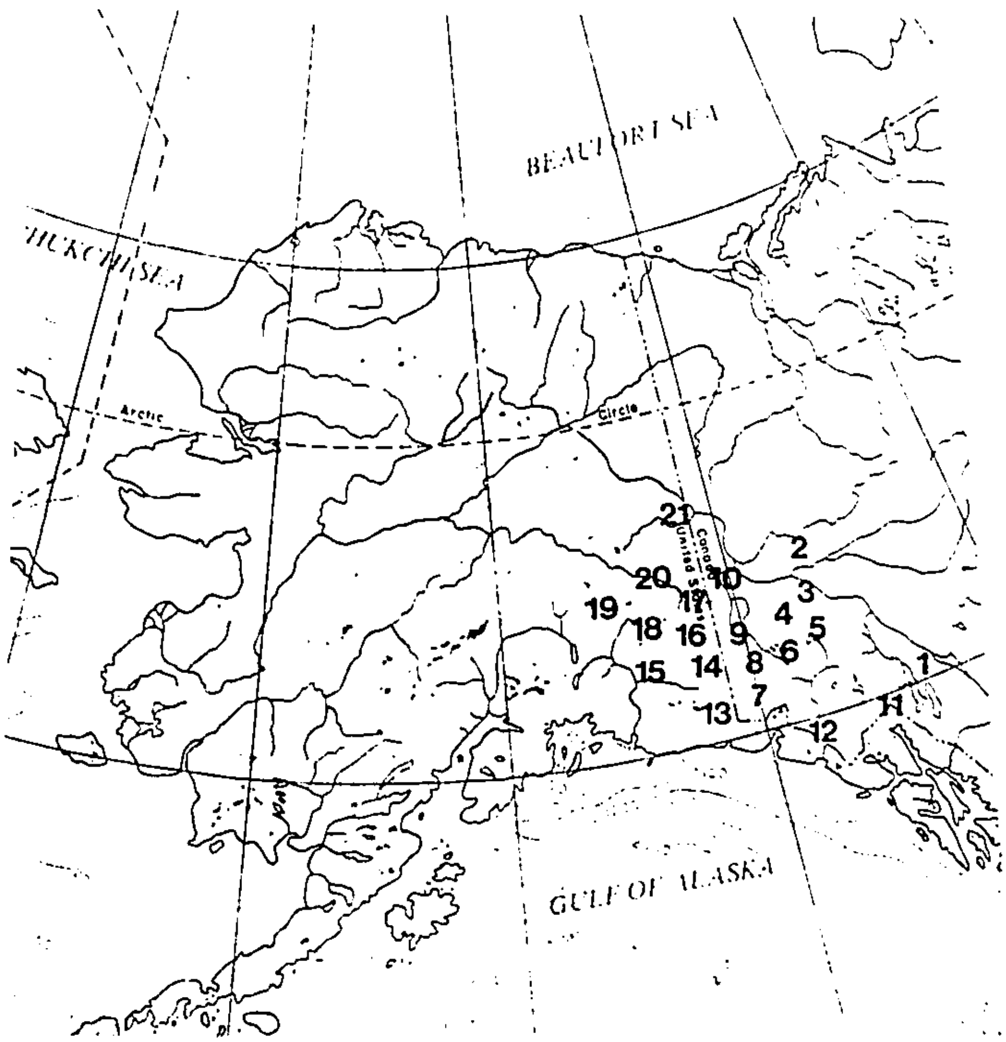
Note: Numbers in brackets appearing at the top of each column identify location of area on Figure 7. Numbers appearing after events in the columns refer to the number of events (i.e. Glaciations (4): four glaciations). Solid lines separating events are divisions proposed by authors, dotted lines represent uncertain or transitional boundaries.

Table 10
GLACIAL SEQUENCES IN SOUTHEASTERN AND CENTRAL ALASKA

Ka	Era	Period	Epoch		Stage	Sub-stage	BOUNDARY RANGES (11) Miller (1964)	SOUTHEAST ALASKA (12) Mann (1986)	NORTHEAST GULF OF ALASKA (13) Molnia (1986)	ST. ELIAS AND WRANGELL MTS. (14) Denton and Armstrong (1969)	SOUTHEAST COPPER RIVER BASIN (15) Nichols (1989)	WHITE RIVER VALLEY-NORTHERN ST. ELIAS MTS. (16) Denton (1974)	UPPER TANANA RIVER VALLEY (17) Fernald (1965)	ALASKA RANGE (18) Hughes (1989)	ALASKA RANGE (19) Péwé (1987)	NORTHERN NENANA VALLEY (20) Thorson (1986)	YUKON TANANA UPLAND (21) Weber (1986)
5, 10	CENOZOIC	QUATERNARY	HOLOCENE				Retracted Icefield Glaciation	Holocene Max. Glaciation	Neoglacial Max. 2000 yr B.P.	Holocene Max. 8290 yr B.P.			Four (4) Advances	Nonglacial	Summit Lake Glaciation		Ramshorn 1&2 Glacial
15, 20			PLEISTOCENE	LATE	WISCONSIN	LATE	Extended Icefield Glaciation	Glaciation	Glaciation	Kluane Glaciation	Glaciation	Macauley Glaciation	Jatahmund Lake Glaciation	Donnelly Glaciation	Donnelly Glaciation	Riley Creek Glacial	Salcha Glacial
25, 30, 35, 40, 45, 50, 55						MIDDLE	Mountain Ice-Sheet Glaciation	Glaciation		Glaciation; Glaciation	Glaciation						
60, 65, 70, 75						EARLY		Glaciation			Glaciation			Nonglacial	Delta Glaciation	Healy Glacial	Eagle Glacial
100, 125					PRE-WISCONSIN		Intermontane Icecap Glaciation			Glaciation	Glaciation	Pre-Macauley Glaciations (5)	Black Hills Glaciation	Delta Glaciation		Lignite Creek Glacial	
				MIDDLE				Glaciation	Glaciation		Glaciation			Nonglacial	Darling Creek Glaciation	Bear Creek Glacial; Browne Glacial	Mount Harper Glacial
340, 1000				EARLY					Glaciation	Glaciations (12)				Darling Creek Glaciation		Teklanika River Glacial	Charley River Glacial
2000		TERTIARY	PLIOCENE						Glaciations (4)			Glaciations				Nenana Glacial	Oldest Drift
7000			MIOCENE														
27000			OLIGOCENE														
38000			EOCENE														
55000, 65000			PALAEOCENE														

Note: Numbers in brackets appearing at the top of each column identify location of area on Figure 7. Numbers appearing after events in the columns refer to the number of events (ie. Glaciations (4), four glaciations). Solid lines separating events are divisions proposed by authors, dotted lines represent uncertain or transitional boundaries.

Figure 7
LOCATIONS OF REPORTED
GLACIER SEQUENCES IN YUKON AND ALASKA



1 LIARD PLAIN
2 CENTRAL YUKON
3 EAST-CENTRAL YUKON
4 YUKON PLATEAU
5 AISHIHIK LAKE
6 KLUANE LAKE
7 KLUANE NATIONAL PARK
8 ST. ELIAS MTS.
9 SNAG-KLUTLAN
10 SOUTHWEST YUKON
11 BOUNDARY RANGES
12 SOUTHEAST ALASKA
13 NORTHEAST GULF OF ALASKA
14 ST. ELIAS AND WRANGELL MTS.
15 SOUTHEAST COPPER RIVER BASIN
16 WHITE RIVER VALLEY/NORTHERN ST. ELIAS MTS.
17 UPPER TANANA RIVER VALLEY
18 ALASKA RANGE
19 ALASKA RANGE
20 NORTHERN NENANA VALLEY
21 YUKON TANANA UPLAND

Kluane Lake Map Area

Muller (1967) delineated Pleistocene glaciations in the Kluane Lake region. Three progressively less extensive ice-sheets are described: i) pre-Wisconsin (Nisling Ice-Sheet), ii) early Wisconsin (Ruby Ice-Sheet), and iii) late Wisconsin (St. Elias Ice-Sheet). The upper limit (elevation) of each succeeding ice-sheet is recognized as being 300 to 450 m below that of the the preceding one. (A detailed discussion of this area is available in the section titled, 'The Pleistocene In The Ruby Range').

Aishihik Lake Map Area

A report on the surficial geology and geomorphology of Aishihik Lake map area is presented by Hughes (1990). Evidence of only two glaciations have been found (McConnell and Reid glaciations), and the lack of firm evidence for pre-Reid glaciation, either in stratigraphic section or at the surface beyond the Reid limit, prohibits speculation on earlier glacial history. However, it is unlikely that the entire area could have escaped earlier glaciation. Although there are no radiocarbon dates from Aishihik Lake that relate to Reid glaciation, Hughes (1990) states that a date from near the type Reid locality indicates retreat of ice from the area before 42,900 yr B.P.

There is also evidence that indicates a minimum date of retreat following Reid glaciation of about 80,000 yr B.P. (Hamilton and Bischoff, 1984). Retreat of the McConnell glaciation probably began before 9,660 yr B.P. Hughes (1990) assigned the Reid glaciation to the early Wisconsin, and the McConnell glaciation to the late Wisconsin.

The eastern portion of the region was glaciated by the Cassiar lobe of the Cordilleran Ice Sheet which flowed to the northwest. The central and southern parts were invaded by ice from the Coast and St. Elias Mountains that followed pathways through the mountain plateau areas.

Kluane National Park

Rampton (1981) described the surficial deposits and landforms of Kluane National Park. The Park was subjected to glaciation throughout the Quaternary, and in all probability, the Tertiary as well. In the St. Clare Creek area, west of Klutlan Glacier, glacial drift has been identified as being Tertiary or early Pleistocene. Although similar evidence has not been found within Kluane Park boundary, the same events undoubtedly affected the Park.

As well as hypothesizing glaciation in early Pleistocene and/or Tertiary, Rampton (1981) identified three glaciations and two nonglacial intervals during the mid and late Pleistocene. These are: i) >100,000 yr B.P. middle - early Pleistocene, Shakwak Glaciation; ii) early - middle Pleistocene, Silver Nonglacial; iii) Illinoian - early Wisconsin, Icefield Glaciation; iv) ending about 30,000 yr B.P. late Wisconsin, Boutillier Nonglacial; v) beginning about 30,000 yr B.P., and ending about 12,500 yr B.P. late Wisconsin, Kluane Glaciation.

Southeastern St. Elias Mountains

Working in the southeastern St. Elias Mountains, Denton and Stuiver (1967) identified four late Pleistocene glaciations separated by three nonglacial intervals. The chronology is as follows: i) Shakwak glaciation, >49,000 yr B.P.; ii) Silver nonglacial interval, >49,000 yr B.P.; iii) Icefield glaciation began >49,000 yr B.P. and ended approx. 37,700 yr B.P.; iv) Boutellier nonglacial interval began about 37,700 yr B.P. and ended <30,100 yr B.P.; v) Kluane Glaciation started <30,100 yr B.P. and ended between 12,500 - 9,780 yr B.P.; vi) Slims nonglacial interval began approx. 12,500 yr B.P. and ended about 2,640 yr B.P.; vii) Neoglaciation began about 2,640 yr B.P. and is current.

Snag-Klutlan Area

Rampton (1971) reported on the Quaternary history of the Snag-Klutlan area. Although vegetational and climatic interpretations of the paleobotanical record were the main focus, Rampton presented some observations on glaciation. Glacial features, which are plentiful except on the upland areas, are used to identify two glaciations; a late Pleistocene, Mirror Creek Glaciation, and a Wisconsin maximum (beginning at >40,000 yr B.P., and ending at about 13,500 yr B.P.), Macauley Glaciation. It is hypothesized however, that the region was glaciated several times during the Pleistocene.

Southwest Yukon: Wellesley Basin

Moraine sequence in southwestern Yukon (centring on the Wellesley Basin, north of St. Elias Mountains), have been described by Krinsley (1965). Four major, successively less extensive glaciations, originating from St. Elias Mountains, are identified. These are: i) a pre-Illinoian/middle - early Pleistocene, Nisling glaciation; ii) an Illinoian/middle - late Pleistocene, Donjek glaciation; iii) a pre-classical Wisconsin/early Wisconsin, Snag glaciation; and iv) a classical Wisconsin/middle - late Wisconsin, Tchawsahmon glaciation. An attempt was also made to correlate glacial sequences in southwest Yukon and Cook Inlet, Alaska.

Southeastern Yukon: Liard Plain

At least four major Quaternary glaciations separated by three nonglacial intervals were identified by Klassen (1978, 1987) in the Liard Plains, southeastern Yukon. These events are based on a Tertiary - Quaternary stratigraphic succession that consists of (from bottom to top); i) early Pleistocene till, ii) early Pleistocene, nonglacial silt and sand, iii) middle - late Pleistocene till, iv) middle - late Pleistocene, nonglacial fossiliferous clay, v) early Wisconsin till, vi) middle Wisconsin, nonglacial silt, gravel and clay, and vii) late Wisconsin till.

East-Central Yukon

Surficial geology and Quaternary stratigraphy in east-central Yukon were the basis of interpretation of three glaciations: i) pre-Wisconsin, Pre-Reid Glaciation; ii) early - middle Wisconsin, Reid Glaciation; iii) middle - late Wisconsin, McConnell Glaciation (Jackson, 1989). Glaciation was from the Selwyn lobe of the Cordilleran Ice Sheet. Former ice-flow directions are evident on plateau surfaces and along valley sides in the form of whaleback ridges and crag-and-tail features.

Dawson, Larson Creek, and Nash Creek Map Areas

Vernon and Hughes (1966) identified at least three glaciations with intervening periods of nonglaciation in the region. The glaciations are referred to as Old, Intermediate, and Last. Limits of the Intermediate and Last Glaciations are marked by moraines. Evidence of the extent of the Old Glaciation is in the form of erratics beyond the limits of the later glaciations. There is also a possibility of multiple Old Glaciations.

The Last Glaciation is tentatively correlated with glaciation in the Alaska Range which culminated about 10,000 yr B.P. The lack of data prevents speculation as to the age of the Intermediate and Old Glaciation(s). However, subsequent investigations have lead Hughes (1989) to assign the Intermediate Glaciation to the middle Pleistocene, and Old Glaciation(s) to the early Pleistocene or Tertiary.

3.5.2 Southeastern and Central Alaska

The landmark study of the Quaternary in Alaska is probably that of Péwé (1975) where a summary of glaciation throughout the Cenozoic was presented. Although Taliaferro (1932) first recognized that glacial deposits in southern Alaska were Pliocene in age, it wasn't until the 1960's that evidence of glacial sediments of Pliocene and Miocene time was found at higher elevations

(Péwé, 1975). Mountain glaciation probably began in late Tertiary (about 10 to 13 Ma B.P.) in the St. Elias Mountains and other ranges along the Gulf of Alaska, including the Boundary Ranges, Wrangell Mountains, and Chugach Mountains.

Southeastern and central regions of Alaska were subjected to repeated glaciation throughout the Pleistocene. However, Péwé (1975) pointed out that because of the extremely rugged and tectonically active terrain in southeastern Alaska, the ranges would present a prominent topographic barrier to moisture-bearing winds from the south before the Pleistocene began. Hence, different or delayed (non-synchronous) Pleistocene glacially related responses would be expected in southeastern and central parts of Alaska. The discussion which follows summarizes a selection of glacial sequences reported for southeastern/central regions of Alaska (Table 10).

Boundary Range

Miller (1964) investigated the morphogenetic classification of Pleistocene glaciations in the Boundary Range. Four glaciations were identified, determined on the basis of maximum extent. These are: i) A pre-wisconsin, Intermontane Icecaps Glaciation covering all summits with flow from local centres generally in a north and south direction; ii) An early - middle Wisconsin, Mountain Ice-Sheet Glaciation typified by local centres of flow with many nunataks. Extent was generally restricted to the mountain ranges; iii) A late Wisconsin, Extended Icefield Glaciation confined entirely within the mountain ranges. Several local centres of outflow produced extensive valley glacier systems; and iv) A Holocene, Retracted Icefield Glaciation restricted to intermediate and high alpine regions.

Southeastern Alaska: Gulf of Alaska

The northeast Gulf of Alaska region, including parts of the St. Elias Mountains, have been subjected to repeated glaciation since at least late Tertiary (Molnia, 1986). He identified Pleistocene glaciations and at least three late Tertiary glaciations. Late Tertiary glaciations (Pliocene) have been determined by data derived from deposits of glacial and marine sequences and are identified as occurring as follows: i) Glaciation 1 (from 6.3 to 6.0 Ma B.P.); Glaciation 2 (from 6.0 to 5.4 Ma B.P.); Glaciation 3 (from 5.4 to 3.2 Ma B.P.); Glaciation 4 (from 3.2 to 1.8 Ma B.P.). Pleistocene glaciations were identified primarily from glacially derived sediments on the continental shelf and are described as: Glaciation 5 (early Pleistocene); Glaciation 6 (middle Pleistocene); Glaciation 7 (late Pleistocene). Neoglacial maximum occurred about 2,000 yr B.P.

Southeastern Alaska

Mann (1986) outlined the Wisconsin glacial history of southeastern Alaska and commented on probable pre-Wisconsin glaciations but points out that because evidence of these is at higher altitudes, there has been rapid tectonic uplift, so assessment of their relative magnitudes is difficult. At least two advances are identified as pre-Wisconsin and/or early-Wisconsin. A middle-Wisconsin advance is dated at about 27,000 yr B.P., followed by a late-Wisconsin advance between 14,000 - 16,000 yr B.P. A subsequent late-Wisconsin or early Holocene advance at about 10,000 yr B.P. is also recognized. A Holocene advance at approximately 6,000 yr B.P. is said to have occurred only in some regions and there is speculation that this could be Neoglacial.

Alaska Range

Three Pleistocene glaciations and one Holocene glaciation were identified by Péwé (1975, 1987) in the Alaska Range: i) an early - middle Pleistocene, Darling Creek Glaciation; ii) a middle - late Pleistocene, Delta Glaciation; iii) a Wisconsin, Donnelly Glaciation; and iv) a Holocene, Summit Lake Glaciation. The placing of Donnelly Glaciation within the Wisconsin is not attempted, however it is implied to be Wisconsin maximum. Hughes's (1989) summary of Quaternary glaciations in the Alaska Range identifies three major glaciations: i) an early Pleistocene, Darling Creek Glaciation; ii) a middle Pleistocene, Delta Glaciation; and iii) a late Wisconsin, Donnelly Glaciation.

Copper River Basin

The stratigraphy of till sheets in the Copper River basin of southeastern Alaska has been interpreted as the result of five major Pleistocene glaciations (Nichols, 1986). Although the ages of the youngest and oldest glaciations has been established at about 9,400 yr B.P. (Holocene), and 200,000 yr B.P. (middle Pleistocene) respectively, the three interceding glaciations are described only as being older than 40,000 yr B.P. The following glacial sequence is implied: i) middle Pleistocene (about 200,000 yr B.P.), Glaciation 1; ii) middle Pleistocene - early Wisconsin, Glaciation 2; iii) early Wisconsin, Glaciation 3; iv) middle Wisconsin (>40,000 yr B.P.), Glaciation 4; v) late Wisconsin - Holocene (about 9,400 yr B.P.), Glaciation 5.

Yukon-Tanana Uplands

Six glacial episodes have been identified by Weber (1986) in the Yukon-Tanana Uplands region of east-central Alaska. The glacial record consists of: i) a Tertiary, Oldest Drift;

ii) an early(?) Pleistocene, Charley River Glacial; iii) a middle(?) Pleistocene, Mount Harper Glacial; iv) an early(?) Wisconsin, Eagle Glacial; v) a late Wisconsin, Salcha Glacial; and vi) a Holocene, Ramshorn I & II Glacials.

St. Elias and Wrangell Mountains

Denton and Armstong (1969) examined tillites of Tertiary age in numerous localities in the St. Elias, Chugach, and Wrangell Mountains of southeastern Alaska and concluded that their widespread occurrence indicates major glaciation beginning in late Tertiary. Overlying the tillites are at least four Pleistocene till sheets. During the past 10 ma, the region has experienced at least sixteen (16) major glacial expansions; twelve of these recorded by tillites and four by till sheets. The evidence suggested that many of the more ancient glaciations occurred 9 to 10 Ma B.P. (Miocene), and several may be older. At least one glaciation occurred about 3.6 Ma B.P. (Pliocene), and at least two occurred between 8.8 to 2.7 Ma B.P. (Miocene - Pliocene). Four additional glaciations are said to have taken place during the past 1.6 Ma (Pleistocene).

Northern Nenana Valley

Seven distinct late Cenozoic glaciations in southeast-central Alaska, in the vicinity of the northern Nenana Valley and northern slopes of the Alaska Range, were identified by Thorson (1986). Evidence supports the possibly of: i) two late Tertiary glaciations (Nenana Gravel Glacial and Teklanika River Glacial); ii) three pre-Wisconsin glaciations (Browne Glacial, early - middle Pleistocene; Bear Creek Glacial, middle Pleistocene; and Lignite Creek Glacial, middle - late Pleistocene), iii) an early-Wisconsin, Healy Glacial, and iv) a late Wisconsin, Riley Creek Glacial.

White River Valley, Northern St. Elias Mountains

Denton (1974) stated that the White River Valley, northern St. Elias Mountains, and adjacent regions were subjected to repeated glaciation throughout the Quaternary and late Tertiary. The most recent Pleistocene glaciation, the Macauley, began about 37,000 yr B.P., reached it's maximum 14,000 yr B.P., and rapidly disintegrated within 1,500 to 2,700 years. As many as five pre-Macauley glaciations are said to have occurred during the Quaternary, along with a long succession of glaciations during the late Tertiary.

Upper Tanana River Valley

In the Nabesna River region (Upper Tanana River Valley), Fernald (1965) found evidence for two major glaciations in moraines; an Illinoian, Black Hills Glaciation, and a Wisconsin, Jatahmund

Lake Glaciation. The assignment of ages of these two glaciations was based on topographic modifications of the moraines, radiocarbon dated nonglacial deposits, and the sequential analogy with other regions of Alaska.

3.6 Quaternary Correlations: Southern Yukon and Southeast/Central Alaska

As an introduction to the correlation of Quaternary events proposed for regions in southern Yukon and southeastern and central Alaska, basis and limitations of these correlations will be discussed. Areas in southern Yukon and southeastern and central Alaska constitute parts of the western Cordillera that were repeatedly glaciated during the Pleistocene (and Tertiary), or escaped glaciation altogether. Since the Quaternary history of these regions has been mainly inferred from glacial stratigraphic and geomorphic evidence, the Quaternary reconstruction emphasizes glacial and nonglacial events.

Péwé (1975) stated that radiometric dates, significant stratigraphic relations with marine sediments and other interglacial deposits, and an improved understanding of the nature and rates of weathering and denudation have made it possible to make correlations with considerable confidence within single mountain ranges, and with somewhat less confidence between ranges. The most reliable correlations are based on the tracing of moraines from one area to another, on radiocarbon dating, and on similar stratigraphic relations. However, placing these events in a Quaternary framework has proven troublesome. Jackson et al., (1989) stated:

> *"The paucity of relevant absolute age determinations has made it extremely difficult to date these events, and the lack of continuity of individual drift sheets from one drainage basin to the next and from one mountain range to another has made regional correlations inferential and tentative at best. Even in a single basin, stratigraphic relationships may not be straightforward, and consequently different conclusions as to the sequence of glacial events have been reached by different workers."*

Fulton (1984) explained that local Quaternary stratigraphic correlations within regions are relatively straight forward, but correlations between regions are speculative, especially for deposits beyond the radiocarbon dating limit. Correlation of late Wisconsin and middle Wisconsin events are generally based upon radiocarbon dates; correlations of older events are based largely on counting downwards and the matching of glacial and nonglacial units (Fulton et al., 1986).

Radiocarbon dating has itself been the subject of discussion with regard to reliability. Brigham-Grette (1989) stated that accurate dating is essential for piecing together the Quaternary history of any area, and for correlating glacial and nonglacial sequences between regions. For example, Carter (1989) reported seven different ages ranging from about 13 to 30 Ka on different size fractions of an organic deposit from Alaska. Denton and Struiver (1969) stated that the temporal relations of events can cause regional correlations to be misleading and that only sequences identified with detailed, controlled C^{14} dates can be adequately correlated. The reliability of datable material and the need for more accurate methods provide the greatest obstacle to determining and correlating Quaternary (late Cenozoic) events. In this regard, Brigham-Grette (1989) argued that:

> *"Problems in correlation will remain unsolved until distinctions can be made between the real gaps in the preserved record and disagreements caused by inadequate age control. Direct matching of series of glacial and nonglacial events may mislead us into false correlations unless they can be supported by adequate age control."*

Local Quaternary chronologies for southern Yukon and southeastern and central Alaska have been evolving over the past decades. Although regional correlations have been presented by many individuals, the first cooperative attempt at correlating Yukon and Alaska events was the result of a joint Canadian and American workshop held in Calgary in 1984 (Heginbottom and Vincent, 1986; Hughes, 1986). The workshop produced a first approximation of a correlation of Quaternary deposits and events in the area adjacent to the Beaufort Sea which focused primarily on northern and central regions of Yukon and Alaska. As well, other individual contributions outlining regional correlations within and between Yukon and Alaska were presented (e.g., Hughes, 1986).

Tables 11 and 12 present some correlations of Quaternary events proposed by various researchers for regions in southern Yukon, and southeastern and central Alaska. Quaternary correlations in southern Yukon along with regions in Alaska appearing in Table 11 are essentially the result of the efforts of the joint Canadian and American workshop (Heginbottom and Vincent, 1986; Hughes, 1986). Quaternary correlations in southeastern and central Alaska along with regions in Yukon appearing in Table 12 have been tabulated for the most part by Péwé (1975), and Hamilton (1986).

In making these correlations, attempts were made to use the major event, ie. glaciation, nonglacial. However, these units have not been established for all regions in Yukon and Alaska, hence the degree of precision of these units is not the same in all areas, or for all intervals of the Quaternary (Fulton et al., 1984). In addition, it is generally agreed that these events are time-transgressive (lacking synchroniety), so that, although they are identified as distinct events, they may not necessarily fit exactly into the divisions identified on the tables. This accounts for the somewhat broad divisions of the Quaternary in the tables. As well, there is relatively little chronological control on events predating early Wisconsin. Much of the older Quaternary record in parts of southwestern Yukon, and southeastern and central Alaska has been destroyed or obscured by tectonism and widespread Wisconsin glaciation. Because of the lack of adequate chronological

Table 11
QUATERNARY CORRELATIONS IN SOUTHERN YUKON WITH REGIONS IN ALASKA

			YUKON PLATEAU	SNAG-KLUTLAN	SILVER CREEK	KLUANE LAKE	OGILVIE MOUNTAINS	LIARD LOWLANDS	ALASKA RANGE
HOLOCENE			Postglacial	Neoglacial	Neoglacial Slims Nonglacial		Postglacial		
PLEISTOCENE	*LATE*	*WISCONSIN*	McConnell Glaciation	Macauley Glaciation	Kluane Glaciation	St. Elias Advance (?) / Ruby Ice Sheet	Last Glaciation	Till D	Donnelly Glaciation
			Reid/McConnell Nonglacial	Macauley/Mirror Creek Nonglacial	Bouteiller Nonglacial		Nonglacial	Intertill Unit C-D	Nonglacial
	MIDDLE	*PRE WISCONSIN*	Reid Glaciation	Mirror Creek Glaciation	Icefield Glaciation	Nisling Ice Sheet	Intermediate Glaciation	Till C	Delta Glaciation
			Klaza/Reid Nonglacial		Silver Nonglacial		Nonglacial	Intertill Unit B-C (?)	Nonglacial
	EARLY		Klaza Glaciation		Shakwak Glaciation (?)		Old Glaciation (?)	Till B (?)	Darling Creek Glaciation
			Nansen/Klaza Nonglacial					Intertill Unit A-B (?)	
			Nansen Glaciation		Shakwak Glaciation		Old Glaciation	Till A (?)	

Compiled from: Denton and Stuiver (1967), Hughes et al. (1969); Rampton (1971); Miller (1976); Fulton et al. (1986); Fulton (1986); Heginbottom and Vincent (1986); Hughes (1989); Jackson et al. (1989).

Note: i) Pleistocene divisions are taken from Hughes (1989).
ii) Correlation(s) with regions in Alaska are outlined in bold.
iii) Solid lines separating events are divisions proposed by authors, dotted lines represent uncertain or transitional boundaries.

Table 12
QUATERNARY CORRELATIONS IN SOUTHEASTERN AND CENTRAL ALASKA WITH REGIONS IN YUKON

	YUKON-TANANA UPLAND	UPPER TANANA RIVER VALLEY	ALASKA RANGE	NORTHERN NENANA VALLEY	ST. ELIAS & WRANGELL MOUNTAINS	WHITE RIVER VALLEY	ALASKA/CANADA BOUNDARY RANGE
LATE WISCONSIN	Salcha Glacial	Jatahmund Lake Glaciation	Donnelly Glaciation	Riley Creek Glaciation	Kluane Glaciation	Macauley Glaciation	Extended Icefield Glaciation
EARLY WISCONSIN AND LATE PLEISTOCENE	Eagle Glacial		Delta Glaciation	Healy Glaciation	Glaciation Glaciation	Pre-Macauley Glaciations (5)	Mountain Ice-Sheet Glaciation
MIDDLE PLEISTOCENE	Mount Harper Glacial	Black Hills Glaciation		Lignite Creek Glaciation	Glaciation		Intermontane Icecap Glaciation
EARLY PLEISTOCENE	Charley River Glacial		Darling Creek Glaciation	Browne Glaciation	Glaciations (12) (Early Pleistocene-Late Tertiary)	Glaciations (?) (Early Pleistocene-Late Tertiary)	
TERTIARY							

Compiled from: Denton and Armstrong (1969); Denton (1974); Péwé (1975); Hamilton (1986); Weber (1986); Molnia (1986); Mann (1986).

Note: i) Correlation(s) with regions in Yukon are outlined in bold.
ii) Pleistocene subdivisions follow Hamilton's (1986) designation except for LATE PLEISTOCENE which Hamilton identifies as "EOWISCONSIN".
iii) Solid lines separating events are divisions proposed by authors; dotted lines represent uncertain or transitional boundaries.

control, the number of pre-Wisconsin glaciations in many regions is uncertain. For this reason, events are often grouped into broad divisions (e.g., pre-Wisconsin, middle Pleistocene, early Pleistocene or Tertiary), and are derived by counting downwards and matching glacial and nonglacial units.

As a result, correlative interpretations ranging from a few distinct glaciations to differing numbers of glaciations within similar time periods have been made. For example, Table 9 shows that during pre-Wisconsin, regions in Yukon experienced one to three glaciations. Similarly, Table 10 indicates that during middle to early Pleistocene, regions in Alaska experienced from one to more than twelve glaciations. This implies that events cannot be readily correlated between regions because the evidence is often equivocal and/or lacking (Jackson et al., 1989; Brigham-Grette, 1989). It is emphasized that most regional Quaternary correlations are in developmental stages, and should be viewed accordingly. The age relationships of the events in Tables 11 and 12 are uncertain and therefore the correlations between regions are tentative, and will no doubt be modified as new information becomes available (Heginbottom and Vincent, 1986).

3.7 Pleistocene Glaciation in the Ruby Range

The most detailed account of Pleistocene glaciation in the Ruby Range is provided by Muller (1967) (Kluane Lake region). The most often referred to interpretations of glacial chronology relating to the Ruby Range have been by Bostock (1966) and Hughes (1989) (Yukon Plateau/central Yukon). Additional descriptions that have relevance to the Ruby Range are presented by Rampton (1971) (Snag-Klutlan); and Denton and Stuiver (1967) (Silver Lake). Since these descriptions have already been reviewed in the two previous sections, only summary comments with regard to the similarities of these interpretations will be reviewed. A more detailed description of Muller's (1967) observations with respect to glaciation in the Ruby Range will also be provided.

Table 13 presents a tentative correlation of glaciations and nonglacials during the Quaternary for the Yukon Plateau, Snag-Klutlan, Silver Creek, and Kluane Lake regions. During the Wisconsin, all regions experienced one glaciation, are nonglacial. The McConnell Glaciation, Macauley Glaciation, Kluane Glaciation, and Ruby Ice Sheet (and perhaps the St. Elias Advance) are all correlated as probable Wisconsin maximum. The Reid/McConnell Nonglacial, Macauley/Mirror Creek Nonglacial, Boutellier Nonglacial, and presumably an unnamed nonglacial in the Kluane Lake region are also correlated as early Wisconsin. During the middle Pleistocene, all regions experienced glaciation, but in only two regions was evidence found for a nonglacial. The Reid Glaciation, Mirror Creek Glaciation, Icefield Glaciation, and Nisling Ice Sheet are described as middle Pleistocene. A middle Pleistocene Klaza/Reid Nonglacial and Silver Nonglacial have been

Table 13
QUATERNARY CORRELATION OF GLACIATIONS AND NONGLACIATIONS IN THE RUBY RANGE REGION

			YUKON PLATEAU	SNAG-KLUTLAN	SILVER CREEK	KLUANE LAKE
HOLOCENE			Postglacial	Neoglacial	Neoglacial Sims Nonglacial	
PLEISTOCENE	LATE	WISCONSIN	McConnell Glaciation	Macauley Glaciation	Kluane Glaciation	St. Elias Advance (?) Ruby Ice Sheet
			Reid/McConnell Nonglacial	Macauley/Mirror Creek Nonglacial	Boutellier Nonglacial	
	MIDDLE	PRE WISCONSIN	Reid Glaciation	Mirror Creek Glaciation	Icefield Glaciation	Nisling Ice Sheet
			Klaza/Reid Nonglacial		Silver Nonglacial	
	EARLY		Klaza Glaciation		Shakwak Glaciation (?)	
			Nansen/Klaza Nonglacial			
			Nansen Glaciation		Shakwak Glaciation	

identified in the Yukon Plateau and Silver Creek regions respectively, however similar nonglacials have not been inferred in the Snag-Klutlan and Kluane Lake regions.

The remainder of the pre-Wisconsin correlations are even more fragmented. Two early Pleistocene glaciations are identified in the Yukon Plateau (Klaza Glaciation and Nansen Glaciation), separated by the Nansen/Klaza Nonglacial. The Shakwak Glaciation is described as probably persisting throughout the early Pleistocene in the Silver Creek region. Events in the Snag-Klutlan and Kluane Lake regions during these times are uncertain. Bostock's (1966) original account of a fourfold glacial sequence of the Cordilleran ice sheet affecting southern and central Yukon, and Hughes's (1989) subsequent reiteration, has been described in a prior section and forms the basis of interpretation for this thesis. Four Pleistocene glaciations are identified, each being progressively less extensive. These are:

i) Nansen Glaciation (late Tertiary - early Pleistocene),
ii) Klaza Glaciation (early - middle Pleistocene),
iii) Reid Glaciation (early - middle Wisconsin),
iv) McConnell Glaciation (middle - late Wisconsin).

Muller's (1967) description of Pleistocene glaciation in the Kluane lake region includes three progressively less extensive ice-sheets:

i) Nisling Ice-Sheet (pre-Wisconsin),
ii) Ruby Ice-Sheet (early Wisconsin),
iii) St. Elias Advance (late(?) Wisconsin).

The upper limit (elevation) of each succeeding ice-sheet is described as being 300 to 450 meters below that of the preceding one. The Nisling Ice Sheet correlates to Bostock's (1966) Reid Glaciation, whereas the Ruby Ice Sheet and probably the St. Elias Advance, are equivalent to the McConnell Glaciation. The identity and extent of the Nisling Ice-Sheet is apparent from traces of glaciation above the level and beyond the extent of the Ruby Ice-Sheet. Specifically, erratic material such as pebbles and granite boulders are scattered on many plateau surfaces. As well, marginal channels, notches, and flat-domed hills attest to the former Nisling Ice-Sheet.

Nisling ice covered much of the present plateau surfaces of the Ruby Range, leaving many nunatak areas. The average ice elevation was about 1800 masl in the northeast part of the St. Elias Mountains and 1500 masl over Yukon Plateau. The main flow of the ice-sheet was northwest, except in the St. Elias Mountains. Major ice flow occurred through the Shakwak Valley, Duke Depression, the depression between Ruby and Nisling Ranges, and other northwest trending valleys.

Muller (1967) suggested that many of the northwest trending valleys may have been excavated mainly after retreat of Nisling ice by streams, and subsequent glaciation.

The occurrence of large erratic boulders on plateau surfaces at about 1500 masl throughout the region appear to be associated with Nisling glaciation. It is interesting to note that in the Alaska Range, the Browne Glaciation (early - middle Pleistocene) is also characterized by large erratic boulders at high elevations. The best defined remains of the Ruby Ice-Sheet are found in the Ruby Range. Glaciated valleys are the most common expression of this glaciation. Based on the occurrence of lateral moraines, it is estimated that Ruby ice reached a maximum elevation 150 - 300 meters below that of Nisling ice. Hughes et al., (1969) stated that during Ruby glaciation, some small independent ice-caps were sustained in areas of the Ruby and Kluane Ranges at elevations above 1950 m.

Ruby ice covered the Icefield Ranges in the form of a mountain ice-sheet, similar to that which currently exists, but at higher elevations. In the valleys of the Icefields Ranges ice reached up to 2250 masl. Shakwak Valley contained a major icefield which reached elevations of about 1500 masl in the southeast and 1200 masl in the northwest. A northwesterly ice flow is indicated by this ice-gradient and by the alignment of drumlins in Shakwak Valley. On the northeast side of the valley, ice flowed north through the gaps of Talbot and Brooks Arms, Kluane and Donjek Rivers, and other gaps through the lower end of Ruby Range.

In the Ruby Range, the invasion of ice from Shakwak Valley combined with local ice centres, and reached an elevation of about 1,500 masl in the southeast, and 1,200 masl in the northwest. From the Ruby Range, the ice entered a network of interconnecting valleys in the Nisling Range, terminating at elevations of about 900 masl. After retreat of Ruby ice, deep stream canyons were incised in many valleys of St. Elias Mountains. However, similar valley incision did not occur in Ruby Range. Valley glaciation associated with the St. Elias glacial advance was generally confined to St. Elias Mountains, and have association with restricted cirque glaciation in Nisling, Ruby, and Kluane Ranges. The upper limit of the cirques in Nisling Range is about 1,650 masl, and approximately 1,800 masl in Ruby and Kluane Ranges.

The regional patterns of glacial flow and limits in the Ruby Range and bordering areas have been described by Bostock (1966), Muller (1967), Denton and Stuiver (1967), Rampton (1971), Hughes (1989) and others. For the most part, these descriptions have focused on the interpretation of glacial landforms with relation to glacial limits. However, since the Ruby Range lies within a region that includes currently glaciated areas in the St. Elias Mountains to the west, areas to the south and southeast that underwent repeated glaciation during the Pleistocene, and regions to the north and northeast that escaped glaciation altogether, it is not surprising that the Ruby Range has presented problems for interpretation. Specifically, although evidence of former glaciation in valleys is abundant, to date, no evidence of major glaciation affecting the plateau regions of the northern Ruby Range has been presented. To demonstrate the uncertainty that has evolved with regard to the question of glaciation of the plateau regions, Table 14 presents interpretations of glacial limits in the Ruby Range and surrounding areas offered by various researchers.

Johnson (1983) stated that the source of complete regional ice cover of the Ruby Range was primarily from the St. Elias Mountains and that the flow of this ice would be determined by ice thickness in relation to the altitudinal range of the underlying topography. Since the maximum extent of glaciation (or limits) are shown to lie just to the north and northeast of the area, the ice thickness could not have been excessively large compared to the altitudinal extent of the Ruby Range, and thus the flow patterns of the ice would be dominated by topography.

Ice moving into the Ruby Range would initially conform to the major valleys (Raft Creek, Gladstone Creek and Cultus Creek), with subsequent overtopping of the plateau surface as ice build-up in the Kluane Lake basin progressed. Since some plateau surfaces are in excess of 2150 masl, and Kluane Lake surface is at 900 masl (with considerable depth), the ice build-up in the Shakwak Trench would have to be at least 1500 m thick before the Ruby Range could be inundated. Therefore, ice build-up in the Kluane Lake basin was crucial to the complete regional glaciation of the Ruby Range. Johnson (1983) speculated that the occurrence of this regional glaciation must have been early Pleistocene.

Other researchers have portrayed major regional ice flow direction in the Ruby Range and surrounding areas as being north, northeast, and/or northwest (Table 14). This situation is not necessarily in conflict with Johnson's (1983) scenario as confluence between ice originating from the St. Elias Mountains and the Coast Mountains likely followed, producing a large centre of Cordilleran flow in southwest Yukon. However, these portrayals are generalized, and as indicated on Table 14 (Incomplete ?), do not necessarily imply complete inundation of the Ruby Range.

Table 14
MAXIMUM ALTITUDE ATTAINED
BY PLEISTOCENE GLACIATION(S) IN THE RUBY RANGE

Author Reference	Maximum Elevation (m)	Comments
McConnell (1905)	1,560	
Bostock (1952)	1,545	Most summit areas escaped glaciation entirely.
Denton and Stuiver (1967)	1,560	
Muller (1967)	1,800	Upper limits determined from lateral moraines on valley sides.
Prest et al., (1967)	Incomplete ?	Maximum extent of pre-Wisconsin glaciation east of Ruby Range in Nisling Range. Main ice flow north.
Hughes et al., (1969)	1,800	Maximum extent of pre-Reid glaciation is uncertain and tentatively placed just east of Ruby Range in Nisling Range. Main ice flow north and northeast.
Tempelman-Kluit, (1974)	Incomplete ?	Maximum extent of glacial limits in Aishihik and Sekulman lakes region.
Dyke and Prest (1986)	Incomplete ?	Maximum extent of late Wisconsin glaciation shown as being in the Ruby Range.
Hughes (1989)	Incomplete ?	Maximum limits of pre-Reid and Reid glaciations uncertain. Tentatively placed east of Ruby Range in the Aishihik and Sekulman lakes region.
Morrison (1989)	Incomplete ?	Reid and McConnell glacial limits placed in the Aishihik and Sekulman lakes region. No pre-Reid limits identified in this area.
Jackson (1989)	Incomplete ?	Pre-Reid and Reid glacial limits uncertain, tentatively placed in Aishihik and Sekulman lakes region. Main ice flow direction shown as north.
Clague (1989)	Incomplete ?	Maximum extent of Pleistocene ice just east of Ruby Range. Main ice flow direction north, northeast and northwest.

Chapter 4
METHODS AND TECHNIQUES

4.1 Introduction

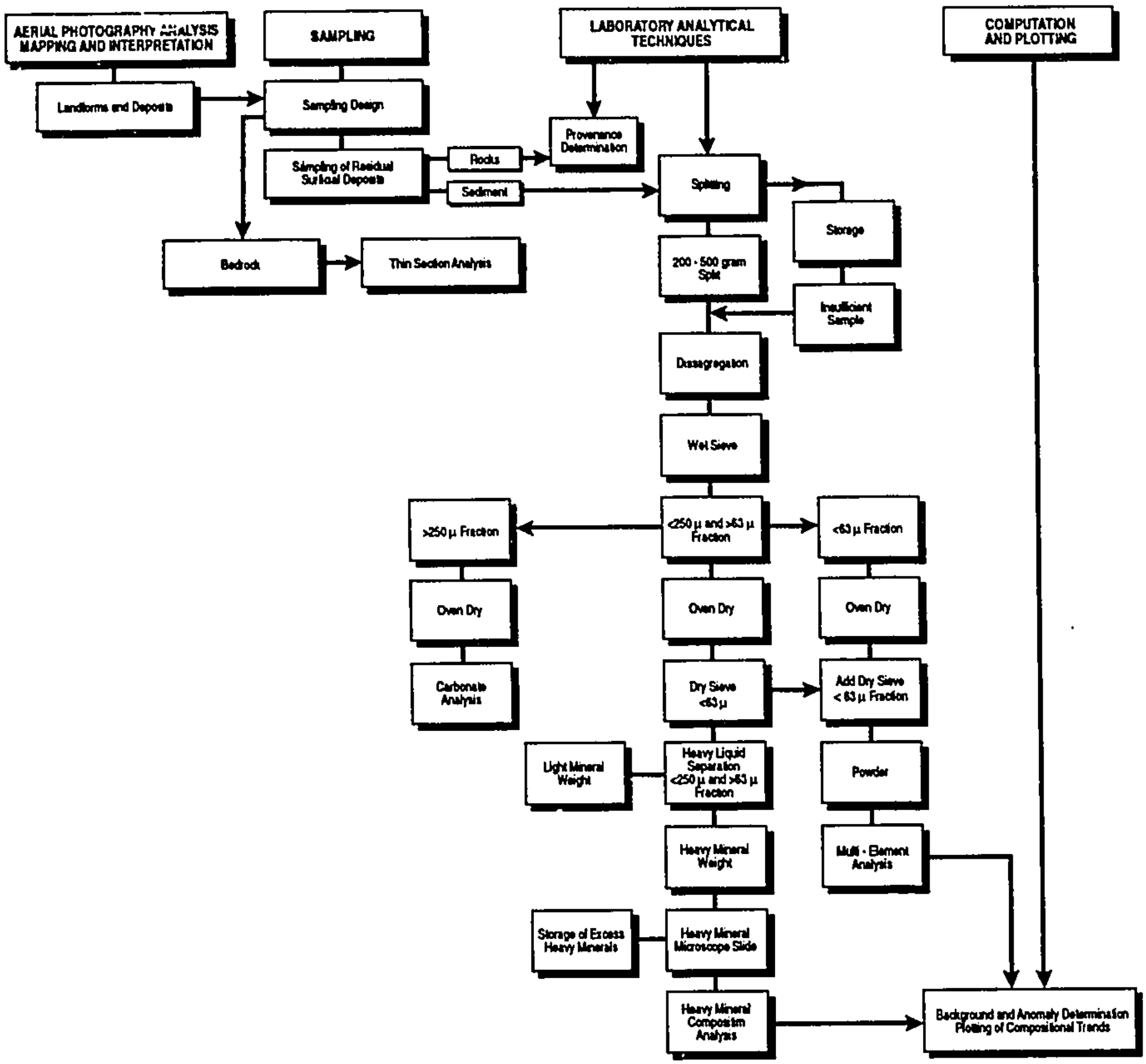
There have been a number of approaches to the interpretation of the Pleistocene history of regions. The Pleistocene history of a region is recorded in the stratigraphy, sediment compositional trends, striations and landforms preserved in the terrain. This study, of an area affected repeatedly by regional glaciation with contemporaneous up-valley and down-valley ice movement, the extent of which is under debate, utilizes a combination of methods and techniques. The research methodology is based on the premise that evidence of glaciation in areas now denuded of glacial landforms and identifiable deposits may be present in the residual deposits in the form of erratics, and heavy mineral and geochemical compositional trends. The data acquisition and analysis consists of; i) aerial photography analysis, including mapping and interpretation of landforms and deposits, ii) sampling of residual surficial deposits, iii) laboratory analytical methods, and iv) computational and plotting methods. Procedures utilized to secure data and enable analysis is demonstrated on Figure 8 in a flow-chart format.

4.2 Aerial Photography Analysis: Mapping and Interpretation

In describing the Pleistocene history of regions, geomorphologists have traditionally adopted a mapping program which includes extensive aerial photography interpretation, supported by field verifications. This method has proven invaluable in identifying and interpreting landform evolution (e.g., Muller, 1967; Prest et al., 1967; Ford, 1981; Geddes, 1981; St. Onge, 1984).

The mapping approach used in this thesis consists of two components. The first was the interpretation of aerial photographs; landforms were identified and delineated on photographs. The second was the field mapping of landforms that necessitated verification in the field and/or were not discernable from the aerial photographs. The mapping was based primarily upon the interpretations of Muller (1967), Rampton (1977), Johnson et al., (1984), and Dewez (1988). (Aerial photographs used for interpretation were 1979 Department of Energy, Mines and Resources, Government of Canada prints at a scale of 1:31,680; A25193, Lines 46-W, 47-E, 48-W, 49-W.). Detailed mapping of morphology was completed on a scale of 1:31,680. The map sheets used were issued as preliminary guide sheets for mineral claims by the Department of Indian Affairs and Northern Development. The identification of landforms enabled a more precise understanding of the geomorphological characteristics of the study area.

Figure 8
METHODS AND TECHNIQUES



4.3 Sampling

Three summer field seasons, each of approximately 60 days duration were carried out between 1982 and 1984. Initial reconnaissance and mapping of the field area was conducted during 1982 fieldwork. Extensive traverses of plateau regions and interconnecting valley systems enabled a better appreciation of bedrock variations as well as the nature and extent of landforms.

A sampling program was initiated during the first field season. Target areas were also identified for detailed study in the following field season. The targeting of the areas involved the delineation of regions which included extensive plateau and valley regions which were readily accessible. Activity during the third field season (1984) centred around sampling in specific areas that were found to present gaps in the 1983 sampling network (ie. missing samples), and/or were insufficient in sample size to allow for laboratory analyses. Samples were taken as sub-surface grabs, below any noticeable soil development and ash deposits. Sampling of rock types, surface and sub-surface, was also carried out during sediment sampling.

Soils in the area consist of Cryosols and Regosols. Horizon development is poor to non existent, and organic layers, where encountered, do not exceed 20 cm. Where permafrost was encountered, samples were collected at the permafrost/active layer boundary. As with the sampling of sediments at the bedrock/sediment interface, sampling at the permafrost/active layer boundary is where the most significant concentration of heavy minerals would be found (Boniwell and Dujardin, 1964; McAnerney, 1967; Coope, 1968; Shilts, 1973; Levinson, 1980; Ridler and Shilts 1974; Shilts et al., 1979; Henderson per comm.; Paré per comm.). Samples were approximately 400 to 1000 grams to allow for laboratory splitting (so that original samples could be utilized later if needed).

Daugherty (1974) stated that the essence of sampling lies in the fact that a large number of items or locations may, within specific limits of acceptance, be represented by a smaller group of samples selected from a larger group. If sampling is carried out effectively, a limited number of samples will be sufficient for making generalizations. The key to success in sampling lies in adopting a procedure which permits satisfactory conclusions to be drawn about a parent unit from a sample of minimal size (Peltier, 1962; Chorley, 1966; Kershaw, 1973; Dixon and Leach, 1977).

Three basic considerations were addressed before the sampling procedure was adopted:

1) The size and extent of what was being sampled was determined.

2) The most appropriate sampling procedure was adopted taking into account time constraints and field conditions.

3) The minimal size of sample both in number and volume for reliable representation of the feature was considered.

Numerous texts have been written solely on the subject of sampling in the environment (e.g., Watterson and Theobold, 1979; Green, 1979; Gy, 1979; Sanders et al., 1983; Keith, 1988; Horowitz, 1991). The design of a sampling program should ideally be predicated on what the researcher is trying to discover (Horowitz, 1991). In attempting to solve most geochemical and mineralogical field problems (Klassen and Shilts, 1977; Bolviken and Gleeson, 1979; Levinson, 1980; Henderson 1985, 1989), and geological problems (Milner, 1976; Jaeger, 1979; Mathewson, 1981), researchers have conventionally collected as many samples as time and laboratory facilities permit and to spot sample localities as evenly as possible over the study area. A sampling design of this nature implies some rather restrictive assumptions about the variability of mineral composition of the rock and/or sampling medium. It assumes that the material sampled is fairly uniform on a local scale as exemplified by the taking of only one sample per locality, and that a more important variation in composition is exhibited on a regional scale as samples from many parts of the parent unit are needed.

Sampling used in terrain analysis or surveys are generally collected at or near the surface to cover a specific area (e.g., map areas). Samples are usually collected on a grid pattern or a linear pattern (DiLabio, 1989). For example, in sampling sediment in Keewatin, Northwest Territories, Shilts et al., (1979) adopted a systematic sampling approach whereby, on an areal basis, even coverage was desired. Likewise, while sampling long continuous eskers, Shilts (1973) systematically sampled at predetermined intervals of 125 m. On a larger scale, Kaszycki (1986) adopted a grid sampling procedure where an area of approximately 1,600 km^2 was being studied for regional distributions of background mineral values in till.

Any decision as to the number of samples needed or distribution of sampling locations can only be based on knowledge of the natural variability in composition of the sampling medium. Visual assessment and interpretation of unit mineralogical composition and physical features are stressed as being essential before any sampling program can be undertaken (Mathewson, 1981). The sampling approach used in this study was designed on the basis of information gathered from the reconnaissance field season. The terrain of the study area varied greatly in the presence and/or absence of glacial landforms and, more importantly, the type and extent of sediment cover. On plateau surfaces, where sediment cover was generally sparse and discontinuous, the sampling procedure adopted was a systematic random approach (Dixon and Leach, 1977). This facilitated the sampling of those areas on plateaus where sediment did exhibit a somewhat uniform areal extent, generally not exceeding 0.5 km^2, and also those areas where sediment was sparse.

In valley locations, where sediment cover was generally widespread and recognized as till, a systematic random sampling approach was also followed. The reasons for this were two fold. Firstly, although sediment cover is widespread, it varies greatly in thickness within and between valleys, hence where sediment was thickest, it was deemed necessary to take a number of samples from that immediate unit. Secondly, on lateral and end moraines, systematic sampling of these features was often interrupted by rock slides or debris flows which overrode and in some cases completely obliterated the moraines. Hence random sampling of these linear features was necessary.

The procedure for identifying "erratic" boulders and smaller rocks consisted of two methods. Firstly, during sampling of sediment throughout the study area, rock samples were collected; this was done randomly. However, special consideration was given to the inclusion of rocks that did not appear to be of local origin. This procedure was adhered to in both plateau and valley locations. Secondly, sampling of rocks was carried out during traverses across the study area. The traverses were designed on the basis of areal coverage and accessibility, hence both the collection of rocks samples and the identification of erratic boulders were done so on a random basis. All sampled locations were plotted on aerial photographs.

4.3.1 Heavy Mineral and Geochemical Considerations

Major programs of sediment sampling for anomalous minerals were first undertaken by the Geological Survey of Finland (Wennervirta, 1973; Kauranne, 1976; Stigzeleus, 1977). The program, upon its completion in 1975, had enabled a total of 100,000 samples to be collected (program was initiated in 1971). This experience brought to light the recognition of various problems associated with sampling, as well as new approaches to sampling methods (Boniwell and Dujardin, 1964).

Similar sampling programs in Canada were underway in the late 1950's (Ermengen, 1957). In the 1960's and 1970's, the development of more efficient drilling techniques enabled sampling at greater depths on the Canadian Shield where thick sediment cover was encountered (Van Tassel, 1969; Gleeson and Cormier, 1971; Sopuck and Lehto, 1978; Veillette and Nixon, 1980). The majority of the sampling programs were initiated to explore for ore deposits and the geomorphological implications of mineral dispersal patterns have often been ignored.

Relatively little is known about the applicability, limitations and potential for successful analysis of heavy minerals in areas of permafrost (continuous and discontinuous). In continuous permafrost zones in Canada, only a few detailed studies have been undertaken in this regard (e.g., Allan and Hornbrook, 1970, 1971; and Allan et al., 1972 in the Coppermine River area; Shilts, 1973; and Shilts et al., 1979 in the Keewatin District).

Numerous mineralogical and geochemical surveys, primarily by mining companies, are also known to have been completed. Studies in discontinuous permafrost zones include areas of Yukon such as the Keno Hills (Boyle, 1965), Casino (Archer and Main, 1971), Whitehorse (Smith, 1971), the Northwest Territories (Paré, 1982), areas in Manitoba (Kaszycki, 1986), Hudson Bay and James Bay (Henderson, 1988). Some studies have also been conducted in Alaska (e.g., Brosge and Reiser, 1972). Fundamental work has also been conducted in the Soviet Union (Ivanov, 1966; Pitulko, 1968; Shvartsev, 1972). In most studies, the active layer is referred to as being a preferred sampling medium.

During the process of cryoturbation, material and moisture in the active layer will move or migrate in a circular fashion above the permafrost. The features associated with this process have been identified as mud boils (Shilts, 1973). They are also known as nonsorted circles (Washburn, 1969), or as clay boils, tundra circles and medallion patches in the Soviet literature (Shvartsev, 1972). For the purpose of this thesis, these features will be referred to as mud boils.

Shilts (1973) suggested that mud boils develop primarily on poorly sorted silt and/or clay rich deposits. They may also form on fine grain marine sediments. They are characterized by a segregation of clays and silts in the centre of the boils and coarser materials including stones around the sides. Allan and Hornbrook (1970) sampled the centres of mud boils and found that at or near the surface, the churning action associated with the boils formation results in a representative sample of mineralization from within the active layer and from the top of the permafrost. Similar conclusions were made by Shilts (1973).

4.4 Laboratory Analytical Methods

Shilts (1973) defined four class sizes with significantly different compositional characteristics to adequately describe sediment composition:

i) Gravel-size clasts (>2 mm) are composed primarily of rock fragments.

ii) Fine sand fraction (0.063 to 0.25 mm) is comprised primarily of mineral grains which include feldspar, calcite, quartz, and the bulk of the heavy mineral assemblage.

iii) The silt fraction (0.002 to 0.063 mm) constituting mainly quartz, calcite, feldspar, and minor proportion of heavy minerals.

iv) The clay fraction (<0.002 mm) consisting of mainly clay and other micaceous minerals as well as minor amounts of quartz, feldspar, calcite, and aggregates of clay-size iron and other oxides

Laboratory procedures centred upon the analyses of sediment samples, including heavy mineral analysis of the fine sand fraction, geochemical analysis of the silt and clay fractions, and carbonate determination of the greater than 2 mm size fraction. Various bedrock samples were also analyzed for constituent minerals. The techniques adopted for laboratory procedures are standard in that they are well documented in most sedimentary petrology textbooks and are adhered to by most laboratories conducting mineral and sediment related studies (Folk, 1968; Hough, 1958; Pettijohn, 1957; Milner, 1962; Parfenoff et al., 1970; Bowles, 1971; Carver, 1971; Geological Survey of Canada, Terrain Sciences Division; Bondar-Clegg Company Ltd., Ottawa). The techniques differ from most heavy mineral studies of sediments in that the >3.3 specific gravity (sg) fraction is analyzed rather than the >2.8 sg fraction, and a stereoscopic (binocular) microscope is used to count the grains instead of a petrographic microscope. These technical variations will be referred to in the appropriate sections.

4.4.1 Sample Preparation

Laboratory analyses of samples required an initial splitting of each sample so that approximately 200 to 500 grams of sample were processed (wet or damp weight). Splitting the samples was done manually (original samples were 400 to 1000 grams). Where the original sample weighed less than 400 grams, the entire sample was utilized. Each sample was placed in a numbered beaker. Dissaggregation and dispersion was then instigated to separate individual grains in the sample. This was accomplished by adding a solution of sodium hexametaphosphate (100 ml of 0.5N) to the sample. It should be noted that a variety of dispersants were tested to see which was most efficient. Tested were ammonium hydroxide, sodium carbonate, sodium oxalate, and sodium hexametaphosphate. It was found that sodium hexametaphosphate was the most effective dispersant and posed few health risks. If organic material was present, 10 ml of hydrogen peroxide (30%) was added to the sample solution. The samples were then left to sit overnight. If, after this procedure, dissaggregation was not complete, the sample solution was placed in a sonic bath.

The sample was then wet-sieved by placing a 60 mesh sieve (250 micron) on top of a 230 mesh sieve (63 micron) (Wentworth 1922, McDonald and Kelly, 1968). Three fractions were then separated and collected; i) >250 micron, ii) <250 micron and >63 micron, iii) <63 micron. After drying overnight in an oven at moderate temperature, the fine sand fraction (<250 micron and >63 micron) was dry-sieved to ensure that the separation was complete. The less than 63 micron fraction was dried and powdered. The greater than 250 micron fraction was dried and later separated with a 2 mm (10 mesh) sieve to produce two fractions; >2 mm, and <2 mm.

4.4.2 Carbonate Analysis: Gravel-Size Clasts

Carbonate determination was carried out on the greater than 2 mm fractions. Generally, the maximum size of grains analyzed did not exceed 3 cm. Samples were spread out in a plastic weigh boat and a solution of hydrochloric acid (1N HCL) was dropped on them. If a reaction with the acid occurred with any number of grains, the sample number was recorded.

4.4.3 Heavy Mineral Separations: Fine Sand Fraction

Heavy mineral separation procedures used in this study are described by in a number of texts and studies (e.g., Krumbein and Pettijohn, 1938; Milner, 1940; Twenhofel and Tyler, 1941; Krumbein and Sloss, 1951; Berry and Mason, 1959; McDonald and Kelly, 1968; Paré, 1982; Peuraniemi, 1990). Depending on the separating liquid, the specific gravity at which heavy and light minerals are separated will vary. For example, bromoform has a specific gravity of 2.89, acetyne tetrabromide has a specific gravity of 2.96, methylene iodide has a specific gravity of 3.3. The separating liquid will determine at which specific gravity heavy minerals are separated from the light minerals. Therefore, this decision necessitated consideration as to whether or not certain minerals whose specific gravity ranges between 2.89 and 3.3 are to be included as heavy or light minerals.

Complicating this concern is the problem of maintaining the separating liquid at the required specific gravity. Dilution with acetone or alcohol (rinsing agents) and impurities in the liquid (fine grain particles, dust) can lower the specific gravity of the separating liquid. In order to avoid inclusion of unwanted minerals in the heavy fraction, primarily quartz (sg 2.65) and feldspar (sg 2.66), the separating liquid used was methylene iodide. Hence the specific gravity of 3.3 was chosen as the separation between heavy and light minerals.

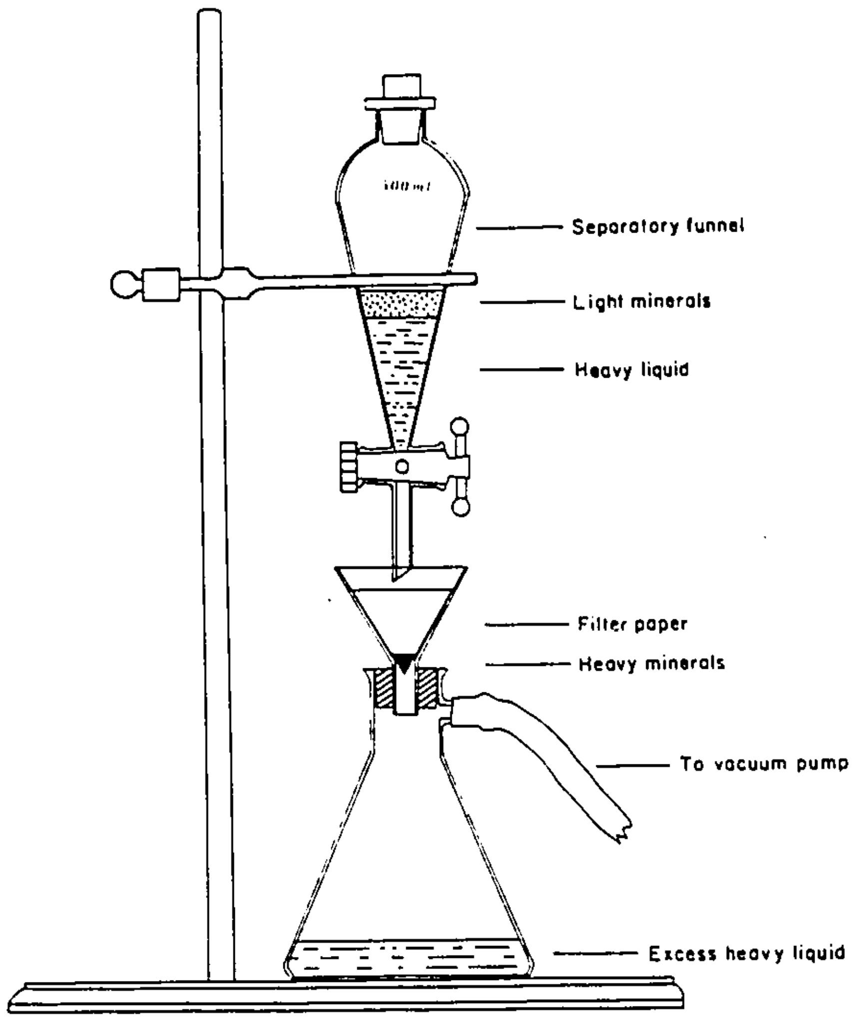
In this regard, it was decided that the use of bromoform might pose problems with respect to unwanted minerals being included in the heavy mineral fraction. As well, as with all heavy liquids, there are difficulties in ensuring a constant specific gravity. For example, if the specific gravity of bromoform was to be inadvertently reduced from 2.89 to 2.65, then various minerals such as quartz (sg 2.65), feldspar (sg 2.66), calcite (sg 2.71) and dolomite (sg 2.85) could be included in the heavy mineral fraction.

Because the geology of the study area is primarily made up of granodiorite, the possible inclusion of quartz and feldspars in the heavy fraction would give the heavy fraction an abnormally high weight percentage compared to the light mineral fraction and these minerals would not serve as suitable indicators. Various authors discuss the importance of keeping separating liquids at their required specific gravity to ensure the exclusion of quartz, feldspars, calcite, and dolomite in the heavy mineral fraction (Feo-Codecido, 1956; Berry and Mason, 1959; Luepke 1985). Heavy minerals are separated from the fine sand fraction (63 microns to 250 microns) by gravity settling in glass separatory funnels (Figure 9).

Approximately 100 ml of methylene iodide is added to a 500 ml separatory funnel, sample is then added to the separating liquid. Usually not more than 100 g of sample are added. If the total sample is larger than 100 g then either more methylene iodide is added or another separatory funnel is used. After approximately 10 minutes of settling the sample is stirred. After another 10 minutes, with vacuum on, the separatory funnel is drained into a funnel and flask containing a #1 Whatman filter paper. Heavy minerals and an amount of methylene iodide is released. After the pure methylene iodide has been drained from the heavy minerals, the filter with the minerals is placed on a washings flask and cleaned with acetone. A second #1 Whatman filter paper is placed on the flask containing the pure Methylene Iodide. The light minerals are released in the same manner as were the heavy minerals. The acetone washing procedure is repeated for the light minerals. Both heavy and light minerals are then placed in the oven at low setting to dry overnight.

The heavy and light mineral fractions are then weighed. The magnetic fraction is then separated. The heavy minerals are spread over a piece of craft or similar smooth paper. A hand controlled release magnet is then passed over the minerals collecting the magnetic fraction. These are released over a second piece of paper. This procedure is repeated several times on each sample. The weight of the magnetic fraction is recorded. The magnetic minerals are then returned to the heavy fraction. The procedures associated with the preparation of grain slides are treated by Krumbein and Pettijohn (1939). The process is straight forward and does not require detail discussion. Heavy minerals were mounted in Araldite 502 (n=1.56) on slides using a representative split where sufficient fraction was available.

Figure 9
HEAVY MINERAL SEPARATION APARATUS



4.4.4 Geochemical Analysis: Silt and Clay Fraction

The distribution of elements can be related to the provenance of the material and be utilized in determining dispersal patterns in sediments, which can be traced back to a source area (Speidal and Agnew, 1982; Peuraniemi, 1990; Saarnisto, 1990). The clay-size fraction (< 2 micron) is the best fraction to use when analysing for many elements because it is more metal-rich than other fractions. The clay-size material is also more chemically reactive because of the larger surface area which also increases the ability of certain minerals to adsorb mobile elements (DiLabio, 1989; Henderson, 1989).

This fraction is composed of detrital material produced from the weathering and erosion of bedrock and other coarser sediment. Transport by fluvial, glacial, aeolian, and other means results in a terrigenous material which is subsequently modified by diagenetic processes (Henderson, 1989). If there is an insufficient amount of clay-size fraction to conduct analysis (as was the case in this thesis), the < 63 micron fraction can be utilized (DiLabio, 1989; Saarnisto, 1990; Horowitz, 1991).

The oven-dried < 63 micron fraction is placed in a mortar and ground by pestle until the sample has the appearance of talcum powder. The sample is then removed with a teflon spatula and placed in a vial. Analysis was contracted to Bondar-Clegg and Company Ltd. of Ottawa. The elements analyzed were part of a suit offered by Bondar-Clegg. The trace element analysis consisted of a quantitative multi-element analysis where the sample is digested with aqua-regia. The sample is then diluted with deionized water and measured on the atomic absorption unit. The following elements were analyzed; Copper (Cu), Lead (Pb), Molybdenum (Mo), Cobalt (Co), Nickel (Ni), Chromium (Cr), Manganese (Mn), Cadmium (Cd), Silver (Ag), Bismuth (Bi), Iron (Fe), Arsenic (As), Zinc (Zn), Vanadium (V), Tellurium (Te), Uranium (U), Tungsten (W), Antimony (Sb), Selenium (Se), Tin (Sn), and Gold (Au).

Gold (Au) is analyzed by fire assay where a pre-concentration is accomplished by fusing 10 grams of finely pulverized sample with a flux of litharge, borax and soda ash. The precious metal is collected in a resulting lead button. The lead is evolved by cupellation and the resulting precious metal is dissolved in aqua-regia. The final measurement is by atomic absorption. Detection limits range from 0.5 ppm for Cadmium to 10 ppm for Uranium. The detection limit for most other elements is 1 ppm.

4.4.5 Heavy Mineral Identification

Heavy minerals were identified and counted under a stereoscopic binocular microscope using reflected light following procedures described by Paré (1982), and Henderson (1989). There is on going discussion as to what counting technique provides best results. There are few papers dealing with the statistical aspects of the counting technique for example; Smithson, 1930; Dryden, 1931; Berthois, 1947; Imbrie and van Andel, 1964; Galloway, 1972; all of which appear in Luepke's (1985), Benchmark Papers in Geology Series.

Van Der Plas (1962) contends that the results of line and point counting analyses lead to results that are biased by the sampling technique. For a line counting technique, the probability of a particle being crossed by a line is proportional to the diameter of the visible surface normal to the line. For a point counting technique, the probability of a particle being touched by the cross lines is proportional to the area of the visible surface. Hence, Van Der Plas (1962) suggests that a ribbon technique be used. This method of sampling may be regarded as stratified random as the sampled grains are determined by a systematic shifting of an area over the surface of the slide. All grains within the ribbon area are identified.

Henderson (1989) advocates the use of a point counting technique for identifying minerals and asserts that if a sufficient number of grains are counted (200-300), the random nature of minerals being counted eliminates any sampling bias. Point counting techniques, with various technical modifications, have traditionally been used by researchers studying grain slides as well as thin sections (Paré, 1982).

It is generally agreed that the more grains identified the more representative or reliable the results will be (Dryden, 1931; Paré, 1982; Henderson, per comm.). The assumption must first be made that the heavy mineral grain slide is an average or representative sample of that fraction. Many researchers state that a count of 500 to 1,000 grains would allow an accurate statement of mineral composition. However, it is also stressed that the law of diminishing returns probably will make us content to count only a few hundred grains, depending on the statistical degree of accuracy required.

In various studies, it was shown that a 200 grain count is required for an accuracy of about 5%, almost 4,500 grains would be required for an accuracy of 1% (Dryden, 1931; Paré, per comm.). In this study, the vast majority of slides had 200 grains identified, on twelve slides 209 to 300 grains were counted, and on three slides 122 to 184 grains were identified. The number of grains counted was based on abundance of grains, and the counting technique; a Swift mechanical stage and automatic counter (Paré, 1982; Henderson, 1989).

4.5 Computational and Plotting Methods

Graphically illustrating variations of heavy minerals and geochemical composition in sediments from the northern Ruby Range is made difficult because of field and laboratory related factors such as; i) inequities in sample density, ii) relatively small sample size, and iii) random fluctuations in specific values. Henderson (1989) experienced similar problems while attempting to illustrate variations in compositional trends in Hudson Bay. Such restraints can lead to extensive extrapolation between sample sites, therefore introduce biases in plotting. In order to minimize this concern, both computer assisted and manual plotting techniques were employed.

The plotting of the heavy mineral and geochemical compositional trends was facilitated by the use of Synagraphic Computer Mapping adapted to the University of Ottawa I.B.M. 360/65 computer. The output consisted of an isopleth map where the isolines shown on the map are for specific values and the values are assumed to smoothly vary over the interval between any two adjacent isolines forming a continuous surface. This technique is widely utilized in heavy mineral and geochemical plotting programs (Agterberg and Chung, 1975; Henderson, 1989; Paré per comm.) and produces similar results as the Appmap programs (Ellwood, 1981).

Since the primary purpose of the heavy mineral and geochemical data analysis was to establish anomalous values and dispersal trends, it was felt that relying solely on the computer generated plots may minimize the effects of local concentrations, and mask anomalous values. Therefore, the computer generated plots were used for initial assessment of regional trends, and later complemented with manual plotting to ensure that isoline intervals indicated general trends, and identified local anomalies.

Chapter 5
ANALYSES AND INTERPRETATION

5.1 Aerial Photographic Analysis: Mapping and Interpretation

The mapping of the geomorphology in the study area (Figure 10) was accomplished by analyzing aerial photographs, incorporating previous interpretations (e.g., Muller, 1967; Rampton, 1977; Johnson et al., 1984; Dewez, 1988) and by field verifications.

In the study of alpine glaciations, both current and past, researchers have for the most part been confronted with relatively simple glaciological problems. For example, most glaciers find their source in cirques and/or icefields of various sizes. The resultant glacier flow is generally controlled by topography, hence down valley or down slope flow results.

Areas peripheral to centres of alpine glaciation such as the northern Ruby Range provide unique environments for Quaternary studies by posing numerous problems with respect to mapping and interpretation. The implications of multiple glaciations throughout the Pleistocene, ice movement up and down valleys contemporaneously from various directions, changes from total ice cover to remnant ice activity, as well as consideration of areas that either escaped glaciation entirely, or the glacial history is uncertain, must all be considered.

Probably the biggest problem is the difficulty of dating all the events identified in the field and hence the presentation of a chronology. Late Pleistocene dating has generally depended upon the occurrence of material which can be processed by the Carbon 14 technique. Material suitable for this technique has not been found in the northern Ruby Range and therefore absolute dates are not available.

5.2 Plateau Regions

The plateau regions in the study area present a challenge with regard to geomorphological interpretation. Initial analysis of aerial photographs, and subsequent field observations, demonstrated that the relatively smooth, undulating terrain of the plateaus was void of any conspicuous landforms except for a few tor-like structures along with felsenmeer, and generally resembled a periglacial landscape. The felsenmeer grade into a system of massive pattern ground formations which, from the lichen cover, appear to have been inactive for a very long period of time (Johnson et al., 1984). Denuded kame and kettle topography is evident at some lower plateau elevations occupying col-like positions (not mapped).

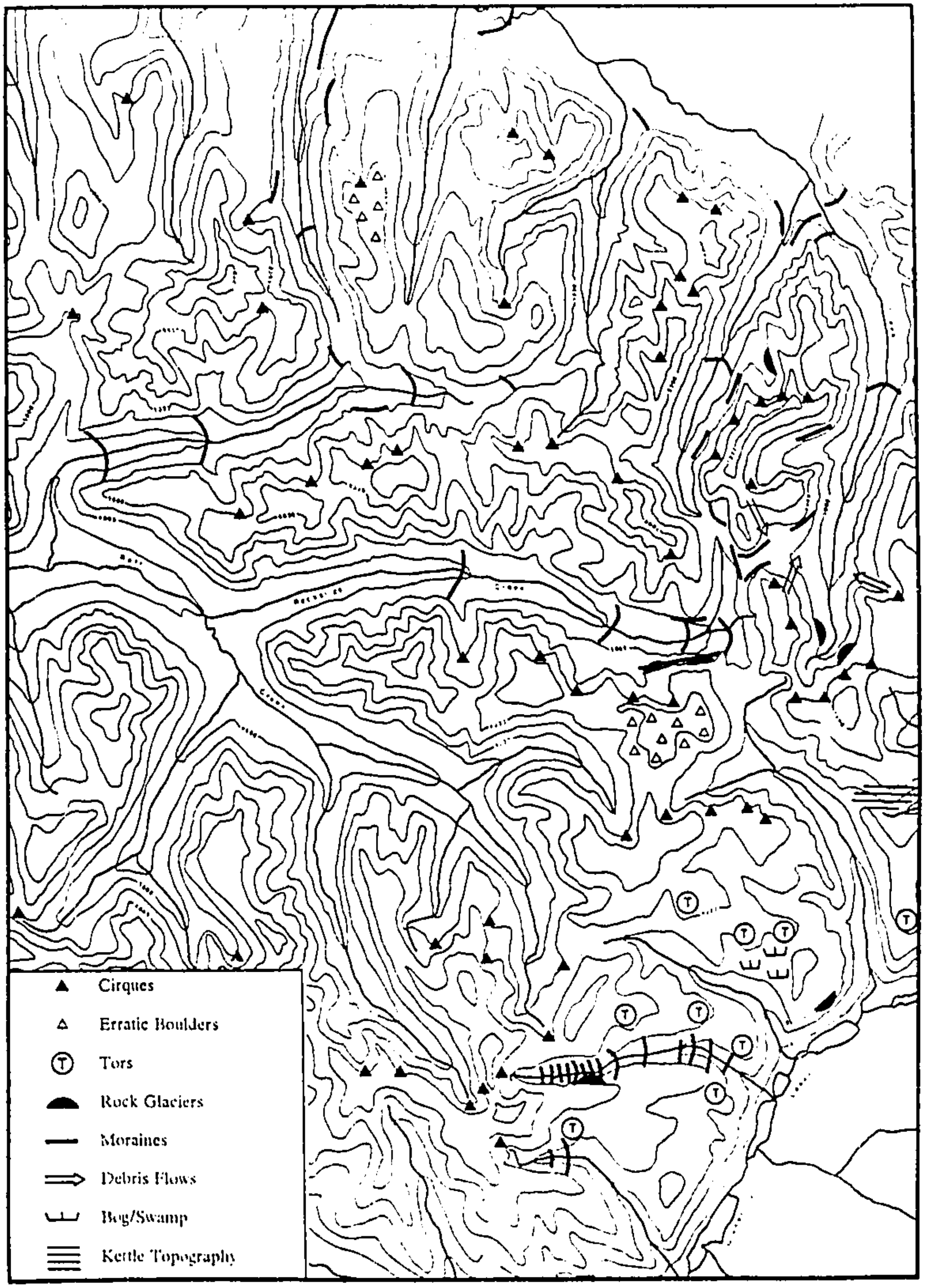


Figure 10
GENERAL GEOMORPHOLOGY OF THE NORTHERN RUBY RANGE

The plateau surfaces have been characterized as an old erosion surface of the Yukon Plateau which has been uplifted and subsequently incised by streams, resulting in narrow valleys which were later modified by glacial activity (Cockfield, 1922; Bostock, 1948; Price, 1973; Tempelman-Kluit, 1974; Hughes, 1990). It is of interest to note that Rampton (1982) provided a similar description of the Yukon Coastal Plain. Although no detailed evolutionary geomorphological scenario is available for the Ruby Range, inferences may be drawn from Rampton's (1977) description of the Burwash Uplands which is identified as sharing similar physiographic characteristics:

> *"The plateau-like surface forming the Burwash Uplands appears to be part of an erosion surface or pediment that formed between Paleocene or Eocene time and Pliocene or early Pleistocene time. Late Cenozoic tectonism and stream dissection have resulted in this surface being elevated above present stream levels."*

The glaciological implications of this description are two-fold; i) glaciation in late Tertiary and/or early Pleistocene could have preceded major stream incision, hence valleys may be post-glacial (i.e., formed after glaciation in late Tertiary and/or early Pleistocene), and ii) significant uplift of plateau regions may have occurred after the late Tertiary and/or early Pleistocene glaciation(s) (i.e., during middle - late Pleistocene).

Vanicek and Nagy (1981) concluded that the maximum rate of current uplift for southwest Yukon is about 2.4 cm/yr. There is no reported data on uplift in this region for the late Tertiary - early/late Pleistocene. However, if the rate of uplift is assumed to have been relatively constant from the later period of early glaciation(s) in late Tertiary - early Pleistocene (approx. 340 Ma), to the middle - late Pleistocene glaciation(s) (approx. 200 Ma) which subsequently modified the valleys, uplift during this 140 Ma period would total about 3,360 m. When denudation rates are taken into account (estimated by Slaymaker and McPherson (1977) to be upwards of 0.008 cm/yr for this region), and assumed to be relatively constant over the same 140 Ma period, net uplift would be about 3,348 m. If however, uplift was not constant during this period (which was likely the case), and to account for this, an uplift rate of 1.2 cm/yr (one-half of the 2.4 cm/yr as reported by Vanicek and Nagy, 1981) is assumed to have occurred, then even with a denudation rate of 0.008 cm/yr, net uplift would be 1,668 m. The existing data therefore suggest a considerable disparity between current rates of denudation and uplift. However, an unknown amount of uplift may be attributed to glacial unloading affects (Evans, 1989).

It is evident that uplift in the Ruby Range and surrounding regions occurred throughout the Pleistocene (Hughes et al., 1972; Vanicek and Nagy, 1981; Parrish, 1981; Naeser et al., 1982; Clague, 1989), but since current elevations in the Ruby Range reach a maximum of about 2200 masl, it is inconceivable that the region could have experienced uplift of 3,348 m during this period, even in consideration of a lower sea level. Conceptually, it would seem more likely that net uplift during this period was considerably less, perhaps on the order of 1,668 m as stated above. Furthermore, Muller (1967) contemplated that it was unlikely that denudation rates were much higher during this period:

> *"Except for the disappearance of the ice, the morphology of this region has changed little since Pleistocene times. The valleys are mostly filled with lakes, swamps, and sluggish streams, and little erosion or aggradation takes place."*

The glaciological implications regarding net uplift bring to light considerations which were previously not accounted for in discussions of complete regional ice cover of the Ruby Range. Johnson (1983) proposed a scenario of ice conditions necessary for regional ice cover of the Ruby Range whereby:

> *"During this situation the flow of ice in the Ruby Range would be determined by the ice thickness in relation to the altitudinal range of the underlying topography. As the glaciated region boundary is hypothesised to lie just to the north of the area one can postulate that the thickness of the ice was not large compared to the topographical range and that the stress patterns within the ice, and thus the flow of the ice, would be dominated by topography with convergent and divergent stress conditions. The ice moving into the Ruby Range would initially conform to the valleys of Raft Creek, Gladstone Creek and Cultus Creek with overtopping of the plateau surface as ice build up in the Kluane Lake basin continued. The Plateau surfaces lie at up to 2,200 m., 10 km. from Kluane Lake whose surface lies at about 900 m and the lake itself is of considerable depth. Thus the ice build up in the Shakwak Trench would have to be at least 1,500 m thick before the Ruby Plateau could be overtopped. Ice build up in the Kluane Lake basin is therefore crucial to the glaciation of the Ruby Range. Ice with an elevation of over 2,200 m would also imply total coverage of the Donjek and Kluane Ranges on the east side of the St. Elias Mountains. The occurrence of this total coverage must have been early in the Pleistocene..."*

However, since the altitudinal extent of the Ruby Range was considerably less in early Pleistocene than what is currently observed, in order to glaciate the Ruby Range, the thickness of ice build up in the Kluane Lake basin may not have had to be as extensive, and therefore not have been as crucial to the glaciation of the Ruby Range as previously thought. Furthermore, the Kluane Lake basin must owe much of its' current altitudinal extent to the affects of glaciation, therefore the basin may also have exhibited less altitudinal range, relative to surrounding areas.

In summary, it is thus conceivable that the northern Ruby Range (and surrounding regions) may have been glaciated (perhaps several times) in late Tertiary and/or early Pleistocene, undergone uplift with stream incision intensifying as uplift progressed, followed by middle - late Pleistocene glaciation(s) which modified the valleys. The plateau surfaces would have experienced periglacial processes continually throughout the Pleistocene, rendering any glacigenic material residual.

5.2.1 Tors

In the present discussion, in order to emphasize the uncertainty with regard to identifying "tors" as such in the northern Ruby Range, the term "tor-like structures" is used.

Rampton (1977) has described the Kluane Plateau (including the Ruby Range) as an environment which:

> *"During the Pleistocene, periglacial processes continually have affected the terrain, especially at high elevation. Tors and altiplanation terraces on ridges at high elevations are confined mainly to unglaciated parts of the Kluane Plateau... The area has been subjected to a number of late Pleistocene glaciations with high parts standing as nunataks above the ice."*

However, although the presence of tors and other periglacial features do not necessarily rule out the possibility of glaciation (Ives, 1974, 1975; Sugden and Watts, 1977), they do present difficulties to interpretation especially in the northern Ruby Range where they occur in close proximity to valleys which exhibit an array of glacially related landforms and deposits.

The geomorphological significance of tors has been an on going debate among researchers (Worth, 1930; Pullan, 1959; Demek, 1964; King, 1958; Price, 1971, 1972; Hughes, 1972, 1990; Sugden and Watts, 1977; Washburn, 1979). The term tor was defined by Pullan (1959) as:

> *"An exposure of rock in situ, upstanding on all sides from the surrounding slopes and it is formed by the differential weathering of the rock bed and the removal of the debris by mass movement."*

They are generally regarded as periglacial features, yet from a review of literature (e.g., Embleton and King, 1968; Washburn, 1979; Hughes, 1990), it is evident that there is considerable diversity of opinion as to the mechanisms of formation, and the climatic conditions under which they form. There are a number of hypotheses relating to the formation of tors which Hughes (1990) characterized as usually involving; i) a single stage in which there is concurrent differential weathering and removal of material, leaving more resistant rock exposed, or ii) a two stage development where deep differential weathering during a warm preglacial or interglacial interval is followed by a stage of exhumation under periglacial conditions.

Tor-like structures on the plateau surfaces of the northern Ruby Range have been observed at elevations over 1600 masl. They are hence above the limit attained by the Ruby Ice-Sheet (McConnell Glaciation), and some within the limit of the Nisling Ice-Sheet (Reid Glaciation). As well, according to Muller's (1967) glacial map, some cirques (attributed to the

St. Elias Glacial Advance and Ruby Ice Sheet), are at higher elevations than many tor-like structures. In other parts of the Ruby Range, Howse (per. comm.) has observed comparable tor-like structures above and below 1600 masl which would clearly place at least some within the limits of the McConnell Glaciation. Hughes (1990) on the other hand, has commented on presumably similar structures suggesting that they commonly appear on steep valley walls and distinguishes them from tors by calling them "castellated outcrops", "tumps", or "outcrops".

It has been the general belief of many geomorphologists that tors were much too fragile to survive glaciation and were therefore indicators of unglaciated areas (e.g., Linton, 1950; Dyke, 1976). However, Sugden and Watts (1977) stated that tors need not necessarily be diagnostic of areas that escaped glaciation. They have documented evidence of tors, felsenmeer, and glacial erratics at the same site and suggest that tors could have protruded above the surface prior to glaciation, and along with the felsenmeer, survived inundation by an ice sheet. Similar observations have been reported in regions of Norway (Dahl, 1963), Scotland (Sugden, 1968), and Baffin Island and Labrador (Ives, 1974, 1975).

Hughes et al., (1989) stated that some tors in southern and central Yukon may have formed by cryoplanation of larger rock masses. However, many tors are at lower elevations than these terraces, and even occur in some glaciated areas where there are no terraces, implying a different origin for these tors. One possible formative mechanism suggested by Hughes et al., (1989) is downwasting by solifluction whereby the tors would be left upstanding where the bedrock is more resistant to weathering and erosion.

Large boulders have been found at the same elevations as tor-like structures on the plateau surfaces of the northern Ruby Range and appear to be fresher in appearance than the structures, even though they are of the same rock type (granodiorite). These boulders may be attributed to ice moving up-valley, and overtopping plateau surfaces in isolated cases. Hughes (1990) has also identified glacial erratics at the base of a tor in the Duke Depression at just over 1500 masl and states that tors may have formed during periglacial or interglacial intervals and survived glaciation as roche moutonnée forms, and later undergone modification.

No new evidence is offered on the origin or age of tor-like structures in the northern Ruby Range. However, it is evident from the literature that the presence of tors, or tor-like structures, does not necessarily rule out glaciation (Dahl, 1963; Sugden, 1968; Ives, 1974, 1975; Sugden and Watt, 1977; Hughes, 1990). The possibility of selective glacial erosion and/or non-eroding (perhaps cold base) ice may have occurred throughout the late Tertiary and early Pleistocene, which could have resulted in minimal modification to the structures.

5.2.2 Other Periglacial Considerations

Most, if not all of the plateau area is underlain by permafrost. Active layer depths have been observed to be at about 40 cm. Although not shown on Figure 10, several features diagnostic of periglacial conditions are evident on these surfaces and include poorly sorted stone circles and stripes, mud boils, and solifluction lobes. Hughes (1990) identified a series of extensive cryoplanation terraces just east of the study area in the Aishihik Lake region, but similar features were not observed in the northern Ruby Range.

Interest in the periglacial landscape focuses on processes which take place in the surficial material left in the zone of seasonal freezing and thawing. This concern will be reviewed in a later section dealing with heavy mineral and geochemical compositional trends.

5.2.3 Erratics

The use of erratic materials as a criterion for determining whether plateau regions experienced glaciation in this study was complicated in part, by the logistical problems of investigating vast plateau areas, determining what constitutes erratic material based upon the current understanding of the local geology, and on the extent of weathering and periglacial activity in the area which may have altered or destroyed material which was erratic.

The problems associated with the covering of vast areas of the plateaus is self-evident. Logistically, not all regions can be examined in detail. Therefore, where erratics were not identified means that either they; i) have not yet been found, ii) have been altered or destroyed, or iii) do not exist. The geological mapping (Muller, 1969; Goodall, 1972) of the Ruby Range was on a large scale and hence is of limited assistance in determining the provenance of 'erratic' boulders on the plateau surfaces. Based on the current knowledge of the local geology, large, relatively fresh appearing 'erratic' boulders identified on the plateau surfaces are probably of local valley origin (granodiorite). Smaller rock fragments also appear to be of local origin.

Much of the erratic material may have been altered or destroyed in the periglacial environment, making identification extremely difficult. Remnant boulder blocks, some continuations of tors demonstrate that the landscape has evolved through the phase of tor formation to an almost completely denuded condition. It is therefore not surprising that erratic material, in the form of boulders or rock fragments are difficult or impossible to find (Johnson et al., 1984).

5.3 Valley Locations

Most valleys in the northern Ruby Range contain an array of glacially related landforms and deposits, while others exhibit little or no evidence of glaciation, a situation which may be attributed to glacier surging. Also evident in some upper valley locations are mass-wasting forms. The most common geomorphological expressions are glacially derived and are manifested in the form of U-shaped valleys, cirques, moraines, and till. Common in Talbot, Rockslide and Gladstone Creeks, and particularly in the down-valley sections, are glacio-lacustrine and glacial-fluvial deposits.

5.3.1 Glacier Surging: Geomorphological Implications

Although there is no apparent evidence for the former occurrence of glacier surging in the northern Ruby Range, the currently observed phenomena in the St. Elias Mountains suggests the possibility of prior occurrences in bordering regions (Johnson et al., 1984). However, though not mapped on Figure 10, it was observed in the field that some smaller U-shaped valleys appeared to be lacking or void of glacially related deposits, while neighbouring valleys (some less than 1 km away), exhibit numerous signs of former glaciation.

Surging glaciers can affect existing glacier deposits in a number of ways. Surging may obliterate morainal evidence of former ice marginal positions, construct new moraines, or incorporate various materials which may have been attributed to a combination of previous glacial events into a single new moraine. It is evident that the geomorphology of a valley that has experienced surging will contrast dramatically with a valley that has not. In the fossil landscape, the implications may be that sequences of glacial events will not necessarily be registered in neighbouring valleys, or even regionally. The occurrence of the surge phenomena in the St. Elias Mountains during the Holocene most certainly indicates similar occurrences throughout the Pleistocene in other regions; surely glacier surging is not an exclusive Holocene and St. Elias Mountain phenomena.

This prompts the question as to why surging occurs, and perhaps more important, whether glaciers in any given region register the same response to climatic events (synchronous and non-synchronous activity). Furthermore, since it is evident that the Holocene exhibits variable fluctuations with respect to climatic events, then similar fluctuations must undoubtedly have occurred throughout the Pleistocene, indicating time-transgressive responses.

5.3.2 U-Shaped Valleys

Without exception, all valleys in the northern Ruby Range exhibit a glacially modified U-shape. The evolution of these valleys has already been reviewed in a prior discussion and therefore only warrants a brief summary. Although no definitive explanation is offered with regard to valley evolution, it does seem likely that much of the glacial activity responsible for their form occurred during middle - late Pleistocene which corresponds to Muller's (1967); Nisling Ice Sheet, Ruby Ice Sheet and St. Elias Advance, or Bostock's (1966) and Hughes's (1989) equivalent; Reid Glaciation and McConnell Glaciation.

The largest valleys (Figure 10) are oriented east-west and include Gladstone, Rockslide, Raft, and Talbot Creeks, all of which are connected through a system of low passes. These valleys are widest at the junctures of the Talbot Arm and Kluane Lake, and narrow as the upper valley sections are approached. It is apparent from the dimensions of these valleys along with morainal evidence, that they were glaciated repeatedly from both up-valley and down-valley directions. Other, less extensive valleys (not formally named), generally oriented north-south, also take on a glaciated U-shape, but not all contain glacial depositional features.

5.3.3 Cirques and Associated Moraines

The cirques identified on Figure 10 have been attributed by Muller (1967) to the Ruby Ice Sheet and the St. Elias Advance (or, the McConnell Glaciation; Bostock (1966); Hughes (1989)). Cirques are most numerous in Raft, Talbot and Alaskite Creeks. Moraines marking the down valley extension of former cirque glaciers are generally well defined but not easily discernable from aerial photographs. There are no cirques in the northern Ruby Range assigned to the Nisling Ice Sheet (Reid Glaciation), although some are identified in the Nisling Range (Muller, 1967).

The Ruby Ice Sheet in the study area is described as having two main sources; i) local ice centres on plateau surfaces above the present 1800 masl interval which produced valley glaciers many kilometres long, and ii) cirques at between the present 1500 to 1800 masl interval that gave rise to alpine glaciers only a few kilometres long at most (Muller, 1967). Cirques associated with the St. Elias Advance are also located at elevations between 1500 to 1800 masl but are described as differing from their Ruby Ice Sheet counterparts by exhibiting well defined ground and end moraines, craggy walls, and in some instances tarns. Conversely, Muller (1967) observed that cirques lacking these features have been modified by relatively long subaerial erosion and are roughly contemporaneous with the Ruby Ice Sheet.

5.3.4 Moraines and Till

Lateral and end moraines in the valleys of the northern Ruby Range along with till deposits were attributed by Muller (1967) to the Ruby Ice Sheet and the St. Elias Advance. The most apparent and continuous moraines are found in the more extensive east-west oriented valleys which include Gladstone, Raft, Rockslide and Talbot Creeks (Dewez, 1988 provides a detailed morphological description of Gladstone Creek which constitutes the approximate southern limit of this study area). Lateral moraines of Ruby Ice Sheet origin are located at elevations between 1200 to 1350 masl. They are often overridden by talus and scree deposits, especially in the upper valley sections, making their delineation difficult.

The lateral and end moraines in the lower valley sections and some upper sections are primarily associated with regional up-valley ice penetration from Kluane Lake. Similarly, lateral and end moraines in the upper valley sections are attributed to local ice centres originating on the plateau surfaces which, probably contemporaneous with up-valley flow, produced valley glaciers many kilometres long. These are particularly evident in Talbot and Alaskite Creeks. In some instances, lateral moraines (of both up-valley and down-valley origin) in the larger valleys appear to have obstructed or blocked some of the smaller tributary valley glaciers (cirque) that are of Ruby Ice Sheet and/or St. Elias Advance origin. This has resulted in some relatively extensive debris buildup at these confluences.

Two rather extensive end moraines, from up-valley flow, are located in the upper reaches of Talbot and Rockslide Creeks, which judging from the dimensions of these moraines, are insufficient to produce major dams to drainage east or west. However, Johnson et al., (1984) identify a lake dammed in Talbot Creek between the moraine and the confluence of Alaskite Creek which drained to the north through Tyrell Creek, and is likely associated with the Ruby Ice Sheet. In the Gladstone Lake system, Dewez (1988) has documented evidence of the existence of ponding of meltwater and has proposed a model for glaciation and deglaciation in the area.

As described above, cirque glaciers, tentatively associated with the St. Elias Advance, exhibit relatively well defined end moraines, and generally, due to their relatively short distance of expansion, did not develop extensive lateral moraines. Based upon the positioning of end moraines, the maximum extent of many of these smaller cirque glaciers is typically not more than 0.5 km. It is of interest to note that in one valley (dubbed Moraine Valley), just to the north of the Gladstone Lakes (see Figure 10), thirteen recessional moraines have been identified, along with an end moraine at the confluence of Gladstone Creek. Although similar recessional moraine sequences have been observed in other valleys, they commonly have not numbered more than two or three. This indicates that glacial responses, with respect to retreat and re-advance are not necessarily synchronous from valley to valley or regionally, or if they are, the same response may not be registered throughout the region.

Till is found throughout the study area on the valley floors but based on field observations, it was evident that textural characteristics were clearly diagnostic of glacier extent. Specifically, till attributed to the more extensive up-valley and down-valley regional flow associated with the Ruby Ice Sheet contains a large amount of sand and pebbles, along with rock fragments, boulders and finer fractions. Cirque glacier derived till attributed to the Ruby Ice Sheet and the St. Elias Advance is dominated by larger fragments of rocks and boulders, with lesser amounts of finer materials. Ruby Ice Sheet and St. Elias Advance cirque glacier derived till can be differentiated as well on the basis of the relatively fresh appearance of the more recent St. Elias Advance deposits. In much of the till found throughout the area, a "popcorn" like characteristic of the coarse sand to granule fraction is evident, and is attributed to the relatively easily erodible granodiorite and stocks of alaskite.

5.3.5 Mass-Wasting Processes and Forms

There are a number of mass-wasting processes and forms evident in the valleys of the northern Ruby Range including, various types of avalanching, slushflows along stream courses, thaw slumping on steeper slopes, taluses, and rock glaciers. Since it is beyond the scope of this study to detail all these processes and related forms, this discussion will focus on talus and rock glaciers which can be addressed with respect to geomorphological and evolutionary significance.

The most common, currently active mass-wasting process contributing debris to, and modifying existing valley features, are talus. As well, widespread evidence debris form stabilized or inactive talus occur throughout the area with an established vegetation cover (indicating relatively little activity in recent time). Currently active talus are most pronounced in the upper valley sections of Talbot, Alaskite, Rockslide, and Raft Creeks, and in other smaller valley locations where bedrock is exposed at a cliff face (These locations correspond to the sudden break in slope where plateau surfaces are dissected by valleys). Since vegetation at, and immediately on top of the cliff faces is scarce, the breaking of rock from the face is primarily a result of frost weathering (expansion and contraction of joints).

Rates of cliff recession in the study area are not well known, but most certainly are related to bedrock structure and lithology. In the upper valley sections of Talbot, Rockslide and Alaskite Creeks, where the dominant bedrock type is alaskite, the talus contributes to slopes that are composed primarily of smaller rock fragments and a "popcorn-like" granule which is characteristic of alaskite when it weathers; a detailed description of alaskite and granodiorite is presented in a following section. The debris constituting the slopes in these regions appear to be supplied on a somewhat continuous basis. Specifically, there seems to be a regular supply of cliff face material which is likely related to the relatively easy weathering of alaskite.

In areas dominated by granodiorite, the slopes are comprised of fractured boulders, rocks, and relatively fewer finer materials. In some instances, extremely large boulders measuring 4 m have broken from the adjoining cliff faces. In contrast to the alaskite bedrock regions, the granodiorite cliff face areas have been observed to be quite catastrophic in the manner in which debris is supplied.

Rates of talus derived debris movement are difficult to measure, however movement is discontinuous and is commonly regarded as occurring by subsidence, creep, rolling or gliding of rock(s), small talus slides, or a combination of these. Talus movements in the study area were not determined, however Rampton and Dugal (1974) have reported talus movements of about 2.5 cm/yr in the northern Yukon. Most of the movement is said to take place within about a half metre of the surface. In some of the smaller valleys, talus derived debris has crossed the valley floor and abutted against the opposite valley side. In the larger east-west oriented valleys, talus debris is generally restricted close to valley sides. However, there are some instances where the debris, perhaps aided by spring and early summer melt, has washed from the valley side and extended past the midpoint of the valley floor. Such is the case in Rockslide Creek where the talus debris has impeded local drainage in the valley, redirecting flow around the debris.

Evident in areas of currently active and fossil talus locations, is the inundation of lateral moraines, especially in the upper sections of Rockslide, Talbot, and Alaskite Creeks. In the upper valley extremity of Rockslide Creek, a series of talus derived cones, extending approximately 1 km down valley, have completely inundated a lateral moraine which is traceable on either sides of the forms. Although no age determination of the talus derived cones was undertaken (and non reported in the literature for the area), most of the fossil cones in the study area are thought to have developed in the Holocene with some possibly active during the Ruby Ice Sheet and St. Elias Advance in late Wisconsin.

There are a number of rock glaciers located in the study area, most are relatively small and occupy either valley sides or the side walls of cirques. The largest concentration of rock glaciers is in the upper valley sections of Talbot, Rockslide and Alaskite Creeks. Vegetation on the rock glaciers is generally restricted to mosses, grasses and sedges. Vegetation at the base of the structures commonly include a variety of flowers and small willows (Campbell, 1985). Except for a few isolated signs of disturbances, the rock glaciers show no evidence of recent major movements. Although the exact age of the rock glaciers in the northern Ruby Range is unknown, they may span the period from the late Wisconsin to Holocene and may be contemporaneous with fossil talus derived forms.

Muller (1967) stated that except for the disappearance of ice, the morphology of the Ruby Range has changed little since Pleistocene time. However, the widespread occurrence of talus activity (currently active and fossil), and to a lesser extent, rock glaciers, indicates that mass-wasting processes have been fairly active since at least late Wisconsin time. Although only speculations can be made with respect to the formation of rock glaciers, it appears that they evolved as a result of debris supplied from adjoining cliff faces onto late Wisconsin ice or perhaps permanent Holocene snow patches; a scenario which is supported by the widespread occurrence of fossil talus derived cones.

5.4 Carbonate Determination

Roof pendants, partly of carbonate metamorphic rocks (primarily limestone), in various stages of migmatization, occur in the northern Ruby Range study area just north of Raft Creek, along the Talbot Arm. The carbonate rocks commonly occur as lenses in the granitic rocks. Although their extents are rather restricted (approximately 10 km^2), they have been mapped by Muller (1967) as occupying plateau regions at elevations ranging from about 1600 to 2100 masl. The carbonate rocks become more abundant west of the Talbot Arm, and are the dominate bedrock type of the Ruby Range west of the Brooks Valley. As well, occurring both to the north (Nisling Range), and south of the study area (south of Gladstone Creek) are two belts of quartz-biotite schist and gneiss, with some minor limestone (see Table 5, Formations of the Yukon Plateau). The most extensive occurrences of carbonate rocks are located west and south of Kluane Lake in the Kluane Ranges, Duke Depression and other regions of the St. Elias Mountains (see Table 4, Formations of the St. Elias Mountains).

Results from the carbonate analysis on the greater than 2 mm size fraction indicated that no carbonate material was present in any of the samples (plateaus or valleys). However, the presence or absence of carbonate material as a criterion for the definition of glaciation or non-glaciation presents a number of difficulties. As with the problems associated with the use of erratic materials as a criterion for determining whether plateau regions experienced glaciation, there were problems related to the investigation of vast regions, which resulted in a sampling design being based, in part, on the logistical realities in the field. This resulted in some regions not being sampled because of the relative inaccessibility. As well, the extent of weathering and periglacial activity may have altered or destroyed much of the carbonate material, if in fact it was present in the sediment. Furthermore, the carbonate material may have been deposited in such a way or form that has not yet been identified. Therefore, the results of this analysis are viewed as somewhat inconclusive, as carbonate material in the sediment samples taken in this study from the northern Ruby Range either; i) has not yet been found, or ii) has been altered or destroyed. However, it is worth noting that Dewez (1988) found carbonate in the Gladstone Lakes region.

5.5 Bedrock Compositional Analysis

The general geology of the Ruby Range has been mapped by Muller (1967) and is summarized on Figure 11. In order to gain a more detailed understanding of local bedrock composition, thin sections of alaskite and granodiorite core samples were prepared following the standard procedures set forth by Dickson (1966).

5.5.1 Granodiorite

The most widespread granitic rock in the study area is granodiorite. Table 15 presents the point count of constituent minerals in three separate granodiorite thin sections. Granodiorite is a medium to coarse grain rock and exhibits a dark colour. This colour is partly due to the amount of feldspar in the rock. Plagioclase is generally light green and orthoclase dark grey or reddish. Their abundance in thin section is between 15% to 20%. Hexagonal biotite flakes, up to about 5 mm in diameter are very common. Some thin sections are visibly heterogeneous and contain megascopic green patches of hornblende, biotite, chlorite and plagioclase within a matrix of pinkish plagioclase, orthoclase, quartz and minor hornblende and biotite.

Under microscopic examination, subhedral plagioclase from 2 mm to 8 mm in length are observed. It is commonly semicitized, the indicated composition being near 50% anorthite. Hornblende in many places with augite nuclei, occurs in clusters with biotite, chlorite, magnetite and some minor apatite. Anhedral orthoclase and quartz, in places as granophyre, fill the spaces between plagioclase and opaque minerals. Embayments of quartz and to a lesser extent orthoclase, penetrate plagioclase or appear to have replaced that mineral entirely, leaving odd shaped fragments of plagioclase in a surrounding of quartz.

5.5.2 Alaskite

Alaskite was subdivided on the basis of initial field observations into a relative classification based on visual assessment of grain size which included fine, medium and coarse. This was accomplished on the basis of relative mineral sizes. Table 15 presents the results of the microscopic point count on the alaskite.

Figure 11
GENERAL GEOLOGY OF THE NORTHERN RUBY RANGE

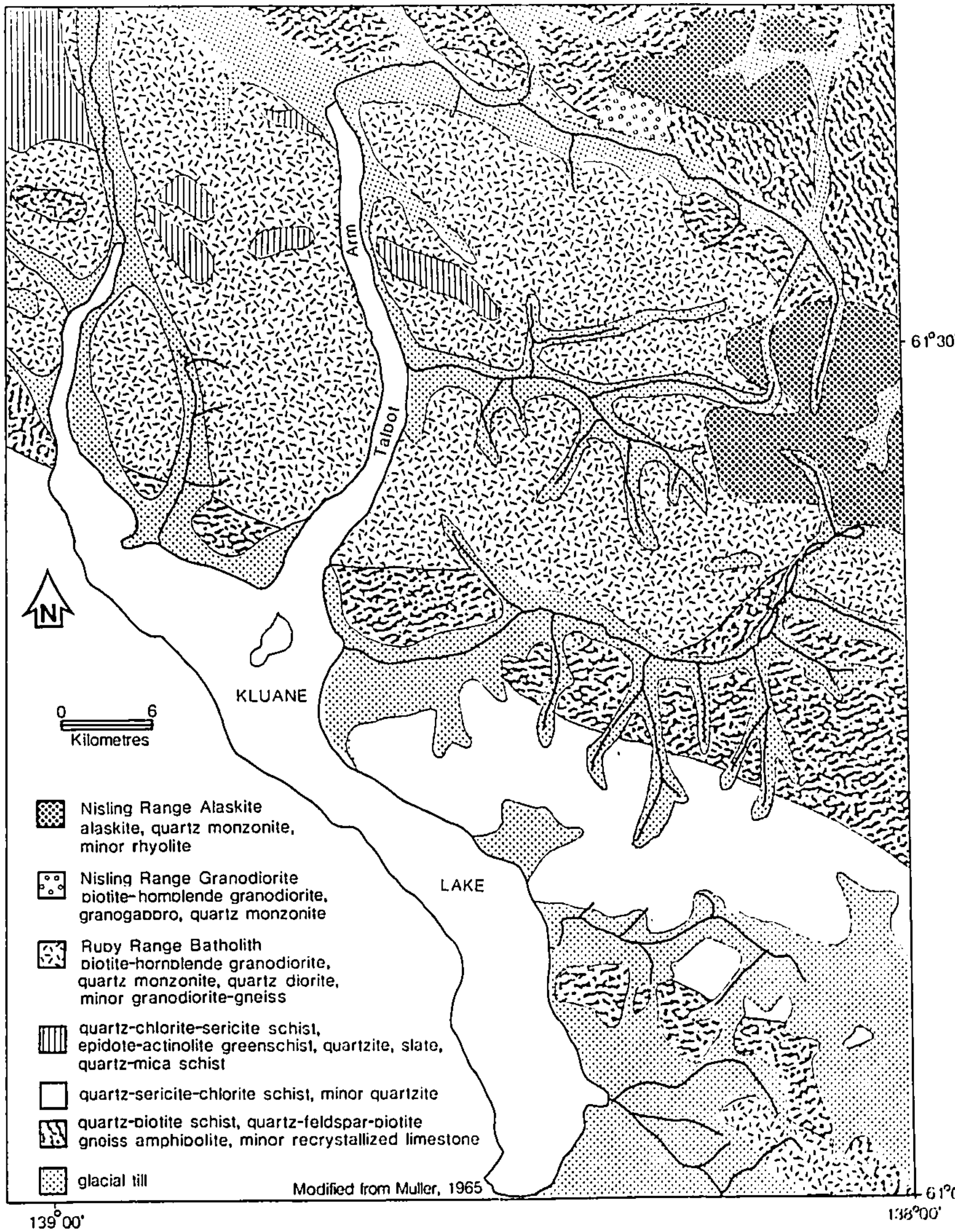


Table 15
GRANODIORITE AND ALASKITE COMPOSITIONAL ANALYSIS

	Granodiorite Section A	Granodiorite Section B	Granodiorite Section C	Granodiorite Average	Granodiorite Section 1	Granodiorite Section 2	Fine Grain Alaskite	Medium Grain Alaskite	Coarse Grain Alaskite	Alaskite Section 1	Alaskite Section 2
Quartz	20%	20%	30%	25%	23.2%	26.5%	50%	50%	15%	25.3%	24.2%
K-feldspar	10%	15%	15%	15%	19.4%	10.2%	40%	40%	70%	64.8%	65.0%
Plagioclase	50%	45%	40%	45%	49.1%	49.5%	10%	<5%	5%	9.0%	10.1%
Biotite	10%	10%	5%	10%	6.2%	13.3%	<5%	<5%	5%	0.3%	0.5%
Chlorite	-	-	-	-	-	-	-	-	<5%	-	-
Hornblende	<5%	>5%	>5%	<5%	1.9%	0.2%	-	-	-	0.3%	-
Pyroxene	<5%	>5%	-	<5%	-	-	-	-	-	-	-
Others	-	-	-	-	0.2%	0.3%	<5%	5%	<5%	0.3%	0.2%

- The mineralogical composition of Granodiorite (Sections A, B and C), and Alaskite (Fine, Medium and Coarse Grain) was determined on the basis of a 200 point count per slide. Results are rounded to the nearest 5% level.
- The mineralogical composition of Granodiorite and Alaskite (Section #'s 1 and 2), outlined in bold, are taken from Muller (1967), and are located in regions bordering the northern Ruby Range.
- Pyroxene primarily includes diopside with some very minor augite. Orthopyroxene is not included in this group.
- "Others" denotes altered grains beyond recognition and/or unknowns.

The fine grain alaskite exhibits a varied grain size. Large composite quartz phenocrysts containing crystals up to 2 mm in diameter are contained in a matrix of weathered feldspars, quartz and some minor biotite averaging approximately 0.3 mm in diameter. Quartz exhibits large variation in crystal size. Quartz occur primarily as fractured anhedral crystals showing some alterations along fractures.

K-feldspar is highly altered and predominantly in the matrix of the rock as small anhedral grains approximately 0.3 mm in diameter. Large K-feldspar crystals are also present exhibiting perthitic texture. Plagioclase feldspar occurs as elongate subhedral highly altered grains showing twinning and some zoning. All specimens of fine grain alaskite exhibit a reddish-brown staining which has been attributed to secondary hematite. The hematite staining is considered to be a diagnostic characteristic of the fine grain alaskite.

Medium grain alaskite consists of crystalline quartz and K-feldspar measuring approximately 2 mm to 3 mm in diameter with a finer crystalline matrix of quartz and K-feldspar averaging 0.3 mm to 0.5 mm in diameter. This section appears to be more altered than the fine grain alaskite. Quartz generally occurs as fractured anhedral crystals although several orthorhombic quartz euhedra (high temperature) are present. Boundaries of these crystals are somewhat diffuse due to alteration by K-feldspar.

K-feldspar is highly altered and occurs as medium subhedral tabular crystals (some exhibiting perthitic texture) and smaller altered anhedral in the matrix. Plagioclase feldspar occurs as elongate subhedral highly altered grains showing twinning and some zoning. Biotite is more common in the medium thin section (5%). It tends to occur as elongate to tabular crystals with ragged edges measuring approximately 1 mm in length.

The coarse grain alaskite exhibits, relative to the other sections, less alteration of feldspars. The section is dominated by very large elongate K-feldspar crystals measuring approximately 1.5 cm in length, in a matrix of primarily quartz and plagioclase averaging 1.5 mm in diameter. Perithic texture is observed in the large K-feldspars.

Biotite occurs as elongate to tabular crystals measuring approximately 1 mm in length. The abundance of biotite in this section is greater than that observed in the fine and medium grain alaskites. Chlorite is present as ragged, tabular crystals, 0.8 mm to 1.0 mm in length, and tends to be associated with biotite. The presence of chlorite is unique to this section and can be described as a diagnostic feature of coarse grain alaskite. The high proportion of K-feldspars is due to the large crystal size of the mineral which tends to dominate the mineralogical appearance of this section.

5.5.3 Summary

For comparative purposes, Table 15 shows modes of some granitic rocks of the Ruby Range batholith and Nisling Range reported by Muller (1967) in neighbouring regions. The average granodiorite composition of the three sections examined in this study show approximately 25% quartz, 15% K-feldspar, 45% plagioclase, 10% biotite, <5% hornblende, and <5% pyroxene. The average of Muller's (1967) two granodiorite sections are quite similar with about 25% quartz, 15% K-feldspar, 50% plagioclase, 10% biotite, 1% hornblende, and 0.2% others.

The composition of fine and medium grain alaskites appears to be relatively similar except for the amount of plagioclase; 10% in fine grain, <5% in medium grain. Coarse grain alaskite exhibits a relatively low amount of quartz (15%) compared to the fine and medium grain alaskites (50%), and a higher percentage of K-feldspar; 70% for coarse grain, 40% for fine and medium grain. Biotite is also more abundant in the coarse grain alaskite (5%). Chlorite is found in the coarse grain alaskite but not in the fine and medium grain alaskites. Comparatively, Muller's (1967) analysis of the two alaskite sections are similar to each other, but differ from the three sections examined in this study. This is interpreted as indicating that although alaskite exhibits a somewhat uniform mineralogical composition, the relative amounts of some minerals vary significantly.

Of particular interest to this study are the occurrences of heavy minerals in the local bedrock. The compositional analysis of granodiorite and alaskite indicates that locally derived heavy minerals include hornblende, pyroxene (diopside and augite), and minerals under the "other" category which may consist of epidote, zircon and pyrite. As well, although not identified by Muller as occurring in samples from the immediate study area, the occurrence of rutile and garnet in other areas of the Ruby Range batholith without doubt indicate similar occurrences in the study area.

Regional geology of the Kluane Lake, Aishihik Lake and Dezadeash map areas were mapped by Muller (1967), Tempelman-Kluit (1974), and Kindle (1953) respectively, with supplementary descriptions made by Campbell and Dodds (1978), Gabrielse et al., (1980), Dodds (1982), Dewez (1988), and Hughes (1990). However, the relative large scale of much of the regional mapping has resulted in a highly generalized description of bedrock variations and composition, especially in the study area. For example, out of sixty-seven bedrock sites sampled and described by Muller (1967), twenty-three were taken from the Ruby Range batholith, and of these, only two were from the study area.

Therefore, detailed analysis of variation in bedrock composition and mineralization on a local scale (i.e., within the study area), are not well known, and it is possible that mineral occurrences, such as the tungsten, molybdenum, copper, and fluorite reported from near the head of Alaskite Creek, and fluorite from the head of Rockslide Creek (Muller, 1967) occur elsewhere. These concerns naturally invoke consideration of heavy mineral compositional modes in bordering regions, which will be examined in the following section.

5.6 Fine Sand Fraction: Heavy Mineral Composition, Assemblage, Provenance and Dispersal Trends

5.6.1 Introduction

The composition of the surficial sediment was analysed to identify and count mineral species, determine sources (bedrock and others), and to assess the nature of sediment transport. In the northern Ruby Range study area, and surrounding regions, the geology has been mapped at a scale which adequately portrays bedrock patterns on a regional basis, but generally lacks the detailed lithological descriptions necessary to determine mineral associations. This is particularly true of heavy mineral compositions. However, within the bounds imposed by the available geological descriptions, coupled with heavy mineral suites common to rock types encountered in the study area and bordering regions (Pettijohn, 1957; Hubert, 1971; Paré, 1982), as well as bedrock analysis carried out in this study, some broad goupings of certain mineral associations have been recognized in the sediment that may be related to bedrock sources, or other origins such as volcanic ash and meteorites.

Henderson (1989) stated that the degree to which sediment composition reflects the bedrock source from which it was derived depends on the erosion process(es) and the extent of sediment modification during transport, deposition, and diagenesis. For example, in glaciated terranes sediments are generally composites of the bedrock traversed during glacier flow. In fluvial or marine environments, sediments are generally sorted aggregates derived from previously deposited debris. Sediments derived in situ will generally reflect the underlying bedrock characteristics. Through erosion and comminution, material is distributed within the sediment in various size fractions, which is primarily dependent upon the nature of the bedrock source. The relationship between sediment texture and composition has been reviewed in numerous studies (e.g., Dreimanis and Vagners, 1971; Shilts, 1973). Of particular interest in this study are characteristics common to igneous and metamorphic rocks. Haldorson (1977) determined that these rocks produce a matrix mode in the very fine to fine sand range, which also corresponds to the heavy mineral fraction, and forms the basis for the ensuing discussions.

5.6.2 Heavy Mineral Composition

Heavy mineral identification and counts were perfomed on a total of 148 mineral grain slides. The data derived from this analysis was arranged on the basis of sample location, ie., plateau data and valley data (Table 16). Table 16 shows that of 97 plateau sediment samples collected, 7 were insufficient in fine-sand fraction to extract heavy minerals, and 7 others were inadvertently destroyed during slide preparation, leaving 83 plateau mineral grain slides for examination. Likewise for the valley samples, 4 were insufficient in the fine-sand fraction to extract heavy minerals, and 13 were accidentally destroyed, leaving 65 slides to examine.

The average proportion (% by weight) of heavy minerals comprising the fine-sand fraction for plateau regions is 1.68%, and 1.93% for valleys. The maximum and minimum percentiles for these proportions are 12.35 to 0.24% for plateau regions, and 31.93 to 0.03% for valley locations. Although no definitive statement can be made with respect to differences between plateau and valley heavy mineral proportions in the fine-sand fraction, the data indicates an overall similar average heavy mineral content in both plateau and valley samples, but a more variable range in this content in valley samples.

Heavy mineral identification was performed on the basis of mineral familiarity and an appreciation of the reported lithology of the study area and bordering regions (see section on General Geology). However, by no means does this suggest that other species of heavy minerals are not present in the sediment samples, but rather, if they were encountered, and identification was not possible due to mineral alterations or unfamiliarity, they were included in the "other" category. Based upon initial identification of the "other" minerals, along with collaborating opinion (Paré, per. comm.), minerals comprising part of this category <u>may</u> include diopside, staurolite, anatase, monozite, siderite, ilmenite, and kyanite. Due to the uncertainty with respect to identification of these minerals, comments pertaining to provenance would be highly speculative, and therefore does not constitute part of the discussion.

In the vast majority of slides, 200 mineral grains were identified. Six individual heavy minerals were identified along with groupings of minerals classified as garnet, orthopyroxene, and opaque. The individual minerals include: i) four silicates, epidote, allanite, hornblende, zircon; ii) one hydroxide, goethite; and iii) one oxide, hematite. The mineral groupings include: i) garnet (silicates); pyrope, almandine, spessartine, grossular, andradite, and uvarivite; ii) orthopyroxene (silicates); enstatite, and hypersthene; and iii) opaques; magnetite (oxide), pyrite, molybdenite, chalcopyrite, pyrrholite, and bornite (sulphides) (both hematite and goethite are opaque minerals but were identified individually). Detailed descriptions of these minerals are

Table 16
HEAVY MINERAL IDENTIFICATION AND POINT COUNT
(PLATEAU SAMPLE SITES)

SAMPLE	FINE SAND FRACTION (63 - 250 U) TOTAL	LIGHT MINERAL	HEAVY MINERAL	HEAVY MINERAL	HEAVY MINERAL COUNT: ORTHO-PYROXENE		GARNET		EPIDOTE		ALLANITE		HEMATITE		GOETHITE		HORNBLENDE		ZIRCON		OPAQUES		OTHERS		TOTAL
(#)	(Wt.)	(Wt.)	(Wt.)	(% Total)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)
82-01P	62.15	61.75	0.40	0.64	8	4.0	0	0	25	12.5	12	6.0	0	0	14	7.0	104	52.0	8	4.0	26	13.0	3	1.5	200
82-03P	38.85	38.00	0.85	2.18	11	5.3	4	1.9	4	1.9	0	0	11	5.3	0	0	158	75.6	0	0	21	10.0	0	0	209
82-04P	Insufficient Sample																								
82-05P	63.72	63.17	0.55	0.86	13	6.5	0	0	4	2.0	0	0	3	1.5	0	0	171	85.5	1	0.5	7	3.5	1	0.5	200
82-06P	Insufficient Sample																								
82-07P	53.75	53.42	0.33	0.61	21	10.5	0	0	52	26.0	25	12.5	5	2.5	0	0	41	20.5	2	1.0	39	19.5	15	7.5	200
82-12P	27.03	26.54	0.49	1.81	5	1.7	13	4.3	92	30.7	19	6.3	16	5.3	3	1.0	65	21.7	6	2.0	2	0.7	34	11.3	300
82-13P	65.66	64.60	1.06	1.90	88	29.3	0	0	55	18.3	22	7.3	15	5.0	14	4.7	58	19.3	3	1.0	13	4.3	32	10.7	300
82-14P	Insufficient Sample																								
82-15P	40.86	39.65	1.21	2.96	4	1.7	0	0	29	12.6	65	28.3	8	3.5	6	2.6	25	10.9	3	1.3	44	19.1	1	0.4	230
82-16P	22.28	21.88	0.40	1.80	31	12.4	2	0.8	12	4.8	1	0.4	7	2.8	7	2.8	43	17.2	5	2.0	14	5.6	2	0.8	250
82-17P	60.18	59.30	0.88	1.46	56	18.7	2	0.7	24	8.0	5	1.7	9	3.0	84	28.0	64	21.3	7	2.3	2	0.7	29	9.7	300
82-18P	57.75	57.13	0.62	1.07	85	28.3	3	1.0	7	2.3	22	7.3	4	1.3	52	17.3	98	32.7	9	3.0	3	1.0	17	5.7	300
82-21P	44.25	43.68	0.57	1.29	36	12.0	5	1.7	21	7.0	28	9.3	0	0	26	8.7	161	53.7	9	3.0	13	4.3	1	0.3	300
82-23P	42.61	41.71	0.90	2.11	19	9.5	6	3.0	21	10.5	5	2.5	5	2.5	15	7.5	64	32.0	4	2.0	33	16.5	28	14.0	200
82-32P	35.96	31.52	4.44	12.35	28	9.0	0	0	13	4.3	0	0	2	0.7	6	2.0	237	79.0	7	2.3	0	0	7	2.3	300
82-33P	66.23	64.63	1.60	2.42																					
82-36P	51.17	49.60	1.57	3.07	21	7.0	0	0	0	0	0	0	6	2.0	225	75.0	28	9.3	0	0	2	0.7	0	0	300
82-37P	59.71	58.76	0.95	1.59																					
82-38P	Insufficient Sample																								
82-40P	Insufficient Sample																								
82-42P	55.06	54.06	1.00	1.82																					
82-43P	40.47	4.00	0.47	1.16	17	8.5	5	2.5	22	11.0	9	4.5	6	3.0	2	1.0	107	53.5	6	3.0	13	6.5	13	6.5	200
82-44P	61.74	61.24	0.50	0.81	46	23.0	6	3.0	24	12.0	3	15.0	2	1.0	11	5.5	31	15.5	12	6.0	35	17.5	3	1.5	200
82-46P	63.56	63.26	0.30	0.47	111	55.5	8	4.0	5	2.5	2	1.0	5	2.5	11	5.5	17	8.5	4	2.0	34	17.0	3	1.5	200
82-50P	42.70	42.24	0.46	1.08	1	5.0	13	6.5	36	18.0	12	6.0	5	2.5	5	2.5	29	14.5	15	7.5	62	31.0	13	6.5	200
82-55P	17.02	16.70	0.32	1.88	23	11.5	13	6.5	17	8.5	26	13.0	8	4.0	14	7.0	44	22.0	13	6.5	16	8.0	26	13.0	200
82-57P	61.05	60.59	0.46	0.76	2	1.0	2	1.0	26	13.0	25	12.5	4	2.0	17	8.5	53	26.5	2	1.0	11	5.5	22	11.0	200
82-60P	71.16	65.66	5.50	7.73	4	2.0	0	0	0	0	0	0	2	1.0	2	1.0	59	29.5	0	0	94	47.0	3	1.5	200
82-62P	19.01	17.77	1.24	6.52	12	6.0	2	1.0	17	8.5	5	2.5	0	0	1	5.0	131	65.5	4	2.0	8	4.0	11	5.5	200
82-63P	62.13	58.73	3.40	5.47	29	14.5	1	0.5	15	7.5	1	5.0	6	3.0	0	0	42	21.0	2	1.0	85	42.5	1	5.0	200
82-65P	65.24	61.24	4.00	6.13	1	5.0	0	0	2	1.0	6	2.5	0	0	1	0.5	179	89.5	0	0	2	1.0	1	0.5	200
82-66P	52.30	51.64	0.66	1.26	1	5.0	1	0.5	1	0.5	1	0.5	0	0	146	73.0	5	2.5	0	0	18	9.0	18	9.0	200
82-67P	50.76	50.57	0.19	0.37	15	7.5	9	4.5	7	3.5	7	3.5	0	0	85	42.5	3	15.0	9	4.5	36	18.0	2	1.0	200
82-69P	52.38	52.05	0.33	0.63	26	13.0	6	3.0	27	13.5	5	2.5	0	0	23	11.5	7	35.0	7	3.5	24	12.0	12	6.0	200
82-70P	73.72	73.18	0.54	0.73	12	6.0	0	0	36	18.0	15	7.5	0	0	1	0.5	111	55.5	2	1.0	21	10.5	2	1.0	200
82-72P	49.96	49.71	0.25	0.60	4	2.0	0	0	36	18.0	13	6.5	5	2.5	16	8.0	53	26.5	13	6.5	15	7.5	9	4.5	200
82-73P	46.75	46.42	0.33	0.71	24	12.0	6	3.0	29	14.5	4	2.0	0	0	12	6.0	6	3.0	13	6.5	52	26.0	0	0	200
82-74P	76.32	75.39	0.93	1.22	1	5.0	9	4.5	15	7.5	5	2.5	0	0	6	3.0	114	57.0	0	0	38	19.0	3	1.5	200
82-75P	40.95	39.89	1.06	2.59																					
82-77P	47.38	46.90	0.48	1.01																					
82-84P	60.31	59.31	1.00	1.66																					
82-85P	50.75	48.36	2.39	4.71	37	18.5	1	0.5	1	5.0	6	3.0	0	0	4	2.0	105	52.5	3	1.5	28	14.0	6	3.0	200
82-86P	Insufficient Sample																								
82-88P	68.00	66.57	1.43	2.10	14	7.0	5	2.5	21	10.5	13	6.5	0	0	0	0	82	41.0	5	2.5	58	29.0	2	1.0	200
82-90P	68.00	65.70	2.30	3.38	74	37.0	2	1.0	9	4.5	0	0	0	0	4	2.0	68	34.0	0	0	43	21.5	0	0	200
82-91P	47.07	46.23	0.84	1.78																					
83-01P	69.33	68.70	0.63	0.91	28	14.0	0	0	0	0	34	17.0	6	3.0	22	11.0	24	12.0	28	14.0	46	23.0	12	6.0	200
83-02P	88.71	87.51	1.20	1.35	18	9.0	0	0	41	20.5	4	2.0	4	2.0	4	2.0	10	5.0	11	5.5	14	7.0	4	2.0	200

Table 16
(PLATEAU SAMPLE SITES)
(continued)

SAMPLE	FINE SAND FRACTION (63 - 250 U) TOTAL	LIGHT MINERAL	HEAVY MINERAL	HEAVY MINERAL	HEAVY MINERAL COUNT ORTHO-PYROXENE		GARNET		EPIDOTE		ALLANITE		HEMATITE		GOETHITE		HORNBLENDE		ZIRCON		OPAQUES		OTHERS		TOTAL
(#)	(Wt.)	(Wt.)	(Wt.)	(% Total)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)
83-03P	143.50	142.70	0.80	0.56	2	1.0	0	0	8	4.0	6	3.0	0	0	12	6.0	8	4.0	8	4.0	145	72.5	11	5.5	200
83-05P	75.59	75.15	0.44	0.58	28	14.0	5	2.5	24	12.0	8	4.0	0	0	16	8.0	18	9.0	12	6.0	61	30.5	28	14.0	200
83-07P	157.70	150.70	7.00	4.44	11	5.5	0	0	1	0.5	2	1.0	0	0	0	0	2	1.0	2	1.0	181	90.5	1	0.5	200
83-08P	476.80	463.33	13.47	2.83	0	0	0	0	0	0	0	0	0	0	0	0	56	28.0	0	0	144	72.0	0	0	200
83-09P	106.01	104.15	1.86	1.75	12	6.0	0	0	4	2.0	0	0	0	0	12	6.0	14	7.0	2	1.0	152	76.0	4	2.0	200
83-26P	173.50	172.50	1.00	0.58	0	0	0	0	11	5.5	15	7.5	0	0	6	3.0	136	68.0	0	0	32	16.0	0	0	200
83-27P	205.78	204.45	1.33	0.65	9	4.5	0	0	17	8.5	28	14.0	0	0	2	1.0	64	32.0	21	10.5	59	29.5	0	0	200
83-30P	118.28	117.70	0.58	0.49	3	1.5	0	0	4	2.0	2	1.0	0	0	0	0	9	4.5	0	0	182	91.0	0	0	200
83-31P	156.08	155.60	0.48	0.31	9	4.5	0	0	1	0.5	0	0	8	4.0	0	0	11	5.5	0	0	168	84.0	3	1.5	200
83-32P	94.42	94.00	0.42	0.44	0	0	0	0	0	0	0	0	0	0	1	5.0	8	4.0	1	0.5	179	89.5	2	1.0	200
83-33P	90.65	90.10	0.55	0.61	0	0	0	0	6	3.0	17	8.5	4	2.0	0	0	4	2.0	1	0.5	167	83.5	1	0.5	200
83-34P	120.97	120.30	0.67	0.55	3	1.5	0	0	6	3.0	0	0	0	0	18	9.0	12	6.0	0	0	153	76.5	8	4.0	200
83-37P	74.36	74.00	0.36	0.48	3	1.5	0	0	5	2.5	6	3.0	0	0	8	4.0	0	0	0	0	168	84.0	1	5.0	200
83-38P	97.10	96.60	0.50	0.51	5	2.5	0	0	3	1.5	14	7.0	9	4.5	22	11.0	4	2.0	1	0.5	136	68.0	6	3.0	200
83-40P	137.95	137.20	0.75	0.54	4	2.0	0	0	14	7.0	19	9.5	0	0	12	6.0	8	4.0	8	4.0	125	62.5	1	5.0	200
83-42P	86.54	85.80	0.74	0.86	0	0	0	0	0	0	6	3.0	0	0	1	0.5	11	5.5	0	0	182	91.0	0	0	200
83-43P	205.72	204.56	1.16	0.56	1	0.5	0	0	3	1.5	0	0	0	0	4	2.0	9	4.5	7	3.5	174	87.0	2	1.0	200
83-46P	182.26	181.50	0.76	0.42	0	0	5	2.5	6	3.0	18	9.0	0	0	9	4.5	9	4.5	1	5.0	125	62.5	18	9.0	200
83-55P	213.70	211.24	2.46	1.15	3	1.5	0	0	0	0	0	0	0	0	5	2.5	8	4.0	0	0	184	92.0	0	0	200
83-56P	180.66	180.20	0.46	0.25	0	0	0	0	12	6.0	16	8.0	0	0	5	2.5	5	2.5	2	1.0	154	77.0	6	3.0	200
83-60P	161.86	161.36	0.50	0.31	3	1.5	0	0	0	0	0	0	0	0	0	0	13	6.5	0	0	184	92.0	0	0	200
83-61P	177.88	176.28	1.60	0.90	2	1.0	0	0	0	0	0	0	0	0	0	0	3	1.5	5	2.5	184	92.0	6	3.0	200
83-62P	127.90	124.90	3.00	2.35	0	0	0	0	1	0.5	0	0	0	0	0	0	6	3.0	0	0	192	96.0	1	0.5	200
83-64P	158.50	156.40	2.10	1.32	0	0	0	0	5	2.5	0	0	0	0	0	0	14	7.0	2	1.0	178	89.0	1	0.5	200
83-70P	120.85	120.50	0.35	0.29	1	0.5	0	0	0	0	0	0	0	0	1	0.5	6	3.0	0	0	19	95.0	2	1.0	200
83-71P	78.84	78.47	0.37	0.47	24	12.0	0	0	13	6.5	16	8.0	0	0	0	0	81	40.5	6	3.0	6	3.0	0	0	200
83-72P	134.30	133.70	0.60	0.45	6	3.0	0	0	41	20.5	4	2.0	0	0	5	2.5	46	23.0	3	15.0	63	31.5	5	2.5	200
83-73P	801.04	796.11	4.93	0.62	35	17.5	0	0	23	11.5	8	4.0	0	0	6	3.0	24	12.0	8	4.0	95	47.5	1	0.5	200
83-75P	168.10	157.30	10.8	6.42	0	0	0	0	0	0	0	0	0	0	0	0	193	96.5	3	1.5	4	2.0	0	0	200
83-76P	143.57	139.57	4.00	2.79	2	1.0	0	0	32	16.0	11	6.5	0	0	4	2.0	88	44.0	61	25.5	11	5.5	1	0.5	200
83-77P	101.30	100.88	0.42	0.41	0	0	0	0	24	12.0	33	16.5	4	2.0	8	4.0	58	29.0	17	8.5	41	20.5	15	7.5	200
83-78P	110.35	109.85	0.50	0.45	22	11.0	0	0	36	18.0	3	1.5	4	2.0	6	3.0	1	5.0	14	7.0	99	49.5	6	3.0	200
83-79P	164.80	164.40	0.40	0.24	19	9.5	0	0	13	6.5	6	3.0	0	0	23	11.5	15	7.5	2	1.0	88	44.0	16	8.0	200
83-81P	180.55	175.25	5.30	2.94	2	1.0	0	0	7	3.5	2	1.0	0	0	0	0	159	79.5	5	2.5	6	3.0	1	0.5	200
83-82P	160.55	159.55	1.00	0.62	4	2.0	0	0	7	3.5	2	1.0	0	0	0	0	59	29.5	8	4.0	116	58.0	4	2.0	200
83-84P	77.32	76.72	0.60	0.78	6	3.0	0	0	2	1.0	0	0	2	1.0	2	1.0	9	4.5	0	0	179	89.5	0	0	200
83-85P	77.74	77.00	0.74	0.95	5	2.5	0	0	0	0	11	5.5	3	1.5	4	2.0	36	18.0	7	3.5	134	67.0	0	0	200
84-15P	275.34	272.64	2.70	0.98	2	1.0	0	0	59	29.5	0	0	0	0	5	2.5	0	0	2	1.0	132	66.0	0	0	200
84-16P	137.65	135.65	2.00	1.45	16	8.0	2	1.0	37	18.5	4	2.0	0	0	0	0	7	3.5	0	0	134	67.0	0	0	200
84-17P	117.77	115.77	2.00	1.70	14	7.0	0	0	29	14.5	2	1.0	0	0	0	0	4	2.0	1	0.5	144	72.0	6	3.0	200
84-19P	131.25	129.80	1.45	1.10	22	11.0	0	0	67	33.5	8	4.0	0	0	0	0	6	3.0	1	0.5	91	45.5	5	2.5	200
84-20P	126.25	124.25	2.00	1.58	2	1.0	2	1.0	26	13.0	1	0.5	0	0	0	0	12	6.0	1	0.5	137	68.5	1	0.5	200
85-06P	165.74	163.20	2.54	1.53	18	9.0	0	0	4	2.0	1	0.5	0	0	2	1.0	5	2.5	1	0.5	169	84.5	0	0	200
85-07P	Insufficient Sample																								
85-08P	76.81	75.28	1.53	1.99	1	5.0	0	0	7	3.5	8	4.0	0	0	2	1.0	57	28.5	3	1.5	113	56.5	0	0	200
85-09P	165.23	163.75	1.48	0.90	15	7.5	0	0	9	4.5	1	0.5	0	0	101	50.5	1	0.5	1	0.5	24	12.0	48	24.0	200
85-10P	94.00	92.40	1.60	1.70	19	9.5	1	0.5	7	3.5	0	0	0	0	0	0	5	2.5	0	0	168	84.0	0	0	200
85-11P	100.49	99.20	1.29	1.28	7	3.5	7	3.5	32	16.0	9	4.5	0	0	9	4.5	8	4.0	0	0	124	62.0	4	2.0	200

Table 16
(VALLEY SAMPLE SITES)
(continued)

SAMPLE	FINE SAND FRACTION (63 - 250 U) TOTAL	LIGHT MINERAL	HEAVY MINERAL	HEAVY MINERAL	HEAVY MINERAL COUNT ORTHO-PYROXENE		GARNET		EPIDOTE		ALLANITE		HEMATITE		GOETHITE		HORNBLENDE		ZIRCON		OPACUES		OTHERS		TOTAL
(#)	(Wt.)	(Wt.)	(Wt.)	(% Total)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)
82-08V	26.09	23.89	1.20	4.78	1	0.5	0	0	8	4.0	1	5.0	0	0	3	1.5	157	78.5	8	4.0	5	2.5	8	4.0	200
82-09V	47.19	45.55	1.64	3.48	11	5.5	1	0.5	93	46.5	6	3.0	0	0	17	8.5	36	18.0	12	6.0	15	7.5	9	4.5	200
82-10V	35.91	35.16	0.75	2.09	14	7.0	0	0	47	23.5	6	3.0	1	0.5	12	6.0	101	50.5	1	0.5	12	6.0	6	3.0	200
82-11V	56.20	55.20	1.00	1.78	4	2.0	0	0	36	18.0	4	2.0	2	1.0	5	2.5	123	61.5	8	4.0	2	1.0	16	8.0	200
82-24V	43.64	43.54	0.10	0.23	37	30.3	6	4.9	8	6.6	5	4.1	1	0.8	36	29.5	8	6.6	2	1.6	5	4.2	14	11.5	122
82-25V	Insufficient Sample																								
82-26V	Insufficient Sample																								
82-27V	32.27	32.00	0.27	0.84																					
82-28V	59.83	59.16	0.67	1.12	51	17.0	4	1.3	55	18.3	5	1.7	2	0.7	46	15.3	32	10.7	15	5.0	78	26.0	12	4.0	300
82-29V	68.74	68.49	0.25	0.36																					
82-30V	10.40	9.00	1.40	13.46	18	6.0	0	0	0	0	19	6.3	0	0	0	0	243	81.0	8	2.7	0	0	12	4.0	300
82-34V	58.37	39.73	18.64	31.93	0	0	0	0	4	1.3	0	0	0	0	0	0	286	95.3	0	0	8	2.7	2	0.7	300
82-35V	47.76	46.20	2.56	6.38	0	0	0	0	27	13.5	6	3.0	0	0	6	3.0	146	73.0	5	2.5	8	4.0	2	1.0	200
82-38V	56.83	56.25	0.58	1.02	27	13.5	0	0	8	4.0	2	1.0	3	1.5	19	9.5	65	32.5	2	1.0	71	35.5	3	1.5	200
82-41V	68.33	68.13	0.20	0.29	18	9.8	7	3.8	23	12.5	8	4.3	29	15.8	14	7.6	43	23.4	5	2.7	32	17.4	5	2.7	184
82-46V	46.73	46.45	0.28	0.60																					
82-48V	23.28	23.12	0.16	0.69	22	11.0	4	2.0	13	6.5	104	52.0	5	2.5	11	5.5	13	6.5	12	6.0	7	3.5	8	4.0	200
82-49V	37.64	37.50	0.14	0.37																					
82-52V	36.52	36.32	0.20	0.55																					
82-53V	55.98	55.64	0.34	0.61	6	3.0	4	2.0	26	13.0	31	15.5	8	4.0	7	3.5	37	18.5	18	9.0	29	14.5	34	17.0	200
82-54V	62.39	61.79	0.60	0.96	14	7.0	13	6.5	42	21.0	61	30.5	0	0	16	8.0	17	8.5	5	2.5	6	3.0	26	13.0	200
82-56V	50.09	49.65	0.44	0.88																					
82-58V	40.58	38.85	1.73	4.26	8	5.9	3	2.2	7	5.1	3	2.2	0	0	21	15.4	25	18.4	6	4.4	15	11.0	21	15.4	136
82-59V	76.13	73.26	2.87	3.77																					
82-64V	57.97	54.76	3.21	5.54	29	14.5	1	0.5	17	8.5	24	12.0	0	0	12	6.0	81	40.5	7	3.5	29	14.5	0	0	200
82-71V	19.04	18.85	0.19	1.00																					
82-76V	39.64	39.35	0.29	0.73	35	17.5	0	0	9	4.5	26	13.0	0	0	24	12.0	28	14.0	11	5.5	57	28.5	1	0.5	200
82-78V	49.81	49.27	0.54	1.08																					
82-79V	56.36	54.02	2.34	4.16	17	8.5	1	0.5	19	9.5	6	3.0	0	0	1	0.5	83	41.5	2	1.0	7	35.0	1	0.5	200
82-80V	47.38	46.93	0.45	0.95	53	26.5	0	0	35	17.5	8	4.0	0	0	12	6.0	41	20.5	5	2.5	43	21.5	3	1.5	200
82-81V	38.92	38.11	0.81	2.08																					
82-82V	57.87	57.02	0.85	1.47																					
82-83V	55.16	53.83	1.33	2.41	13	6.5	3	1.5	3	1.5	2	1.0	0	0	0	0	12	6.0	2	1.0	57	28.5	0	0	200
82-87V	62.95	61.39	1.56	2.48	11	5.5	0	0	12	6.0	7	3.5	0	0	0	0	11	66.0	9	4.5	41	20.5	1	0.5	200
82-89V	Insufficient Sample																								
82-92V	63.76	61.87	1.89	2.96																					
82-93V	48.20	47.30	0.90	1.87																					
82-99V	50.99	48.57	2.42	4.75	12	6.0	0	0	0	0	0	0	0	0	9	4.5	1	0.5	0	0	136	68.0	42	21.0	200
83-04V	107.50	107.30	0.20	0.19	3	1.5	0	0	59	29.5	12	6.0	0	0	27	13.5	21	10.5	12	6.0	66	33.0	0	0	200
83-06V	104.50	104.30	0.20	0.19	38	19.0	0	0	19	9.5	74	37.0	0	0	0	0	7	3.5	16	8.0	34	17.0	12	6.0	200
83-10V	163.97	163.80	0.17	0.10	2	1.0	0	0	19	9.5	51	25.5	0	0	3	15.0	8	4.0	36	18.0	26	13.0	1	0.5	200
83-12V	231.21	231.00	0.21	0.09																					
83-15V	85.04	84.83	0.21	0.25	3	15.0	4	2.0	6	3.0	26	13.0	0	0	9	4.5	11	5.5	36	18.0	46	23.0	32	16.0	200
83-16V	165.54	163.92	1.62	0.98	1	0.5	0	0	2	1.0	0	0	0	0	0	0	6	3.0	1	0.5	18	9.0	1	0.5	200
83-17V	171.31	166.43	4.88	2.85	7	3.5	1	0.5	2	1.0	0	0	0	0	4	2.0	18	9.0	0	0	166	83.0	2	0.1	200
83-18V	276.51	274.88	1.63	0.59	3	1.5	0	0	13	6.5	6	3.0	1	0.5	6	3.0	117	58.5	12	6.0	4	2.0	2	0.1	200
83-19V	132.99	132.59	0.40	0.30	17	8.5	0	0	4	2.0	28	14.0	0	0	14	7.0	12	6.0	14	7.0	93	46.5	18	9.0	200
83-20V	132.57	132.28	0.29	0.22	13	6.5	0	0	13	6.5	1	5.0	0	0	7	3.5	3	15.0	7	3.5	119	59.5	1	0.5	200
83-21V	279.59	276.75	2.84	1.02	0	0	0	0	0	0	9	4.5	0	0	4	2.0	38	19.0	8	4.0	139	69.5	2	0.1	200

Table 16
(VALLEY SAMPLE SITES)
(continued)

SAMPLE	FINE SAND FRACTION (63 - 250 U) TOTAL	LIGHT MINERAL	HEAVY MINERAL	HEAVY MINERAL	HEAVY MINERAL COUNT ORTHO-PYROXENE		GARNET		EPIDOTE		ALLANITE		HEMATITE		GOETHITE		HORNBLENDE		ZIRCON		OPAQUES		OTHERS		TOTAL
(#)	(Wt.)	(Wt.)	(Wt.)	(% Total)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)
83-22V	407.77	400.97	6.80	1.67	0	0	0	0	0	0	2	1.0	0	0	1	0.5	72	36.0	0	0	125	62.5	0	0	200
83-23V	362.73	358.77	3.96	1.09	12	6.0	0	0	5	2.5	1	0.5	0	0	7	3.5	136	68.0	12	6.0	18	9.0	9	4.5	200
83-24V	291.26	287.78	3.48	1.19	4	2.0	0	0	6	3.0	6	3.0	0	0	0	0	171	85.5	2	1.0	11	5.5	0	0	200
83-25V	122.18	117.64	4.54	3.72	0	0	0	0	0	0	2	1.0	0	0	4	2.0	17	8.5	5	2.5	169	84.5	3	1.5	200
83-38V	129.03	128.80	0.23	0.18	9	4.5	0	0	6	3.0	0	0	0	0	4	2.0	15	7.5	0	0	159	79.5	7	3.5	200
83-44V	83.99	83.74	0.25	0.30	2	1.0	0	0	8	4.0	39	19.5	4	2.0	1	5.0	29	14.5	6	3.0	98	49.0	4	0.2	200
83-49V	116.35	116.00	0.35	0.30	11	5.5	0	0	0	0	9	4.5	0	0	1	5.0	14	7.0	3	1.5	152	76.0	1	0.5	200
83-50V	143.05	142.80	0.25	0.17	2	1.0	0	0	16	8.0	13	6.5	0	0	14	7.0	15	7.5	11	5.5	118	59.0	11	5.5	200
83-63V	136.35	136.25	0.10	0.07	2	1.0	0	0	0	0	0	0	0	0	0	0	9	4.5	2	1.0	185	92.5	2	0.1	200
83-65V	111.74	111.50	0.24	0.21	2	1.0	0	0	24	12.0	6	3.0	0	0	2	1.0	27	13.5	11	5.5	119	59.5	9	4.5	200
83-66V	87.43	87.40	0.03	0.03	2	1.0	0	0	5	2.5	21	10.5	4	2.0	7	3.5	6	3.0	6	3.0	145	72.5	4	0.2	200
83-80V	65.10	65.00	0.10	0.15	12	6.0	0	0	11	5.5	1	5.0	0	0	2	1.0	0	0	7	3.5	158	79.0	0	0	200
83-83V	169.14	169.00	0.14	0.08	3	1.5	0	0	15	7.5	46	23.0	6	3.0	0	0	1	0.5	12	6.0	114	57.0	3	1.5	200
84-01V	167.88	167.25	0.63	0.40	7	3.5	0	0	5	2.5	0	0	0	0	4	2.0	13	6.5	0	0	168	84.0	3	1.5	200
84-02V	97.10	95.00	2.10	2.18	4	2.0	0	0	9	4.5	0	0	0	0	12	6.0	1	5.0	0	0	165	82.5	0	0	200
84-03V	234.60	233.10	1.50	0.64	58	29.0	0	0	39	19.5	12	6.0	0	0	0	0	3	15.0	1	0.5	54	27.0	6	3.0	200
84-04V	145.84	144.00	1.84	1.26	8	4.0	0	0	2	1.0	4	2.0	0	0	6	3.0	13	6.5	4	2.0	145	72.5	0	0	200
84-05V	129.28	128.80	0.48	0.37	7	3.5	0	0	32	16.0	1	0.5	0	0	0	0	8	4.0	1	0.5	148	74.0	3	1.5	200
84-08V	99.38	98.16	1.22	1.23	0	0	0	0	5	2.5	0	0	0	0	1	0.5	8	4.0	0	0	186	93.0	0	0	200
84-09V	133.97	132.34	1.63	1.22	19	9.5	1	0.5	23	11.5	2	1.0	0	0	1	0.5	0	0	0	0	153	76.5	1	0.5	200
84-11V	148.38	148.00	0.38	0.26	2	1.0	0	0	1	5.0	8	4.0	0	0	0	0	19	9.5	0	0	161	80.5	0	0	200
84-13V	141.38	14.00	1.38	0.98	3	1.5	0	0	5	2.5	0	0	0	0	0	0	12	6.0	0	0	18	9.0	0	0	200
84-21V	154.20	152.20	2.00	1.30	9	4.5	0	0	31	15.5	3	1.5	0	0	2	1.0	9	4.5	1	0.5	144	72.0	1	0.5	200
84-23V	152.98	150.40	2.58	1.69	7	3.5	0	0	29	14.5	3	1.5	0	0	0	0	7	3.5	5	2.5	147	73.5	2	1.0	200
84-24V	141.40	139.00	2.40	1.70	4	2.0	0	0	3	15.0	4	2.0	0	0	0	0	4	2.0	0	0	158	79.0	0	0	200
84-25V	172.97	170.20	2.77	1.60	0	0	0	0	28	14.0	7	3.5	0	0	0	0	18	9.0	0	0	146	73.0	1	0.5	200
84-26V	104.00	103.00	1.00	0.96	0	0	0	0	81	40.5	8	4.0	0	0	6	3.0	2	1.0	5	2.5	96	48.0	2	1.0	200
84-27V	Insufficient Sample																								
85-01V	213.29	209.57	3.72	1.74	15	7.5	19	9.5	24	12.0	4	2.0	0	0	4	2.0	8	4.0	0	0	122	61.0	4	2.0	200
85-02V	213.21	21.00	3.21	1.51	18	9.0	2	1.0	36	18.0	0	0	0	0	2	1.0	0	0	0	0	135	67.5	7	3.5	200
85-03V	433.50	43.00	3.50	0.81	38	19.0	4	2.0	26	13.0	0	0	0	0	0	0	0	0	4	2.0	112	56.0	16	8.0	200
85-04V	159.45	155.10	4.35	2.73	23	11.5	9	4.5	35	17.5	0	0	0	0	6	3.0	15	7.5	0	0	96	48.0	16	8.0	200
85-05V	222.03	218.68	3.35	1.51	5	2.5	1	5.0	19	9.5	0	0	0	0	1	0.5	6	3.0	2	1.0	148	74.0	9	4.5	200

available from a number of texts (e.g., Williams et al., 1954; Kerr, 1959; Berry and Mason, 1959; Deer et al., 1982; Edwards and Atkinson, 1986; Blackburn and Dennen, 1988). Table 17 presents a compilation of some of some of the more distinguishable mineral characteristics.

A summary of the relative proportions of heavy minerals for both plateau and valley data shows the following comparisons:

	PLATEAU SAMPLES		VALLEY SAMPLES	
	Average Proportion (%)	Range (%)	Average Proportion (%)	Range (%)
Orthopyroxene	7.2	55.5	6.3	30.3
Garnet	0.8	6.5	0.8	9.5
Epidote	7.7	33.5	9.3	46.5
Allanite	4.1	28.3	6.0	52.0
Hematite	1.0	5.3	0.5	15.8
Goethite	6.2	75.0	3.9	29.5
Hornblende	22.6	96.5	19.4	95.3
Zircon	2.7	25.5	3.0	18.0
Opaques	41.1	96.0	42.9	93.0
Other	3.3	24.0	3.7	21.0

The average proportion of the heavy minerals identified in this study is almost the same for both plateau and valley sediment samples. However, the range of the proportions (i.e. difference between minimum and maximum) varies considerably in most minerals. Specifically, the range of orthopyroxene (55.5%), goethite (75.0%), and zircon (25.5%) for plateau samples is considerably higher than the valley samples (30.3%, 29.5%, and 18.0% respectively). On the other hand, the range in valley samples for epidote (46.5%), allanite (52.0%), and hematite (15.8%) is greater than that occurring in plateau samples (33.5%, 28.3%, and 5.3% respectively).

The significance of these comparisons is not clear, however, preliminary indications are that the sediments from both plateau and valley sites do not vary greatly in heavy mineral species and proportions, but do exhibit a variable range. This leads to the question of what was the origin of much of the sediment that covers the plateaus and valleys; was it locally or externally derived, and more specifically can the data obtained from the heavy mineral analysis help determine this origin? To help answer this question it is necessary to consider and contrast heavy mineral occurrences in bordering regions.

Table 17
HEAVY MINERAL DESCRIPTIONS

MINERAL	COLOUR / LUSTER	DENSITY / HARDNESS	SYSTEM AND HABIT	CLEAVAGE	ASSOCIATION AND OCCURRENCE
EPIDOTE $Ca_2Fe^{+3}Al_2O\cdot OH[Si_2O_7][SiO_4]$	Brown, yellow, green / Vitreous	3.3 - 3.5 / 6	Monoclinic; prismatic to acicular crystals, massive, fibrous or granular	{001} perfect	Formed during metamorphism of calcium rich rocks or by hydrothermal alterations of plagioclase. Associated with iron oxides, quartz, feldspar, chlorite, and hornblende.
ALLANITE $(Ca,Ce)_2(Fe^{+2},Fe^{+3})Al_2O\cdot OH[Si_2O_7][SiO_4]$	Brown to black / Vitreous to submetallic	3.4 - 4.6 / 5.5 - 6	Monoclinic; tabular crystals	{001} poor	Accessory mineral in many granites, granodiorites, monzonites and syenites. Also in limestone skarns and in pegmatites.
GOETHITE $HFeO_2$	Yellow, brown, red / Submetallic to dull	3.3 - 4.3 / 5 - 5.5	Orthorhombic; usually massive, also stalactitic	{010} perfect {100} fair	Occurs as the oxidation product of iron-bearing minerals. Associated with hematite, calcite, quartz, and pyrolusite.
HORNBLENDE $NaCa_2(Mg,Fe,Al)_5(Al,Si)_8O_{22}(OH)_2$	Black to green / Vitreous	2.9 - 3.4 / 5 - 6	Monoclinic; prismatic crystals, also columnar, bladed, and fibrous	{110} perfect	Occurs in igneous rocks and associated with quartz, feldspar, pyroxene, and chlorite. Component of metamorphosed basaltic rocks.
HEMATITE Fe_2O_3	Red brown to black / Metallic to earthy	5.2 - 5.3 / 5 - 6.5	Hexagonal; tabular crystals and massive	None	Common and widespread mineral. Found, in contact metamorphic deposits, as minor accessory in igneous rocks, as coatings and nodules derived by alteration from other iron minerals, in hydrothermal veins, and in the oxidized zone above metalliferous veins.
ZIRCON $ZrSiO_4$	Generally transparent / Vitreous	4.5 - 4.7 / 7.5	Tetragonal; prismatic crystals terminated by pyramids	{110} poor	Accessory mineral in igneous rocks, particularly in the plutonic rocks.
GARNETS					Characteristic of metamorphic rocks and found in some igneous rocks.
Pyrope $Mg_3Al_2Si_3O_{12}$	Pink red to purple / Vitreous to resinous	3.5 / 7 - 7.5	Isometric; commonly crystals	None	Common accessory mineral in ultramafic rocks.
Almandine $Fe_3^{+2}Al_2Si_3O_{12}$	Dark red to black / Vitreous to resinous	4.3 / 7 - 7.5	Isometric; commonly crystals	None	Common mineral in schists and gneisses. Used as gems.
Spessartine $Mn_3Al_2Si_3O_{12}$	Red black, orange yellow / Vitreous to resinous	4.2 / 7 - 7.5	Isometric; commonly crystals	None	Common mineral in schists and gneisses.
Grossular $Ca_3Al_2Si_3O_{12}$	Brown red, green / Vitreous to resinous	3.6 / 7 - 7.5	Isometric; commonly crystals	None	Common in metamorphosed limestone.
Andradite $Ca_3(Fe^{+3},Ti)_2Si_3O_{12}$	Red black, brown yellow / Vitreous to resinous	3.8 / 7 - 7.5	Isometric; commonly crystals	None	Commoly occurs with ore deposits in calcareous rocks.
Uvarovite $Ca_3Cr_2Si_3O_{12}$	Dark green / Vitreous to resinous	3.8 / 7 - 7.5	Isometric; commonly crystals	None	Common in ultramafic rocks and frequently occurs with chromite.
ORTHOPYROXENE					
Enstatite $MgSiO_3$	Grey, green white, green or brown / Vitreous to pearly	3.2 - 3.4 / 5 - 6	Orthorhombic; usually massive, fiberous or granular	{210} good	Common constituent of calcium-poor mafic and ultramafic igneous rocks.
Hypersthene $(Mg,Fe)SiO_3$	Brown black / Vitreous to pearly	3.4 - 4.0 / 6	Orthorhombic; usually massive, platy	{210} good	Found in mafic igneous rocks, especially in lavas. Also in high-grade metamorphic rocks.
OPAQUES					
Pyrite FeS_2	Brass yellow / Metallic	5.0 / 6 - 6.5	Isometric; pyritohedrons, cubes, or octahedrons, also massive, compact or granular	{100} poor	Most widespread and abundant of the sulfides.
Molybdenite MoS_2	Blue to lead grey / Metallic	4.7 / 1 - 1.5	Hexagonal; hexagonal plates, short prisms or massive	{0001} perfect	Found with scheelite, wolframite, topaz, fluorite, chalcopyrite, cassiterite, and epidote in veins, pegmatites, granite, and contact metamorphic deposits.
Chalcopyrite $CuFeS_2$	Brass yellow / Metallic	4.2 - 4.3 / 3.5 - 4	Tetragonal; sphenoidal crystals, usually massive	{011} poor	Found in most rock types. Most widely occuring copper-bearing mineral. Found with pyrite, galena, sphalerite, bornite, chalcocite, quartz, and calcite.
Pyrrholite Fe_7S_8-FeS	Bronze yellow / Metallic	4.6 - 4.7 / 3 - 4	Hexagonal; usually massive, granular or compact	None	Found with pyrite, chalcopyrite, pentlandite, galena, and magnetite in mafic igneous rocks, pegmatites, metamorphic rocks, and hydrothermal veins.
Bornite Cu_5FeS_4	Bronze brown to purple / Metallic	5.06 - 7.6 / 2.5	Isometric; usually massive	{111} traces	Widespread in copper deposits, usually with chalcopyrite and quartz.
Magnetite Fe_3O_4	Black / Metallic (dull)	5.5 / 5.5 - 6.5	Isometric; usually massive	{111} good	Accessory mineral in all igneous rocks.

Sources: Williams et al., (1954); Kerr (1959); Berry and Mason (1959); Deer et al., (1982); Edwards and Atkinson (1986); Blackburn and Dennen (1988).

5.6.3 Heavy Mineral Assemblages

Individual heavy minerals in sediment are rarely diagnostic of provenance (Henderson, 1989). However, the presence of the mineral(s) in the sediment of regions where they are either not known to occur, or are in such minute quantities in the underlying bedrock, have been important in determining source areas (Paré, 1982). The most successful application of heavy mineral analysis (with respect to provenance related studies), lies in identifying certain suites of heavy minerals which are charateristic of specific bedrock lithologies. Such knowledge can aid in the identification of the source region (Pettijohn, 1975). A comparative listing of mineral suites characteristic of source rock types is presented in Table 18.

The observed presence or absence of minerals in sediments is also related to mineral stability. Pettijohn (1975) stated that mineral stability is primarily associated with resistance to solution and decomposition. The presence, absence or relative amount of heavy mineral content in sediment deposits of different ages (such as the plateau and valley samples in this study), may be the result of selective loss of certain minerals after deposition rather than a different provenance. Table 19 presents a listing of the relative stability of heavy minerals as reported by various authors (modified from Pettijohn, 1975). As well, a list based on the relative order of mineral persistence is provided in this table. The order of persistance was based on a collection of minerals in sediments from various locations and ages. Pettijohn (1975) described the list as being derived by considering the frequency of occurrence of minerals in recent (presumably Holocene) sediments, and the average frequency of occurrence in ancient sediments (pre-Holocene?). The ratio of these two frequencies is regarded as the survival ability of the minerals.

Because of the large area covered by the samples (plateau and valley), along with the sampling density, and the sketchy geological information available for much of the study area and surrounding regions, discussions of the results of mineral assemblages are limited to regional patterns. Although a general modal analysis of Ruby Range bedrock was reported by Muller (1967), a detailed heavy mineral assemblage has yet to be collated. Based upon the descriptions from Table 5 (Formations of the Yukon Plateau), the mineral assemblages provided in Table 18, and analysis of bedrock conducted in this study, a heavy mineral assemblage for the Ruby Range (study area) has been compiled (Table 20). As well, an attempt was made at compiling similar assemblages for bordering regions, based on the descriptions of formations in the Dezadeash (south of study area), Aishihik Lake (east of study area), Yukon Plateau (north of study area), and St. Elias regions (west of study area) (Tables 7, 6, 5, and 4 respectively). However, due to the fragmented and generalized lithological descriptions in many of these regions, it is not currently possible to identify the major heavy mineral constituents.

Table 18
SOME HEAVY MINERAL SUITES
IN VARIOUS SOURCE ROCKS

<table>
<tr><th>Rock Source</th><th>Pettijohn (1975)</th><th>Hubert (1971)</th><th>Paré (1982)</th><th>Thorleifson (1989)</th><th>Berry & Mason (1959)</th></tr>
<tr><td>Low-Grade Metamorphic</td><td></td><td>Spessarite
Garnet</td><td></td><td></td><td></td></tr>
<tr><td>High-Grade Metamorphic</td><td>Garnet
Hornblende
Staurolite
Epidote
Zoisite
Magnetite</td><td>Actinolite
Apatite
Almandine
Garnet
Epidote
Clinozoisite
Hornblende
Ilmenite
Kyanite
Magnetite
Staurolite
Tremolite
Zircon</td><td>Garnet
Epidote
Titanite
Zircon
Orthopyroxene
Ilmenite</td><td>Garnet
Epidote
Ilmenite
Pyrite</td><td>Orthopyroxene
Almandite
Pyrope
Diopside
Augite
Hornblende</td></tr>
<tr><td>Acid-Igneous</td><td>Hornblende
Monazite
Sphene
Zircon
Magnetite</td><td>Hornblende
Ilmenite
Monazite
Rutile
Sphene
Zircon</td><td rowspan="2">Undifferentiated Igneous:
Garnet
Epidote
Titanite
Zircon
Ilmenite
Orthopyroxene</td><td rowspan="2">Undifferentiated Igneous:
Garnet
Epidote
Ilmenite
Pyrite</td><td>Zircon
Corundum
Sphene</td></tr>
<tr><td>Basic-Igneous</td><td>Anatase
Augite
Hypersthene
Ilmenite
Magnetite
Chromite
Leucoxene
Olivine
Rutile</td><td>Augite
Epidote
Hornblende
Hypersthene
Ilmenite
Magnetite
Olivine
Pyrope
Garnet</td><td>Olivine
Orthopyroxene
Augite
Hornblende
Ilmenite
Magnetite
Pyrite
Pyrrhotite</td></tr>
<tr><td>Pegmatite</td><td>Garnet
Monazite
Topaz</td><td>Apatite
Garnet
Monazite
Rutile</td><td></td><td></td><td>Columbite
Tantalite
Wolframite</td></tr>
</table>

- Mineral suits identified by Paré (1982), and Thorleifson (1989) are from specific suits in Quaternary deposits. These were also reported by Henderson (1989).

Table 19
RELATIVE STABILITY AND PERSISTENCE
OF SPECIFIC HEAVY MINERALS

STABILITY OF SELECTIVE HEAVY MINERALS
(IN DECREASING ORDER)

Pettijohn (1941)	Smithson (1941)	Sindowski (1949)	Goldrich (1938)	Dryden (1946)
Rutile				
Zircon	Zircon	Zircon		Zircon
Monazite	Monazite			Monazite
Garnet	Garnet			
Staurolite	Staurolite	Staurolite		
Kyanite	Kyanite	Kyanite		Kyanite
Hornblende		Hornblende	Hornblende	Hornblende
		Garnet		Garnet
Augite		Augite	Augite	
Olivine		Olivine	Olivine	

- Modified from Pettijohn (1975). Spacing reflects similarities between series. Stability listed in decreasing order.

ORDER OF PERSISTENCE OF SELECTIVE HEAVY MINERALS
(IN DECREASING ORDER)

1- Anatase	5- Garnet	9- Staurolite	13- Topaz	17- Hypersthene
2- Rutile	6- Apatite	10- Kyanite	14- Sphene	18- Diopside
3- Zircon	7- Ilmenite	11- Epidote	15- Zoisite	19- Actinolite
4- Monazite	8- Magnetite	12- Hornblende	16- Augite	20- Olivine

- Modified from Pettijohn (1975). Persistence is in decreasing numerical order.

Table 20
HEAVY MINERAL ASSEMBLAGES
FOR THE RUBY RANGE STUDY AREA AND BORDERING REGIONS

Rock Type	Ruby Range Study Area	East of Study Area	South of Study Area	West of Study Area	North of Study Area
Undifferentiated Metamorphic Rocks	Epidote Hornblende Garnet Pyrite Rutile Magnetite	Garnet Grossularite Idocrase Epidote Staurolite	Hornblende Actinolite Garnet Magnetite Apatite Zircon Staurolite	Epidote Hornblende Garnet Rutile Magnetite	Epidote Hornblende Garnet Rutile Magnetite
Undifferentiated Igneous Rocks	Apatite Hornblende Sphene Pyrite Zircon	Hornblende Sphene Pyrite Pyrrhotite Hematite Epidote Magnetite Augite	Magnetite Hornblende Zircon Pyrite	Tremolite Epidote Hornblende Magnetite Pyrite Sphene Hematite	Apatite Pyrite Zircon Hornblende Sphene
Non-metamorphed Volcanic and Sedimentary Rocks		Epidote Chalcopyrite Augite	Chalcocite Malachite Bornite Hematite Limonite	Epidote Magnetite Pyrite Hematite Pyroxene Apatite	Hornblende Magnetite
Known Mineralizations	Molybdenite Wolframite Chalcopyrite Pyrrhotite Pyrite Gold	Chalcopyrite Scheelite Molybdenite Magnetite Epidote Garnet Idocrase Pyrrhotite Malachite Pyrite Sphalerite	Gold Pyrite Chalcocite Chalcopyrite Galena Bornite Chromite Copper	Gold Bornite Epidote Copper Platinum Nickel Cobalt Molybdenite Wolframite Pyrite Chalcopyrite Hematite	

- In the Ruby Range Study Area, the underlined minerals represent the major constituents of the heavy mineral fraction as reported by Muller (1967), Smith (1971) and in this study. The other minerals are regarded as minor constituents. Due to the fragmented and generalized lithological descriptions in other regions, it is not possible at this time to similarly identify major constituents in these regions.

- Sources of data include: Kindle, 1945, 1953; Muller, 1967; Smith, 1971; Tempelman-Kluit, 1974; Hughes, 1990; and this study.

Opinions differ as to what percentage of a mineral in the heavy fraction of sediment constitutes a significant amount. Henderson (1989) considered the regional distribution of the >10% portion as being meaningful, whereas Paré (1982) regarded the >5% level as significant. In reality, knowing the natural abundance of a mineral in the source area would help establish this level. However, the presence of a few grains of particular mineral in sediment found far from any known source areas may be notable; the implications of which will be addressed later. For the purposes of this study, the >5% proportion is considered meaningful. Therefore, based on the analysis performed earlier to determine the relative proportions of heavy minerals for both plateau and valley data, the abundance of minerals from plateau regions considered significant are; orthopyroxene (7.2%), epidote (7.7%), goethite (6.2%), hornblende (22.6%), and opaques (41.1%). Similarly, the proportion of minerals from valley sites considered significant are; orthopyroxene (6.3%), epidote (9.3%), allanite (6.0%), hornblende (19.4%), and opaques (42.9%).

The lack of detailed geological information for the study area and surrounding regions, does not allow for definitive conclusions to be made with respect to the relationship between mineral assemblages for sediment samples and rock types. However, observation of a provisional nature can be presented. The major heavy mineral constituents of all the bedrock comprising the Ruby Range study area are tentatively described as including epidote, hornblende, garnet, apatite, pyrite, sphene, and opaques. The assemblage associated with the undifferentiated metamorphic rock (which is confined to one small outcrop in the northwestern extremity of the study area), consists of epidote, hornblende, garnet, and pyrite. The mineral assemblage identified with the undifferentiated igneous rocks (which dominates the study area), consists of apatite, hornblende, and sphene. As well, assembledges associated with known zones of mineralization, occurring within the dominant rock type, consist of molybdenite, wolframite, and chalcopyrite (all opaque).

The heavy mineral assemblage derived from plateau sediments consists of epidote, orthopyroxene, goethite, hornblende, and opaques; the valley sediment assemblage includes epidote, orthopyroxene, allanite, hornblende, and opaques. Apart from the relative lack of allanite on plateaus and goethite in valleys, the assemblages are similar. Although it is evident that epidote, hornblende, and opaques also constitute part of local bedrock mineral assemblages, the source(s) of orthopyroxene, allanite, and goethite in sediment are less clear. (This issue will be addressed in a later discussion).

The data do not support a clear, or exclusive association between heavy mineral assembleges derived from sediments (plateau and valley samples), and those derived from bedrock sources within or outside the study area. However, the assemblages for both plateau and valley samples may not be exclusively related to local bedrock sources. These assemblages appear to demonstrate varying affinities to most bedrock heavy mineral assemblages outside the study area. As well, although the sediment assemblages do not show an overwhelming preference to any particular rock source identified in Table 18, they certainly do not resemble assemblages reported by Hubert (1971) for low-grade metamorphic rocks. This is significant since most of the undifferentiated metamorphic rock in the study area is described as low-grade metamorphic (Muller, 1967; Gabrielse et al., 1980). In fact, there is a more striking correspondence of the heavy minerals identified in this study (orthopyroxene, garnet, epidote, allanite, hematite, goethite, hornblende, zircon, and opaques), with those identified on Table 18 under the high-grade metamorphic, and combined igneous categories.

5.6.4 Provenance Implications of Selected Minerals

As alluded to in the previous discussion, the source of some amounts of minerals identified from the sediment samples of both plateau and valley sites is uncertain, and may be attributed to an external source. Based upon the reported geology of the region, and analysis preformed on bedrock thin sections in this study, evidence will be presented to support the belief that at least some amount of the heavy minerals may be of non-local origin. The minerals in question consist of orthopyroxene, goethite, and allanite. Provenance related discussion will focus primarily on unidentified local bedrock sources, and bedrock outside the study area. As well, other sources such as meteorite fall-out, and volcanic ash will be evaluated.

Orthopyroxene

Berry and Mason (1959) identified orthopyroxene (enstatite and hypersthene) as being associated with basic-igneous rocks and high-grade metamorphic rocks (Table 18). Paré (1982) included orthopyroxene in the heavy mineral assemblage of undifferentiated igneous rocks, while both Hubert (1971), and Pettijohn (1975) described hypersthene as being included in the assemblage of basic-igneous rocks along with augite (pyroxene). As well, both entastite and hypersthene are found in stony meteorites (Blackburn and Dennen, 1988). The bedrock thin section anaylsis conducted for this study indicated there were no occurrences of orthopyroxene in the study area. In addition, Muller (1967) was unequivocal in stating that no hypersthene or enstatite was noted in the modal analyses of the northern Ruby Range.

Furthermore, the occurrence of orthopyroxene is described as being associated with olivine (Blackburn and Dennen, 1988). Berry and Mason (1959) considered this association essential (see mineral assemblages in Table 18). The only reported occurrence of olivine in the immediate vicinity of the study area is in a small intrusive mass of dunite located approximately 50 km northwest of the Talbot Arm (Muller, 1967). It is imperative to note that based on; i) Muller's (1967) modal analysis on bedrock, ii) Smith's (1970) description of bedrock constituent minerals, and iii) this study's analysis on bedrock, not a single olivine mineral grain was identified in the northern Ruby Range study area. Moreover, the identification and point count of heavy minerals for both plateau and valley sediment has not yielded a single olivine mineral count. Until evidence is presented to the contrary, the heavy mineral assemblage of bedrock within the study area does not include orthopyroxene. A more detailed analysis of bedrock throughout the area would substantiate this claim.

Since it is well documented that enstatite and hypersthene are identified with high-grade metamorphic rocks, basic rocks (e.g., basalt, dolerite, gabbro), and ultrabasic rocks (e.g., peridotites, dunites), it follows that regions exhibiting these rock types may serve as possible source areas for these minerals. Tempelman-Kluit (1974) identified biotite-schist in southeastern Snag and northwestern Aishihik Lake map-areas, which are located north, and southeast of the northern Ruby Range study area respectively. These rocks are equivalent to Muller's (1967) quartz-biotite schist (Unit 1, Table 5), which can be generally described as high-grade metamorphic rocks. They also correspond to the Yukon Group which occurs south of the study area in the Dezadeash map region (Kindle, 1953). Although Table 20 does not specifically identify orthopyroxene as constituting part of the bedrock heavy mineral assemblages in these regions, minor undifferentiated pyroxenes are present.

Based on current geological reports, the most prolific sources of orthopyroxenes are the basic and ultabasic rocks that border the northern Ruby Range (see Tables 4 and 5). South of the study area (Dezadeash region), Kindle (1952), and Gabrielse et al., (1980) have reported on the occurrence of peridotite, dunite, basalt and gabbro, all rich in olivine and orthopyroxene. West of the study area (St. Elias Mountains), Muller (1967) has identified basalt (part of St. Clare and Mush Lake groups), some gabbro, and peridotite. Although orthopyroxene was not specifically noted in the basalt, Muller stated that the gabbros did not include enstatite and hypersthene. The peridotites were reported to contain about 10 - 60% undifferentiated pyroxene. To the east of the study area (Aishihik Lake), basalt is identified as occurring in the following units; Little Ridge Volcanics, Carmacks Group, and Massive Green Volcanics (Tempelman-Kluit, 1974; Hughes, 1990). However, no indication of the presence of orthopyroxene is provided.

The possibility that some of the orthopyroxene is of meteorite origin is quite plausible since the occurrence of this phenomenon is ongoing, although isolated. However, olivine is also a common mineral in meteorites, but unlike orthopyroxene, none was identified in the sediment samples. The absence of olivine in sediment samples may be attributed to the fact that the mineral weathers readily and alters to goethite and hematite (Deer et al., 1982), both of which have been observed in the study area. Still, meteorite impact is generally regarded as an isolated phenomena both temporally and spatially. Thus, although a meteoritic origin of at least some of the orthopyroxene cannot be dismissed, it is unlikely that the widespread occurrence of the mineral throughout the study area can be attributed to this phenomena.

Another source of orthopyroxene (hypersthene) may be attributed to the White River Ash that inundated southwestern Yukon about 1,230 yrs BP (Downes, 1985). The heavy minerals that comprised part of the deposit included, hornblende, hypersthene, ilmenite, and magnetite. Lerbekmo and Campbell (1969) reported the mineral composition (Wt %) of the ash as; quartz (23.2%), orthoclase (14.8%), albite (34.7%), anorthite (13.9%), hornblende (9.6%), hypersthene (1.0%), ilmenite (0.9%), and magnetite (1.8%). The heavy mineral proportion of the composition is 13.3%.

Dewez (1988) determined that the heavy minerals minerals could be differentiated on the basis of a pitted and glassy, perlitic texture. With this textural feature in mind, care was taken to record such occurrences during the identification and counting of heavy minerals. With respect to hypersthene, no grains were identified with these textural traits. This may be attributed to a number of factors:

i) Dewez (1988) used bromoform (S.G. 2.89) to separate the heavy mineral fraction; in this study, methylene iodide (S.G. 3.3) was utilized. Since the specific gravity of hypersthene varies between 3.1 - 3.9 (depending on Fe content), only hypersthene >3.3 S.G. was identified in this study. Therefore, a portion of the hypersthene derived from volcanic ash, S.G. <3.3, was not encountered. It may be that much of this volcanic derived hypersthene was of lower specific gravity, although there is no data to substantiate this speculation.

ii) In order to avoid contamination of samples with volcanic ash, care was taken to sample below any noticable ash layer. However, although this may have eliminated much of the ash, natural mixing and migratory processes would certainly result in some of the deposit being incorporated into the samples.

iii) The textural traits, particularly glassy, perlitic texture, may be on such an infinitesimal scale that it was not evident under examination with a stereoscopic binocular microscope.

Goethite

Goethite accounts for 6.2% of the heavy mineral fraction from plateau sediments, and 3.9% from valley sediments. Goethite usually results from the weathering of iron minerals such as siderite, pyrite, magnetite, etc., under oxidizing conditions. It is also an alteration product of olivine.

Andrews-Jones (1968) has empirically estimated the relative chemical mobilities of elements in the secondary environment which has application to the Ruby Range. The classification is based on environmental conditions which include oxidizing potential (Eh), and acid to alkaline (pH). The Eh of a solution is a measure of the oxidizing or reducing tendency, or the potential of a system (Cloke, 1966; Hem, 1970). Since oxidation and reduction involve the transfer of electrons, it is basically an electrical property which can be measured in volts. Measurements are referred to the standard hydrogen electrode, the potential of which is taken as zero at pH 7 and a pressure of 1 atmosphere. The more positive the potential, the more oxidizing it is relative to the hydrogen half cell.

At virtually all sample locations, and in particular at valley sites, a hematite staining was observed on the undersides of numerous rocks and boulders. The staining is a good indication of strong oxidation. Barakso (1970), after taking numerous readings of pH and Eh of streams in the Talbot\Rockslide Creeks area, noted that the environmental conditions were extremely acid and rich in oxidizing material (average pH 2.5, Eh 0.8). Evidence of oxidation was noticed on plateaus, but not to the same extent as in the valleys. This is likely due to restricted drainage and the presence of permafrost. Reducing conditions are noted to be more predominant on the plateau regions (Pollard per. comm.).

The greater proportion of goethite in plateau samples as compared to valley samples, is probably related to the occurrence of pyrite which constitutes part of the mineral assemblage of bedrock within the study area, and specifically may be related to the reported mineralizations on the plateaus. It is also possible that pyrite (and perhaps other minerals which alter to form goethite), may have been more abundant in plateau sediment which could have resulted in more goethite formation, even though the oxidizing environment is less severe on plateaus. In addition to local bedrock sources, other source areas may include regions to the south where the bedrock mineral assemblages include limonite (goethite), pyrite, and magnetite. As well, significant iron content in some peridorite and dunite bodies were reported by Kindle (1952). Similar occurrences were reported in regions of the St. Elias Mountains (Muller, 1967).

Since goethite is related to the weathering of siderite, pyrite, magnetite, etc., it is possible that the occurrence of some goethite can be attributed to the weathering of volcanic ash derived magnetite (which constituted 1.8% of the White River ash deposit). The proportion of magnetite in the ash deposit appears to be far greater than any reported proportion in the bedrock mineral assemblages for the study area and surrounding regions (except for some areas in the Burwash Uplands where combined magnetite and pyrite proportions were reported at 2.0 - 2.8% by Muller, 1967). Meteoritic olivine may also constitute a source for goethite in the study area.

Allanite

Allanite constitutes 4.1% of plateau, and 6.0% of valley sediment heavy mineral fraction. Muller (1967) reported some minor epidote (0.1%) in the modal analysis of Ruby Range batholith rocks, but did not distinquish allanite from other epidotes. Furthermore, of the 23 rock specimens Muller examined (none of which were taken from the study area), only 2 specimens contained epidote. Minerals of the epidote group characteristically occur in metamorphic rocks except for allanite. Although small amounts of allanite are found in granitic rocks, it is more abundant in granite pegmatites (vein-like bodies distinguished by unusually coarse-grained texture of the mineral aggregates; Berry and Mason, 1959), and syenites. There are no known occurrences of these in the study area (Muller, 1967; Smith, 1971).

Since the proportion of allanite in the plateau and valley sediment heavy mineral fraction appears to be considerably higher than in any known local bedrock occurrence, the provenance of much of the allanite may be attributed to an external source. Kindle (1959) reported numerous granitic pegmatites associated with the Yukon Group, in the Dezadeash Lake region (south of the northern Ruby Range study area), as well as syenite rock. Although pegmatites and syenites have not been separately mapped in other regions of the St. Elias Mountains, they must certainly occur (Kindle, 1953; Muller, 1967; Gabrielse et al., 1980). The association of allanite with the White River volcanic ash (Lerbekmo and Campbell, 1969), and meteoritic sources have not been reported (Berry and Mason, 1959; Whitten, 1982).

5.6.5 Heavy Mineral Composition, Assemblages and Provenance: An Overview

The average proportion of heavy minerals (% wt) comprising the fine-sand fraction for plateau regions is 1.68%, and 1.93% for valleys. The maximum and minimum percentiles for these proportions are 12.35% to 0.24% for plateau regions, and 31.93% to 0.03% for valley locations. Although Muller (1967) presented bedrock modes based on 500 point counts, some general observations pertaining to these proportions can be made. The average proportion of heavy minerals constituting the mineral assemblage of 23 granitic rocks sampled by Muller from the Ruby Range batholith is 6.6%. The maximum and minimum percentiles are 38.6% to 0%. The minerals included in Muller's heavy mineral group (S.G. 3.3) are; amphibole (undifferentiated), pyroxene (undifferentiated), epidote, apatite, magnetite, pyrite, sphene and zircon. Although Muller does not differentiate the amphiboles (S.G. 2.9 - 3.6), and the pyroxenes (S.G. 3.1 - 3.9), it is clear from his report that hornblende is the major amphibole, and diopside and augite are the main pyroxenes.

Muller (1967) identified amphibole (hornblende) as the predominant mineral, generally comprising about 80% of the heavy mineral portion of the bedrock assemblage. This corresponds to about 5.3% of the average total heavy mineral component (6.6%) in the rock samples. By contrast, Muller (1967) showed that, on average, the heavy minerals constituting the mineral assemblage of 10 alaskite rocks was 0.18%. The proportional range of the heavy mineral content was reported as 0% to 0.6%. The only heavy minerals suites constituting the assemblage of these rock samples were magnetite, pyrite, and in one sample, zircon.

Another interesting comparison concerns the proportion of opaques. The opaques identified in plateau and valley sediment samples included; magnetite, pyrite, molybdenite, chalcopyrite, pyrrholite, and bornite. By comparison, Muller's opaques include magnetite and pyrite. The average opaque portion of the heavy mineral fraction from plateau and valley sediment samples is virtually identical; 41.1% for plateaus, and 42.9% for valleys. By contrast, Muller (1967) reported the average opaque portion of the heavy mineral content for the 23 granitic rocks (6.6%) of the Ruby Range as 0.17%. For alaskite, the average opaque portion of the heavy mineral content (1.8%) was also 0.17%. The huge difference between the average opaque content of sediment samples and bedrock is likely the result of local mineralizations (that are known to be abundant in opaques), from which Muller did not analyse for constituent minerals. Apart from the known zones of mineralizations, it is probable that other mineralizations occur which have not yet been identified (Discussion pertaining to this will be addressed in a later section). These zones may be responsible for many of the opaques found in the sediment samples from plateaus and valleys.

The sketchy nature of the geological information within and outside the study area prohibits any definitive statements with respect to the relationship between heavy mineral assemblages for sediment samples and rock types. However, based upon the available geological data, along with the analyses conducted in this study, some tentative observations are presented.

i) There does not appear to be an appreciable difference in the heavy mineral assemblages between plateau and valley sediment samples, except for the relative lack of allanite from plateaus, and goethite from valleys.

ii) The data do not support a clear association between heavy mineral assemblages from plateau and valley sediments, and those derived from bedrock within or outside the study area. However, the sediment assemblages may not be exclusively related to local bedrock sources.

iii) The provenance of orthopyroxene, goethite, and allanite may be attributed, in part, to bedrock from outside the study area, volcanic ash, and meteoritic impacts, or a combination of these. The occurrence of some amounts of these minerals may also be associated with mineral alterations.

5.6.6 Heavy Mineral Dispersal Trends

5.6.6.1 Introduction

The discussion on heavy minerals to this point has dealt with composition, counts, assemblages, and provenance implications. To gain an appreciation of the distribution of the heavy minerals in the sediment, it is advantageous to graphically illustrate the variations in composition. The analysis of dispersal trends, together with the preceding discussion on composition, assemblages, and provenance will facilitate discussion on the extent, nature and direction of sediment transport and deposition.

Mineral dispersals trends were graphically illustrated for plateau sediments data only. The extensive nature of the plateau surfaces, along with the sampling procedure facilitated interpolation between sample sites. Conversely, in valley locations, the rather restricted confines, and sampling of linear features (lateral moraines) was not conducive to interpolation. Therefore, discussion on heavy mineral dispersals in valleys will focus on the compositional data presented in the previous section.

5.6.6.2 Background Values and Anomalies

Being confronted with the general lack of reported detailed lithological descriptions in the study area and surrounding regions (particularly with respect to heavy mineral composition), along with the knowledge that some minerals are of volcanic origin (ash-fall), and the possibility that others may be of meteoritic origin, and simply not being able to establish the genesis of much of the surficial sediment (particularly on plateaus), it was decided that a somewhat unorthodox approach to interpreting the empirical data was necessary to help account for these "unknowns". In consideration of the above mentioned complexities, it was determined that one way to account for the "unknowns" was to consider that the heavy minerals are associated with normal background values, which in turn may aid in the detection of anomalies; analogous to geochemical distributions (to be discussed in a following section).

By establishing background values, the presence of unusually high values will be revealed by anomalous values. The background value is usually taken as the median value for the data distribution (Levinson, 1980). The median is defined as the value for which one half the values in the distribution are less and one half are greater. The background value is naturally likely to vary since the distribution is rarely uniform. Hence background values should be viewed as estimates which fall within a range for any given area.

There is no universally accepted rule for what constitutes an anomaly. However, for all practical purposes, anomalies are characterized only by abnormally high concentrations. For example, Boyle (1974) suggested that samples which contain amounts double the background or greater should be considered anomalous. Other researchers consider anything over background to be anomalous (Barakso and Tarnocai, 1970; Levinson, 1980). Hawkes and Webb (1962) describe two methods of determining anomalies. They claim the background may be considered as the mean plus twice the standard deviation, which is generally equivalent to 1 in 40 samples exceeding the background. Alternatively, one of the most commonly used method is to consider the median value to be the background and only deem about the highest 2.5% of values above background as being anomalous. It is this latter method of approximation that is used in this study. (The determination of background and anomaly values is the same for the geochemical data).

The background and anomaly values for heavy minerals in the surficial sediments from plateaus and valleys are as follows:

	PLATEAU DATA		VALLEY DATA	
	Background (%)	Anomaly (%)	Background (%)	Anomaly (%)
Orthopyroxene	5.0	29.3	4.0	29.0
Garnet	0.0	4.3	0.0	6.5
Epidote	4.5	29.5	6.5	40.5
Allanite	2.5	16.5	3.0	37.0
Hematite	0.0	4.5	0.0	4.0
Goethite	2.0	50.5	2.0	15.4
Hornblende	12.0	85.5	7.5	85.5
Zircon	1.0	14.0	2.5	9.0
Opaques	29.5	92.0	46.5	92.5

A review of various texts reveals that there is not a commonly practiced method for determining at what measure background and anomaly values between areas (such as plateau and valley samples in this study) constitutes a meaningful difference (e.g., Hawkes and Webb, 1962; Krumbein and Graybill, 1965; Levinson, 1980; Mathewson, 1984). However, it has often been advocated that a qualitative appraisal of these differences produces the most meaningful results (Hawkes and Webb, 1962; McCammon, 1974; Shilts per. comm.). Specifically, since background values are considered estimates which fall within a range for any given area, the qualitative approach to determining what constitutes a meaningful difference in values is acceptable. In this respect, it was decided that if the variation in background or anomaly value of a mineral was twice as much between the plateau and valley sites, it constituted a meaningful difference.

Under these guidelines, except for minor differences, the background values for most heavy minerals in both plateau and valley samples are quite similar; the exception being zircon, 1.0% for plateaus, 2.5% for valleys. The anomaly values are also relatively similar for most minerals (ie. orthopyroxene, garnet, epidote, hematite, hornblende, zircon, and opaques), but vary for allanite (16.5% for plateaus, compared to 37.0% for valleys), and goethite (50.5% for plateaus, compared to 15.4% for valleys). These values indicate there is a relatively uniform concentration of minerals in the sediment from both plateaus and valleys, as exemplified by background values, but because of the differences in some anomaly values, demonstrate a somewhat varying dispersal pattern. This finding is similar to that presented in the discussion on heavy mineral composition where, the average proportion of heavy minerals were found to be similar in both plateau and valley sediment samples, but a significant range in some values was observed, particularly with respect to epidote, allanite, and goethite for valley samples.

5.6.6.3 Valley Dispersal Trends

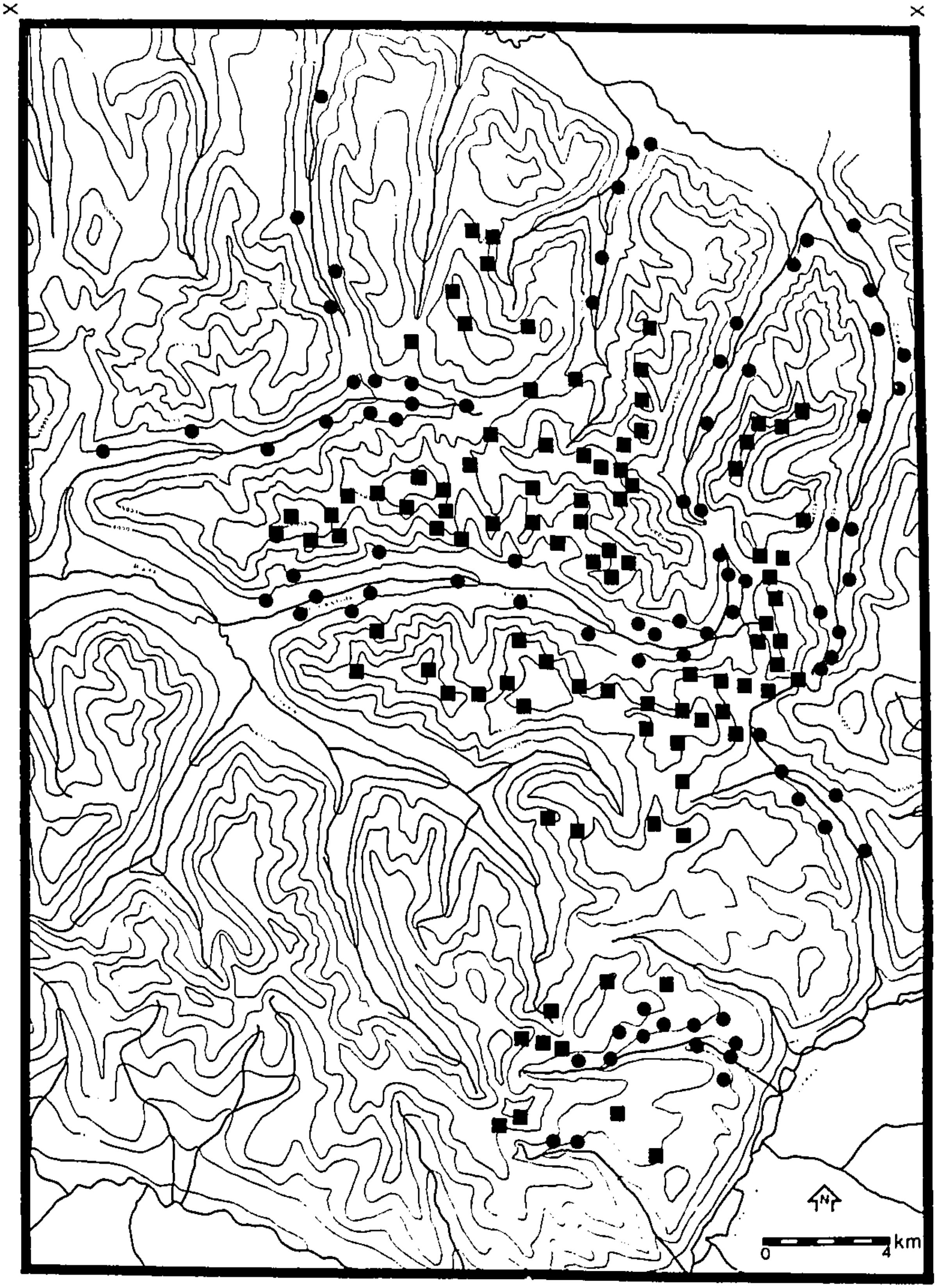
As stated earlier, due to the rather restricted confines, and sampling of linear features (moraines) in the valleys, it was not possible to graphically portray dispersal trends of minerals. Therefore, discussion on mineral dispersals will focus on compositional data presented in the previous section. Table 21 provides a listing of heavy mineral concentrations in the valleys in the northern Ruby Range which include; Talbot Creek, Alaskite Creek, Rockslide Creek, and informally named valleys which include, Ice Valley, Pass Creek, Moraine Valley, Gladstone North, Caribou Creek and Camp Creek (see Figure 12). The data in Table 21 is arranged from an upper to lower valley order. The interpretations made with respect to mineral dispersals in valleys in this discussion are based in many instances on a limited number of samples, and therefore should be viewed as provisional. In order to avoid confusion with respect to the discussion on mineral trends, the reader is reminded that glacier deposits were sampled in valleys.

Table 21
HEAVY MINERAL DISPERSALS IN VALLEYS
(UPPER TO LOWER VALLEY LOCATIONS)

SAMPLE	FINE SAND FRACTION (63 - 250 U) TOTAL	LIGHT MINERAL	HEAVY MINERAL	HEAVY MINERAL	HEAVY MINERAL COUNT: ORTHO-PYROXENE		GARNET		EPIDOTE		ALLANITE		HEMATITE		GOETHITE		HORNBLENDE		ZIRCON		OPAQUES		OTHERS		TOTAL
(#)	(Wt.)	(Wt.)	(Wt.)	(% Total)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)
TALBOT CREEK																									
82-24V	43.64	43.54	0.10	0.23	37	30.3	6	4.9	8	6.6	5	4.1	1	0.8	36	29.5	8	6.6	2	1.6	5	4.2	14	11.5	122
82-25V	Insufficient Sample																								
82-26V	Insufficient Sample																								
82-27V	32.27	32.00	0.27	0.84																					
82-28V	59.83	59.16	0.67	1.12	51	17.0	4	1.3	55	18.3	5	1.7	2	0.7	46	15.3	32	10.7	15	5.0	78	26.0	12	4.0	300
83-04V	107.50	107.30	0.20	0.19	3	1.5	0	0	59	29.5	12	6.0	0	0	27	13.5	21	10.5	12	6.0	66	33.0	0	0	200
82-29V	68.74	68.49	0.25	0.36																					
82-49V	37.64	37.50	0.14	0.37																					
82-48V	23.28	23.12	0.16	0.69	22	11.0	4	2.0	13	6.5	104	52.0	5	2.5	11	5.5	13	6.5	12	6.0	7	3.5	8	4.0	200
82-46V	46.73	46.45	0.28	0.60																					
83-06V	104.50	104.30	0.20	0.19	38	19.0	0	0	19	9.5	74	37.0	0	0	0	0	7	3.5	16	8.0	34	17.0	12	6.0	200
82-30V	10.40	9.00	1.40	13.46	18	6.0	0	0	0	0	19	6.3	0	0	0	0	243	81.0	8	2.7	0	0	12	4.0	300
82-41V	68.33	68.13	0.20	0.29	18	9.8	7	3.8	23	12.5	8	4.3	29	15.8	14	7.6	43	23.4	5	2.7	32	17.4	5	2.7	184
ALASKITE CREEK																									
83-36V	129.03	128.80	0.23	0.18	9	4.5	0	0	6	3.0	0	0	0	0	4	2.0	15	7.5	0	0	159	79.5	7	3.5	200
82-38V	56.83	56.25	0.58	1.02	27	13.5	0	0	8	4.0	2	1.0	3	1.5	19	9.5	65	32.5	2	1.0	71	35.5	3	1.5	200
82-35V	47.76	45.20	2.56	5.36	0	0	0	0	27	13.5	6	3.0	0	0	6	3.0	146	73.0	5	2.5	8	4.0	2	1.0	200
82-34V	58.37	39.73	18.64	31.93	0	0	0	0	4	1.3	0	0	0	0	0	0	286	95.3	0	0	8	2.7	2	0.7	300
82-11V	56.20	55.20	1.00	1.78	4	2.0	0	0	36	18.0	4	2.0	2	1.0	5	2.5	123	61.5	8	4.0	2	1.0	16	8.0	200
82-10V	35.91	35.16	0.75	2.09	14	7.0	0	0	47	23.5	6	3.0	1	0.5	12	6.0	101	50.5	1	0.5	12	6.0	6	3.0	200
82-08V	25.09	23.89	1.20	4.78	1	0.5	0	0	8	4.0	1	5.0	0	0	3	1.5	157	78.5	8	4.0	5	2.5	8	4.0	200
82-09V	47.19	45.55	1.64	3.48	11	5.5	1	0.5	93	46.5	6	3.0	0	0	17	8.5	36	18.0	12	6.0	15	7.5	9	4.5	200
ROCKSLIDE CREEK																									
83-44V	83.99	83.74	0.25	0.30	2	1.0	0	0	8	4.0	39	19.5	4	2.0	1	5.0	29	14.5	6	3.0	98	49.0	4	0.2	200
83-49V	116.35	116.00	0.35	0.30	11	5.5	0	0	0	0	8	4.5	0	0	1	5.0	14	7.0	3	1.5	162	78.0	1	0.5	200
83-63V	136.35	136.25	0.10	0.07	2	1.0	0	0	0	0	0	0	0	0	0	0	9	4.5	2	1.0	185	92.5	2	0.1	200
83-60V	143.05	142.80	0.25	0.17	2	1.0	0	0	16	8.0	13	6.5	0	0	14	7.0	15	7.5	11	5.5	118	59.0	11	5.5	200
83-65V	111.74	111.50	0.24	0.21	2	1.0	0	0	24	12.0	6	3.0	0	0	2	1.0	27	13.5	11	5.5	119	59.5	9	4.5	200
83-66V	87.43	87.40	0.03	0.03	2	1.0	0	0	5	2.5	21	10.5	4	2.0	7	3.5	6	3.0	6	3.0	145	72.5	4	0.2	200
83-23V	362.73	358.77	3.96	1.09	12	6.0	0	0	5	2.5	1	0.5	0	0	7	3.5	138	69.0	12	6.0	18	9.0	9	4.5	200
83-22V	407.77	400.97	6.80	1.67	0	0	0	0	0	0	2	1.0	0	0	1	0.5	72	36.0	0	0	125	62.5	0	0	200
83-24V	291.26	287.78	3.48	1.19	4	2.0	0	0	6	3.0	6	3.0	0	0	0	0	171	85.5	2	1.0	11	5.5	0	0	200
83-80V	65.10	65.00	0.10	0.15	12	6.0	0	0	11	5.5	1	5.0	0	0	2	1.0	0	0	7	3.5	158	79.0	0	0	200
83-83V	169.14	169.00	0.14	0.08	3	1.5	0	0	15	7.5	46	23.0	6	3.0	0	0	1	0.5	12	6.0	114	57.0	3	1.5	200
83-21V	279.59	276.75	2.84	1.02	0	0	0	0	0	0	9	4.5	0	0	4	2.0	38	19.0	8	4.0	139	69.5	2	0.1	200
83-20V	132.57	132.28	0.29	0.22	13	6.5	0	0	13	6.5	1	5.0	0	0	7	3.5	3	15.0	7	3.5	119	59.5	1	0.5	200
83-25V	122.18	117.64	4.54	3.72	0	0	0	0	0	0	2	1.0	0	0	4	2.0	17	8.5	5	2.5	169	84.5	3	1.5	200
83-19V	132.99	132.59	0.40	0.30	17	8.5	0	0	4	2.0	28	14.0	0	0	14	7.0	12	6.0	14	7.0	93	46.5	18	9.0	200
83-16V	165.54	163.92	1.62	0.98	1	0.5	0	0	2	1.0	0	0	0	0	0	0	6	3.0	1	0.5	18	9.0	1	0.5	200
83-17V	171.31	166.43	4.88	2.85	7	3.5	1	0.5	2	1.0	0	0	0	0	4	2.0	18	9.0	0	0	166	83.0	2	0.1	200
83-18V	276.51	274.38	1.63	0.59	3	1.5	0	0	13	6.5	6	3.0	1	0.5	6	3.0	117	58.5	12	6.0	4	2.0	2	0.1	200
83-10V	163.97	163.80	0.17	0.10	2	1.0	0	0	19	9.5	51	25.5	0	0	3	15.0	8	4.0	36	18.0	26	13.0	1	0.5	200
83-12V	231.21	231.00	0.21	0.09																					
83-15V	85.04	84.83	0.21	0.25	3	15.0	4	2.0	6	3.0	26	13.0	0	0	9	4.5	11	5.5	36	18.0	46	23.0	32	16.0	200

SAMPLE	TOTAL	LIGHT MINERAL	HEAVY MINERAL	HEAVY MINERAL	ORTHO-PYROXENE		GARNET		EPIDOTE		ALLANITE		HEMATITE		GOETHITE		HORNBLENDE		ZIRCON		OPAQUES		OTHERS		TOTAL
	FINE SAND FRACTION (63 - 250 U)				HEAVY MINERAL COUNT																				
(#)	(Wt.)	(Wt.)	(Wt.)	(% Total)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)	(%)	(#)
PASS CREEK																									
82-59V	78.13	73.26	2.87	3.77																					
82-53V	55.98	55.64	0.34	0.61	6	3.0	4	2.0	26	13.0	31	15.5	8	4.0	7	3.5	37	18.5	18	9.0	29	14.5	34	17.0	200
82-54V	62.39	61.79	0.60	0.96	14	7.0	13	6.5	42	21.0	61	30.5	0	0	16	8.0	17	8.5	5	2.5	6	3.0	26	13.0	200
82-56V	50.09	49.65	0.44	0.88																					
82-58V	40.58	38.85	1.73	4.28	8	5.9	3	2.2	7	5.1	3	2.2	0	0	21	15.4	25	18.4	6	4.4	15	11.0	21	15.4	136
82-62V	36.52	36.32	0.20	0.55																					
ICE VALLEY																									
84-24V	141.40	139.00	2.40	1.70	4	2.0	0	0	3	16.0	4	2.0	0	0	0	0	4	2.0	0	0	158	79.0	0	0	200
84-21V	154.20	152.20	2.00	1.30	9	4.5	0	0	31	15.5	3	1.5	0	0	2	1.0	9	4.5	1	0.5	144	72.0	1	0.5	200
84-25V	172.97	170.20	2.77	1.60	0	0	0	0	28	14.0	7	3.5	0	0	0	0	18	9.0	0	0	146	73.0	1	0.5	200
84-28V	104.00	103.00	1.00	0.96	0	0	0	0	81	40.5	8	4.0	0	0	6	3.0	2	1.0	5	2.5	96	48.0	2	1.0	200
84-27V	Insufficient Sample																								
84-05V	129.28	128.80	0.48	0.37	7	3.5	0	0	32	16.0	1	0.5	0	0	0	0	8	4.0	1	0.5	148	74.0	3	1.5	200
84-23V	152.98	150.40	2.58	1.69	7	3.5	0	0	29	14.5	3	1.5	0	0	0	0	7	3.5	5	2.5	147	73.5	2	1.0	200
84-04V	145.84	144.00	1.84	1.26	8	4.0	0	0	2	1.0	4	2.0	0	0	6	3.0	13	6.5	4	2.0	145	72.5	0	0	200
84-03V	234.60	233.10	1.50	0.64	58	29.0	0	0	39	19.5	12	6.0	0	0	0	0	3	15.0	1	0.5	54	27.0	6	3.0	200
84-02V	97.10	95.00	2.10	2.16	4	2.0	0	0	9	4.5	0	0	0	0	12	6.0	1	5.0	0	0	165	82.5	0	0	200
84-01V	157.88	157.25	0.63	0.40	7	3.5	0	0	5	2.5	0	0	0	0	4	2.0	13	6.5	0	0	168	84.0	3	1.5	200
MORAINE VALLEY																									
82-76V	39.64	39.35	0.29	0.73	35	17.5	0	0	9	4.5	26	13.0	0	0	24	12.0	28	14.0	11	5.5	57	28.5	1	0.5	200
82-83V	55.16	53.83	1.33	2.41	13	6.5	3	1.5	3	1.5	2	1.0	0	0	0	0	12	6.0	2	1.0	57	28.5	0	0	200
82-78V	49.81	49.27	0.54	1.08																					
82-79V	56.36	54.02	2.34	4.15	17	8.5	1	0.5	19	9.5	6	3.0	0	0	1	0.5	83	41.5	2	1.0	7	35.0	1	0.5	200
82-89V	Insufficient Sample																								
82-82V	57.87	57.02	0.85	1.47																					
82-87V	62.95	61.39	1.56	2.48	11	5.5	0	0	12	6.0	7	3.5	0	0	0	0	11	55.0	9	4.5	41	20.5	1	0.5	200
82-99V	50.99	48.57	2.42	4.75	12	6.0	0	0	0	0	0	0	0	0	9	4.5	1	0.5	0	0	136	68.0	42	21.0	200
82-71V	19.04	18.85	0.19	1.00																					
82-64V	57.97	54.76	3.21	5.64	29	14.5	1	0.5	17	8.5	24	12.0	0	0	12	6.0	81	40.5	7	3.5	29	14.5	0	0	200
82-80V	47.38	46.93	0.45	0.95	53	26.5	0	0	35	17.5	8	4.0	0	0	12	6.0	41	20.5	5	2.5	43	21.5	3	1.5	200
82-81V	38.92	38.11	0.81	2.08																					
GLADSTONE NORTH																									
82-92V	63.76	61.87	1.89	2.96																					
82-93V	48.20	47.30	0.90	1.87																					
CARIBOO CREEK																									
84-11V	148.38	148.00	0.38	0.26	2	1.0	0	0	1	5.0	8	4.0	0	0	0	0	19	9.5	0	0	161	80.5	0	0	200
84-13V	141.38	14.00	1.38	0.98	3	1.5	0	0	5	2.5	0	0	0	0	0	0	12	6.0	0	0	18	9.0	0	0	200
84-09V	133.97	132.34	1.63	1.22	19	9.5	1	0.5	23	11.5	2	1.0	0	0	1	0.5	0	0	0	0	153	76.5	1	0.5	200
84-08V	99.38	98.16	1.22	1.23	0	0	0	0	5	2.5	0	0	0	0	1	0.5	8	4.0	0	0	186	93.0	0	0	200
CAMP CREEK																									
85-05V	222.03	218.68	3.35	1.51	5	2.5	1	5.0	19	9.5	0	0	0	0	1	0.5	6	3.0	2	1.0	148	74.0	9	4.5	200
85-04V	159.45	155.10	4.35	2.73	23	11.5	9	4.5	35	17.5	0	0	0	0	6	3.0	15	7.5	0	0	96	48.0	16	8.0	200
85-03V	433.50	43.00	3.50	0.81	38	19.0	4	2.0	26	13.0	0	0	0	0	0	0	0	0	4	2.0	112	56.0	16	8.0	200
85-02V	213.21	21.00	3.21	1.51	18	9.0	2	1.0	36	18.0	0	0	0	0	2	1.0	0	0	0	0	135	67.5	7	3.5	200
85-01V	213.28	209.57	3.72	1.74	15	7.5	19	9.5	24	12.0	4	2.0	0	0	4	2.0	8	4.0	0	0	122	61.0	4	2.0	200

Figure 12
LOCATION OF VALLEY AND PLATEAU SAMPLE SITES



● Valley Sample Sites ■ Plateau Sample Sites

Talbot Creek

In Talbot Creek, sample sites extended from the head of the valley to the Alaskite Creek confluence. Orthopyroxene, garnet, epidote, allanite, hematite, hornblende, zircon and opaques do not exhibit an increasing or decreasing tendency in either an up-valley or down-valley direction. Goethite displays a slight decreasing concentration down-valley. In only three minerals are anomalies found. These include; orthopyroxene and goethite in the upper reaches of the valley; allanite and hematite in the middle and lower parts of the valley.

Alaskite Creek

Sampling in Alaskite Creek extended down the entire valley to the junction of Talbot Creek. The concntration of orthopyroxene, allanite, hematite, goethite and hornblende exhibit no obvious trends. Garnet is virtually absent altogether, while epidote and zircon increase slightly in the down-valley direction. Opaques show a decrease in tendency down-valley. Only two minerals display anomaly values; epidote in a lower-valley sample, and hornblende in a middle-valley location.

Rockslide Creek

In Rockslide Creek, samples were taken from the head of the valley to just before the Raft Creek confluence. There are no discernable dispersal trends associated with orthopyroxene, epidote, allanite, hematite, goethite, hornblende and opaques. Garnet is generally absent. Zircon may be interpreted as demonstrating a slight increase in the down-valley direction. An anomaly value is detected for zircon in the two lower-valley sample sites, and opaques in an upper-valley site.

Pass Creek

In Pass Creek, virtually all the minerals show no preferred dispersal trends. Hematite is virtually absent, whereas goethite increases down-valley. However, because of the limited number of samples (3) from this valley, these observations are highly speculative. Anomaly values are recorded for hematite and zircon in upper-valley locations, garnet in mid-valley, and goethite in a lower-valley sample site.

Ice Valley

Samples in Ice Valley were collected from the upper reaches of the valley to just east of the Raft Creek junction. There do not appear to be discernable dispersal trends associated with any of the minerals analysed for. Garnet and hematite do not constitute part of any sample. Two minerals exhibit anomaly values; epidote in a mid-valley site, and orthopyroxene in a lower-valley location.

Moraine Valley

Dispersal trends in Moraine Valley do not demonstrate an up-valley or down-valley preference. Hematite is not present in any sample, and garnet is scarce. No anomaly values are recorded for any of the minerals.

Gladstone North

No interpretation.

Caribou Creek

There are no decernable trends in the mineral dispersals. Hematite and zircon were found to be absent, while garnet only appeared in one sample. An anomaly for opaques is located in a lower-valley location.

Camp Creek

All the minerals analysed for, with the exception of hematite which is absent entirely, exhibit no trends in dispersal. Allanite was identified in one sample. Garnet displays an anomaly in a lower-valley sample site.

Although these observations may be somewhat tentative, the data as a whole indicates that there is no evidence to suggest minerals demonstate an overwhelming dispersal trend in either an up-valley or down-valley direction. As well, mineral anomaly values only occur occasionally, and are not concentrated in a preferred valley location (i.e., in upper, middle, or lower valley sites). These findings are however not unexpected. Valleys in the northern Ruby Range exhibit a classical glaciated U-shape with evidence of glaciation on a local scale (small ice cap, cirque, and valley glaciers) with down-valley ice movement, and regional ice penetrating the area in an up-valley direction. Since these glaciological conditions were ongoing since at least the middle Pleistocene (Nisling Ice Sheet; Muller, 1967), and even simultaneously during the Ruby Ice Sheet advance, much of the glacial sediments in the valleys must have been reworked numerous times by up-valley and down-valley glacier flows.

Based upon the heavy mineral data pertaining to composition, assemblages, background, and anomalies, it is evident that at least in the heavy mineral fraction of plateau and valley sediments, similarities do exist. Specifically, the data indicate that:

i) The average proportion of heavy minerals identified in the sediment from plateau and valley locations were similar. Therefore, the sediment from both plateau and valley sites do not vary greatly in heavy mineral species and proportions.

ii) The heavy mineral assemblages derived from plateau and valleys sediments were alike.

iii) Except for minor differences, the background values for most heavy minerals in both plateau and valley samples are quite similar. As well, the anomaly values are also relatively comparable for most minerals.

Furthermore, in comparing the heavy mineral assemblage for the bedrock found within and outside the study area with that of the surficial sediments (plateau and valleys), it was determined that the data do not support a clear, exclusive association between heavy mineral assembleges derived from sediments (plateau and valley samples), and those derived from bedrock sources within or outside the study area. However, assemblages for both plateau and valley samples were not exclusively related to local bedrock sources. Therefore, since sediment from the plateaus and valleys share similar heavy mineral compositional attributes, which are not exclusively related to local bedrock, it follows that at least some of the glacial sediment sampled in the valleys is of similar origin as the surficial sediment on plateau surfaces. Whether in fact the plateau sediments are glaciogenic will be addressed in later discussion.

5.6.6.4 Plateau Dispersal Trends

The plotting of the heavy mineral compositional trends was facilitated by the use of Synagraphic Computer Mapping. The output consisted of an isopleth map where the isolines shown on the map are for specific values and the values are assumed to smoothly vary over the interval between any two adjacent isolines forming a continuous surface. However, since this may have resulted in minimizing the effects of local concentrations, and masking anomalies, manual plotting was also used to ensure that isolines indicated local anomalies.

There is also another disadvantage associated with this plotting method which is associated with the "fitting" and "closing" of isolines. Specifically, in areas where two or more values are in close proximity, and if the required number of isolines representing the surface between the points cannot fit between them, the plot tends to exaggerate the isolines beyond the limits set by the points. Since there is however, a compositional trend in these regions, it is emphasized that the trends depicted on the plots where this occurs is somewhat a false representation, and will be identified as such accordingly. As well, in plotting, all isolines are closed even if data points are not located in these areas.

The ensuing observations are made with reference to the individual heavy mineral compositional trends illustrated by plots, and arrived at with the prior knowledge of the "false" or "exaggerated" patterns that the computer mapping generates. The inferred direction of the predominant trends are described with reference to "head and tail" portions of the dispersals. The "head" refers to the shorter, steep part of the trend, up to the defined area of highest concentration, whereas the "tail" includes the less steep area from the highest concentration to the limit of the plot. Lines characterizing the broad, general dispersal trends appear on the plots. As well, reference is be made to three known mineralizations occurring in the study area, identified at the heads of Alaskite Creek, Rockslide Creek, and on the ridge that divides Rockslide and Talbot Creeks (Muller, 1967; Smith, 1971). (The topographic transparent overlay, provided in the back pocket, should be aligned with the registration points on the individual dispersal plots.)

Orthopyroxene

Figure 13 shows a northern and possibly a western dispersal of orthopyroxene. There are three centres of high concentration (>16%); neither are associated with known mineralization. The northern dispersal trend (tail) from the main centre of highest concentration (>16%), located between Talbot and Alaskite creeks, to the 2% concentration interval is over 16 km. The head of this trend over the same concentration range is less than 2 km. A possible west and northwest dispersal pattern is also evident from the plot but does not appear to be as pronounced as the northern trend.

Garnet

The dispersal trend for garnet (Figure 14) has an area of highest concentration (>4%) in a region which is not associated with known mineralization. The prevailing northwest dispersal trend (tail) is mapped from the area of high concentration to the 0.1% interval, a distance of about 12 km. The head of the trend over the same concentration range is less than 4 km. However, due to the lack of sample sites, much of the head portion is exaggerated.

Epidote

Figure 15 displays the epidote dispersal trend. Two centres of highest concentration (>25%) are evident; one is located on the ridge between Alaskite and Talbot creeks, the second centre is about 8 km to the west of the first. The centres are not associated with known mineralizations. The overall trend of the dispersal appears to be in a northwest direction, centred on the area of high concentration located on the ridge between Alaskite and Talbot creeks. The approximate distance of dispersal of the tail portion is approximately 14 km. The head dispersal is about 6 km, but is largely exaggerated because of lack of sample sites in this area. The second area of high concentration exhibits a western dispersal trend.

Allanite

The dispersal trend for allanite is shown on Figure 16. Two areas of high concentration (>6%) are shown in close proximity to one another. They are both located in the vicinity of a zone of known mineralization at the head of Alaskite Creek. A dominant northwest dispersal has been plotted to the 0.1% interval over approximately 14 km (tail). The head portion of this dispersal over the same concentration range is less than 4 km.

Hematite

The dispersal pattern for hematite (Figure 17) identifies four centres of high concentration in excess of 5%. One centre occurs in the vicinity of the mineralization found at the head of Alaskite Creek. A second centre occupies an area in the vicinity of known mineralization along the ridge that separates Rockslide and Talbot creeks. A third centre is located in the area of mineralization at the head of Rockslide Creek. The general direction of dispersal is difficult to discern but appears to be north to northwest. The tail portion of the dispersal is approximately 20 km. The head dispersal distance is about 4 km.

Goethite

The general direction of the dispersal of goethite (Figure 18) is north and northwest. The two centres of high concentration (>42%) are not associated with known mineralizations. The tail portion of the dispersal from the most easterly centre of high concentration, which is shown as a northwestern trend, is approximately 12 km. The distance of the head portion appears as about 4 km, but has been exaggerated. The northeast bulging of this trend is attributed to an exaggeration due to computer interpolation. Also apparent from the second centre of high concentration is a north and northwest dispersal trend.

Hornblende

The dispersal pattern of hornblende show two areas of high concentration of more than 69% (Figure 19). One of these, at the head of Alaskite Creek, is in the proximity to the mineralization at the head of Alaskite Creek. The other area, to the west, is not associated with known mineralization. The general north to northwest dispersal of the tail portion from the greater than 69% concentration area in Alaskite Creek to the 1% interval is over 12 km. The head portion of this dispersal is about 4 km.

Zircon

The dispersal of zircon (Figure 20) is difficult to discern but does seem to have a northwest to west trend. Two centres of high concentration (>17%) are identified, none of which are associated with known mineralizations. The tail portion of the dispersal from the eastern centre of high concentration to the 1% interval is approximately 12 km. The head portion is about 5 km.

Opaques

A northwest dispersal trend in opaques evident from Figure 21. The area of highest opaque concentration (>87%) in the southeast portion of the dispersal pattern and corresponds to an area of known mineralization at the head of Rockslide Creek. The dispersal distance of the tail portion from this area to the 7% interval is approximately 16 km. The head portion distance over the same concentration range is less than 6 km. The second centre of high concentration is not associated with known mineralization.

As mentioned previously, the descriptions provided above were made with reference to the mineral plots, but were done so with a view to compensate for computer generated "false" patterns. The following is a synopsis of the heavy mineral dispersal trends.

	DISPERSAL TREND DIRECTION	DISPERSAL DISTANCE HEAD	TAIL	PROXIMITY TO KNOWN MINERALIZATIONS
Orthopyroxene	North and West	2 km	16 km	None
Garnet	Northwest	4 km	12 km	None
Epidote	Northwest	6 km	14 km	None
Allanite	Northwest	4 km	14 km	Alaskite Creek
Hematite	Northwest and North	4 km	20 km	1) Alaskite Creek 2) Rockslide Creek/ Talbot Creek Ridge 3) Rockslide Creek
Goethite	Northwest and North	4 km	12 km	None
Hornblende	Northwest and North	4 km	12 km	Alaskite Creek
Zircon	Northwest and West	5 km	12 km	None
Opaques	Northwest	6 km	16 km	Rockslide Creek

Since it was not the intent of this analysis to detail minor variations in dispersal trends, but rather to interpret regional patterns within the study area, it can guardedly be stated that there is a preferred northwest dispersal trend with respect to heavy mineral concentration in the plateau surficial sediments. Futhermore, these trends are continuous, and have been mapped over extensive plateau surfaces that are dissected by numerous valleys. The regional implications of these trends will be further discussed with reference to sediment transport in a later section.

Figure 13
DISPERSAL PATTERN OF ORTHOPYROXENE

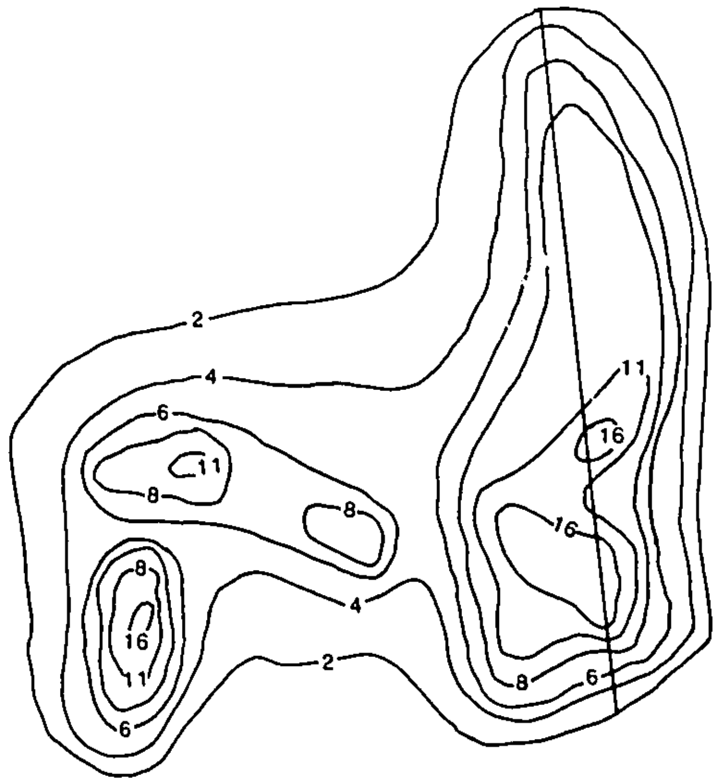


Figure 14
DISPERSAL PATTERN OF GARNET

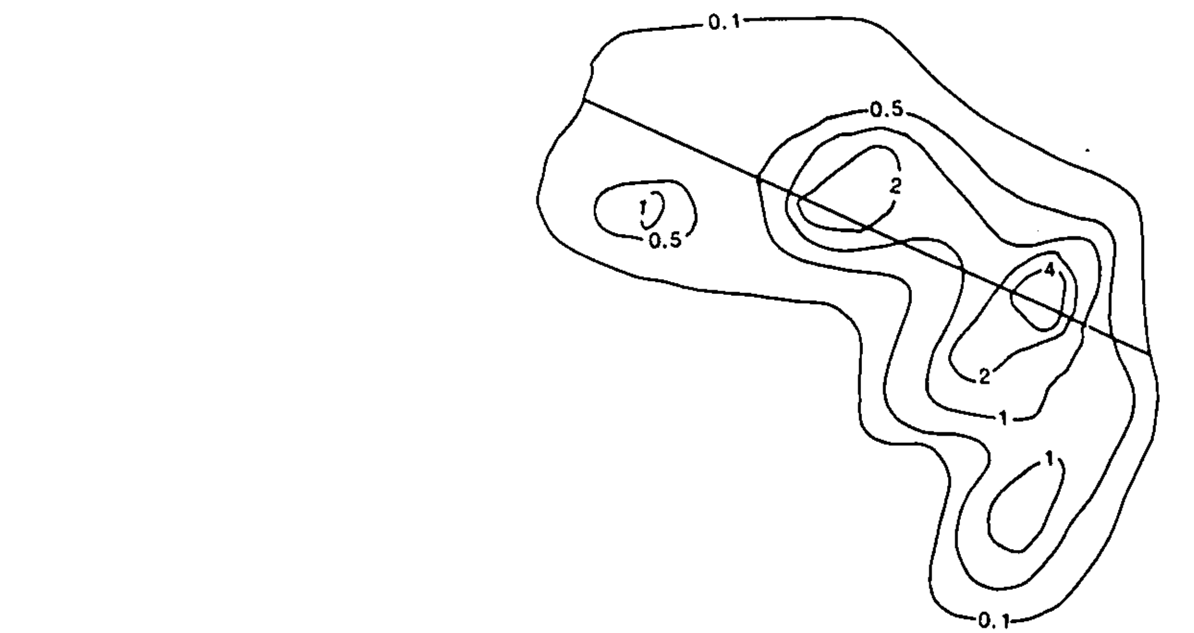


Figure 15
DISPERSAL PATTERN OF EPIDOTE

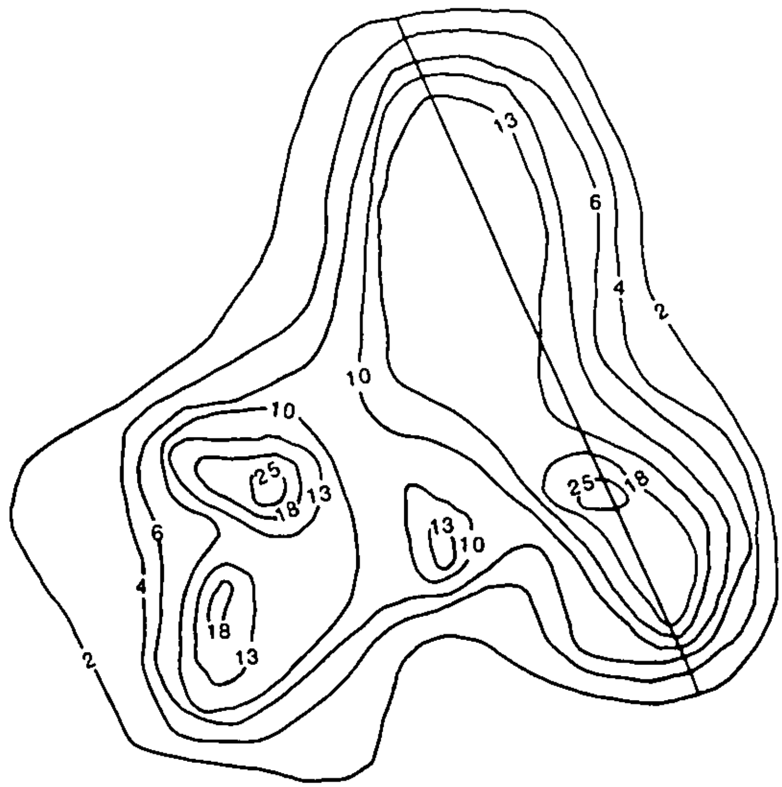


Figure 16
DISPERSAL PATTERN OF ALLANITE

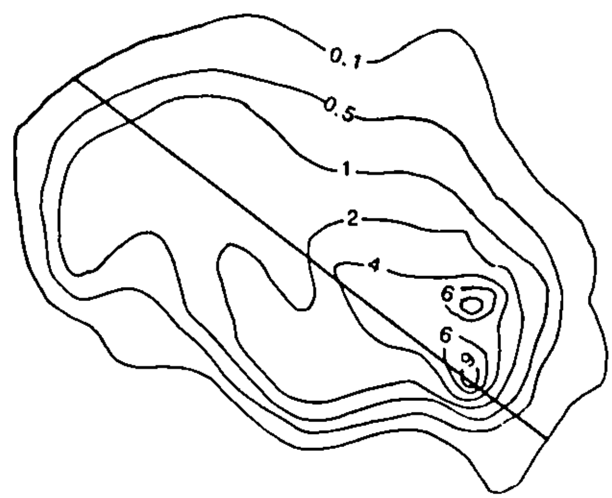


Figure 17
DISPERSAL PATTERN OF HEMATITE

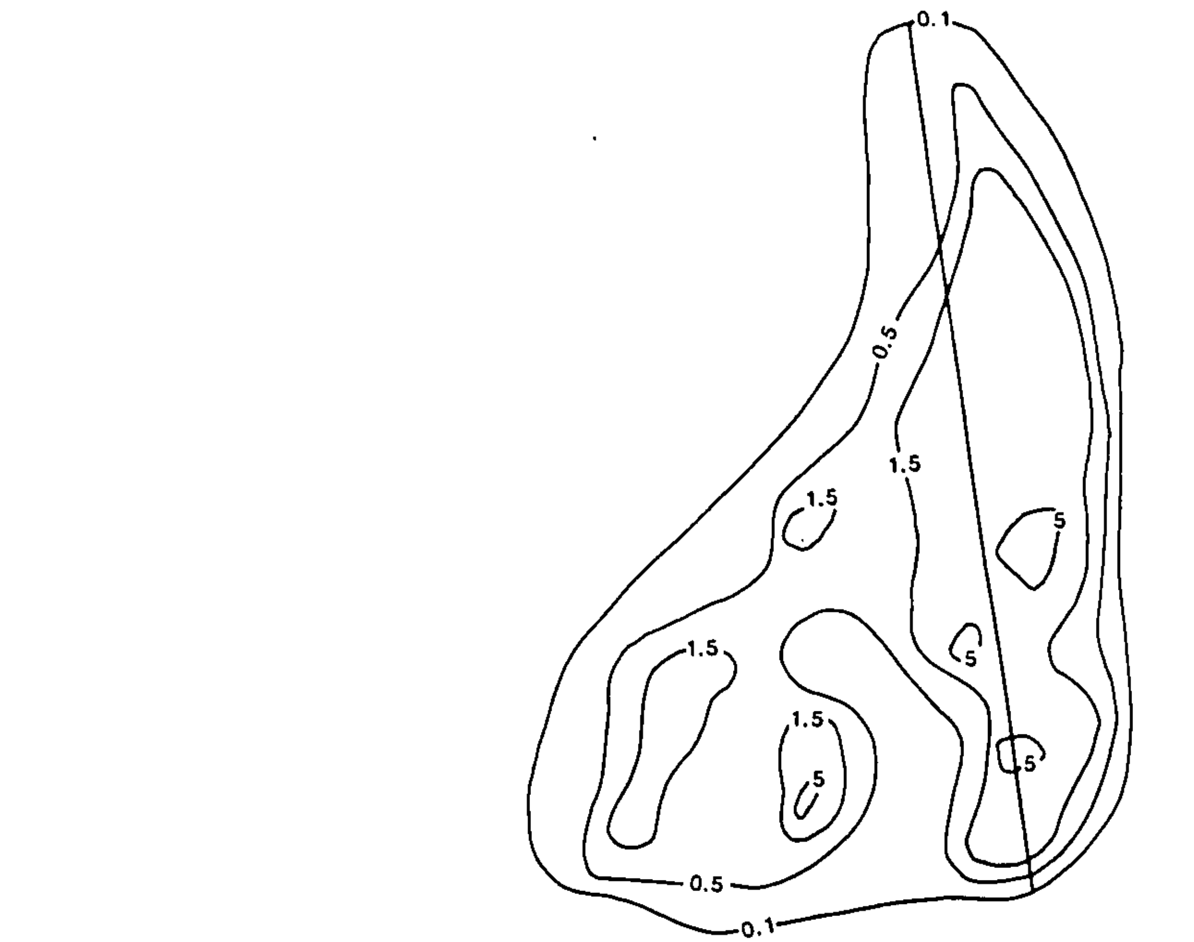


Figure 18
DISPERSAL PATTERN OF GOETHITE

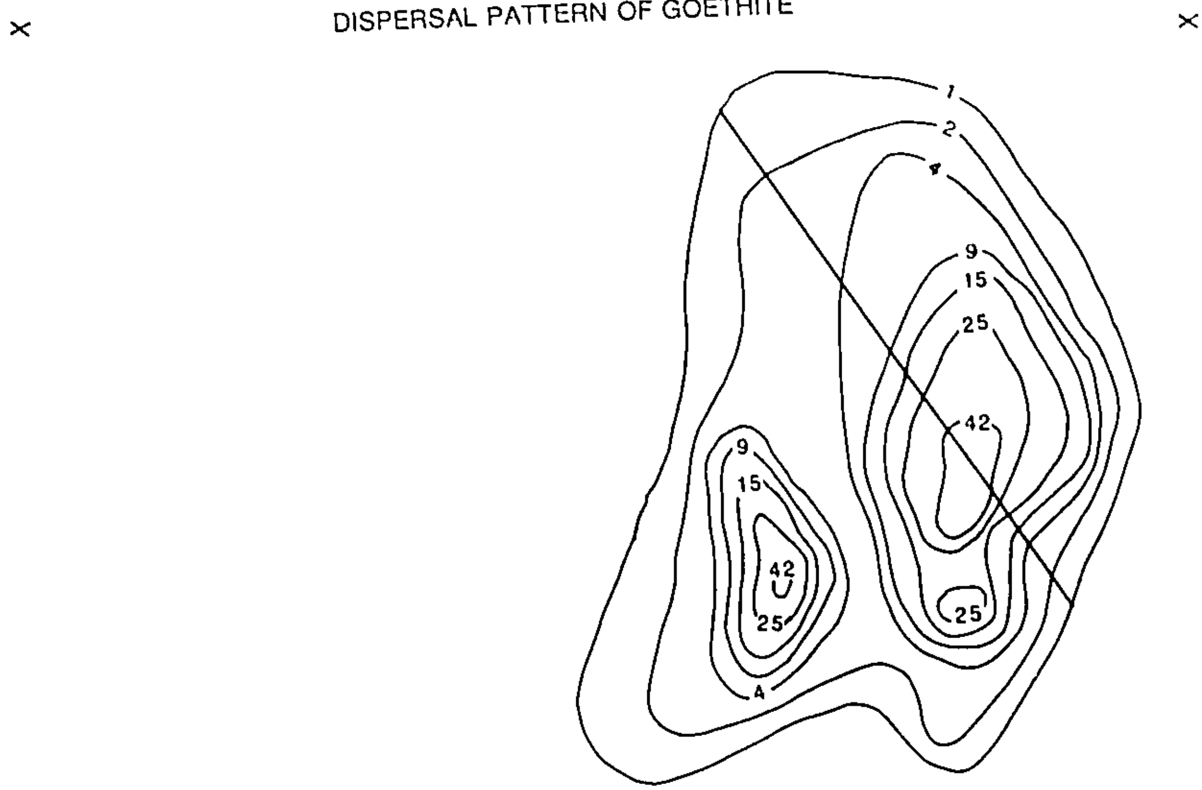


Figure 19
DISPERSAL PATTERN OF HORNBLENDE

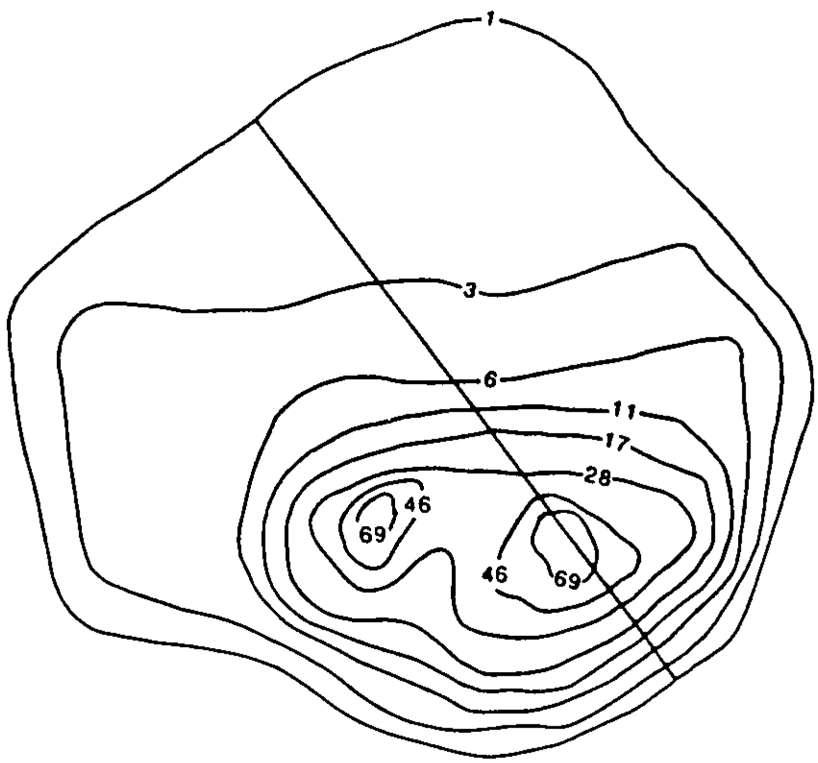


Figure 20
DISPERSAL PATTERN OF ZIRCON

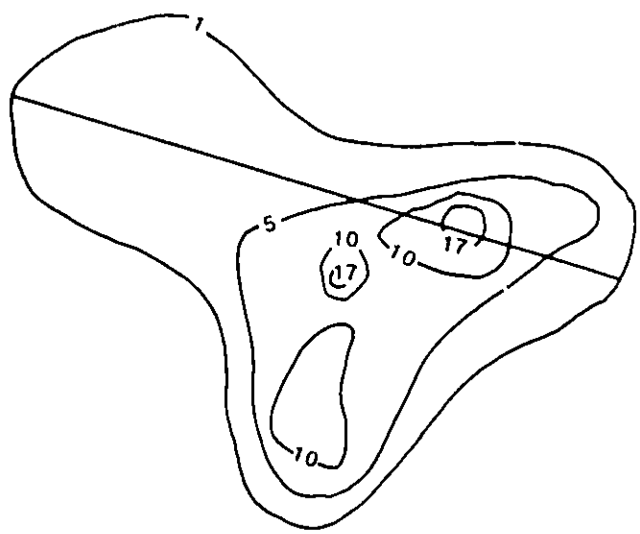
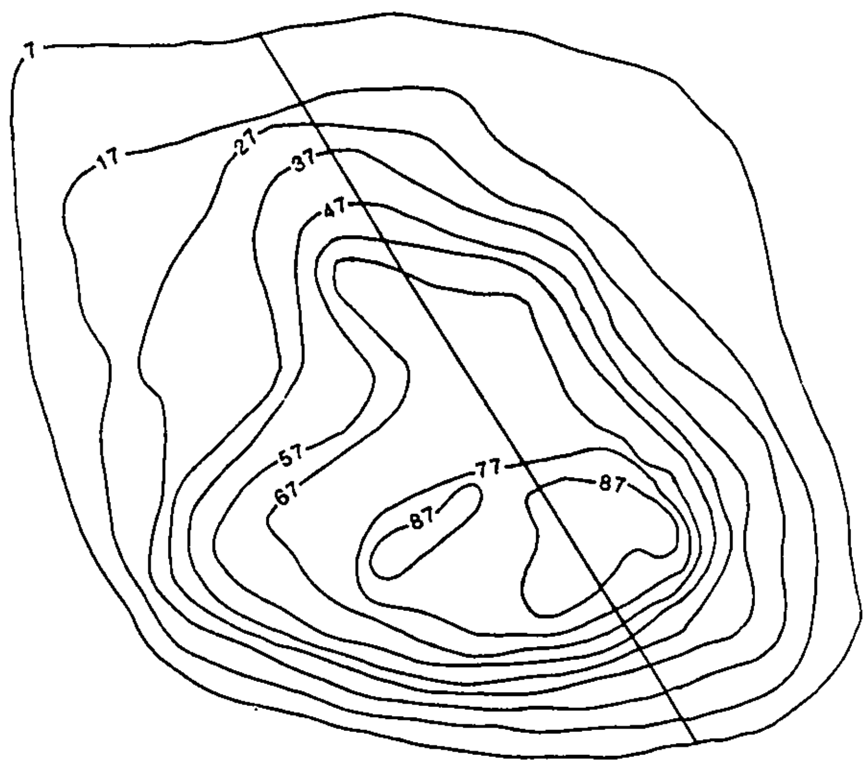


Figure 21
DISPERSAL PATTERN OF OPAQUES



5.7 Silt and Clay Fraction: Geochemical Analysis

As with the heavy mineral analysis, the multi-element data are reported separately under plateau and valley sample sites. A total of 94 plateau, and 53 valley silt and clay fraction samples were submitted for analysis. The trace elements analysed for included: copper (Cu), lead (Pb), molybdenum (Mo), cobalt (Co), nickel (Ni), chromium (Cr), manganese (Mn), cadmium (Cd), silver (Ag), bismuth (Bi), iron (Fe), arsenic (As), zinc (Zn), vanadium (V), tellurium (Te), uranium (U), tungsten (W), antimony (Sb), selenium (Se), tin (Sn), and gold (Au). Table 22 present the results of the multi-element analysis for plateaus and valleys respectively. All values are in ppm except for Fe (%), and Au (ppb). Background and anomaly levels were calculated in the manner described previously (detailed by Hawkes and Webb, 1962).

In order to set the stage for the ensuing discussion on these values, it is again emphasised that since background values are considered estimates which fall within a range for any given area, if variations in background or anomaly values of a mineral was twice as much between the plateau and valley sites, it constituted a meaningful difference. For comparative purposes, the average abundance of elements in granodiorite are also reported.

The background values for sixteen of the twenty-one elements analysed in both plateau and valley sites are relatively similar. These include: Cu, Pb, Mo, Ni, Cr, Mn, Bi, Fe, Zn, V, Te, W, Sb, Se, Sn, and Au. The remaining five elements (Co, Cd, Ag, As, and U), exhibit a distinct difference in background values. In most cases, except for Co, background values for these elements in valleys are reported as zero.

A somewhat different pattern emerges from the anomaly data. Twelve elements (Co, Ni, Cr, Mn, Fe, Zn, V, Te, U, Sb, Se, and Sn) do not show an appreciable difference in anomaly levels. However, nine elements (Cu, Pb, Mo, Cd, Ag, Bi, As, W, and Au) demonstrate considerable differences, with plateau anomaly values in all cases being at least twice the corresponding valley levels. The relatively similar background values for the majority of elements in the sediments from plateau and valley sites probably indicates that there is a relatively uniform concentration of elements, but because of profound differences in nine element anomaly values, clearly, distribution of these elements is far from uniform.

Table 22
MULTI-ELEMENT ANALYSIS
(PLATEAU SAMPLE SITES)

Sample	Cu	Pb	Mo	Co	Ni	Cr	Mn	Cd	Ag	Bi	Fe	As	Zn	V	Te	U	W	Sb	Se	Sn	Au
82-01P	150	320	17	0	8	84	6277	6.3	0.0	0	40.0	128	793	526	0	18	0	21	0	102	3580
82-03P	33	158	2	93	46	96	2226	0.0	0.0	0	39.4	0	624	1640	0	0	0	0	0	0	5
82-04P	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•
82-05P	31	149	3	90	45	88	2151	0.0	0.0	0	38.5	0	560	1640	0	0	0	0	0	0	2
82-06P	19	129	4	56	32	59	1594	0.0	0.0	0	25.8	0	336	928	0	0	0	0	0	0	•
82-07P	113	318	79	0	39	76	6622	0.0	0.5	0	50.2	13	1498	233	0	46	0	0	0	122	•
82-12P	75	524	30	32	50	192	5560	2.9	0.0	0	41.4	33	944	923	0	17	0	7	0	107	•
82-13P	45	332	33	28	61	253	5200	0.0	0.0	0	41.9	12	841	1071	0	15	0	5	0	39	•
82-14P	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	1100
82-15P	24	238	109	0	23	50	6929	0.0	0.0	0	44.8	0	1663	86	0	39	0	0	0	97	5780
82-16P	50	234	68	0	39	97	6352	0.0	0.0	0	43.2	0	908	250	0	39	0	9	0	133	1370
82-17P	218	230	27	19	73	229	5398	16.2	0.8	4	23.9	125	1032	608	0	13	900	10	0	132	192
82-18P	238	214	32	5	65	224	6086	15.4	0.7	29	20.2	126	1055	549	0	15	0	0	0	51	830
82-21P	155	389	28	0	45	134	7690	30.3	0.0	21	21.1	6	940	346	0	48	0	0	0	174	1920
82-23P	82	160	10	91	47	103	2521	0.0	0.0	0	33.4	18	579	1260	0	0	12	0	6	0	1790
82-32P	92	121	4	0	28	201	5360	2.6	0.0	0	16.8	10	793	165	0	0	28	6	5	46	3
82-33P	84	203	39	0	24	69	6210	1.7	0.0	0	29.4	9	1169	124	0	18	50	0	6	111	50
82-36P	391	12960	176	9	16	31	4033	70.7	13.2	520	22.5	2000	3438	63	0	73	40	18	9	130	50
82-37P	113	586	24	34	40	120	3406	13.8	1.0	0	30.7	223	679	977	0	17	220	0	0	53	•
82-39P	114	145	4	67	23	74	1939	0.0	0.0	0	31.0	19	401	113	0	0	0	0	0	0	•
82-40P	70	395	5	60	58	74	2858	0.0	1.2	0	42.0	0	610	583	0	11	0	0	0	0	•
82-42P	48	199	12	0	14	97	8557	2.7	0.0	0	31.5	0	770	464	0	13	0	0	0	16	•
82-43P	90	232	13	0	37	152	5831	3.4	0.7	0	24.7	17	694	469	0	15	0	0	0	149	•
82-44P	43	470	8	0	136	275	6267	28.0	0.7	0	19.6	96	489	593	0	17	0	0	0	318	•
82-45P	29	69	4	0	16	74	1922	0.0	0.0	0	6.3	7	121	137	0	13	40	0	0	19	•
82-50P	114	238	7	67	57	156	3455	8.6	0.0	0	38.8	9	581	1421	0	0	0	0	0	13	•
82-55P	196	192	2	24	51	175	3657	11.0	0.0	0	31.5	11	509	1036	0	0	0	38	0	32	•
82-57P	146	324	15	0	33	287	6385	17.0	0.6	0	21.7	33	595	556	0	26	0	0	0	74	•
82-60P	55	179	5	87	43	135	3608	0.0	0.6	0	39.8	0	613	1420	0	0	0	0	0	0	•
82-62P	39	183	10	0	1	41	6300	1.6	0.5	0	26.7	0	1011	117	0	0	0	0	0	18	50
82-63P	71	98	8	16	37	238	8346	0.0	0.0	0	26.6	0	348	932	0	0	0	0	0	14	•
82-65P	33	105	7	0	27	57	6470	1.5	0.0	0	20.5	0	455	184	0	0	0	0	0	14	50
82-66P	100	192	8	118	50	90	4905	0.0	5.0	0	31.7	1724	904	140	0	11	0	48	0	0	50
82-67P	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	297
82-69P	103	131	12	0	38	249	8340	2.4	0.0	0	18.2	116	347	454	0	34	0	0	0	47	•
82-70P	263	118	8	0	8	118	6520	2.2	0.0	0	16.4	0	376	403	0	0	0	0	0	64	•
82-72P	123	125	3	6	90	317	4439	0.0	0.0	0	16.5	51	271	424	0	17	0	0	0	0	•
82-73P	122	352	4	0	73	177	7611	3.2	0.0	0	14.2	8	155	154	0	0	0	0	0	15	•
82-74P	79	91	4	27	22	87	3934	0.0	0.0	0	19.0	10	541	384	0	0	0	0	0	0	•
82-75P	51	406	8	7	37	133	5048	5.6	0.0	0	23.2	34	667	381	0	12	0	0	0	0	52
82-77P	45	150	6	4	48	149	6011	0.0	0.0	0	15.8	0	120	328	0	0	0	0	0	50	•
82-84P	46	59	7	4	28	80	5435	1.0	0.0	0	15.2	30	271	304	0	0	0	10	0	34	•
82-85P	31	43	6	6	24	174	8830	1.3	0.0	0	22.2	9	388	302	0	0	18	0	0	12	10
82-86P	41	45	6	5	35	189	6931	1.1	0.0	0	17.3	5	405	305	0	0	0	0	0	25	•
82-88P	72	59	5	18	26	138	6891	1.1	0.0	0	23.1	0	424	558	0	0	0	0	0	12	50
82-90P	31	36	4	41	122	365	4568	1.1	0.0	0	15.5	0	290	271	0	0	0	0	0	13	10
82-91P	78	65	6	27	40	153	4845	0.0	0.0	0	14.6	0	330	0	0	0	0	0	0	44	•
83-01P	345	458	18	0	148	252	8590	46.8	55.6	0	22.6	42	818	502	74	34	150	0	0	120	•
83-02P	399	357	4	0	166	206	7933	27.9	17.5	0	22.1	0	814	397	67	38	113	0	13	213	•
83-03P	585	445	3	73	462	192	4120	3.9	53.0	0	49.2	61	719	1451	50	33	125	9	9	35	•
83-05P	214	310	2	6	176	408	7894	3.0	19.4	0	26.2	168	434	646	58	33	65	0	16	42	•
83-07P	58	415	0	119	63	141	3014	1.4	13.1	0	54.0	0	642	2059	55	0	0	0	7	0	5
83-08P	87	204	0	86	54	91	2201	0.0	0.7	0	26.9	0	407	1104	20	0	0	0	0	0	2
83-09P	227	459	2	85	59	122	3164	2.6	1.1	0	35.2	26	787	1339	40	17	90	0	7	36	4

Table 22
(PLATEAU SAMPLE SITES)
(continued)

Sample	Cu	Pb	Mo	Co	Ni	Cr	Mn	Cd	Ag	Bi	Fe	As	Zn	V	Te	U	W	Sb	Se	Sn	Au
83-26P	146	276	3	0	27	118	8690	6.6	24.8	0	26.8	32	865	233	59	11	200	0	5	77	*
83-27P	199	316	4	0	18	117	8400	5.5	17.8	0	22.1	36	857	255	72	12	225	0	0	81	*
83-30P	118	368	2	108	50	128	3006	2.0	1.0	0	47.2	0	648	1808	38	0	0	0	0	18	*
83-31P	94	352	4	94	48	124	2964	0.0	0.8	0	41.8	0	546	1572	50	13	0	6	0	14	*
83-32P	290	440	2	98	54	112	2870	0.0	10.8	0	51.2	56	682	1756	70	31	12	16	0	12	*
83-33P	392	3464	730	58	54	110	3278	16.4	2.6	0	46.0	42	694	1688	78	43	0	0	0	62	*
83-34P	580	2752	610	52	48	100	3008	18.0	11.4	0	47.2	54	730	1540	64	35	160	0	0	30	*
83-37P	456	2814	610	44	110	112	3436	18.4	32.2	0	46.6	106	750	1572	88	66	0	0	0	42	*
83-38P	368	5042	1440	4	50	96	3086	47.2	4.8	0	40.6	148	620	1360	82	38	0	0	0	30	*
83-40P	454	7270	1500	14	51	112	3398	46.9	15.4	0	30.8	104	693	1444	91	29	0	0	7	19	*
83-42P	248	2348	400	84	118	118	2964	16.2	8.2	9	44.6	48	648	1810	70	12	80	0	0	24	*
83-43P	324	664	30	87	44	110	2745	25.3	2.8	0	40.3	11	576	1549	53	0	24	0	0	20	33
83-46P	1710	2135	260	0	40	123	3823	35.4	18.4	46	32.6	688	1124	1247	96	68	550	0	0	123	*
83-55P	8	611	8	92	43	107	2428	1.2	4.4	0	36.8	0	557	1587	49	18	32	0	0	10	178
83-56P	3200	2338	1590	0	1	96	4788	40.2	22.8	630	36.2	596	668	1276	96	33	2000	0	10	262	*
83-60P	426	480	42	86	40	86	2236	1.8	2.2	0	36.2	38	496	1444	52	0	100	20	26	20	*
83-61P	810	692	73	19	49	104	3650	13.6	14.4	190	28.8	100	697	1495	34	12	2000	0	0	57	28
83-62P	65	179	4	98	51	107	2534	0.0	1.0	0	34.9	0	621	1674	27	0	28	10	0	0	7
83-64P	93	186	7	90	53	122	2756	1.1	24.3	0	37.8	0	621	1584	32	0	50	0	0	0	9
83-70P	140	260	16	110	52	118	2778	1.0	3.2	0	41.0	0	676	1842	28	0	220	12	0	0	*
83-71P	494	228	14	0	46	172	8210	7.4	0.8	0	23.0	8	836	290	38	36	150	0	0	104	*
83-72P	406	230	6	0	22	248	9130	14.8	36.0	0	23.0	10	898	290	38	12	180	0	0	98	*
83-73P	217	185	6	0	20	124	5278	7.3	5.2	0	20.2	0	514	792	39	15	20	0	0	58	140
83-75P	13	105	3	0	4	20	9875	3.1	0.8	0	24.1	0	1137	121	21	0	50	0	0	34	2
83-76P	71	151	5	0	26	181	8304	6.5	18.3	0	26.3	0	852	301	17	13	280	0	0	46	5
83-77P	276	416	18	0	48	100	6492	18.4	61.0	0	24.4	0	768	274	0	18	200	0	0	102	*
83-78P	200	356	50	0	108	386	6892	5.2	42.6	0	31.0	0	610	620	20	58	1250	0	0	110	*
83-79P	378	342	32	0	112	422	6652	5.6	19.0	0	23.8	8	560	550	0	29	750	0	0	72	*
83-81P	69	192	4	0	13	79	9435	10.3	7.5	0	26.1	0	1125	299	34	11	40	0	0	74	3
83-82P	299	227	6	0	38	90	8663	15.5	21.7	0	27.3	0	882	337	50	19	180	0	0	83	*
83-84P	125	213	3	87	47	128	2930	1.2	0.6	0	32.8	0	626	1471	32	0	0	0	0	0	*
83-85P	153	263	9	0	57	89	5958	20.4	15.7	25	25.3	0	900	923	48	26	600	0	0	51	*
84-15P	41	305	8	3	14	206	4772	0.0	0.0	0	34.8	0	238	985	0	0	0	6	0	46	10
84-16P	27	303	8	5	15	200	4334	0.0	0.0	0	35.9	0	278	1080	0	40	0	0	0	45	10
84-17P	68	341	7	26	23	204	4423	0.0	0.0	0	37.2	0	351	1130	0	0	0	14	10	59	10
84-19P	86	298	6	0	19	205	3838	1.3	0.0	0	36.7	0	270	831	0	14	0	0	0	48	10
84-20P	36	283	9	7	26	243	5021	0.0	0.0	0	37.8	0	278	993	0	20	0	0	0	32	194
85-06P	0	340	7	19	36	312	3100	0.0	0.0	0	45.0	0	359	1114	0	33	0	0	0	16	10
85-07P	0	414	4	26	34	278	4152	0.0	0.0	0	47.2	0	353	1320	0	12	18	14	0	6	4
85-08P	6	341	37	0	67	139	8850	0.0	0.0	0	44.2	0	948	609	0	10	24	7	0	77	10
85-09P	536	759	10	326	291	133	2755	1.8	0.6	0	48.8	616	426	591	0	0	0	24	29	90	13
85-10P	10	374	7	30	53	277	3522	0.0	0.0	0	35.5	0	253	1030	0	13	0	14	0	49	10
85-11P	107	304	6	20	62	233	9850	0.0	0.0	0	50.0	0	253	899	0	25	55	11	0	18	10

Table 22
(VALLEY SAMPLE SITES)
(continued)

Sample	Cu	Pb	Mo	Co	Ni	Cr	Mn	Cd	Ag	Bi	Fe	As	Zn	V	Te	U	W	Sb	Se	Sn	Au
82-08V	35	258	48	0	8	53	8497	2.9	0.0	0	38.7	0	1308	269	0	18	28	0	0	103	161
82-09V	47	320	58	0	12	63	6827	1.3	0.0	0	39.4	0	1349	254	0	18	0	0	0	118	5065
82-10V	62	282	43	0	27	108	9573	2.7	0.0	0	37.8	0	1092	361	0	12	0	0	0	141	110
82-11V	59	244	37	0	12	75	7450	6.6	0.0	0	37.8	0	945	424	0	18	90	0	0	98	•
82-24V	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	173
82-25V	232	1191	12	74	41	101	4619	12.9	2.9	0	42.6	400	1208	1441	0	102	0	9	0	27	•
82-26V	63	156	7	71	39	90	2375	0.0	0.0	0	34.0	6	483	1173	0	0	0	0	0	0	•
82-27V	93	196	11	81	64	156	3415	0.0	0.0	0	34.3	30	594	1220	0	12	260	15	0	12	•
82-28V	110	223	20	87	75	176	3715	0.0	0.5	0	45.4	29	675	1815	0	0	0	32	0	0	•
82-29V	63	169	23	0	30	124	6284	2.3	0.0	0	24.2	37	663	322	0	48	0	0	0	110	•
82-30V	103	157	7	20	36	76	3781	3.6	0.0	0	13.9	26	724	280	0	0	0	0	0	42	•
82-34V	27	65	4	65	74	110	2060	0.0	0.0	0	12.6	0	220	540	0	0	0	0	0	0	2
82-35V	70	197	35	0	53	166	5924	1.2	0.0	0	28.8	23	1378	243	0	16	55	0	7	108	•
82-38V	98	215	6	65	35	85	2156	1.5	0.0	0	30.8	43	469	1084	0	0	0	0	0	0	•
82-53V	73	329	5	0	15	128	6123	14.3	0.0	0	18.8	28	338	470	0	26	0	0	0	320	•
82-76V	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	50
82-78V	280	91	5	0	35	164	7081	2.0	0.0	0	15.9	15	324	234	0	0	0	0	0	30	•
82-79V	33	36	8	0	10	52	8225	1.4	0.0	0	24.8	0	513	395	0	0	0	0	0	10	50
82-80V	250	80	5	0	101	114	7770	2.5	0.0	0	17.1	0	348	56	0	0	0	0	0	0	•
82-81V	70	67	6	71	45	90	2485	0.0	0.0	0	32.8	0	430	1104	0	0	0	0	0	0	•
82-82V	59	107	9	2	29	105	6281	1.0	0.7	0	16.7	7	361	294	0	0	0	5	0	53	•
82-83V	32	60	4	93	45	85	2308	0.0	0.0	0	39.8	0	500	1620	0	0	0	0	0	0	50
82-87V	24	51	7	0	13	71	7831	1.3	0.0	0	18.6	0	485	249	0	0	0	0	0	18	50
82-89V	102	66	7	12	36	190	7821	1.8	0.0	0	16.5	0	331	359	0	0	0	0	0	22	186
82-92V	38	49	7	22	44	245	7548	0.0	0.0	0	17.4	0	330	278	0	0	0	5	0	17	50
82-93V	114	194	5	21	41	182	5492	2.7	0.0	0	16.1	10	639	273	0	0	0	13	0	19	•
83-16V	155	278	8	88	118	138	3234	1.6	0.7	0	42.8	0	640	1438	31	0	40	0	0	11	13
83-17V	72	274	0	62	50	92	4265	0.0	0.6	0	35.2	0	715	1200	40	0	0	0	7	13	4
83-18V	151	250	3	8	23	103	5867	9.0	2.1	0	22.3	0	781	407	38	15	120	9	0	63	•
83-19V	110	372	14	0	50	170	8064	13.0	21.8	0	33.8	50	838	418	68	32	140	0	0	96	•
83-20V	342	404	16	0	54	186	7060	21.0	83.0	0	35.4	88	856	378	68	35	120	6	8	80	•
83-21V	207	360	5	0	17	99	8197	11.2	6.5	0	36.4	34	962	570	68	0	240	0	0	84	6
83-22V	60	328	1	26	19	70	6721	4.3	0.5	0	28.3	11	837	732	50	0	70	5	0	45	3
83-23V	66	262	2	0	9	66	8000	8.7	0.0	0	21.9	11	1020	209	41	11	60	0	0	78	4
83-24V	85	281	6	0	6	67	7730	9.0	0.9	0	20.8	33	1032	211	56	0	200	7	0	82	5
83-25V	105	354	1	67	35	93	4411	1.9	4.2	0	31.2	0	675	1306	48	0	320	0	5	24	4
84-01V	160	180	5	85	50	125	2990	0.0	0.0	0	41.1	0	540	1500	0	0	0	0	0	35	•
84-02V	204	262	5	81	46	181	5500	0.0	0.0	0	42.5	0	675	1780	0	0	0	13	0	18	10
84-03V	86	135	27	0	17	143	7383	0.0	0.0	0	16.8	0	201	579	0	16	32	0	6	119	10
84-04V	278	198	9	21	19	130	3931	0.0	0.0	0	37.6	0	426	894	0	0	0	11	6	51	10
84-05V	54	98	0	35	17	65	2010	0.0	0.0	0	23.3	0	249	660	0	80	0	0	0	26	•
84-08V	33	216	3	107	50	146	2874	0.0	0.0	0	39.7	0	684	1940	0	0	0	13	7	0	3
84-09V	74	310	5	19	16	207	3404	0.0	0.0	0	47.6	0	271	1040	0	15	0	10	10	20	10
84-13V	29	260	5	87	36	171	3320	0.0	0.0	0	44.2	0	627	1870	0	0	0	0	0	0	10
84-21V	342	295	7	9	86	257	3806	0.0	0.0	0	39.4	0	416	994	0	13	55	0	0	64	10
84-23V	14	276	7	3	14	197	4182	0.0	0.0	0	38.9	0	248	991	0	18	16	0	0	32	10
84-24V	35	290	6	4	15	208	4082	0.0	0.0	0	44.5	0	239	952	0	38	0	0	0	36	10
84-25V	21	360	8	0	6	182	3460	0.0	0.0	0	46.3	0	250	1088	0	21	0	7	5	46	10
84-26V	36	294	7	0	23	144	3622	0.0	0.0	0	38.4	0	191	897	0	26	0	0	0	74	•
84-27V	28	436	5	21	22	269	3091	0.0	0.0	0	48.8	0	248	1400	0	0	0	13	0	16	10
85-01V	32	317	4	27	71	279	9500	0.0	0.0	0	35.4	0	272	1024	0	12	16	0	0	26	4
85-02V	25	294	8	24	72	250	8040	0.0	0.0	0	40.1	0	257	998	0	15	45	11	12	30	5
85-03V	34	255	6	24	81	220	8372	0.0	0.0	0	36.8	0	286	813	0	14	70	0	0	20	4
85-04V	25	169	6	37	67	258	7669	0.0	0.0	0	32.7	0	245	555	0	15	40	6	0	18	3
85-05V	0	326	7	20	48	268	6373	0.0	0.0	0	44.4	0	239	1121	0	16	90	0	0	0	5

5.7.1 Valley Dispersal Trends

The same restrictions which prohibited the plotting of mineral dispersal trends in valleys applies also applies to the geochemical data. Table 23 provides a listing of element concentrations in the valleys of the northern Ruby Range. The data in Table 23 is arranged from an upper to lower valley order.

Talbot Creek

None of the element dispersal concentrations in Talbot Creek demonstrate a distinct preference for increasing or decreasing in either an up-valley or down-valley direction. Three elements, Bi, Te, and Se are not detected in any of the samples. Gold is only detected in an upper valley location. The highest concentration of Cu, Pb, Cd, Ag, As, and U are also detected in upper valley locations. The only elements to record concentrations equalling or exceeding anomaly levels are Pb, As, V, U, W, and Sb; all occurring in upper valley locations.

Alaskite Creek

Most of the elements analysed for show no preferred concentration trend. The only element that appears to be demonstrating a trend is gold, which is increasing in concentration down-valley. Concentrations for Ag, Bi, Te, and Sb were not detected in any samples. Anomaly levels for Mo, Mn, Sn, and Au are observed in down-valley locations; anomaly levels for zinc occur in both upper and lower-valley locations.

Rockslide Creek

Samples in Rockslide Creek demonstrate no concentration trends in either a down or up-valley direction. Bismuth (Bi) has not been detected in any sample. Anomaly values are detected for Cu, Ni, Cd, Ag, As, Te, and W. All of these occurrences are generally located midway down the valley.

Pass Creek

There are insufficient data to demonstrate preferred concentration trends. Anomaly levels are recorded for Cd and Sn in an upper-valley location.

Ice Valley

None of the element dispersal concentrations in Ice Valley demonstrate a distinct preference for increasing or decreasing in either an up-valley or down-valley direction. Five elements (Cd, Ag, Bi, As, and Te) are not detected in any of the samples. Gold (Au) exhibits a consistent concentration down the entire valley (10 ppb). An anomaly for Cu is detected in an upper-valley site, and anomalous values for Pb, Cr, and Fe are observed in middle-valley locations.

Table 23
MULTI-ELEMENT ANALYSIS IN VALLEYS
(UPPER TO LOWER VALLEY LOCATIONS)

Sample	Cu	Pb	Mo	Co	Ni	Cr	Mn	Cd	Ag	Bi	Fe	As	Zn	V	Te	U	W	Sb	Se	Sn	Au
TALBOT CREEK																					
82-24V	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	173
82-25V	232	1191	12	74	41	101	4619	12.9	2.9	0	42.6	400	1208	1441	0	102	0	9	0	27	•
82-26V	63	156	7	71	39	90	2375	0.0	0.0	0	34.0	6	483	1173	0	0	0	0	0	0	•
82-27V	93	196	11	81	64	156	3415	0.0	0.0	0	34.3	30	594	1220	0	12	260	15	0	12	•
82-28V	110	223	20	67	75	176	3715	0.0	0.5	0	45.4	29	675	1815	0	0	0	32	0	0	•
83-04V	INSUFFICIENT SAMPLE																				
82-29V	63	169	23	0	30	124	6284	2.3	0.0	0	24.2	37	663	322	0	48	0	0	0	110	•
82-49V	INSUFFICIENT SAMPLE																				
82-48V	INSUFFICIENT SAMPLE																				
82-46V	INSUFFICIENT SAMPLE																				
83-06V	INSUFFICIENT SAMPLE																				
82-30V	103	157	7	20	38	76	3781	3.6	0.0	0	13.9	26	724	280	0	0	0	0	0	42	•
82-41V	INSUFFICIENT SAMPLE																				
ALASKITE CREEK																					
83-36V	INSUFFICIENT SAMPLE																				
82-38V	98	215	6	65	35	85	2156	1.5	0.0	0	30.8	43	489	1084	0	0	0	0	0	0	•
82-35V	70	197	35	0	53	166	5924	1.2	0.0	0	28.8	23	1378	243	0	16	55	0	7	108	•
82-34V	27	65	4	65	74	110	2060	0.0	0.0	0	12.6	0	220	540	0	0	0	0	0	0	2
82-11V	59	244	37	0	12	75	7450	6.6	0.0	0	37.8	0	945	424	0	18	90	0	0	98	•
82-10V	62	282	43	0	27	108	9573	2.7	0.0	0	37.8	0	1092	361	0	12	0	0	0	141	110
82-08V	35	258	48	0	8	53	8497	2.9	0.0	0	38.7	0	1308	269	0	18	28	0	0	103	161
82-09V	47	320	58	0	12	63	6827	1.3	0.0	0	39.4	0	1349	254	0	18	0	0	0	118	5065
ROCKSLIDE CREEK																					
83-44V	INSUFFICIENT SAMPLE																				
83-49V	INSUFFICIENT SAMPLE																				
83-63V	INSUFFICIENT SAMPLE																				
83-50V	INSUFFICIENT SAMPLE																				
83-65V	INSUFFICIENT SAMPLE																				
83-66V	INSUFFICIENT SAMPLE																				
83-23V	66	262	2	0	9	66	8000	8.7	0.0	0	21.9	11	1020	209	41	11	60	0	0	78	4
83-22V	60	328	1	26	19	70	6721	4.3	0.5	0	28.3	11	837	732	50	0	70	6	0	45	3
83-24V	85	261	6	0	6	67	7730	9.0	0.9	0	20.8	33	1032	211	56	0	200	7	0	82	5
83-80V	INSUFFICIENT SAMPLE																				
83-83V	INSUFFICIENT SAMPLE																				
83-21V	207	360	5	0	17	99	8197	11.2	6.5	0	36.4	34	962	570	68	0	240	0	0	84	6
83-20V	342	404	16	0	54	186	7060	21.0	83.0	0	35.4	68	856	376	66	35	120	6	8	80	•
83-25V	105	354	1	67	35	93	4411	1.9	4.2	0	31.2	0	675	1305	48	0	320	0	5	24	4
83-19V	110	372	14	0	50	170	8064	13.0	21.8	0	33.8	50	838	418	68	32	140	0	0	96	•
83-16V	155	278	8	88	118	138	3234	1.6	0.7	0	42.8	0	640	1436	31	0	40	0	0	11	13
83-17V	72	274	0	62	50	92	4265	0.0	0.6	0	35.2	0	715	1200	40	0	0	0	7	13	4
83-18V	151	250	3	8	23	103	5867	9.0	2.1	0	22.3	0	781	407	38	15	120	9	0	63	•
83-10V	INSUFFICIENT SAMPLE																				
83-12V	INSUFFICIENT SAMPLE																				
83-15V	INSUFFICIENT SAMPLE																				

Table 23 (continued)

Sample	Cu	Pb	Mo	Co	Ni	Cr	Mn	Cd	Ag	Bi	Fe	As	Zn	V	Te	U	W	Sb	Se	Sn	Au
PASS CREEK																					
82-59V	INSUFFICIENT SAMPLE																				
82-53V	73	329	5	0	15	128	6123	14.3	0.0	0	18.8	28	338	470	0	26	0	0	0	320	•
82-54V	INSUFFICIENT SAMPLE																				
82-56V	INSUFFICIENT SAMPLE																				
82-58V	INSUFFICIENT SAMPLE																				
82-52V	INSUFFICIENT SAMPLE																				
ICE VALLEY																					
84-24V	35	290	6	4	15	208	4082	0.0	0.0	0	44.5	0	239	952	0	38	0	0	0	36	10
84-21V	342	295	7	9	86	257	3808	0.0	0.0	0	39.4	0	416	994	0	13	55	0	0	64	10
84-25V	21	360	8	0	6	182	3460	0.0	0.0	0	46.3	0	250	1088	0	21	0	7	5	48	10
84-26V	36	294	7	0	23	144	3622	0.0	0.0	0	38.4	0	191	897	0	26	0	0	0	74	•
84-27V	28	436	5	21	22	269	3091	0.0	0.0	0	48.8	0	248	1460	0	0	0	13	0	18	10
84-O5V	54	98	0	35	17	65	2010	0.0	0.0	0	23.3	0	249	660	0	80	0	0	0	26	•
84-23V	14	276	7	3	14	197	4182	0.0	0.0	0	38.9	0	248	991	0	18	16	0	0	32	10
84-O4V	278	198	9	21	19	130	3931	0.0	0.0	0	37.6	0	426	894	0	0	0	11	6	51	10
84-O3V	86	135	27	0	17	143	7383	0.0	0.0	0	16.8	0	201	579	0	16	32	0	6	119	10
84-O2V	204	262	5	81	46	181	5500	0.0	0.0	0	42.5	0	675	1760	0	0	0	13	0	18	10
84-O1V	160	180	5	85	50	125	2990	0.0	0.0	0	41.1	0	540	1500	0	0	0	0	0	35	•
MORAINE VALLEY																					
82-76V	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	50
82-83V	32	60	4	93	45	85	2308	0.0	0.0	0	39.8	0	500	1620	0	0	0	0	0	0	50
82-78V	280	91	5	0	35	164	7081	2.0	0.0	0	15.9	15	324	234	0	0	0	0	0	30	•
82-79V	33	36	8	0	10	52	8225	1.4	0.0	0	24.8	0	513	395	0	0	0	0	0	10	50
82-89V	102	66	7	12	36	190	7821	1.8	0.0	0	16.5	0	331	359	0	0	0	0	0	22	186
82-82V	59	107	9	2	29	105	6281	1.0	0.7	0	16.7	7	361	294	0	0	0	5	0	53	•
82-87V	24	51	7	0	13	71	7831	1.3	0.0	0	18.6	0	485	249	0	0	0	0	0	18	50
82-99V	INSUFFICIENT SAMPLE																				
82-71V	INSUFFICIENT SAMPLE																				
82-64V	INSUFFICIENT SAMPLE																				
82-30V	250	80	5	0	101	114	7770	2.5	0.0	0	17.1	0	348	56	0	0	0	0	0	0	•
82-81V	70	67	6	71	45	90	2485	0.0	0.0	0	32.8	0	430	1104	0	0	0	0	0	0	•
GLADSTONE NORTH																					
82-92V	38	49	7	22	44	245	7548	0.0	0.0	0	17.4	0	330	278	0	0	0	5	0	17	50
82-93V	114	194	5	21	41	182	5492	2.7	0.0	0	16.1	10	639	273	0	0	0	13	0	19	•
CARIBOU CREEK																					
84-11V	INSUFFICIENT SAMPLE																				
84-13V	29	260	5	87	36	171	3320	0.0	0.0	0	44.2	0	627	1670	0	0	0	0	0	0	10
84-O9V	74	310	5	19	16	207	3404	0.0	0.0	0	47.6	0	271	1040	0	15	0	10	10	20	10
84-O8V	33	216	3	107	50	146	2874	0.0	0.0	0	39.7	0	684	1940	0	0	0	13	7	0	3
CAMP CREEK																					
85-O5V	0	326	7	20	48	268	6373	0.0	0.0	0	44.4	0	239	1121	0	16	90	0	0	0	5
85-O4V	25	169	6	37	67	258	7689	0.0	0.0	0	32.7	0	245	555	0	15	40	8	0	18	3
85-O3V	34	255	6	24	81	220	8372	0.0	0.0	0	36.8	0	286	813	0	14	70	0	0	20	4
85-O2V	25	294	8	24	72	250	8040	0.0	0.0	0	40.1	0	257	998	0	15	45	11	12	30	5
85-O1V	32	317	4	27	71	279	9500	0.0	0.0	0	35.4	0	272	1024	0	12	16	0	0	26	4

Moraine Valley

There are no prominent element dispersal trends in either an up or down-valley direction. Ag and Sb are only detected in one middle-valley sample. No values were detected for Bi, Te, U, W, and Se. Anomaly levels for Cu and Co are found in upper-valley locations. An anomalous value for Au is found in a middle-valley sample site, and Ni exhibits an anomaly in a lower-valley location.

Gladstone North

There are insufficient data to demonstrate preferred concentration trends. There are no element anomaly levels detected.

Caribou Creek

There are insufficient data to demonstrate preferred concentration trends. Three elements exhibit anomaly levels; Co and V in a lower-valley location, and Se in a middle-valley site.

Camp Creek

The element data do not demonstrate a favoured up-valley or down-valley direction. Five elements (Cd, Ag, Bi, As, and Te) are not detected in any sample. Anomaly values for Cr, Mn and Se are detected in lower-valley locations.

The general observations made with respect to mineral dispersal trends apply to patterns observed with element dispersals as well. The data as a whole indicates there are no overwhelming dispersal trends in either an up-valley or down-valley direction. As well, element anomaly values only occur occasionally, and are not concentrated in a preferred valley location (i.e., upper, middle, or lower valley sites). These findings would be consistent with valley regions that experienced glaciation on a local scale (small ice cap, cirque, and valley glaciers) with down-valley ice movement, and regional ice penetrating the area in an up-valley direction, resulting in much of the glacial sediments in the valleys being reworked numerous times by this up-valley and down-valley flow, and subsequently modified by processes acting in the post-glacial alpine environment.

5.7.2 Plateau Dispersal Trends

The procedures and explanations for plotting the element concentration trends are analogous to those described in the plotting of mineral trends in plateau regions. As with the heavy mineral trends, the ensuing observations are made with reference to the individual element compositional trends illustrated by plots, and arrived at with the prior knowledge of the "false" or "exaggerated" patterns that the computer plots generate. (The topographic transparent overlay, provided in the back pocket, should be aligned with the registration points on the individual dispersal plots.)

Of the twenty-one elements analyzed for in the silt and clay fraction of the plateau surficial sediment, eight elements exhibited prominent dispersal trends that facilitated plotting, and provides the basis for the the ensuing descriptions. These include; lead (pb), molybdenum (Mo), chromium (Cr), cadmium (Cd), silver (Ag), arsenic (As), uranium (U), and tin (Sn). (The plotting of element concentration trends followed the same procedure outlined in the plotting of heavy mineral trends). The dispersal patterns of the remaining eleven elements; copper (Cu), cobalt (Co), nickel (Ni), manganese (Mn), bismuth (Bi), iron (Fe), zinc (Zn), vanadium (V), tellurium (Te), tungsten (W), antimony (Sb), selenium (Se), and gold (Au), displayed a random trend which did not facilitate plotting. In many of the silt and clay fraction plateau samples, some of these elements were not detected at all, or there was insufficient sample to make a determination.

Lead

There are two distinct dispersal patterns shown on Figure 22. The dispersal associated with the area of high concentration (4970 ppm) on the southern plateau ridge at the head of Rockslide Creek is associated with a zone of known mineralization. This dispersal demonstrates a preferred northwest to west trend, with the tail of the pattern traced for about 8 km. The head portion of the trend extends less than 2 km. The second pattern is centred on the ridge that separates Talbot and Alaskite creeks, and is associated with an anomaly area of greater than 10610 ppm. This centre is not identified with any known zone of mineralization. The trend exhibits a northwest to north dispersal trend, with the tail portion of the dispersal traced for over 12 km. The head of this trend extends less than 2 km, however, much of this is attributed to computer "fitting".

Molybdenum

The dispersal of Molybdenum on plateaus is shown on Figure 23. The area of highest concentration (> 1300 ppm) corresponds to an area of known mineralization on the southern plateau ridge at the head of Rockslide Creek. The general dispersal trend of the tail portion from the area of highest concentration (> 1300 ppm) in a northwest direction is mapped over a distance of about 8 km to a concentration of 173 ppm. The head portion of this trend is less than 2 km, and much of this is attributed to computer "fitting".

Chromium

Chromium displays a northwest to north dispersal trend, with the area of highest concentration located on the plateau surface between the upper-valley reaches of Ice Valley, Alaskite Creek, and Rockslide Creek (Figure 24). This centre is in the vicinity of known mineralization at the head of Alaskite Creek. The tail portion of the dispersal trend, from the area of highest concentration to the 129 ppm interval in the northwest or north direction is approximately 12 to 16 km. The tail portion over the same concentration range is less than 1 km.

Cadmium

Figure 25 shows a rather complex dispersal pattern with a general northwest trend, along with some north to northeast patterns emerging. The centre of highest concentration (an anomaly of >57 ppm) is located in an area of known mineralization at the head of Rockslide Creek. The tail of the general northwest dispersal pattern extends about 10 km, while the head portion is traced for less than 2 km. The widening of the dispersal trend in a northeast to north direction may be attributed to; the lack of plateau sample sites in these locations, unknown mineralizations, or secondary dispersals.

Silver

The dispersal trend of Silver is shown on Figure 26. Two centres of high concentration are evident, each with a northwest to north dispersal trend. The centre (>31 ppm) on the plateau ridge separating Rockslide, Talbot and Alaskite creeks is located in an area of known mineralization. It is difficult to discern whether the main dispersal trend (tail) is in a northwest or north direction. In all cases, the tail portions would extend at least 4 km. The head portions would corresponingly extend up to about 2 km. The second dispersal pattern is centred in the vicinity of another known zone of mineralization at the head of Alaskite Creek. The tail portion of this trend is tracable for at least 10 km in a northwesterly direction. The head portion appears to extend for about 4 km, but is strongly compromised by the plotting technique.

Arsenic

Arsenic demonstrates a distinct northwest dispersal trend (Figure 26) with the area of highest concentration (anomaly >1636 ppm) not associated with any known zone of mineralization. The tail portion of the trend is mapped for over 18 km while the head was traced for less than 1 km. What appears to be a northern dispersal pattern is attributed to computer interpretation, since no plateau samples are located in this area.

Uranium

Uranium exhibits an apparent multitude of dispersal trends from northwest, north, to northeast (Figure 28). However it appears, albeit not distinct, that a northwest trend predominates. The two areas of highest concentration (>59 ppm) are located in the vicinity of known mineralization on the plateau ridge at the head of Rockslide Creek. The tail portions of these trends can be traced approximately 6 to 10 km in a northwest direction. The corresponding heads of these trends extend less than 2 km. A third centre of high uranium concentration (>37 ppm) is located on the plateau surface near the upper-valley portion of Ice Valley and is not associated with a zone of known mineralization. A weak northwest to north dispersal trend (tail) is trace for approximately 8 km. The head portion of this trend extends for less than 2 km.

Tin

The main dispersal pattern of tin (Figure 29) is difficult to determine but appears to be in a northwest to north direction with the area of highest concentration (>206 ppm) being located in the vicinity of a zone of known mineralization on the plateau ridge that separates Talbot, Rockslide, and Alaskite creeks. The tail portion of the dispersal pattern extending northwest is about 14 km; the tail portion for a northern trend extends about 12 km. In both instances, the head portion of the dispersals extend to about 2 km. Because of the greater abundance of plateau sample sites in the northwestern portion of this pattern, as compared with fewer sites in the northern portion of the dispersal pattern, the northwest trend probably respresents the main direction of dispersal.

As was the case with the heavy mineral descriptions, the dispersal patterns described above were made with reference to the element concentration plots, but were done so with a view to compensate for computer generated "false" patterns. The following is a summary of element dispersal trends.

	DISPERSAL TREND DIRECTION	DISPERSAL DISTANCE HEAD	TAIL	PROXIMITY TO KNOWN MINERALIZATIONS
Lead	Northwest and West	2 km	8 km	Rockslide Creek
	Northwest and North	2 km	8 km	None
Molybdenum	Northwest	2 km	8 km	Rockslide Creek
Chromium	Northwest and North	1 km	16 km	Alaskite Creek
Cadmium	Northwest, North, and Northeast	2 km	10 km	Rockslide Creek
Silver	Northwest and North	2 km	4 km	Rockslide Creek/ Talbot Creek Ridge
	Northwest and North	4 km	10 km	Alaskite Creek
Arsenic	Northwest	1 km	18 km	None
Uranium	Northwest, North, and Northeast	2 km	10 km	Rockslide Creek
	Northwest and North	2 km	8 km	None
Tin	Northwest and North	2 km	14 km	Rockslide Creek/ Talbot Creek Ridge

Although as indicated above, and illustrated in the individual element plots, the dispersal trends may be interpreted as being in more than one direction, it is evident that there is a regionally preferred northwest dispersal trend. Other directional trends may be attributed to a number of factors including; i) inadequate sampling density on some plateau regions, ii) plotting technique, iii) processes affecting dispersals on a local scale, iv) dispersals attributed to unknown mineralizations, and v) multiple regional trends of varying extent, some possibly relic.

Futhermore, the northwest trends are generally more continuous, and have been mapped over extensive plateau surfaces that are dissected by numerous valleys. The other directional trends are generally less extensive, and are not observed to span extensive plateau surfaces dissected by valleys.

Figure 22
DISPERSAL PATTERN OF LEAD

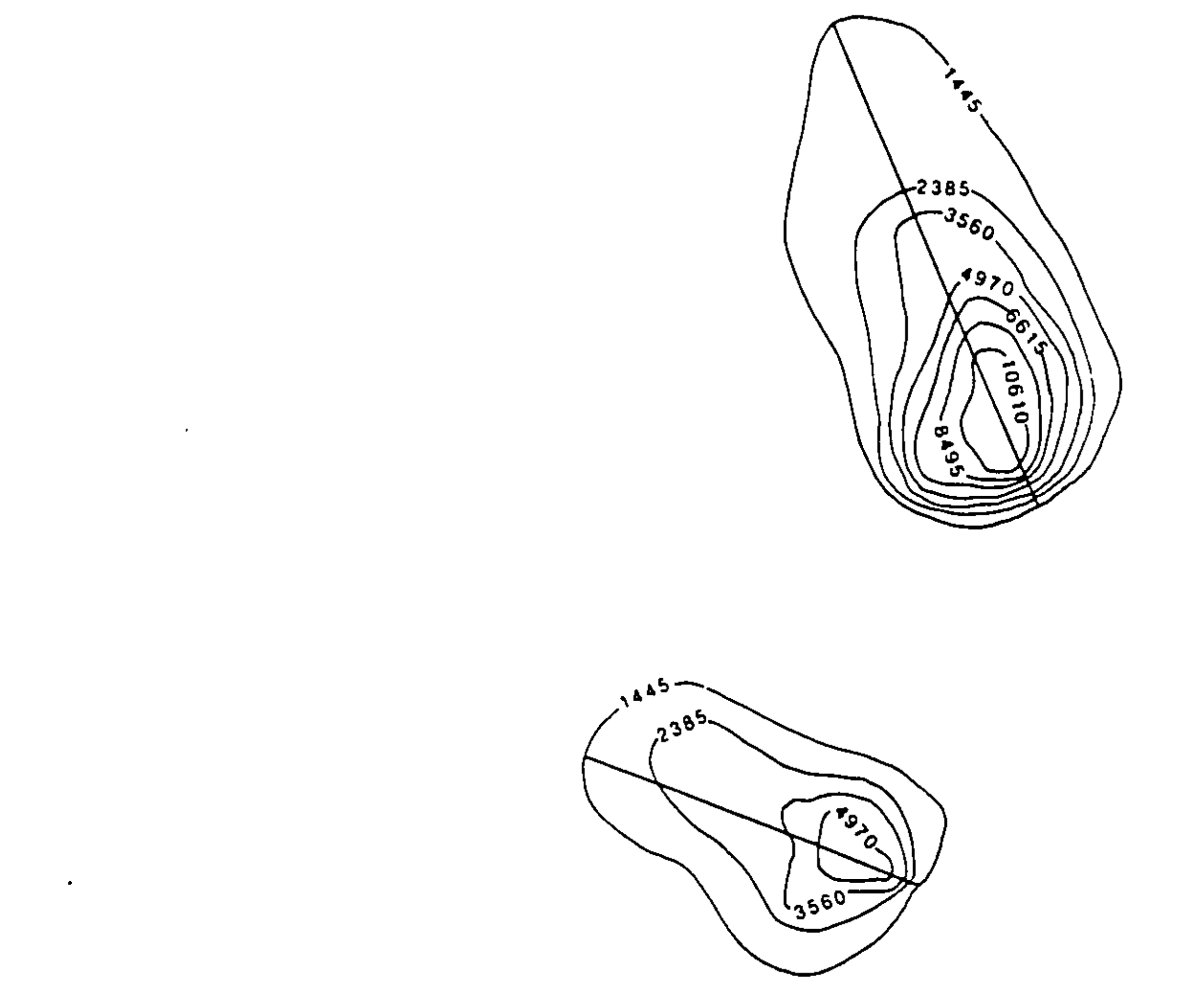


Figure 23
DISPERSAL PATTERN OF MOLYBDENUM

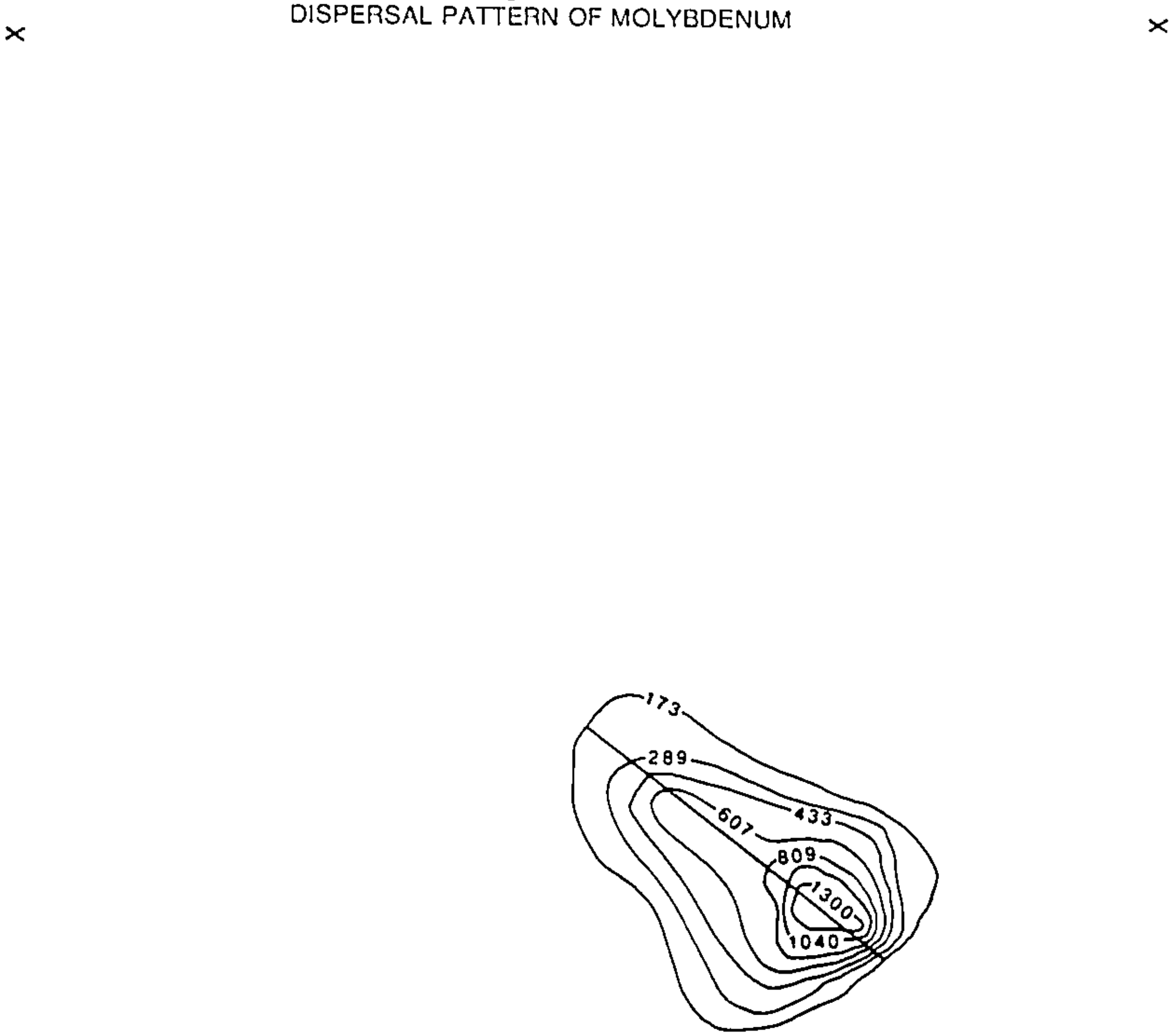


Figure 24
DISPERSAL PATTERN OF CHROMIUM

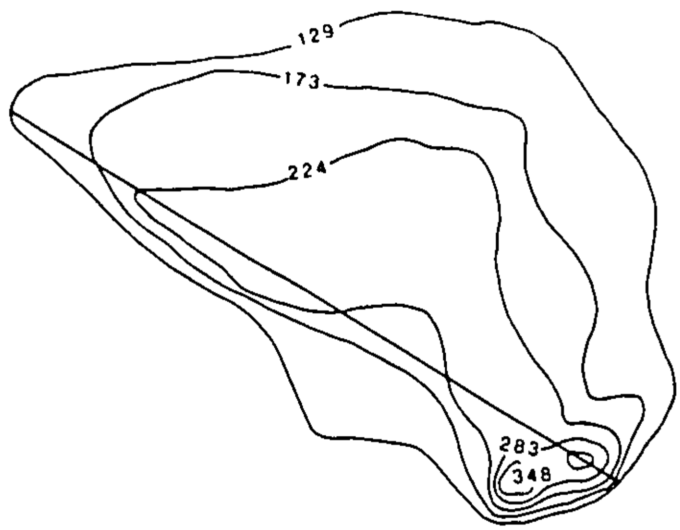


Figure 25
DISPERSAL PATTERN OF CADMIUM

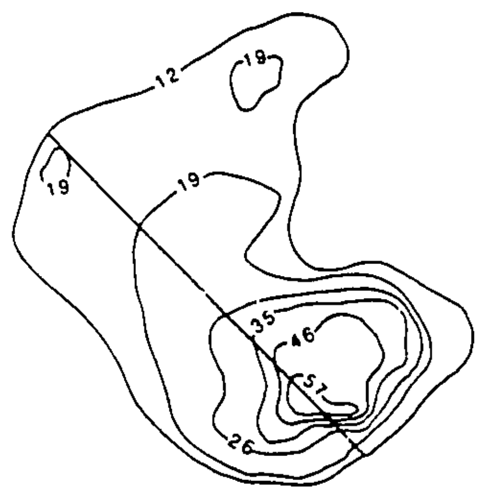


Figure 26
DISPERSAL PATTERN OF SILVER

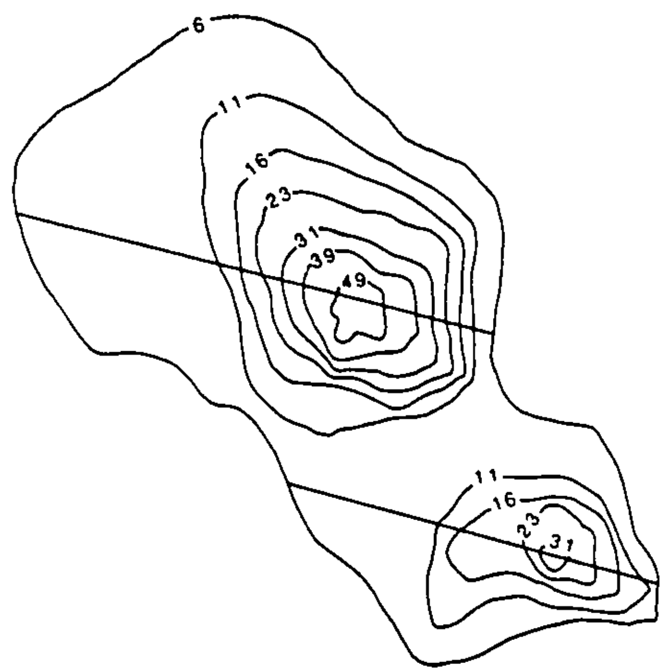


Figure 27
DISPERSAL PATTERN OF ARSENIC

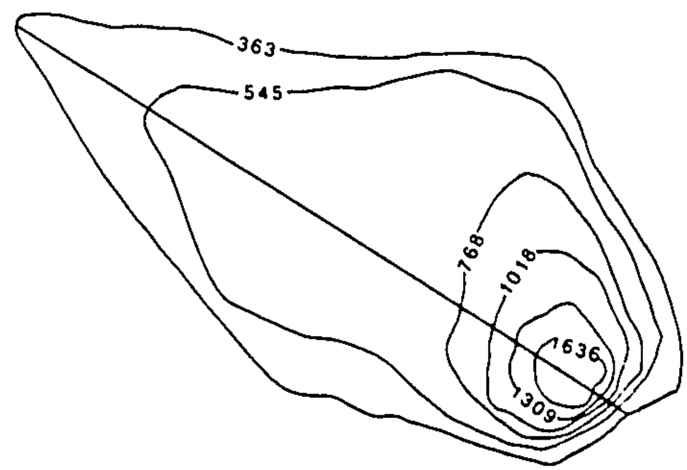


Figure 28
DISPERSAL PATTERN OF URANIUM

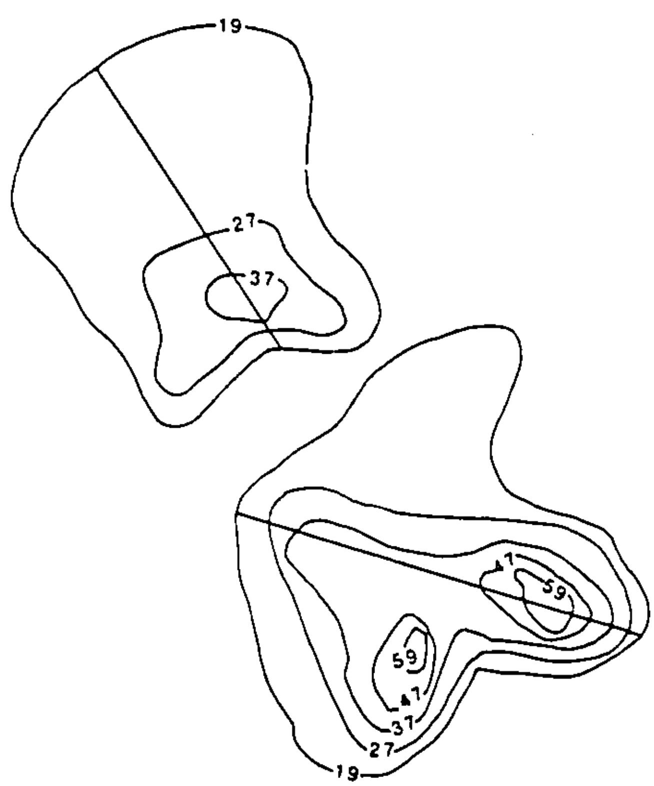
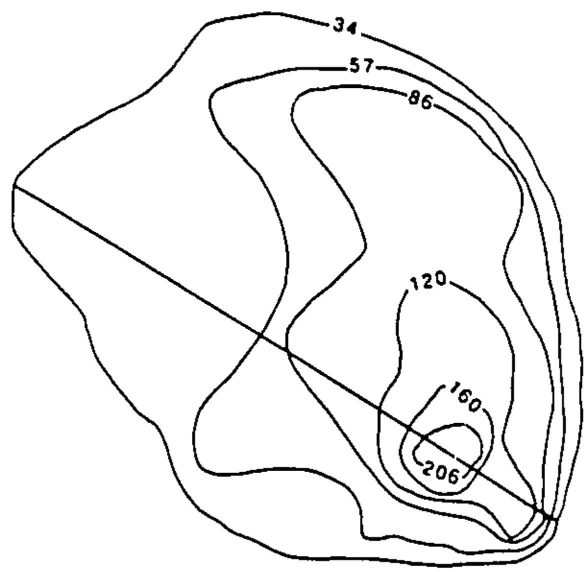


Figure 29
DISPERSAL PATTERN OF TIN



5.7.3 Element Mobility in the Secondary Environment: Implications for Valley and Plateau Dispersal Patterns

Element mobility in the secondary environment has already been briefly referred to in previous discussions relating to provenance implications of selected minerals. Since this topic has important implications to element dispersal trends, a more detailed discription will follow. There are numerous texts relating to the circulation and movement of material within and between the primary and secondary environment (e.g., Rankama and Sahama 1950; Goldschmidt, 1954; Bear 1964; Mitchell, 1964, 1972; Garrels and Christ, 1965; Krauskopf, 1967; Saxby, 1969; Loughnan, 1969; Hunt, 1972; Levinson, 1980). Although it is beyond the scope of this discussion to detail these aspects, some relative degrees of mobility of elements in the northern Ruby Range environment will be addressed.

The secondary environment comprises the surficial processes of weathering, soil formation and sedimentation on the bedrock surface (Hawkes 1957; Hawkes and Webb 1962). Therefore, the secondary environment includes all unconsolidated material on the bedrock surface along with associated processes of weathering. It is characterized by relatively lower temperatures and pressures (as compared to those of the deep seated processes of igneous differentiation and metamorphism), abundant free oxygen and other gases, and the relatively free flow of fluids (Hawkes and Webb 1962; Levinson 1980). The distribution of elements in the secondary environment is dependent upon chemical and physical weathering acting simultaneously. Chemical weathering involves the breakdown of rocks and minerals and dispersion of the released elements, usually by water. Hence, abundant water, carbon dioxide, and oxygen are essential in this type of weathering. Physical weathering includes those processes of rock disintegration which do not employ chemical reactions or mineralogical changes. This process is generally most effective in cold environments, and alpine regions.

The distribution of elements can be attributed to chemical and physical mobility. Mobility is defined as the ease with which an element can be dispersed within a specific environment. Determining the mobility of elements in the secondary environment requires an appreciation of many factors, for example:

1) The nature of the medium; mobility in groundwater will vary from that in surface water flow, and water migration within and through sediment.

2) Variations of rock types in surficial sediment; certain heavy metals form acid water precipitates when the water is neutralized by contact with carbonates.

3) The presence and affects of micro-organisms which can reduce sulfate to sulfides.

4) The presence of dissolved gases such as CO_2 which control the solubility of certain compounds. (These and other factors are summarized by Levinson, 1980).

As was previously mentioned, Andrews-Jones (1968) empirically estimated the relative chemical mobilities of elements in the secondary environment which has application to the northern Ruby Range study area. His classification is based on environmental conditions which include an oxidizing and reducing tendency. Table 24 summarizes the relative mobility of elements under various environmental conditions. The relative mobilities and environmental conditions are adapted from Andrews-Jones (1968). Eh values of elements are given in the oxidizing column.

At virtually all sample locations, and in particular in valley sampled sites, a hematite staining was noticed on many rocks and boulders. The staining is a good indication of strong oxidation. Barakso (1970) after taking numerous readings of the pH and Eh of streams in the Talbot Creek and Rockslide Creek areas noted that the environmental conditions were extremely acid and rich in oxidizing material (average pH 2.5, Eh 0.8). In addition to abundant oxygen, the free drainage in valleys makes oxidation a significant process of element dispersal. Similarly, but to a much lesser extent, oxidation was noticed to occur on plateaus. As discussed in the provenance implications of selected minerals section, the intensity is not as great as that in the valleys because of restricted free drainage and the presence of permafrost. Reducing conditions were noted to be more predominant on the plateau surfaces.

From Table 24, it is evident that under oxidizing conditions the following elements exhibit medium to high mobilities; Zn, Mo, V, U, Se, Cu, Co, Ni, Ag, As, Au, and Cd. No elements analysed for are identified under the medium, high, or very high relative mobility column in reducing conditions. In fact, most elements are identified as being relatively immobile. Under acid conditions, Mo, V, U, Se, Zn, Cu, Co, Ni, Ag, Au, As, and Cd exhibit medium to high mobilities. Since the valleys of the study area are environmentally acidic and oxidizing, most of the elements analyzed for would be expected to be widely dispersed. This was in fact observed in the dispersal patterns of elements in valleys, where it was concluded that the data as a whole indicated there were no discernable dispersal trends in either an up-valley or down-valley direction, but rather a scattered or random pattern.

On plateau surfaces, where reducing conditions dominate, virtually all elements are identified as exhibiting very low to immobile mobilities (Table 24). In this study, eight of the twenty-one elements analyzed demonstrated discernable dispersal trends, and of the remaining eleven elements,

Table 24
RELATIVE MOBILITY OF ELEMENTS
IN THE SECONDARY ENVIRONMENT

RELATIVE MOBILITIES	ENVIRONMENTAL CONDITIONS			
	Oxidizing	Reducing	Acid	Neutral/Alkaline
Very High				Mo, Se, U, V
High	Zn, Se, Mo, V, U		Mo, Au, Ag, Se, Zn, Cu, Co, Ni, U, V	
Medium	Cu, Co, Ni, Ag, Au, As, Cd		As, Cd	As, Cd
Low	Bi, pb, Sb	Fe, Mn	Bi, pb, Sb, Fe, Mn	Bi, pb, Fe, Mn
Very Low to Immobile	Fe, Sn, Mn, Te, Cr, W	Sn, Sb, Te, Cr, Bi, Mo, Se, pb, Zn, Co, Cu, Ni, Ag, Au, As, Cd, U, V, W	Sn, Cr, Te, W	Sn, Au, Te, Cr, Zn, Cu, Co, Ni, Ag, W

Modified from Andrews-Jones (1968)

in many of the silt and clay fraction plateau samples, some elements were not detected at all, or there was insufficient sample to make a determination (these included, Bi, Te, W, Sb, Se, and Au). Therefore, over half of the elements for which continuous data was available, demonstrated discernable dispersal trends on the plateau surfaces, which cannot be exclusively related to element mobility in this environment.

5.7.4 Dispersal Models: Distance Decay Analysis

To facilitate discussion on the nature of sediment transport, it is useful to characterize mineral and element dispersal trends on plateau surfaces with respect to distance decay curves (Figure 30), which represent the decrease in concentration with distance (Taylor, 1975; Harries, 1982). The slope and dimensions of the curves are determined by the physical characteristics of the components being dispersed, and by the inferred mode of transport.

The dispersal trends of heavy minerals and elements plotted on the plateau surfaces are described with reference to dispersal curves. A straight line was drawn through the dispersal plots, which is taken to represent the main directional trend. Although in most cases this line spans the furthest extent of the dispersal plots, it is recognized that in some instances alternative lines representing the main directional trend may be inferred. However, for the purposes of this analysis, it was decided not to pursue this aspect further. A profile view of the actual dispersal of elements was plotted so that the X axis represents distance, the Y axis represent concentration. (Graphs are not shown.)

The procedure was carried out twice for each dispersal trend. The first operation represented the actual concentration over distance of the computer generated plots so that the intersections of the isolines with the line representing the dispersal directions were taken as data points. The second operation utilized the actual sample sites as data points with relation to the line representing dispersal direction. This was accomplished by joining concentration values represented by sample sites perpendicular to the line of dispersal. The rational for using isoline values and actual sample site concentration values was to see what, if any affect computer smoothing had on the dispersal trends.

The isolines for the dispersal trends are for specific concentrations and the values are assumed to vary smoothly over the intervals between any two isolines such that midway between any two isolines, the concentration will be half the difference of the two isolines.

Figure 30
DISTANCE DECAY CURVES

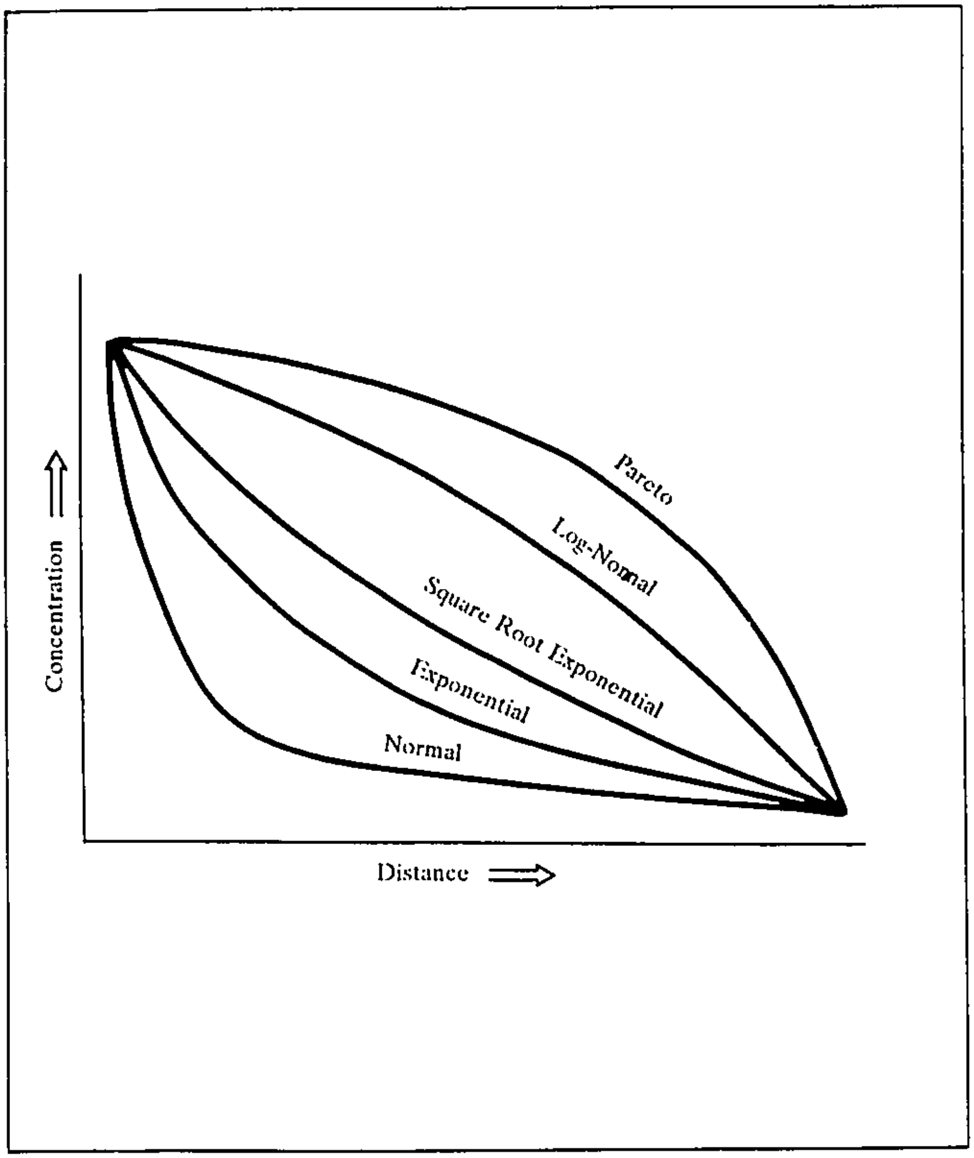


Table 25
HEAVY MINERAL DISTANCE DECAY ANALYSIS

HEAVY MINERAL	EXPONENTIAL MODEL	NORMAL MODEL	SQUARE ROOT EXPONENTIAL MODEL	LOG NORMAL MODEL	PARETO MODEL
ALLANITE					
Isoline Data	.614	.685	.549	.563	.482
Actual Data	.259	.347	.234	.198	.163
GOETHITE					
Isoline Data	.957	.846	.886	.929	.961
Actual Data	.908	.673	.977	.808	.940
EPIDOTE					
Isoline Data	.695	.736	.667	.607	.611
Actual Data	.263	.148	.442	.179	.299
ZIRCON					
Isoline Data	.915	.839	.929	.796	.929
Actual Data	.626	.508	.656	.551	.617
HORNBLENDE					
Isoline Data	.932	.748	.914	.913	.979
Actual Data	.862	.679	.892	.867	.954
ORTHOPYROXENE					
Isoline Data	.807	.814	.810	.763	.6[illegible]5
Actual Data	.741	.627	.773	.673	.717
HEMATITE					
Isoline Data	.903	.975	.801	.847	.695
Actual Data	.138	.071	.258	.071	.172
GARNET					
Isoline Data	.726	.773	.646	.643	.555
Actual Data	.682	.653	.612	.662	.551
OPAQUES					
Isoline Data	.727	.895	.551	.688	.554
Actual Data	.745	.892	.535	.727	.549

- Isoline and Actual Data values are correlation coefficients.
- Best fit correlation coefficients are underlined.

Table 26
ELEMENT DISTANCE DECAY ANALYSIS

ELEMENTS	EXPONENTIAL MODEL	NORMAL MODEL	SQARE ROOT EXPONENTIAL MODEL	LOG NORMAL MODEL	PARETO MODEL
CHROMIUM					
Isoline Data	.802	.797	.797	.774	.791
Actual Data	.463	.451	.416	.443	.457
ARSENIC					
Isoline Data	.911	.877	.926	.882	.936
Actual Data	.772	.782	.746	.798	.697
SILVER (1)					
Isoline Data	.941	.871	.964	.875	.974
Actual Data	.205	.225	.206	.211	.187
SILVER (2)					
Isoline Data	.865	.768	.897	.810	.910
Actual Data	.011	.010	.038	.012	.049
CADMIUM					
Isoline Data	.954	.891	.972	.914	.973
Actual Data	.744	.785	.714	.750	.670
LEAD (1)					
Isoline Data	.973	.986	.955	.975	.929
Actual Data	.871	.790	.890	.853	.888
LEAD (2)					
Isoline Data	.948	.871	.957	.931	.936
Actual Data	.844	.782	.864	.827	.867
MOLYBDENUM					
Isoline Data	.908	.942	.889	.917	.857
Actual Data	.840	.845	.817	.853	.775
URANIUM (1)					
Isoline Data	.862	.901	.837	.869	.812
Actual Data	.396	.385	.395	.430	.382
URANIUM (2)					
Isoline Data	.954	.900	.976	.890	.994
Actual Data	.929	.968	.889	.959	.830
TIN					
Isoline Data	.920	.807	.957	.885	.971
Actual Data	.389	.288	.441	.360	.478

- Isoline and Actual Data values are correlation coefficients.
- Best fit correlation coefficients are underlined.

In order to facilitate the adaptation of these curves to the distance decay curves, the data are transformed so that it has a linear pattern. Then a simple linear regression line is fitted to the transformed data. On all the plots, the logarithm of the concentration is plotted against various transformations of the distance to achieve the desired model. These transformations are;

i) Normal Model, distance squared (d^2),
ii) Exponential Model, distance untransformed (d),
iii) Square Root Exponential Model, square root of distance ($\sqrt{d}$),
iv) Log-Normal Model, logarithm of distance squared ($(\log d)^2$),
v) Pareto Model, logarithm of distance (log d).

A comparison of correlation coefficients should reveal the affects of smoothing, as well as indicate which distance decay curve best fits the dispersal trends. Tables 25 and 26 present the heavy mineral and element distance decay correlation coefficient (best fit) data respectively. Initial examination of the dispersal graphs of heavy minerals and elements shows that computer smoothing has had an affect on the distance decay curves representing dispersal trends. This is demonstrated in the correlation coefficients representing both isoline and actual data. Generally, the heavy mineral distance decay curves appear to be best represented by the normal, and square root exponential models. The element distance decay curves seem to favour the normal, and pareto models.

5.7.5 Heavy Mineral and Geochemical Analysis: An Overview

The use of element and heavy mineral occurrences for determining composition, provenance and dispersal trends have for the most part been exploratory. The technique has mainly been applied in Canada to those areas affected by the Laurentide Ice Sheet which because of its extent included areas of varied terrain (Shilts et al., 1979; Klassen, 1982). It has also been utilized in Finland (e.g., Virkkala, 1971; Perttunen, 1977; Taipale et al., 1986) and Sweden (e.g., Linden, 1975). The technique has in the past proven invaluable in determining compositional variations in rock types, provenance of mineral suits, and processes and rates of erosion (Flint, 1957; Dreimanis, 1971; Shilts et al., 1979; Klassen, 1982, 1984; Paré, 1982; Henderson, 1989).

Although this technique has been shown to apply both theoretically and practically to alpine valley situations, its adaptation to alpine and sub-alpine regions with extensive plateaus and deeply incised glaciated valleys such as those in the northern Ruby Range have not been attempted (Bradshaw et al., 1975; Bradshaw, 1975; Levinson, 1980).

A number of observations and interpretations have been made concerning the heavy mineral fraction (fine sand), relating to composition, assemblages, and dispersal trends. Similarly, analysis of the silt and clay fraction has generated geochemical data with respect to background and anomaly values, as well as dispersal trends. The purpose of this section is to succinctly summarize these observations and conclusions in order to set the stage for a discussion on the nature of sediment transport.

Fine Sand Fraction: Heavy Mineral Analysis

i) The average proportion of heavy minerals identified in the sediment from plateau and valley locations were similar. Therfore, the sediment from both plateau and valley sites do not vary greatly in heavy mineral species and proportions.

ii) Except for minor differences, the background values for most heavy minerals in both plateau and valley samples are quite similar. As well, the anomaly values are also relatively comparable for most minerals.

iii) The heavy mineral assemblages derived from plateau and valleys sediments were alike.

iv) The data do not support a clear association between heavy mineral assemblages from plateau and valley sediments, and those derived from bedrock within or outside the study area. However, the mineral assemblages do not appear to be exclusively related to local bedrock sources.

v) The provenance of orthopyroxene, goethite, and allanite may be attributed to bedrock from outside the study area, volcanic ash, and meteoritic impacts, or a combination of these. The occurrence of some of these minerals may also be associated with mineral alterations.

vi) Heavy mineral dispersals in valleys do not demonstrate an overwhelming trend in either an up-valley or down-valley direction.

vii) Heavy mineral dispersals on the plateaus display a preferred northwest trend. Futhermore, these trends are continuous, spanning many kilometres, and have been mapped over extensive plateau surfaces dissected by numerous valleys.

viii) The heavy mineral distance decay curves (tail portion of the dispersals) appear to be best represented by the normal, and square root exponential models.

Silt and Clay Fraction: Geochemical Analysis

i) Similar background values for the majority of elements in the sediments from plateau and valley sites indicates that there is a relatively uniform concentration of elements, but because of profound differences in nine element anomaly values, the distribution of these elements is far from uniform.

ii) There are no clear dispersal trends in either an up-valley or down-valley direction.

iii) Although some element dispersal trends on plateaus may be interpreted as being in more than one direction, it is evident that there is a regionally preferred northwest dispersal trend. The northwest trends are generally more continuous, spanning many kilometres, and have been mapped over extensive plateau surfaces that are dissected by numerous valleys. The

other directional trends are generally less extensive, and are not observed to span extensive plateau surfaces dissected by valleys.

iv) Since the valleys of the study area are environmentally acidic and oxidizing, most of the elements analyzed for would be expected to be widely dispersed, which was observed.

v) On plateau surfaces where reducing conditions dominate, element mobility is very low to immobile. Over half of the elements for which continuous data was available demonstrated discernable dispersal trends which cannot be exclusively related to element mobility in this environment.

vi) Element distance decay curves seem to favour the Normal, and Pareto models.

Chapter 6
A DISCUSSION ON THE EVOLUTION OF
THE PLATEAU SURFACES OF THE NORTHERN RUBY RANGE

6.1 Introduction

The primary objective of this thesis was to offer explanations on the evolution of the plateau surfaces in the northern Ruby Range, Yukon Territory. In addressing this objective two research questions were posed:

i) Can the compositional trends observed in the heavy mineral and geochemical data on the plateaus provide geological evidence (lacking in previous descriptions of these regions) which can contribute to the solution to the question of how the plateau landscape evolved? Specifically, depending upon the genesis of the plateau sediments (i.e., local, glacial, etc.), what processes were responsible for these dispersals (glaciation, periglacial, aeolian, tectonic)? and,

ii) Are similar compositional trends observed in the valleys of the Ruby Range, and what are the implications of these patterns?

Based on the observations and interpretations derived from data accrued from the examination of surficial deposits from valleys and plateaus, specifically with the use of heavy mineral and geochemical composition analysis, a discussion on whether the geological and geomorphological evidence from the study area can support the belief that plateau regions were subjected to glaciation, or other evolutionary processes, can ensue.

6.2 Mineral and Geochemical Primary and Secondary Dispersal Patterns

The general environmental (geological) classification of dispersals are grouped into primary and secondary. Primary dispersal is characterized by deep-seated conditions of relatively high pressure and temperature. The products of primary dispersal are the mineralogical and geochemical patterns that are preserved in rocks that later, due to uplift and erosion, become exposed at the surface. The genetic classification of primary dispersals patterns include syngenetic and epigenetic patterns. Syngenetic patterns are formed at the same time as the enclosing rock, whereas epigenetic patterns are formed by material introduced in some way into a pre-existing matrix (Hawkes and Webb, 1962).

Secondary dispersal prevails at the surface under conditions of relatively low pressure and temperature and involves the disintegration of rocks by weathering, which are made available for further erosion and deposition away from the place of origin (Hawkes and Webb, 1962; Kauranne, 1976). During the course of transportation, selective deposition or sedimentation often results in the redistribution of the dispersed products. The dispersal pattern or trends, is regulated by chemical and physical properties of the various dispersed constituents, and of the mode of transportation. These dispersal patterns are a direct consequence of dispersal processes. As was the case with primary dispersal, the dispersal patterns are classified according to when they formed, relative to the sediment they are in, and how they formed. On this basis, patterns formed contemporaneously with the sediment deposit are called syngenetic. Patterns that formed after sediment deposition are called epigenetic. Patterns are further classified as; i) clastic, where the dispersal is mainly by movement of solid particles, ii) hydromorphic, aqueous solutions, and iii) biogenic, where the patterns are the result of biological activity (Hawkes and Webb, 1962).

6.3 Valley Implications

Since the implications of mineral and element compositional trends found in till samples from the valleys of the northern Ruby Range have been addressed in the previous chapter, only a brief summary of these implications will be presented here with reference to plateaus.

i) The average proportion of the heavy minerals is similar for both plateau and valley sediment samples. However, the range of the proportions (ie. difference between minimum and maximum) varies considerably in most minerals. This indicates that the sediment from both plateau and valley sites do not vary greatly in heavy mineral species and proportions, but do exhibit a variable range. This may imply more reworking of sediment in the valleys.

ii) Apart from the relative lack of allanite on plateaus and goethite in valleys, the heavy mineral assemblages derived from valley and plateau sediments are similar. The data do not support a clear association between the heavy mineral assemblage derived from valley sediments (as well as plateau sediments) and those derived from bedrock sources within or outside the study area. Clearly, the heavy mineral assemblage for valley samples is not exclusively related to local bedrock sources.

iii) Except for minor differences, the background values for most heavy minerals in both plateau and valley samples are quite similar; anomaly values are also relatively similar with some minor variations. This indicates there is a relatively uniform concentration of minerals in the sediment from both plateaus and valleys, as exemplified by background values, but because of the differences in some anomaly values, demonstrate a varying dispersal pattern.

iv) An examination of heavy mineral concentration patterns in the valleys in the northern Ruby Range which include; Talbot Creek, Alaskite Creek, Rockslide Creek, Ice Valley, Pass Creek, Moraine Valley, Gladstone North, Caribou Creek and Camp Creek, indicated that there is no evidence to suggest minerals demonstrate an overwhelming dispersal trend in either an up-valley or down-valley direction. These conditions were not unexpected since valleys in the northern Ruby Range exhibit a classical glaciated U-shape with evidence of glaciation on a local scale (small ice cap, cirque, and valley glaciers) with down-valley ice movement, and regional ice penetrating the area in an up-valley direction. Therefore, much of the glacial sediments in the valleys must have been reworked numerous times by up-valley and down-valley glacier flows.

v) The background values for sixteen of the twenty-one elements analyzed in both plateau and valley sites are relatively similar. A somewhat different pattern emerges from the anomaly data. Twelve elements do not show an appreciable difference in anomaly levels. However, nine elements demonstrate considerable differences, with plateau anomaly values in all cases being at least twice the corresponding valley levels. The similar background values for the majority of elements in the sediments from plateau and valley sites indicates that there is a uniform concentration of elements, but because of profound differences in some element anomaly values, the distribution of these elements is variable.

vi) The geochemical data indicates there are no discernable dispersal patterns in either an up-valley or down-valley direction. This may be attributed to the mobility of elements in this environment which is acidic and oxidizing.

Based on the analysis of the fine sand fraction (heavy minerals), and the silt and clay fraction (geochemical analysis) found in till from the valleys of the northern Ruby Range, and the surficial sediments from plateau samples, it is concluded that in both valley and plateau sediments, heavy mineral composition and assemblages are similar, and the geochemical nature of the silt and clay fractions are comparable. Therefore, it would seem likely that some of the glacial sediment in the valleys may be of similar origin as the surficial sediment on plateau surfaces, as well as being derived from plateau sediment.

Since both the heavy mineral and geochemical data show no discernable up-valley or down-valley increase or decrease in concentration, and because the valleys were subjected to a number of up and down valley glacial flows, as well as being affected by other processes, the random dispersals of minerals and elements would be best described as secondary, epigenetic dispersals. The dispersals could also be classified as clastic, hydromorphic, and biogenic patterns.

6.4 Plateau Evolution

There have been few environment related studies focusing on the Ruby Range, and even fewer research endeavours pertaining to glaciation. The plateau regions of the northern Ruby Range have, for the most part, only been referred to in vague generalities as an old, uplifted, unglaciated, erosional surface of unknown age. The presence of tor-like structures and other periglacial features, along with the apparent absence of any glacially related features indeed does promote this portrayal. Analyses performed on surficial sediments taken from the plateau surfaces indicate that at least some of this material may not be of local origin, and discernable dispersal trends are evident in certain heavy minerals and elements which can be traced to known sources of mineralizations. As well, other trends have been mapped which may be associated with mineralizations which have not yet been delineated. The most striking characteristics of these trends are demonstrated by their continuous form, spanning many kilometres, and their patterns extending over extensive plateau surfaces that are dissected by numerous valleys.

In order to account for these findings and observations, the central research question which must be addressed is; "can the compositional trends observed in the heavy mineral and geochemical data on the plateaus provide geological evidence (lacking in previous descriptions of these regions) which can contribute to the solution to the question of how the plateau landscape evolved? Specifically, depending upon the genesis of the plateau sediments (ie. local, glacial, etc.), what processes were responsible for these dispersals (glaciation, periglacial, aeolian, tectonic)?"

6.4.1 Geological Terrane Considerations

The Geological terranes in southwest Yukon are described as allochthonous, that is, displaced over a considerable distance from their place of formation. Generally, terranes of southwestern Yukon are aligned and trending predominately north to northwest. The Ruby Range is part of the Coast Belt which is described as a metamorphic and plutonic belt, consisting of Jurassic to Tertiary granitic rocks and varied metamorphosed sedimentary and volcanic strata ranging in age from Paleozoic to early Tertiary (Clague, 1991). Specifically, the northern Ruby Range constitutes part of the Tracey Arm Terrane which consist of schist, paragneiss, amphibolite, marble, serpentine, and other metamorphosed sedimentary and igneous rocks of probable pre-early Triassic age (Gareau, 1992).

The Tracey Arm Terrane (northern Ruby Range), and indeed all terranes comprising southwestern Yukon have been subjected to uplift, but the patterns of movement are spatially variable within different terranes. Vanicek and Nagy (1981) conclude that the maximum rate of uplift for southwest Yukon is currently about 2.4 cm/yr, but it is not currently possible to distinguish vertical crustal movements due to tectonic processes and those due to glacial unloading (Hicks and Shofnos, 1965; Evans, 1991). However, since contemporary uplift is evident in other parts of the Cordillera, these movements are mainly attributed to tectonic activity (Clague et al., 1982; Riddihough, 1982; Clague, 1991).

The question which arises from the consideration of the geological features associated with the Tracey Arm Terrane is, what are the implications with respect to heavy mineral and element dispersal patterns? Specifically, are the geological and structural characteristics of the Tracey Arm Terrane responsible for the heavy mineral and element dispersal patterns observed in the surficial sediment on the plateau surfaces? In order to pursue this theme, it would seem reasonable (if not essential), to consider the surficial sediment on the plateau surfaces as a product of a progressive disintegration and decomposition of rock material in situ.

Primary dispersal has been described previously as being characterized by deep-seated conditions of relatively high pressure and temperature. The products of primary dispersal are the mineralogical and geochemical patterns that are preserved in rocks that later, due to uplift and erosion, become exposed at the surface. However, when the rocks and minerals typical of the deeper zone are exposed at or near the surface, they are brought into a vastly different environment which is characterized by lower temperatures and pressure, and high concentrations of water, free oxygen, and carbon dioxide. Most minerals that are formed under the deep-seated conditions are not stable in the surface environment, and therefore tend to be reconstituted to new forms that are stable with the environment. This is particularly true of igneous and metamorphic rocks (Hawkes and Webb, 1962). The product of this process is a mixture of resistant primary minerals with a suite of new mineral constituents that are stable in the surface environment, collectively forming the regolith. The regolith will therefore reflect, to a certain extent, the mineralogical and geochemical patterns of the underlying rock, and would be indicative of epigenetic patterns.

Attempts have been made to characterize various types of primary dispersal patterns which have ultimately been exposed to the surface environment, and which find expression in epigenetic patterns in the regolith. Two such epigenetic patterns have been identified as being generated by hydrothermal activity, and pressure-temperature effects (e.g., Hawkes and Webb, 1962; Kauranne, 1976). The complex processes involved in hydrothermal activity, and pressure-temperature effects with respect to heavy mineral and geochemical zonation are reported in a number of texts. The concern with hydrothermal activity and pressure-temperature effects in this study relates specifically to the form of dispersal patterns associated with these processes in regions of mineralizations.

In all genetic classifications of hydrothermal and pressure-temperature generated mineral and geochemical dispersals, the patterns are described as either resembling aureoles, halos, systematic variation with distance, or concentric isotherms (Hawkes and Webb, 1962). In a "static regolith environment", the mineral and geochemical patterns would "fingerprint" the patterns associated with the mineralization in the underlying rock. Hence, depending on the extent of the mineralization, the dispersal patterns would constitute an outward, decreasing trend. However, since "static regolith environment" is somewhat of an oxymoron, it is not likely that these patterns would appear in the regolith as such. The patterns in the regolith would not only reflect mineralization, but local fracturing, etc. Add to this any secondary dispersal patterns resulting from the subsequent weathering of the regolith, mineral and geochemical dispersal trends may vary from those described above.

Therefore, the fact that the heavy mineral and geochemical dispersal trends mapped on plateau surfaces resemble fan-like patterns and not aureoles, halos, systematic variations with distance, or concentric isotherms, does not entirely rule out an exclusive association with the underlying geology. As well, since the current geological descriptions of the northern Ruby Range are somewhat generalized, the identification and delineation of mineralizations in the region is incomplete, and hence the dimensions of patterns cannot be accurately estimated.

With respect to regional geological and structural concerns, whether the overall north to northwest alignment of the Tracey Arm Terrane, and the more restricted but relatively well defined northwest mineral and element dispersal trends are directly related is uncertain, but are more likely coincidental. Although reported evidence of overburden displacement due to earthquakes has been minimal, some observations suggest that both vertical and lateral displacement has occurred throughout the Quaternary. However, the displacements appear to be very localized, of the order of only a few hundred meters, and would probably not account for the mineral and element dispersal trends.

To imply that the general geological and structural characteristics of the Tracey Arm Terrane are entirely responsible for the heavy mineral and element dispersal trends observed on plateau surfaces would be speculative. If terrane attributes did play a major role in these dispersals it is more probable that they are related to varying terrane conditions on a local as opposed to a macro scale. Moreover, the plateau surfaces of the Ruby Range (and entire Yukon Plateau) have been characterized as remnants of an old, uplifted erosion surface, which has experienced uplift throughout the Quaternary and probably earlier (Bostock, 1948; Tempelman-Kluit, 1974; Hughes, 1990). If one is to assume the surficial sediments on the plateau surfaces are exclusively a product of a progressive disintegration and decomposition of rock material in situ, then surely over the span the last 2 Ma this material must have been subjected to secondary dispersal processes.

6.4.2 Secondary Dispersal Patterns

A general introduction to secondary dispersals has been presented in the beginning of this chapter. What follows is a discussion on principal transporting agents and subsequent dispersal patterns. Secondary dispersal is controlled by chemical (including biochemical) and mechanical factors. The products of weathering are partitioned between the relatively immobile solid phase that constitutes overburden, and the mobile fluid phase. The solid phase is comprised of the insoluble products of weathering (clastic fragments), which are transported and dispersed by mechanical processes, whereas the fluid phase consists of weathering products that are either soluble, or that occur in forms that can be readily suspended and transported in water (Hawkes and Webb, 1962).

6.4.2.1 Chemical Factors

Discussion on some aspects of chemical factors has been previously presented with reference to element mobility in the secondary environment. The following is a summary of that discussion:

i) The distribution of elements were attributed to chemical and physical mobility. Mobility was defined as the ease with which an element can be dispersed within a specific environment.

ii) The mobility of elements in the secondary environment is dependent upon:

- The nature of the medium; mobility in groundwater will vary from that in surface water flow, and water migration within and through sediment.

- Variations of rock types in surficial sediment; certain heavy metals form acid water precipitate when the water is neutralized by contact with carbonates.

- The presence and affects of micro-organisms which can reduce sulfate to sulfides.

- The presence of dissolved gases such as $C0_2$ which control the solubility of certain compounds. (These and other factors are summarized by Levinson, 1980).

iii) On plateau surfaces, limited oxidation was noticed to occur because of restricted drainage and the presence of permafrost. Reducing conditions were noted to be more predominant on these surfaces.

iv) Virtually all elements on plateau surfaces were identified as exhibiting very low to no mobility. Over half of the elements for which continuous data was available, demonstrated discernable dispersal trends on the plateau surfaces, which cannot be exclusively related to element mobility in this environment.

The principal transporting agents of elements in solution are ground water, and surface water. Element dispersal patterns associated with ground water transport often manifest themselves in the form of fans. Patterns associated with surface water transport produce ribbon-like or lateral patterns. In both cases, the dispersal patterns would be generally confined in extent by drainage conditions. Since element dispersal patterns observed on plateau surfaces are relatively continuous, spanning many kilometres, and extending over extensive plateau surfaces dissected by numerous valleys, it is unlikely they are the result of ground or surface water transport. A possible scenario that would account for such extensive patterns would involve the plateau surfaces of the northern Ruby Range, prior to stream incision, sloping in a preferred northwest direction, with established ground and surface water drainage in a similar direction.

6.4.2.2 Mechanical Factors

There are a number of mechanical transporting agents that have been described in the literature as redistributing regolith and other surficial sediment, as well as eroding and transporting bedrock, which result in various mineral dispersal patterns (Mackie, 1923; Hawkes and Webb, 1962; Washburn, 1980; Harris, 1986; Makarov, 1988; Bennett and French, 1988). Residual soil, colluvium, alluvial sediments, and till are the most common media for dispersal patterns. The degree to which sediment composition reflects the bedrock source from which it was eroded depends on the erosion process, and the extent of sediment modification during transportation, deposition, and diagenesis (Henderson, 1989). Ultimately dispersal patterns will reflect these conditions. Although this discussion centres primarily on clastic fragments (and in particular heavy minerals), geochemical connotations are also inferred. The principal transporting agents include wind, surface water, gravity, and glaciers.

<u>Wind</u>

Erosion and transport by wind is most effective in environments where there is little vegetative cover to protect the sediment. There is a certain proportion of aeolian material in all superficial deposits, but generally the material does not constitute an appreciable amount, and no discernable dispersal patterns are associated with it. The only exception to this is with regard to the deposition of volcanic ash. As was reported earlier, two volcanic eruptions in southeastern Alaska gave rise to two distinct ash-fall deposits (White River Ash). In the Ruby Range, the thickness of ash was measured at 2 - 7.5 cm (Lerbekmo and Campbell, 1968; Downes, 1985). However, with eruptions of this magnitude, coupled with the distance from the source and prevailing winds, the dispersal patterns of the ash-fall material would probably take the form of a relatively uniform blanket dispersal. Local variations in volcanic ash dispersal patterns could be influenced by topography, and local wind patterns.

Detectable aeolian dispersals of minerals derived from known mineralizations exposed at the surface have rarely been reported. However, Hawkes and Webb (1962) speculated that the probable form of dispersal pattern would be fan-like. The aeolian dispersal patterns would likely be limited in extent, and unlike the mineral and element dispersal patterns observed on plateau surfaces, would likely not be continuously traceable over such distances.

Surface Water

The dispersal of minerals by surface water takes place under three main conditions; runoff or sheetwash, stream channels, and bogs or other standing bodies of water. Although there is little evidence to attest to the recent significance of these three conditions, their effect with respect to mineral dispersals in earlier times may have been significant. In the case of runoff or sheetwash, dispersal patterns would tend to be fan-like. In stream channels, the form of dispersal pattern would be ribbon-like. Dispersal patterns in bogs or other standing water would take on a delta fan form. In both sheetwash and stream channel dispersals, the patterns would be limited by the slope of the surface. Dispersal patterns in bogs or other standing water would be confined to the extent of the features.

It is inconceivable that sheetwash and bog (or other standing water) dispersals could account for the patterns observed on the plateau surface unless, i) plateau surfaces of the northern Ruby Range, prior to stream incision, sloped in a northwest direction so as to not inhibit regional-scale sheetwash, or ii) on the plateau surface of the northern Ruby Range, prior to stream incision, a body of water lay in a basin that roughly corresponds to the extremities of the mapped dispersal patterns; at best, highly speculative.

Gravity

Under the influence of gravity, the overburden tends to move downslope either by slow creep, or by more rapid landsliding. Even on more moderate slopes, there is a continual flow of material downslope. Alternate freezing and thawing, or wetting and drying of the overburden tend to facilitate the down-slope movement of materials. In the periglacial environment (both current and relic), numerous mass-wasting processes have been described (e.g., French, 1976;

Washburn, 1980). Mass-wasting is described as "the movement of regolith downslope by gravity without the aid of a stream, a glacier, or wind" (Washburn, 1980). The most common mass-wasting processes occurring in periglacial environments include avalanching, slumping, frost creep, and gelifluction. Price (1973) measured the rates of mass-wasting in the northern Ruby Range at about 1 - 3 cm/year. The data was derived from large solifluction terraces with complete vegetation cover.

Gravity movement generally takes place near the surface but can include all material above bedrock. Displacement by creep may occur on any slope. Its magnitude is influenced by factors such as moisture, freeze-thaw cycles and relative permeable material. As the angle of slope increases, accelerated rates of movements are likely, resulting in debris flows, avalanche, slumping, landslides, etc. In general, the rate of gravity movement on slopes decreases from the surface with depth to bedrock. This results in elements and minerals on the surface being displaced farther from the source of mineralization than those closer to the bedrock. Where slumping, debris flows or landslides occur, dispersal patterns are disrupted as well as displaced (a feature particularly evident in alpine environments). It is possible that a residual dispersal pattern may develop on top of a slump anomaly which in turn may be buried by further slumping.

There is growing research, particularly in the Soviet Union, devoted to the cryogenetic weathering and dispersal of minerals and elements in permafrost regions (Shumilov and Sukhoroslov, 1969; Yarg, 1974; Christotinov and Shur, 1980; Shumilov, 1981, 1983). The occurrence of mineralizations in the permafrost zone represents a virtually unexplored field from the point of view of mineral deposits and dispersal patterns. Currently available information only refer to the formation of unproductive deposits largely due to the multitude of variables encountered in the determination of dispersal patterns. It has become apparent that dispersals of minerals and elements in the permafrost zone proceeds from the earliest stages of weathering simultaneously at three levels;

1) crude dispersal into particles of crushed stone, gravel size and larger,
2) fine dispersal up to the sand fraction, and
3) chemical extraction of elements from minerals.

Within the layer of secular and seasonal temperature fluctuations, cryogenetic weathering dominates which produces pronounced physical dispersal of fragments. Dispersal proceeds most strongly in the active layer where minerals and elements are liberated and dispersed. The elements and minerals are then usually moved downslope by some mass-wasting process where dispersal patterns are formed (Shumilov, 1983). The most common form of dispersal pattern associated with mass-wasting processes are fans. However, due to the gravity driven nature of mass-wasting processes, associated dispersal patterns would be confined to down-slope extents, and therefore be relatively restricted.

Therefore, the question posed is, can mass-wasting processes account for the general northwest trending dispersal patterns mapped on the plateau surfaces of the northern Ruby Range? The answer to this question is a qualified yes, if the plateau surfaces, prior to stream incision, sloped in a prevailing northwest direction, and if mass-wasting processes in that environment could have been relatively uninterrupted over the distances covered by the dispersals. It is more likely that mass-wasting processes served to subsequently modify pre-existing dispersal patterns on local scales, which may account for some of the overprinting of other minor dispersal directions.

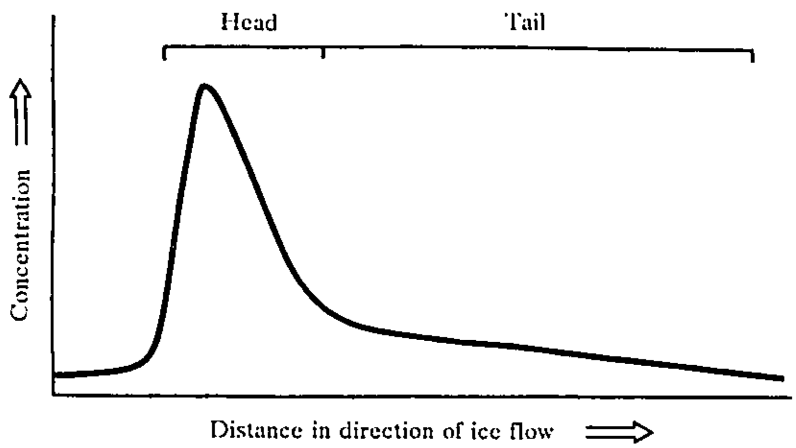
Glaciation

In glaciated terranes, sediments are, in part, a composite of the bedrock traversed, preglacial regolith, preceding weathered glacial deposits, and alluvial material. Therefore, at any given site, glacially derived sediment is not an in situ weathering product, but a lithological summation of source units "up-ice" from the site. It has been concluded in many studies on the transport distance of various size fractions, that the fine-sand fraction represent material from much wider areas than the coarser fractions. This has been demonstrated in Finland by Virkkala (1971), Perttunen (1977), and Taipale et al., (1986), in Sweden by Linden (1975), and in Canada by Harrison (1960), Dreimanis and Vagners (1971), and Shilts (1971, 1973). Debris from any size of source unit is dispersed down-ice to produce a ribbon-shaped or fan-shaped dispersal pattern (DiLabio, 1989 calls the fan-shaped dispersal pattern flame-shaped). Glaciofluvial dispersal patterns are more ribbon-shaped or irregular, whereas patterns in till are fan-shaped.

Shilts (1976) has shown that a plot of the abundance of glacially dispersed debris versus the distance down-ice approximates a negative exponential curve (Figure 31). The highest concentration of a component reaches a peak near the source, and then declines exponentionally down-ice. By analogy, the negative exponential curve proposed by Shilts (1976) can be described in terms of distance decay curves described earlier which represent the decrease of concentration with distance (Taylor 1975, Harries 1982).

Glacially derived dispersal patterns have also been described as having abrupt lateral edges, with a sharp contrast over a short distance between low concentrations outside the dispersal and high concentrations inside (DiLabio, 1989), not unlike those observed in the mineral and element dispersals in this study. The blending, or overriding of patterns derived from up-ice sources produces the mixed lithology which is described as a normal feature of till. The size and shape of a glacially derived dispersal pattern are controlled by the orientation of the source relative to the ice flow, by the size and erodibility of the source, and by the topography of the source and dispersal areas, which can result in broken patterns in rough terrain.

Figure 31
NEGATIVE EXPONENTIAL CURVE



Idealized dispersal curve showing the relationship of the head and tail of the negative exponential curve.

In areas of known mineralization, such as those identified in the study area, the dispersal patterns of minerals and elements identified from the source would be such that the greatest concentration is closest to the mineralization, and decreases down-ice. Hence, the direction of glacial flow can be inferred (Levinson, 1980; Shilts, 1976; Bolviken and Gleeson, 1979). Evidence of glacial dispersal from such zones would include;

i) fan-like dispersal patterns as described above,

ii) a very high concentration of minerals or elements associated with the mineralization directly down-ice from the zone,

iii) low concentration of minerals or elements associated with the mineralization directly up-ice from the zone,

iv) a sharp contrast over a short distance between low concentrations outside the dispersal and high concentrations inside, and

v) overriding of patterns derived from up-ice sources.

In this study, some mineral and element dispersal patterns were reported to be emanating from known zones of mineralization, and furthermore, these minerals and elements were known to be associated with the zones. Other mineral and element dispersals that demonstrate discernable fan-like dispersal patterns may also be associated with currently unknown zones of mineralization. These aspects will be further addressed in the next chapter. Although other mineral and element dispersals did not demonstrate discernable fan-like patterns, this does not necessarily mean they were not glacially derived. As was stated earlier, the blending, or overriding of patterns derived from up-ice sources produces a mixed lithology which is described as a normal feature of glacial dispersal.

Is it possible that glaciation was responsible for the observed mineral and element dispersal patterns observed on the plateau surfaces of the northern Ruby Range? Based on the data presented in this study, and on other studies (case-study and conceptual) reported in the literature (e.g., Levinson, 1980; Shilts, 1976; Bolviken and Gleeson, 1979; DiLabio, 1989), there is some evidence to support the contention that the plateau surfaces of the northern Ruby Range experienced glaciation.

Chapter 7
MINERALIZATION IN THE NORTHERN RUBY RANGE

7.1 Introduction

Although the main focus of this thesis was to determine if the geological evidence from the plateau surfaces could support the belief that these regions were subjected to glaciation and/or other evolutionary processes, the data also made it possible to address an ancillary objective. Specifically, an opportunity at attempting preliminary identification and delineation of mineralized zones that have not yet been identified. Before commencing with this discussion, it is important to again emphasize that detailed geology descriptions of the northern Ruby Range are not currently available. Therefore, the comments and interpretations presented here should be viewed as tentative. Notwithstanding, the observations presented here may add to the geological knowledge of the area.

7.2 Zones of Known Mineralizations

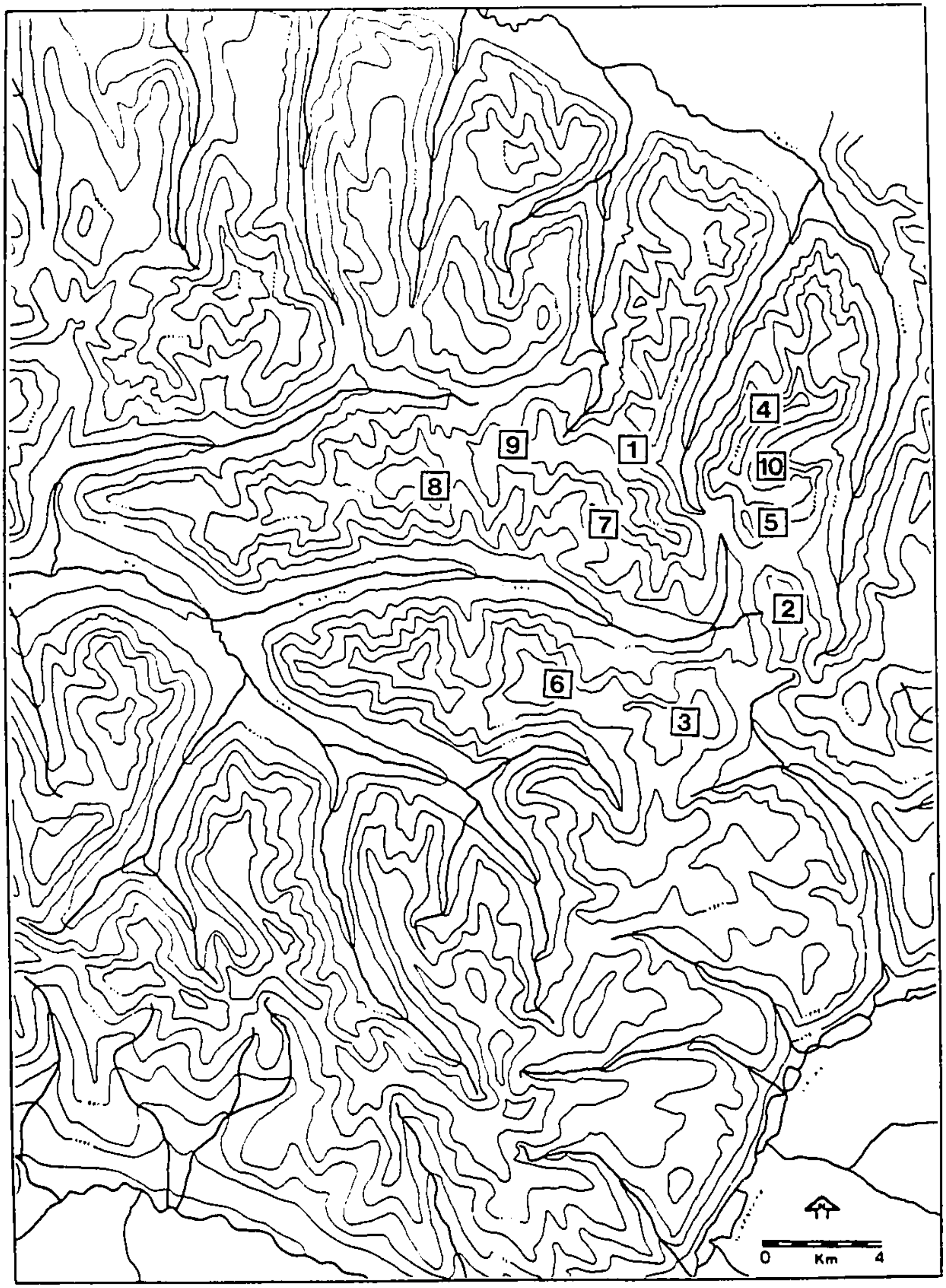
Alaskite has been described by Muller (1967), and further subdivided on the basis of mineral grain size (this study). The alaskite is a major intrusive and has been tentatively dated by Muller as early Tertiary. All of the known mineralizations in the northern Ruby Range study area are associated with alaskite, or in contacts between granitic rocks of the Ruby Range Batholith, and the alaskite.

Location #1 (Figure 32)
At the head of Alaskite Creek, near the divide that separates Rockslide Creek, Muller has identified mineralization that includes tungsten, molybdenite, copper, and fluorite.
Also identified in the vicinity is chalcopyrite and pyrrhotite (Smith, 1972).

Location #2 (Figure 32)
The plateau region that separates Rockslide and Talbot Creek is identified as an alaskite porphyry. It is generally finer grained than the surrounding rock. The porphyry area constitutes a zone of mineralization which has been identified by Smith (1972) as including pyrite, chalcopyrite, molybdenite and bornite.

Location #3 (Figure 32)
On the southern plateau surface at the head of Rockslide Creek, at or near the contact between alaskite and granitic rocks associated with the Ruby Range Batholith, Muller has identified the occurrence of fluorite. Also associated with the mineralization are chalcopyrite, molybdenite and bornite.

Figure 32
ZONES OF KNOWN AND POTENTIAL MINERALIZATIONS



7.3 Geochemical Exploration

In the broad geological sense, geochemical analysis is the practical application of theoretical mineralogical and geochemical principles of exploration. The specific aim is to identify, delineate, and to some extent, attempt to quantify deposits of metals and non-metals (or alternatively, to identify specific types of deposits, map the extent, and estimate the abundance). It is advantageous to have information on local rock types, and associated element occurrences before any interpretation of mineralizations can proceed. These aspects are covered in detail in the literature (e.g., Goldschmidt, 1937, 1954; Warren and Deavault, 1956; Hawkes, 1957; Hawkes and Webb, 1962; Mason, 1966; Krauskopf, 1967; Andrews-Jones, 1968; Saxby, 1969; Holland, 1972), and will not be addressed in any detail. However, a summary of rock types, and associated elements, as well as various pathfinders used to detect mineralizations that are germane to this study area will be presented. The following list presents selected rocks types or occurrences, and element associations (modified from Andrews-Jones, 1968; Levinson, 1980).

ROCK TYPE OR OCCURRENCE	ELEMENT ASSOCIATIONS
Plutonic	
Ultramafic	Cr, Co, Ni, Cu
Mafic	Ti, V, Sc
Alkaline	Ti, Nb, Ta, Zr, RE, F, P
Granitic	Ba, Li, W, Mo, Sn, Zr, Hf, U, Th, Ti
Pegmatites	Li, Rb, Cs, Be, RE, Nb, Ta, U, Th, Zr, Hf, Sc
Hydrothermal Sulfide Ores	
Porphyry Copper Deposits	Cu, Mo, Re
Complex Sulfides	Hg, As, Sb, Se, Ag, Zn, Cd, pb
Low Temperature Sulfides	Bi, Sb, As
Base Metal Deposits	Pb, Zn, Cd, Ba
Precious Metals	Au, Ag, Cu, Co, As, Te, Hg

An extension of these associations is the use of pathfinders or indicators to detect mineralizations. A number of studies have been conducted to associate various element indicators to types of mineral deposits (e.g., Warren and Delavault, 1953, 1956; Ovchinnikov and Grigorvan, 1971; Learned and Boissen, 1973; Coope 1973). The listing below identifies various element indicators and associated types of mineral deposits (modified from Levinson, 1980).

TRACE ELEMENT(S)	TYPE OF MINERAL DEPOSIT
As	Au, Ag; Vein Deposits
As	Au, Ag, Cu, Co, Zn; Complex Sulfide Ores
Mo	W, Sn; Contact Metamorphic Deposits
Mn	Ba, Ag; Vein Deposits; Porphyry Copper
Mo, Te, Au	Porphyry Copper
pb, Cr, Cu, Ni, Co	Platinum in Ultramafic Rocks
Zn	Ag, pb, Zn; Sulfide Deposits
Zn, Cu	Cu, pb, Zn; Sulfide Deposits

With the aid of the geochemical dispersal patterns, it was possible to identify areas that exhibited pronounced anomalous values (Figure 32). Only three element dispersal patterns were shown to exhibit anomalous values. The dispersal pattern for cadmium identifies the area of highest concentration (anomaly of >57 ppm) as being associated with an area of known mineralization at location #3. The dispersal patterns for arsenic and lead show anomaly values (>1636 ppm, and >10610 respectively) occurring in an area not identified with known mineralization at location # 4 on Figure 32. The importance of these anomalies with respect to being able to associate rock types or occurrences is inconsequential. However, the occurrence of anomalous arsenic values in the dispersal pattern may indicate an Au, Ag vein deposit, or Au, Ag, Cu, Co, Zn complex sulfide ore deposit in the general vicinity of the anomaly. If all geochemical dispersal patterns are interpreted in a similar manner, regardless of whether the areas of highest concentrations are associated with anomalous values (as demonstrated below), relationships with respect to rock types or mineral deposits are still negligible (Figure 32).

LOCATION ON FIGURE 32	ELEMENT(S)
Location #1	Ag
Location #2	Ag
Location #3	Cd, U, Mo
Location #4	As, pb
Location #5	Sn
Location #7	Cr

Therefore, based on the geochemical concerns addressed in this section, the only conclusion that can be made is with regard to the occurrence of anomalous arsenic values in the dispersal pattern, which may indicate an Au, Ag vein deposit, or Au, Ag, Cu, Co, Zn complex sulfide ore deposit in the general vicinity of the ridge that separates Alaskite and Talbot Creeks.

7.4 Heavy Mineral Considerations

Anomalous values displayed in the heavy mineral dispersal trends were also identified (Figure 32) and are as follows:

LOCATION ON FIGURE 32	HEAVY MINERAL(S)
Location #2	Hematite
Location #3	Hematite, Opaques
Location #5	Hematite
Location #6	Opaques
Location #8	Zircon
Location #9	Zircon
Location #10	Garnet, Hematite

In only two instances mineral dispersal patterns demonstrate more than one anomaly in the same vicinity, Location #3 and #9. The occurrence of anomalous hematite and opaques at Location #3 is not unexpected since it is a zone of known mineralization. Although mineralization is not reported at Location #9 where anomalous occurrences of garnet and hematite are evident, the anomalies may be indicative of mineralization.

If all mineral dispersal patterns are considered, regardless of whether the areas of highest concentrations are associated with anomalous values or not, numerous multiple-mineral centres of highest concentration are evident. These are shown on Figure 32, and are detailed as follows:

LOCATION ON FIGURE 32	HEAVY MINERAL(S)
Location #1	Goethite, Hornblende
Location #2	Hematite
Location #3	Hematite, Opaques
Location #4	Goethite
Location #5	Hematite, Orthopyroxene
Location #6	Opaques
Location #7	Allanite
Location #8	Zircon, Orthopyroxene
Location #9	Zircon, Hornblende
Location #10	Garnet, Hematite, Epidote

7.5 Conclusions

All of the reported mineralizations (Location #'s 1, 2, and 3) are associated with alaskite, or in contacts between granitic rocks of the Ruby Range Batholith, and alaskite. The ridge that separates Rockslide and Talbot Creek (Location #2) is identified as an alaskite porphyry, and it is likely that the porphyry extends further north to include Location #'s 5, 10, and 4. The textural characteristics of the alaskite in these area are similar, consisting of fine- grained material.

The area of contact between granitic rocks of the Ruby Range Batholith, and alaskite on the south plateau surface at the head of Rockslide Creek (Location #3) has not been clearly defined. Location #6 is probably situated in the vicinity of a similar contact. Based in part on the reported mineralization in the area of Location #3 (chalcopyrite, molybdenite and bornite), and on the occurrence of what appeared to be isolated areas of oxidation on some exposed rock surfaces, a few small gossans may be located in the area.

The extremely gossan-rich area in the vicinity of Location #1 at the head of Alaskite Creek may be indicative of other gossan areas in the environs of Location #'s 7, 8, and 9. Hand specimens from Location #1 displayed mineralization on a sheared surface, which may be indicative of the mineralization in this area.

Although it is highly speculative to make conclusions with respect to other mineralizations based on the inconclusive evidence regarding element and mineral dispersals reported in this chapter, apart from the three known zones of mineralization, there is a possibility that all locations identified (Location #'s 1 - 10) may be indicative of mineralized areas. In the context of this discussion, this evaluation should therefore be viewed as preliminary, with the intent of delineating areas that may warrant more rigorous geological study.

Chapter 8
SYNOPSES AND CONCLUSIONS

8.1 Introduction

The main objective of this thesis was to offer explanations on the evolution of the plateau surfaces in the northern Ruby Range, Yukon Territory. Specifically, to determine whether the geological and geomorphological evidence from the plateau surfaces could support the belief that these regions were subjected to glaciation, or did the evidence indicate other evolutionary processes. The techniques employed consisted of an examination of surficial deposits, with the use of heavy mineral and geochemical composition analysis, and other provenance related studies (erratics) as well as geomorphological mapping and interpretation.

The more traditional geological techniques of geochemical and heavy mineral compositional analyses have been used previously in both geological and glacial related studies in the Canadian Shield and other regions of Canada, and have aided in the description of environments. However these types of analyses have had limited application in alpine environments with extensive plateau regions, such as those experienced in the northern Ruby Range.

In pursuit of the thesis objective, the following research questions were posed:

i) Can the compositional trends observed in the heavy mineral and geochemical data on the plateaus provide geological evidence (lacking in previous descriptions of these regions) which can contribute to the solution to the question of how the plateau landscape evolved? Specifically, depending upon the genesis of the plateau sediments (ie. local, glacial, etc.), what processes were responsible for these dispersals (glaciation, periglacial, aeolian, tectonic)?

ii) Are similar compositional trends observed in the valleys of the Ruby Range, and what are the implications of these patterns?

Although the main focus of this thesis was to determine whether the geological and geomorphological evidence from the plateau surfaces could support the belief that these regions were subjected to glaciation, or other evolutionary processes, the data also afforded the opportunity to address mineralizations within the study area. This theme constituted an ancillary objective in this thesis.

8.2 Geomorphology

The geomorphology of the study area is quite diverse, and is demonstrated by the variation in terrains between valleys and plateaus. The most common geomorphological expressions in valleys are glacially derived, and are manifested in the form of U-shaped valleys, cirques, moraines, and till. However, although most valleys contain an array of glacially related landforms and deposits, others exhibit little evidence of glaciation, a situation which may be attributed to glacier surging. The major valleys in the study area are oriented east-west, and include Gladstone, Rockslide, Raft, and Talbot creeks, all of which are connected through a system of low passes. It is apparent from the dimensions of these valleys along with morainal evidence, that they were glaciated repeatedly from both up-valley and down-valley directions. Other, less extensive valleys, generally oriented north-south, aiso take on a glaciated U-shape, but not all contain glacial depositional features. Also conspicuous in some upper valley locations are mass-wasting processes and forms, both fossil and current.

The plateau regions have generally been regarded as an old erosion surface of the Yukon Plateau which has been uplifted and incised by streams, resulting in narrow valleys which were later modified by glacial activity. Although the evolution of the plateau regions have not been described in any great detail in the literature, accounts of glaciation in the Ruby Range are accompanied by descriptions that identify the plateaus as generally being an unglaciated, weathered landscape of unknown age. The terrain features of the plateau regions are in sharp contrast to those in valleys. The plateaus are characterized by relatively smooth, undulating surfaces, void of any conspicuous landforms except for a few tor-like structures along with felsenmeer, and generally resemble a periglacial landscape. Felsenmeer grades into a system of massive pattern ground formations which, from the lichen cover, appear to have been inactive for a very long period of time. Some denuded kame and kettle topography is evident at lower plateau elevations occupying col-like positions.

The fact that uplift has occurred throughout the region is beyond doubt. However, the net amount of uplift, and the sustainable nature of the uplift during the Pleistocene can only be inferred. The existing data suggest a considerable disparity between current rates of denudation and uplift. Clearly, uplift exceeded denudation in the Ruby Range, and it was suggested that net uplift during the Pleistocene may have been on the order of 1,668 m. As well, it was noted that an unknown amount of uplift, throughout the Pleistocene, was likely attributed to glacial unloading affects.

Glaciological implications with regard to uplift brought to light considerations which were previously not accounted for in discussions of complete regional ice cover of the Ruby Range. Specifically, since the altitudinal extent of the Ruby Range was considerably less in early Pleistocene than what is currently observed, in order to glaciate the Ruby Range, the thickness of ice build up in the Kluane Lake basin may not have had to be as extensive as previously estimated, and therefore not have been as crucial to the glaciation of the Ruby Range. In view of these considerations, an evolutionary scenario was suggested whereby the northern Ruby Range (and surrounding regions) may have been glaciated (perhaps several times) in late Tertiary and/or early Pleistocene, undergone uplift with stream incision intensifying as uplift progressed, followed by middle - late Pleistocene glaciation(s) which modified the valleys. The plateau surfaces would have experienced periglacial processes continually throughout the Pleistocene, rendering any glacigenic material residual.

8.3 Heavy Mineral Analysis: Composition

Heavy mineral analysis of the fine-sand fraction was performed to identify and count mineral species, determine sources (bedrock and others), and to assess the nature of sediment transport. The geology of the study area and surrounding regions has been mapped at a scale which generally lacks the detailed lithological descriptions necessary to determine mineral associations, and in particular heavy mineral compositions. Within the bounds imposed by the available geological descriptions, coupled with heavy mineral suites common to rock types encountered in the study area and bordering regions, as well as bedrock analysis carried out in this study, some broad groupings of heavy mineral associations were recognized in the surficial sediment that may have association with bedrock sources from within and outside the study area, or from other origins such as volcanic ash and meteorites. Heavy mineral identification was performed on the basis of mineral familiarity and an appreciation of the reported lithology of the study area and bordering regions. They included; groupings of minerals classified as garnet, orthopyroxene and opaque, as well as epidote, allanite, hornblende, zircon, goethite, and hematite.

Because of the sketchy nature of the geological information within and outside the study area, conclusions with respect to the relationship between heavy mineral assemblages in sediment samples, and rock types were somewhat qualified. Clearly, a more thorough geological evaluation of the northern Ruby Range and surrounding regions would serve to substantiate or refute the findings presented here. These are:

i) The average proportion of heavy minerals identified was similar for both plateau and valley sediment samples. The range of the proportions was found to vary considerably in most minerals. The significance of these comparisons were not clear, but probably indicated that the sediments from both plateau and valley sites did not vary greatly in heavy mineral species and proportions, but demonstrated a variable range.

ii) There was no appreciable difference in the heavy mineral assemblages between plateau and valley sediment samples, except for the relative lack of allanite from plateaus, and goethite from valleys. The presence of goethite on the plateaus and absence in the valleys may be related to the altered bedrock.

iii) The data did not support a clear association between heavy mineral assemblages from plateau and valley sediments, and those derived from bedrock within or outside the study area. However, the sediment assemblages may not be exclusively related to local bedrock sources. Heavy mineral input from outside the study area has been demonstrated by Dewez (1988)

iv) The provenance of orthopyroxene, goethite, and allanite may be attributed, in part, to bedrock from outside the study area, volcanic ash, and meteoritic impacts, or a combination of these. The occurrence of these minerals may also be associated with mineral alterations.

v) It is certain that mineralizations occur within the study area that have not yet been identified.

8.3.1 Dispersal Trends

Background and anomaly values were determined for heavy minerals. Except for minor differences, the background values for most heavy minerals in both plateau and valley samples were generally similar. The anomaly values were also relatively similar for most minerals. These values were taken to indicate a relatively uniform concentration of minerals in the sediment from both plateaus and valleys, as exemplified by background values, and somewhat varying dispersal patterns as demonstrated by the differences in some anomaly values.

The data as a whole showed no evidence to suggest minerals demonstrate an overwhelming dispersal trend in either an up-valley or down-valley direction. As well, mineral anomaly values only occurred occasionally, and were not concentrated in a preferred valley location (i.e., in upper, middle, or lower valley sites). These findings were attributed to glaciation on both a local scale (small ice-caps, cirque, and valley glaciers) with down-valley ice movement, and regional ice penetrating the area in an up-valley direction. Since these glaciological conditions were ongoing since at least the middle Pleistocene (Nisling Ice Sheet), and even simultaneously during the Ruby Ice Sheet advance, much of the glacial sediments in the valleys must have been reworked numerous times by up-valley and down-valley glacier flows.

Heavy mineral compositional trends were plotted with the aid of Synagraphic Computer Mapping. Computer generated false, or exaggerated patterns were noted and accounted for. The trends were described with reference to "head and tail" portions of the dispersals. Directional lines described as characterizing the broad, general dispersal trends were drawn on the plots. The purpose of this analysis was to interpret regional dispersal patterns within the study area. Overall, the heavy minerals demonstrated a preferred northwest dispersal trend. The areas of highest concentration of four mineral dispersals (allanite, hematite, hornblende, and opaques) were associated with known zones of mineralization. The dispersal patterns were mapped from 12 to 20 km over extensive, undulating plateau surfaces that were dissected by numerous valleys from their areas of highest concentration.

8.4 Geochemical Analysis: Dispersal Trends

Geochemical examination of the silt and clay fraction consisted of a multi-element analysis of twenty-one elements. Background values for sixteen of the elements analyzed in both plateau and valley sites were relatively similar. Anomaly values for twelve elements did not show an appreciable difference. Nine elements demonstrated considerable differences, with plateau anomaly values in all cases being at least twice the corresponding valley levels. The relatively similar background values for the majority of elements in the sediments from plateau and valley sites was interpreted as indicating a relatively uniform concentration of elements, but because of differences in nine element anomaly values, distribution of these elements was not uniform. The data indicated there were no discernable dispersal trends in either an up-valley or down-valley direction. Element anomalies only occurred occasionally in valley sites, and were not concentrated in preferred valley location (i.e., upper, middle, or lower valley sites).

Eight elements exhibited prominent dispersal trends on plateau surfaces that facilitated plotting. These include; lead (pb), molybdenum (Mo), chromium (Cr), cadmium (Cd), silver (Ag), arsenic (As), uranium (U), and tin (Sn). The dispersal patterns of the remaining elements, copper (Cu), cobalt (Co), nickel (Ni), manganese (Mn), bismuth (Bi), iron (Fe), zinc (Zn), vanadium (V), tellurium (Te), tungsten (W), antimony (Sb), selenium (Se), and gold (Au), displayed a random trend. In many of the silt and clay fraction plateau samples, some elements were not detected at all, or were of insufficient sample to make a determination. The dispersal trends were interpreted as demonstrating a regionally preferred northwest direction, however, other directional trends were noted and were attributed to a number of factors including; i) inadequate sampling density on some plateau regions, ii) plotting technique, iii) processes affecting dispersals on a local scale, iv) dispersals attributed to unknown mineralizations, and v) multiple regional trends of varying extent, some possibly relic.

8.5 Dispersal Processes

The discussion on dispersal processes addressed the central research question posed in the introduction to this thesis: "Can the compositional trends observed in the heavy mineral and geochemical data on the plateaus provide geological evidence (lacking in previous descriptions of these regions) which can contribute to the solution to the question of how the plateau landscape evolved? Specifically, depending upon the genesis of the plateau sediments (ie. local, glacial, etc.), what processes were responsible for these dispersals (glaciation, periglacial, aeolian, tectonic)? In addressing this question, a number of processes were examined in association with dispersal forms. Included were considerations of processes relating to; geological terranes, chemical mobilities, wind, surface water, gravity, and glaciation.

Both the heavy mineral and geochemical data indicated there were no discernable dispersal patterns in either an up-valley or down-valley direction. This may be attributed to two factors;

i) mobility of elements in the acidic and oxidizing valley environment, and

ii) the up-valley and down-valley glacial flows that occurred throughout the Pleistocene.

The random dispersals of minerals and elements in the valleys can be characterized as secondary, epigenetic dispersals, and can be further classified as resembling clastic, hydromorphic, or biogenic patterns.

Heavy mineral and element analyses performed on surficial sediments from the plateau surfaces indicated that some of the material may not be of local origin, and discernable dispersal trends were evident in certain heavy minerals and elements which were traced to known sources of mineralizations. The most striking characteristics of these trends were their continuous form, spanning many kilometres, and their patterns which extending over extensive plateau surfaces that were dissected by numerous valleys. The following is an summary of the discussion on forms of dispersals pertaining to processes.

i) The fact that the north to northwest alignment of the Tracey Arm Terrane, and the more restricted, but relatively well defined northwest mineral and element dispersal patterns are similar in trend, is likely coincidental.

ii) It is doubtful that the overall terrane attributes played a major role in mineral and element dispersal patterns. It is more probable that the patterns are related to varying terrane conditions on a local (bedrock related) as opposed to a macro scale.

iii) Virtually all elements on plateau surfaces were identified as exhibiting very low to immobile mobilities. Over half of the elements for which continuous data was available, demonstrated discernable dispersal trends on the plateau surfaces, which cannot be exclusively related to element mobility in this environment.

iv) The probable form of local wind induced dispersal patterns would be fan-like. However, the aeolian dispersal patterns would likely be limited in extent, and unlike the mineral and element dispersals observed on plateau surfaces, may not be continuously traceable over such large distances.

v) Mass-wasting processes could account for the general northwest trending dispersal patterns mapped on the plateau surfaces, if the surfaces, prior to stream incision, sloped in a prevailing northwest direction, and if mass-wasting processes in that environment were relatively uninterrupted over the distances covered by the dispersals. However, it is more likely that mass-wasting processes served to subsequently modify pre-existing dispersal patterns on local scales, which may account for some of the overprinting of other minor dispersal directions.

vi) Based on the data presented in this study, and from other studies (case-study and conceptual), there is evidence to support the contention that the general northwest dispersal patterns observed on the plateau surfaces of the northern Ruby Range may have been glacially derived.

8.6 Probable Mode of Dispersals

After carefully appraising the prevalent dispersal forms indicative of various dispersal agents or processes, as well as considering the environmental (geological terrane) context in which the heavy mineral and element patterns were observed, it is concluded that the regional patterns could be attributed to local terrane characteristics, and glaciation, with glaciation being the main dispersal mode.

Terrane Characteristics

It has been reiterated throughout this thesis that the relative large scale mapping of the geology has resulted in a highly generalized description of bedrock variations and composition, especially in the study area. Therefore, detailed analysis of variation in bedrock composition and mineralization on a local scale (ie. within the study area), are not well known. The discussion in this thesis relating to mineralizations helped to illustrate this point by identifying potential zones of mineralization. The lack of detailed geological information in the study area and surrounding regions, did not allow for definitive conclusions to be made with respect to the relationship between mineral assemblages for sediment samples and rock types. Observation of a provisional nature were made which included a listing of major heavy mineral constituents in various bedrock, and sediment samples. The source of some minerals identified from the sediment samples of both plateau and valley sites was uncertain, and were described as possibly being attributed to external sources. The minerals in question consisted of orthopyroxene, goethite, and allanite. Although evidence was presented to support the belief that at least some amount of these heavy minerals may be of non-local origin, alternative evidence was also put forth to dispute this contention.

If the observed mineral and element dispersal patterns on the plateau surfaces were derived from in situ weathering of bedrock, then clearly, the form of dispersal patterns would generally reflect superjacent patterns. Any subsequent secondary epigenetic processes would eventually alter these patterns. The question is, what process(es) were involved that resulted in such extensive patterns? There are possibilities, but as pointed out in earlier discussions, they would likely involve certain conditions. For example, gravity mass-wasting processes could account for the general northwest trending dispersal patterns, if the surfaces, prior to stream incision, sloped in a prevailing northwest direction, and if mass-wasting processes in that environment were relatively uninterrupted over the distances covered by the dispersals. However, it was also pointed out that it was more likely that gravity induced mass-wasting processes served to subsequently modify pre-existing dispersal patterns on local scales. Could the dispersal patterns be an exclusive reflection of mineralizations? Although there are, without doubt, other mineralized zones in the study area, the ones that are known, have been delineated, and it does not seem feasible their extent could account for the fan-like dispersals.

Clearly, a more detailed geological understanding of the northern Ruby Range would serve to substantiate many of the assertions made with respect to dispersal trends, specifically with regard to mineralizations, and would provide the data necessary to verify the heavy mineral assemblages collated for rock types in the region.

Glaciation

The purpose for plotting mineral and element concentrations from the surficial sediment on plateau surfaces was to interpret regional dispersal patterns. Although minor variations in dispersal direction were apparent in some patterns, it was evident that most of these patterns showed a northwest dispersal trend. The most striking characteristics of these trends are their continuous form, spanning many kilometres, and patterns extending over extensive plateau surfaces that are dissected by numerous valleys. After careful consideration of various modes of dispersion, it was concluded that glaciation was the most likely process that would account for these patterns.

Part of the rational for this conclusion is that in areas of known mineralization (as those identified in the study area), the dispersal patterns of minerals and elements identified from the source would be such that the greatest concentration is closest to the mineralization, and decreases down-ice. Evidence of glacial dispersal from such zones would include;

i) fan-like dispersal patterns,

ii) a very high concentration of minerals or elements associated with the mineralization directly down-ice from the zone,

iii) low concentration of minerals or elements associated with the mineralization directly up-ice from the zone,

iv) a sharp contrast over a short distance between low concentrations outside the dispersal and high concentrations inside, and

v) possible overriding of patterns derived from up-ice sources.

The dispersal patterns of some minerals and elements in this study emanated from known zones of mineralization. As well, other mineral and element dispersals demonstrate discernable fan-like patterns and may be associated with unmapped zones of mineralization. Although other mineral and element dispersals did not demonstrate discernable fan-like patterns, this did not necessarily mean they were not glacially derived. The blending, or overriding of patterns derived from up-ice sources produces a mixed lithology which is described as a normal feature of glacial dispersal. The minor variations in dispersal trends may also be attributed to any one of a number of processes which served to subsequently modify dispersal patterns on local scales.

There are two major ramifications with respect to these assertions:

i) The surficial sediment on the plateau surfaces is comprised of locally and externally derived material, with at least some of the externally derived material being glacigenic. Other externally derived deposits are aeolian, and perhaps meteoritic.

ii) Regional glaciation of the plateau surfaces of the northern Ruby Range occurred at some time.

The statement with regard to the glaciogenic origin of at least some of the surficial sediment on plateau surfaces is naturally based on the premise that the area was glaciated, and as such, stands by itself. However, suggesting regional glaciation of the plateau surfaces of the northern Ruby Range requires further consideration.

The current depiction of the evolution of the plateau regions of the Ruby Range is one of an old, uplifted, weathered landscape of unknown age. The plateaus of the northern Ruby Range have also been depicted as unglaciated. The glacial history of the area has been described by Bostock (1966), and Muller (1967). Bostock (1966) identified four glaciations, each being progressively less extensive:

i) Nansen Glaciation (late Tertiary - early Pleistocene),
ii) Klaza Glaciation (early - middle Pleistocene),
iii) Reid Glaciation (early - middle Wisconsin),
iv) McConnell Glaciation (middle - late Wisconsin).

Muller's (1967) described three progressively less extensive ice-sheets during the Pleistocene in the Kluane lake region:

i) Nisling Ice-Sheet (pre-Wisconsin),
ii) Ruby Ice-Sheet (early Wisconsin),
iii) St. Elias Advance (late(?) Wisconsin).

The Nisling Ice Sheet correlates to Bostock's (1966) Reid Glaciation, whereas the Ruby Ice Sheet and probably the St. Elias Advance, are equivalent to the McConnell Glaciation. The extent of the Nisling Ice-Sheet is apparent from traces of glaciation above the level and beyond the extent of the Ruby Ice-Sheet. The regional patterns of glacial flow and limits in the Ruby Range and bordering areas have already been reported. However, although evidence of former glaciation in valleys is abundant, to date, no evidence of major, regional glaciation affecting the plateau regions of the northern Ruby Range has been presented.

It is proposed in this study that the heavy mineral and element dispersal patterns observed on the plateau regions of the northern Ruby Range may be the evidence lacking in previous studies which can confirm that the plateau surfaces were glaciated. The glaciation (or glaciations) would, in accordance to the reported glacial history of the area, represent pre-Nisling Ice (pre-Reid) glaciation(s) (late Tertiary or early Pleistocene). Glaciation(s) in late Tertiary and/or early Pleistocene could have preceded major stream incision, hence valleys may be post-glacial (i.e., formed after glaciation in late Tertiary and/or early Pleistocene). Significant uplift of plateau regions may have occurred after the late Tertiary and/or early Pleistocene glaciation(s) (i.e., during middle - late Pleistocene). Whether uniform uplift rates were experienced throughout the entire Tracey Arm Terrane is uncertain.

In summary, it is conceivable that the northern Ruby Range (and surrounding regions) may have been glaciated (perhaps several times) in late Tertiary and/or early Pleistocene, undergone uplift with stream incision intensifying as uplift progressed, followed by middle - late Pleistocene glaciation(s) which modified the valleys. The plateau surfaces would have experienced periglacial processes (perhaps severe) continually throughout the Pleistocene rendering any glacigenic material residual.

Alternatively, if the dispersal patterns are attributed to some small independent ice-caps which were sustained on plateau areas of the Ruby Range during the Ruby glaciation as Hughes et al., (1969) have identified, then it may be that these ice caps were much more extensive than originally reported. However, if this was plausible, the plateau surfaces would have had to experience extreme periglacial conditions during the later part of the Wisconsin, and it is uncertain if these conditions, over such a short time period, would have been sufficient to account for the periglacial terrain.

BIBLIOGRAPHY

Agterberg, F.P., and Chung, C.F., (1975): A computer program for polynomial trend-surface analysis. Geological Survey of Canada Paper 75-21, 51 pp.

Alexander, C.S., and Price, L.W., (1980): Radiocarbon Dating of the Rate of Movement of Two Solifluction Lobes in the Ruby Range, Yukon Territory. Quaternary Research, Vol. 13, pp 365-379.

Allan, R.J., and Hornbrook, E.H.W., (1970): Development of geochemical techniques in permafrost terrain. Canadian Mining Journal, Vol. 91, pp 45-49.

Allan, R.J., et al., (1972): Regional geochemical exploration in the Coppermine River area, District of Mackenzie; a feasibility study in permafrost terrain. Geological Survey of Canada Paper 71-33.

Allen, J.F., Findlay, A., (1972): Macmillan tungsten deposit. Proceedings, 4th Northern Resources Conference, Whitehorse, pp 97-101.

Andrews-Jones, D.A., (1968): The application of geochemical techniques to mineral exploration. Colorado School of Mines, Mineral Industry, Bulletin 11(6).

Archer, A.R., and Main, C.A., (1971): Casino Yukon - A Geochemical Discovery of an Unglaciated Arizona - Type Porphyry, Geochemical Exploration. CIM Special Vol. 11, pp 67-77.

Barasko, J., (1970): Referenced in Smith (1970).

Barakso, J.J., and Tarnocai, C., (1970): A mercury determination method and its use for exploration in British Columbia. Canadian Institute of Mining and Metallurgy, Bulletin (63), pp 501-505.

Beach, H.H., (1943): Moose Mountain and Morley map-areas, Alberta. Geological Survey of Canada Memoir 236, 74 pp.

Bear, F.E., (1964): Chemistry of the soil. American Chemical Society, Monogram Series 160, Reinhold, 515 pp.

Belyea, H.R., (1960): The story of the mountains in Banff National Park. Geological Survey of Canada, Misc. Report #1, 42 pp.

Benedict, J.B., (1973): Chronology of cirque glaciation, Colorado Front Range. Journal of Quaternary Research, Vol. 3, pp 584-599.

Berggren, W.A., Burke, L.H., Cita, M.B., Cooke, H.B.S., Funnell, B.M., Gartner, S., Hays, J.D., Kennett, J.P., Opdyke, N.D., Pastouret, L., Shackleton, N.J., and Takayanagi, Y., (1980): Toward a Quaternary time scale. Quaternary Research, 13, pp 277-302.

Berry, L.G., and Mason, B., (1959): Mineralogy. W.H. Freeman and Co., San Francisco, 630 pp.

Berthois, L., (1947): Nouvelle methode d'tude quantitative des assemblages de minraux lourds. Institut Ocanographique de Monaco, Bulletin 44 (912), 36 pp.

Blackburn, W.H. and Dennen, W.H., (1988): Principles of Mineralogy. C. Brown Publishers, Iowa, 413 pp.

Blow, W.H., (1969): Late Middle Eocene to Recent planktonic foraminiferal biostratigraphy. In, P. Bronnermann and H.H. Renz (eds.), Proceedings of International Conference on Planktonic Micro-fossils, Geneva, 1967, pp 199-421.

Boellstorff, J.D., (1973): Correlating and dating some older "Pleistocene" tills in the midcontinent. Geological Society of America Abstracts, Vol. 5(4), 301 pp.

Boellstorff, J.D., (1974): Fission-track ages of Pearlette family ash beds: comment. Geology, Vol. 2, pp 21.

Boellstorff, J.D., (1976): The succession of late Cenozoic volcano ashes in the Great Plains: a Progress Report. Kansas Geological Survey Guidebook Series 1, pp 37-71.

Boellstorff, J.D., (1978): North American Pleistocene stages reconsidered in light of probable Pliocene-Pleistocene continental glaciation. Science, 202, pp 305-307.

Bolviken, B., and Gleeson, G.F., (1979): Focus on the use of soils for geochemical exploration in glaciated terrain. In, P.J. Hood (ed.), Geophysics and Geochemistry in the search for Metallic Ores, Geological Survey of Canada, Report 31, pp 295-326.

Boniwell, J.B., and Dujardin, R.A., (1964): Discovery and exploration of the Poirier ore deposit. CIM Bulletin, Vol. 57(629), pp 945-951.

Bostock, H.S., (1948): Physiography of the Canadian Cordillera, with special reference to the area north of the fifty-fifth parallel. Geological Survey of Canada Memoir 247, 106 pp.

Bostock, H.S., (1952): Geology of the northwest Shakwak Valley, Yukon Territory. Geological Survey of Canada Memoir 267, 54 pp.

Bostock, H.S., (1966): Notes of glaciation in central Yukon Territory. Geological Survey of Canada Paper 65-36, 18 pp.

Bowen, D.Q., (1978): Quaternary Geology: A Stratigraphic Framework For Multidisciplinary Work. Pergamon Press, Oxford, 221 pp.

Bowles, J.E., (1975): Engineering Properties of Soils and Their Measurements. McGraw-Hill Book Company, pp 33-51.

Boyle, R.W., (1965): Geology, geochemistry and origin of the Lead-Zinc-Silver deposits of Keno Hill-Galena Hill area, Yukon Territory. Geological Survey of Canada Bulletin 111.

Boyle, R.W., (1974): Elemental associations in mineral deposits and indicator elements of interest in geochemical prospecting (revised). Geological Survey of Canada Paper 74-45.

Bradshaw, P.M.D., et al., (1971): Exploration Geochemistry, Part 6, Areas of Continental Glaciation. Canadian Mining Journal, Vol. 92(12), pp 29-38.

Bradshaw, P.M.D. (ed.), (1975): Conceptual models in exploration geochemistry. The Canadian Cordillera and Canadian Shield. Journal of Geochemical Exploration, Vol. 4(1), pp 1-207.

Brazier, S., Sparks, R.S.J., Carey, N.S., Sigurdsson, H., and Westgate, J.A., (1983): Bimodal grain size distrubition and secondary thickening in air-fall ash layers. Nature, Vol. 301, pp 115-119.

Brigham-Grette, J., (1989): Geochronology. In, L.D. Carter, et al. (eds), Late Cenozoic History of the Interior Basins of Alaska and the Yukon, United States Geological Survey, Circular 1026, pp 107.

Brooks, A.H., (1906): The geography and geology of Alaska. United States Geological Survey Professional Paper 45.

Brooks, R.R., (1972): Geobotany and Biogeochemistry in Mineral Exploration. Harper and Row, 290 pp.

Brosge, W.P., and Reiser, H.N., (1972): Geochemical reconnaissance in the Wiseman and Chandalar districts and adjacent region southern Brooks Range, Alaska. United States Geological Survey, Professional Paper 709.

Brown, R.J.E., (1978): Permafrost. In, Hydrological Atlas of Canada, Canadian National Committee for the International Hydrological Decade, Ottawa Ministry of Supply and Services, Plate 32.

Burrows, C.J., and Lucas, J., (1967): Variation in two New Zealand glaciers during the past 800 years. Nature, 216, pp 467-468.

Cairnes, D.D., (1907): Recent developments in mining in the southern Yukon. Canadian Mining Journal, Vol. 28, pp 87-88 and pp 121-122.

Cairnes, D.D., (1915): Upper White River District, Yukon. Geological Survey of Canada Memoir 50.

Campbell, R.B., (1967): Glenlyon map-area, Yukon Territory. Geological Survey of Canada Memoir 352.

Campbell, I., (1985): Two Pollen Diagrams from the Ruby Range, Southwestern Yukon Territory, Canada. Unpublished B.Sc. Thesis, Dept. of Geology, University of Ottawa, 81 pp.

Campbell, R.B. and Dodds, C.J., (1978): Operation Saint Elias, Y.T. Geological Survey of Canada Paper 78-1A, pp 35-41.

Campbell, R.B. and Dodds, C.J., (1982): Open Files 829, 830 and 831. Geological Survey of Canada.

Camsell, C., (1906): The Peel River and Tributaries. In, H.S. Bostock (ed.), Yukon Territory, Geological Survey of Canada Memoir 284, pp 173-205.

Capps, S.R., (1910): Rock glaciers in Alaska. Journal of Geology, 18, pp 359-375.

Chamberlain, T.C., (1894): Proposed genetic classification of Pleistocene glacial formations. Journal of Geology, Vol. 2, pp 515-538.

Chorley, R., (1966): The application of statistical methods. In, G.Dury (ed.), Essays in Geomorphology, Heinemann, London, pp 275-387.

Clague, J.J., (1979): The Denali Fault System in southwest Yukon Territory - A geological hazard? Geological Survey of Canada Paper 79-1A, pp 169-178.

Clague, J.J., (1987): Quaternary Stratigraphy and History, Williams Lake, British Columbia. Canadian Journal of Earth Sciences, Vol. 24(1), pp 147-158.

Clague, J.J., (1989): Quaternary Geology of the Canadian Cordillera. In R.J. Fulton (ed.), Quaternary Geology of Canada and Greenland Geology of Canada No. 1, pp 17-26.

Clague, J.J., Armstrong, R.W., and Warner, B.G., (1982): Late Quaternary Geology of eastern Graham Island, Queen Charlotte Islands, British Columbia. Canadian Journal of Earth Sciences, Vol. 19, pp 1786-1795.

Clayton, L., and Moran, S.R., (1982): Chronology of late Wisconsinian glaciation in middle North America. Quaternary Sciences Review, Vol. 1, pp 55-82.

Cloke, P.L., (1966): The geochemical application of Eh-pH diagrams. Journal of Geology, Education 14, pp 140-148.

Cockfield, W.E., (1922): Silver-lead deposits of Davidson Mountains, Mayo District, Yukon. Geological Survey of Canada, Summary Report for 1921 A, pp 1-6.

Cooke, H.B.S., (1984): Recognizing different Quaternary chronologies: A multidisciplinary problem. In, W.C. Mahaney (ed.), Correlation of Quaternary Chronologies, Geobooks, England, pp 1-14.

Cooke, H.B.S., (1972): Pleistocene Chronology: Long or Short? Maritime Sediments, Vol. 8(1), pp 1-12.

Coope, J.A., (1968): The application of geochemical prospecting principles to mineral exploration in Alaska and Yukon Territory. Science and the North, Proceedings of the 19th Alaskan Science Conference, Whitehorse, Alaska Division of the American Association for the Advancement of Science, pp 19.

Coope, J.A., (1973): Geochemical prospecting for Porphyry - Copper type mineralization: A review. Journal of Geochemical Exploration, Vol. 2, pp 81-102.

Copps, S.R., (1915): An ancient volcanic eruption in the Upper Yukon Basin. United States Geological Survey, Professional Paper 95-D, pp 59-64.

Crandell, D R., (1965): The glacial history of western Washington and Oregon. In H.E. Wright and D.G. Frey (eds.), The Quaternary of the United States, Princeton University Press, Princeton, pp 341-355.

Currie, L.D., (1992): Metamorphic rocks in the Tagish Lake area, northern Coast Mountains, British Columbia: a possible link between Stikinia and parts of Yukon-Tanana Terrane. In, Current Research, Part A: Geological Survey of Canada Paper 92-1A, pp 199-208.

Czudek, T., (1964): Periglacial slope development in the area of the Bohemian Massif in Northern Moravia. Builetyn Peryglacjalny, (14), pp 169-194.

Dahl, R., (1963): Shifting ice culmination, alternating ice covering, and ambulant refuge organisms. Geografiska Annaler, 49(2-3), pp 122-138.

Dansgaard, W., et al., (1971): Climate record revealed by the Camp Century ice core. In, K.K. Turekian (ed.), Late Cenozoic Glacial, Ages, Yale University Press, New Haven, pp 37-56.

Daugherty, R., (1974): Science in Geography 2, Data Collecting. Oxford University Press, London, 70 pp.

Davis, J.C., (1973): Statistics and Data Analysis in Geology. Wiley, New York, 550 pp.

Davis, N.F.G. and Mathews, W.H., (1944): Four phases of glaciation with illustrations from southwestern British Columbia. Journal of Geology, Vol. 52, pp 403-413.

Dawson, G.M., (1877): Note on some of the more recent changes in level of the coast of British Columbia and adjacent regions. Canadian Naturalist, New Ser., Vol. 8, pp 241-248.

Dawson, G.M., (1878): On the superficial geology of British Columbia. Geological Society of London, Quarterly Journal, Vol. 34, pp 89-123.

Dawson, G.M., (1879): Notes on the glaciation of British Columbia. Canadian Naturalist, New Ser., Vol. 9, pp 32-39.

Dawson, G.M., (1981): On the later physiographical geology of the Rocky Mountain region of Canada. Royal Society of Canada, Transactions, Vol. 8, Sect. 4, pp 3-74.

Dawson, G.M., (1881): Additional observations on the superficial geology of British Columbia and adjacent regions. Geological Society of London, Quarterly Journal, Vol. 37, pp 272-285.

Dawson, G.M., (1886): Preliminary report on the physical and geological features of that portion of the Rocky Mountains between latitudes 49° and 50°30'. Geological Survey of Canada, Annual Report, New Ser., Vol. 1 (1855), Report B, 169 pp.

Dawson, G.M., (1888): Recent observations on the glaciation of British Columbia and adjacent regions. Geological Magazine, New Ser., Vol. 5, pp 347-350.

Dawson, G.M., (1889): Report on an exploration in the Yukon District, Northwest Territories, and adjacent northern portion of British Columbia, 1887. Geological Survey of Canada, Annual Report 1887-88, Part 1, Report B.

Dawson, G.M., (1889): Glaciation of high points in the southern interior of British Columbia. Geological Magazine, New Ser., Vol. 6, pp 350-352.

Dawson, G.M., (1890): On the glaciation of the northern part of the Cordillera, with an attempt to correlate the events of the glacial period in the Cordillera and Great Plains. American Geologist, Vol. 6, pp 153-162.

Dawson, G.M., (1891): On the later physiographical geology of the Rocky Mountain region in Canada, with special reference to changes in elevation and to the history of the glacial period. Royal Society of Canada, Transactions (1890), Vol. 5, sec. 4, pp 3-74.

DeBlij, H.J., (1977): Human Geography Culture, Society and Space. John Wiley and Sons, Inc., New York, 444 pp.

Deer, W.A., Howie, R.A., and Zussman, J., (1982): An Introduction to the Rock Forming Minerals. Longman House, England, 528 pp.

Demek, J., (1964): Castle koppies and tors in the Bohemian Highland (Czechoslovakia). Builetyn Peryglacjalny, (14), pp 195-216.

Denton, G.H., (1974): Quaternary Glaciations of the White River Valley, Alaska, with a Regional Synthesis for the Northern St. Elias Mountains, Alaska and Yukon Territory. Geological Society of America Bulletin, Vol. 85, pp 871-892.

Denton, G.H., and Stuiver, M., (1967): Late Pleistocene Glacial Stratigraphy and Chronology, Northeastern St. Elias Mountains, Yukon Territory, Canada. Geological Society of America Bulletin, Vol. 78, pp 485-510.

Denton, G.H. and Stuiver, M., (1969): Neoglacial Chronology, Northeastern St. Elias Mountains, Canada. Icefield Ranges Research Project Scientific Results, Volume 1, pp 173-186.

Denton, H.G. and Armstrong, R.L., (1969): Miocene-Pliocene Glaciations in Southern Alaska. American Journal of Science, Vol. 267, pp 1121-1142.

Dewez, V., (1988): Sources, limites d'extension et fusion des glaciers au Wisconsiniensuperieur dans la chaine Ruby et le bassin d'Aishihik, Territoire du Yukon. Unpublished Ph.D Thesis, Department of Geography, University of Ottawa, 320 pp.

Dickson, J.A.D., (1966): Carbonate identification and genesis as revealed by staining. Journal of Sedimentary Petrology, Vol. 36(2), pp 491-505.

Dilabio, R.N.W., (1989): Terrain Geochemistry in Canada. In R.J. Fulton (ed.), Quaternary Geology of Canada and Greenland, Geological Society of America, pp 647-663.

Dixon, C., and B. Leach., (1977): Sampling methods for geographical research. Institute of British Geographers, Vol. 17, 47 pp.

Dodds, C.J., (1982): Operation Saint Elias, Yukon Territory. Geological Survey of Canada.

Douglas, R.J.W., et al., (1980): Macmillan River Yukon District of Mackenzie-Alaska. Geological Survey of Canada, Map 1393 A.

Downes, H., (1985): Evidence for magma heterogeneity in the White River Ash (Yukon Territory). Canadian Journal of Earth Sciences, Vol. 22, pp 929-934.

Dreimanis, A., (1958): Tracing ore boulders as a prospecting method in Canada. Canadian Institute of Mining and Metallurgy, Vol. 51, pp 73-80.

Dreimanis, A., (1971): Procedures of till investigation in North America. In, Goldthwait R.P. (ed.), Till - A Symposium, Ohio State University Press, Columbus, pp 27-37.

Dreimanis, A. and Vagners, V.J., (1971): Bimodal distribution of rock and mineral fragments in basal tills. In, R.P. Goldwaith (ed.), Till - A Symposium, Ohio State University Press, Ohio, pp 27-37.

Dreimanis, A., and Hicock, S.R., (1985): Glaciotectonic structures as useful ice-movement indicators in glacial deposits: Four Canadian case studies. Canadian Journal of Earth Sciences, Vol. 22(3), pp 339-346.

Dryden, A.L., (1931): Accuracy in percentage representation of heavy mineral frequences. National Academy of Sciences Proceedings, 17, pp 233-238.

Dunbar, C.O., and Rodgers, J., (1957): Principles of Stratigraphy. John Wiley and Sons, Inc., New York, 356 pp.

Dyke, A.S., (1976): Tors and associated weathering phenomena, Somerset Island, District of Franklin. Geological Survey of Canada Paper 76-1B, pp 209-216.

Dyke, A.S., et al., (1982): Configuration and dynamics of the Laurentide Ice Sheet during the late Wisconsin maximum. Geographie physique et Quaternaire, Vol. 36(1-2), pp 5-14.

Dyke, A.S, (1984): Quaternary geology of Boothia Peninsula and northern District of Keewatin, central Canadian Arctic. Geological Survey of Canada Memoir 407, 26 pp.

Dyke, A.S., Dredge, L.A., and Vincent, J.S., (1982): Configuration and dynamics of the Laurentide Ice Sheet during the Late Wisconsin maximum. Géographie Physique et Quaternarie, pp 5-14.

Dyke, A.S. and Prest, V.K., (1987): Late Wisconsinan and Holocene history of the Laurentide Ice Sheet. Geographie physique et Quaternarie, Vol. 41, pp 237-263.

Ellwood, D., (1981): Appmap Users Manual. Geological Survey of Canada, Internal Document.

Embleton, C., and King, C.A.M., (1968): Glacial and Periglacial Geomorphology. Arnold, London, 608 pp.

Emiliani, C., (1955): Pleistocene temperatures. Journal of Geology, 63, pp 538-578.

England, J.H., and Andrews, J.T., (1973. Broughton Island: a reference area for Wisconsin and Holocene chronology and sea level changes on eastern Baffin Island. Boreas, 2, pp 17-32.

Erdmer, P., (1989): The Nisling Schist in eastern Dezadeash map area, Yukon. In, Current Research Part E, Geological Survey of Canada Paper 89-1E, pp 139-144.

Erdmer, P., (1990): Studies of the Kluane and Nisling assemblages in Kluane and Dezadeash map areas, Yukon. In, Current Research Part E, Geological Survey of Canada Paper 90-1E, pp 107-111.

Erdmer, P., (1991): Metamorphic terrane east of Denali fault between Kluane Lake and Kusawa Lake, Yukon Territory. In, Current Research, Part A, Geological Survey of Canada Paper 91-1A, pp 37-42.

Ermengen, S.V., (1957): Geochemical prospecting in Chibougamau. Canadian Mining Journal, Vol. 68(4), pp 99-104.

Evans, S.G., (1989): Geomorphic response to endogenic processes. In, R.J. Fulton (Ed.), Chapter 9, Quaternary Geology of Canada and Greenland, Geological Survey of Canada, Geology of Canada, No. 1.

Fairbridge, R.W., and Michel, J.P., (1980): Dictionary of Earth Sciences. Mason Pub., New York, 411 pp.

Fareau, S.A., (1992): Report of fieldwork in the southern Big Salmon metamorphic complex, Teslin map area, Yukon Territory. In Current Research Part A, Geological Survey of Canada Paper 92-1A, pp 267-277.

Feo-Codecido, G., (1956): Heavy-mineral techniques and their application to Venezuelan stratigraphy. American Association Petroleum Geologist, Bulletin 40, pp 984-1000.

Fernald, A.T., (1965): Glaciation in the Nabesna River Area, Upper Tanana River Valley, Alaska. Geological Survey Professional Paper 525-C, pp 120-127.

Flint, R.F., (1947): Glacial Geology and the Pleistocene Epoch. John Wiley and Sons, Inc., London, 489 pp.

Flint, R.F., (1957): Glacial and Pleistocene Geology. John Wiley and Sons, Inc., New York, 553 pp.

Flint, R.F., (1965): The Pliocene-Pleistocene boundary. In, Wright H.E. and Frey, D.G., (eds.), International Studies on the Quaternary, Geological Society of America Special Paper 84.

Flint, R.F., (1971): Glacial and Quaternary Geology. John Wiley and Sons, Inc., New York, 892 pp.

Folk, R.L., (1968): Petrology of Sedimentary. University of Texas, pp 16-43.

Forbes, E., (1846): On the connection between the distribution of the existing fauna and flora of the British Isles. Geological Survey of Great Britain, Memoir 1, pp 336-342.

Ford, T.D., (1981): Dolomite Tors and Sandfilled Sink Holes in the Carboniferous Limestone of Derbyshire, England. University of Leicester, pp 387-397.

Foscolos, A.E., Rutter, N.W., and Hughes, O.L., (1977): The use of pedological studies in interpreting the Quaternary history of central Yukon Teritory. Geological Survey of Canada Bulletin 271, 48 pp.

French, H.M., (1976): The Periglacial Environment. Longmans Pub. Ltd., London and New York, 309 pp.

Frye, J.C., (1973): Pleistocene succession of the central interior United States. Journal of Quaternary Research, Vol. 3(2), pp 275-286.

Fullerton, D.S., (1980): United States Geological Survey Professional Paper 1089, 52 pp.

Fulton, R.J., (1967): Deglaciation Studies in Kamloops Region; An Area of Moderate Relief, British Columbia. Geological Survey of Canada Bulletin 154, 36 pp.

Fulton, R.J., (1984): Correlation of Quaternary events in Canada. In, A. Billard, O. Conchon, and F.W. Shotton (eds.), Quaternary Glaciations in the Northern Hemisphere, Geological Survey, Prague, Czechoslovakia, IGCP Project 73/1/24, Report #9.

Fulton, R.J., Fenton, M.M., and Rutter, N.W., (1984): Summary of Quaternary Stratigraphy and History, Western Canada. In, R.J. Fulton (Ed.), Quaternary Stratigraphy of Canada, A Canadian Contribution to IGCP Project 24, Geological Survey of Canada Paper 84-10, pp 69-83.

Galloway, M.C., (1972): Statistical analyses of regional heavy mineral variation, Hawkesbury Sandstone and Narrabeen group (Triassic), Sydney Bassin. Journal of Geological Society of Australia, 19, pp 65-76.

Garrels, R.M., and Christ, C.L., (1965): Minerals, Solutions and Equilibria. Harper and Row, New York, 450 pp.

Geddes, J.D., (1981): Ground Movements and Structures. Proceedings of the II International Conference, Institute of Science and Technology, University of Wales, John Wiley and Sons, New York, 964 pp.

Gleeson, C.F., and Cormier, R., (1971): Evaluation by geochemistry of geophysical anomolies and geological targets using overburden sampling at depth. In, R.W. Boyle and J.I. McGerrigle (eds.), Geochemical Exploration, Canadian Institute of Mines and Metallurgy, Special Volume 11, pp 159-165.

Goldschmidt, V.M., (1937): The principles of distribution of chemical elements in minerals and rocks. Chemical Society Journal, pp 655-673.

Goldschmidt, V.M., (1954): Geochemistry. Oxford University Press, Oxford, 730 pp.

Goodell, B.C., (1972): Water quantity and flow regime - influences of land use, especially forestry. In, H.O. Slaymaker and H.J. McPherson (ed.), Mountain Geomorphology Geomorphological Processes in the Canadian Cordillera, Tantalus Research Ltd., Vancouver, B.C. Geographical Series No. 14, pp 197-206.

Granier, C.L., (1973): Introduction la prospection gochimique d gites metallifre. Masson et Cie., Paris.

Gravenor, C.P., and Bayrock, L.A., (1956): Stream-trench systems in east-central Alberta. Research Council of Alberta, Preliminary Report 56-4.

Green, L.H., (1972): Geology of Nash Creek, Larsen Creek and Dawson map-areas, Yukon. Geological Survey of Canada Memoir 364, (Includes Maps 1282A, 1283A, 1284A).

Green, P., (1979): Tracking down the past. New Scientist, 22, pp 624-629.

Griffiths, J.C., (1967): Scientific Method in Analysis of Sediments. McGraw-Hill, New York, 508 pp.

Gunn, G.B., (1967): The origin of diamonds in the drift of the north central United States. Journal of Geology, Vol. 75, pp 232-233.

Gy, P., (1979): Sampling of Particulate Materials: Theory and Practice. Elsevier, New York, 431 pp.

Hallberg, G.R., (1980): Pleistocene stratigraphy in east- central Iowa. Technical Information Series # 10, Iowa Geological Survey, 168 p.

Hamilton, T.D., (1986): Late Cenozoic glaciations of the central Brooks Range. In, T.D. Hamilton, et al. (eds.), Glaciation in Alaska, Alaska Geological Survey, pp 9-50.

Hamilton, T.D., (1986): Correlation of Quaternary Glacial Deposits in Alaska. In, V. Sibrava, D.Q. Brown, and G.M. Richmond (eds.), Quaternary Glaciations in the Northern Hemisphere, Report of the International Geological Correlation Programme Project 24, pp 171-180.

Hamilton, T.D. and Thorson, R.M., (1983): The Cordilleran ice sheet in Alaska. In, Porter. S.C. (ed.), Late-Quaternary Environments of the United States, Volume 1. The Late Pleistocene, University of Minnesota Press, Minneapolis pp 38-52.

Hamilton, T.D. and Bischoff, J.L., (1984): Uranium-series dating of fossil bones from the Canyon Creek vertebrate locality in central Alaska. In, Reed, K.M. and Bartsch-Winkler, S. (eds.), United States Geological Survey in Alaska - Accomplishments during 1982, United States Geological Survey Circular 939, pp 26-29.

Hamilton, T.D., Reed, K.M., and Thorson, R.M., (1986): Glaciation in Alaska - Introduction and overview. In T.D. Hamilton et al., (eds), Glaciation in Alaska, Alaska Geological Survey, pp 1-8.

Hansen, V.L., Mortensen, J.K., and Armstrong, R.L., (1989): U-Pb, Rb-Sr and K-Ar isotope constraints for ductile deformation and related metamorphism in the Teslin sature zone, Yukon-Tanana terrane, south-central Yukon. Canadian Journal of Earth Sciences, Vol. 26, pp 2224-2225.

Hansuld, J.A., (1967): Eh and pH in geochemical prospecting. In, E.M. Cameron (ed.), Proceedings, Symposium on Geochemical Prospecting, Geological Survey of Canada Paper 66-54, 282 pp.

Harries, S.A., (1982): Identification of permafrost zones using selected permafrost landforms. In H.M. French (ed.), Proceedings of the Fourth Canadian Permafrost Conference, National Research Council of Canada, Ottawa, pp 45-58.

Harrison, P.W., (1960): Original bedrock composition of Wisconsin till in central Indiana. Journal of Sedimentary Petrology, 30, pp 432-446.

Hawkes, H.E., (1957): Principles of geochemical prospecting. United States Geological Survey Bulletin 1000-F, pp 225-355.

Hawkes, H.E., and Webb, J.S., (1962): Geochemistry in Mineral Exploration. Harper and Row, New York, 415 pp.

Hayes, C.W. (1892): An expedition through the Yukon District. National Geographical Magazine, 4, pp 117-162.

Hays, T, et al., (1969): Pliocene - Pleistocene sediments of the Equatorial Pacific; their paleomagnetic, biostratigraphic and climatic record. Bulletin of Geological Society of America, 80, pp 1481-1514.

Heginbottom, J.A. and Vincent, J.-S. (eds.) (1986): Correlation of Quaternary deposits and events around the margin of the Beaufort Sea: Contributions from a joint Canadian-American workshop, April 1981. Geological Survey of Canada, Open File 1237, 125 pp.

Hem, J.D., (1970): Study and interpretation of the chemical characteristics of natural water. United States Geological Survey, Water Supply Paper 1473, 363 pp.

Henderson, P.J., (1983): A study of the heavy mineral distribution the bottom sediments of Hudson Bay. In, Current Research, Geological Survey of Canada Paper 83-1A, pp 347-351.

Henderson, P.J., (1989): Provenance and Depositional Facies of Surficial Sediments in Hudson Bay, A Glaciated Epeiric Sea. Unpublished Ph.D. Thesis, Department of Geology, University of Ottawa, 288 pp.

Henderson, P.J., Kaszycki, C.A., Klassen, R.A., Shilts, W.W., and Thorleifson, L.H., (1987): Drift composition and glacial ice flow patterns beneath the central portion of the Laurentide Ice Sheet (Abst.). In, Programme with Abstracts, International Union for Quaternary Research, XII International Congress, National Research Council of Canada, Ottawa, pp 185.

Hicks, S.D. and Shofnos, W., (1965): The determination of land emergence from sea level observations in southeast Alaska. Journal of Geophysical Research, Vol. 70, pp 3315-3320.

Holland, H.D., (1972): Granites, solutions and base metal deposits. Economical Geology, 67, pp 281-301.

Hopkins, D.M. (ed.), (1967): The Bering Land Bridge. Stanford, Stanford University Press, 495 pp.

Hornbrook, E.H.W., and Allan, R.J., (1970): Geochemical exploration feasibility study within the zone of continuous permafrost: Coppermine River Region, Northwest Territories. Geological Survey of Canada Paper 70-36.

Horner, R., (1983): Seismicity in the St.Elias region of northwestern Canada and southeastern Alaska. Bulletin of the Seismological Society of America, Vol. 73, No. 4, pp 1117-1137.

Horowitz, A.J., (1991): A Primer on Sediment-Trace Element Chemistry. Lewis Pub., Chelsea, Michigan, 136 pp.

Hough, J.L., (1958): Geology of the Great Lakes. University of Illinois Press, Urbana, Illinois, 313 pp.

Howarth, R.J. (ed.), (1983): Handbook of Exploration Geochemistry, Statistics and Data Analysis in Geochemical Prospecting, Volume 2. Elsevier Scientific Publishing Company, Amsterdam, 437 pp.

Hubert, J.F., (1971): Analysis of Heavy-Mineral Assemblages. In, R.C. Carver (ed.), Procedures in Sedimentory Petrology, Wiley-Interscience, New York, pp 453-478.

Hughes, O.L., (1969): Distribution of open-system pingos in Central Yukon Territory with respect to glacial limits. Geological Survey of Canada Paper 69-34, 8 pp.

Hughes, O.L., (1972): Surficial geology of northern Yukon Territory and northwestern District of Mackenzie, Northwest Territories. Geological Survey of Canada Paper 69-36, 11 pp.

Hughes, O.L., (1986): Quaternary Chronology of Yukon Territory and Western District of MacKenzie. In, J.A. Heginbottom, et al. (eds), Correlation of Quaternary Deposits and Events Around the Margin of the Beaufort Sea; Contribution From a Joint Canadian-American Workshop, April 1984, Geological Survey of Canada Open File Report 1237, pp 13-17.

Hughes, O.L., (1989): Quaternary chronology, Yukon and Western District of Mackenzie. In, L.D. Carter, et al. (eds.), Late Cenozoic History of the Interior Basins of Alaska and the Yukon, United States Geological Survey, Circular 1026, pp 25-29.

Hughes, O.L., (1990): Surficial Geology and Geomorphology, Aishihik Lake, Yukon Territory. Geological Survey of Canada Paper 87-29, 23 pp.

Hughes, O.L., Campbell, R.B., Muller, J.E., and Wheeler, J.O., (1969): Glacial limits and flow patterns, Yukon Territory, south of 65 North Latitude. Geological Survey of Canada Paper 68-34, 9 pp.

Hughes, O.L., Rampton, V.N., and Rutter, N.W., (1972): Excursion A-11, Quaternary Geology and Geomorphology, Southern & Central Yukon. 24th International Geological Congress, Montreal, 59 pp.

Hughes, O.L., et al., (1972): Quaternary geology and geomorphology, southern and central Yukon (northern Canada). XXIVth International Geological Congress, Guidebook Excursion A, 59 pp.

Hughes, O.L., et al., (1981): Upper Pleistocene stratigraphy, paleoecology and archaelogy of the Northern Yukon interior, Eastern Beringia. Arctic, Vol. 34, pp 329-365.

Hughes, O.L., et al., (1983): Northern Yukon Territory and Mackenzie Delta, in French and Heginbotton (eds.). Field Guide to Excursion B-3, Fourth International Conference on Permafrost, Fairbanks, Alaska, pp 1-57.

Hughes, O.L., Rutter, N.W., and Clague, J.J., (1989): Yukon Territory Quaternary stratigraphy and history, Cordilleran Ice-Sheet. In, R.J. Fulton (ed.), Geology of Canada and Greenland, Geology of Canada No. 1, pp 58-63.

Hunt, C.B., (1972): Geology of Soils Their Evolution, Classification and Uses. W.H. Freeman, San Francisco, 344 pp.

Hyvarinen, L., et al., (1973): Modern boulder tracing in prospecting. In, M.J. Jones (ed.), Prospecting in Areas of Glaciated Terrain, Institute of Mining and Metallurgy, London, pp 87-95.

Imbrie, J., and van Andel, T.H., (1964): Vector analysis of heavy-mineral data. Geological Society of America, Bulletin 75, pp 1131-1155.

INQUA, (1979): Symposium on Genesis and Lithology of Quaternary Deposits. Balkema, Rotterdam, 441 pp.

Ivanov, O.P., (1966): Basic factors in the development of sulfide deposit oxidation zones under permafrost conditions. Geokhim., (9), Geochemistry Int., 3, pp 875-884.

Ives, J.D., (1958): Glacial geomorphology of the Torngat Mountains, northern Labrador. Geographical Bulletin, 12, pp 47-75.

Ives, J.D., (1974): Biological refugia and the munatak hypothesis. In, J.D. Ives, and R.G. Barry (eds.), Arctic and Alpine Environments, Methuen & Co. Ltd., London, England, pp 604-636.

Ives, J.D., (1975): Delimitation of surface weathering zones in eastern Baffin Island, northern Labrador, and Arctic Norway: A discussion. Geological Society of America Bulletin, 86, pp 1096-1100.

Ives, J.D., Nichols, H., and Short, S., (1976): Clacial history and palaeoecology of North eastern Nouveau-Québec and Northern Labrador. Arctic, 29(1), pp 48-52.

Jackson, L.E. Jr., (1987): Terrain Inventory and Quaternary History of Nahanni Map Area, Yukon Territory and Northwest Territories. Geological Survey of Canada Paper 86-18, 23 pp.

Jackson, L.E., Jr. (co-ordinator) (1989): The influence of the Quaternary geology of Canada on man's environment. In, R.J. Fulton (ed.), Chapter 12, Quaternary Geology of Canada and Greenland. Geology of Canada No. 1.

Jackson, L.E., Rutter, N.W., Hughes, O.L., and Clague, J.J., (1989): Glaciated fringe Quaternary stratigraphy and history, Canadian Cordillera. In, R.J. Fulton (ed.), Quaternary Geology of Canada and Greenland. Geology of Canada No. 1, pp 63-68.

Jaeger, C., (1979): Rock Mechanics and Engineering Cambridge University Press, New York, 523 pp.

James, C.H., (1970): A rapid method for calculating the statistical precision of geochemical prospecting analysis. Institute of Mining and Metallurgy, Transactions, Section B(79), pp 91-96.

James, P.A., (1972): The periglacial geomorphology of the Rankin Inlet area, Keewatin, Northwest Territories, Canada. Biuletyn Peryglacjalny, (21), pp 127-151.

Johnson, P.G,, (1981): The structure of a talus derived rock glacier as deducted from its hydrology. Canadian Journal of Earth Sciences, Vol 18, pp 1422-1430.

Johnson, P.G., (1983): Glaciological conditions and geomorphological interpretation. Research Notes, University of Ottawa, 43, 20 pp.

Johnson, P.G., and Kodybka, R.J., (1984): Problems of Quaternary glacial correlation due to regional ice/local ice interactions, Ruby Range, Southwest Yukon. In, W.C. Mahaney (ed.), Correlation of Quaternary Chronologies, Geobooks, England, pp 411-422.

Johnson, S.T., (1988): The Tectonic Setting of the Aishihik Battolith, SW Yukon. In, G. Abbott (ed.), Yukon Geology, Volume 2, pp 37-41.

Kaszycki, C.A., (1987): Glacial geomorphology of the southern Canadian Shield. In, W.L. Graf (ed.), Geomorphic systems of North America. Geological Society of America, Centennial Special Volume 2, pp 150-155.

Kaszycki, C.A. and Dilabio, R.N.W., (1986): Surficial Geology and Till Geochemistry, Lynn Lake - Leaf Rapids region, Manitoba. In, Current Research Part B, Geological Survey of Canada Paper 86-1B, pp 245-256.

Kauranne, L.K. (ed.), (1976): Conceptual models in exploration geochemistry. Journal of Geochemical Exploration, Vol. 5, pp 173-420.

Kearney, M.S., and Luckman, B.H., (1981): Evidence for late Wisconsin/early Holocene climate change in Jasper National Park. In, W.C. Mahaney (ed.), Quaternary Paleoclimate, Geo Abstracts Ltd., Norwick, England, pp 85-106.

Keithy, L., (1988): Principles of Environmental Sampling. American Chemical Society, Washinton, D.C., 458 pp.

Kendrew, W.G., and Kerr, D.P., (1955): The Climate of British Columbia and the Yukon Territory. Queens Printer, Ottawa, 222 pp.

Kerr, P.F., (1959): Optical Mineralogy. McGraw-Hill Book Company, New York, 442 pp.

Kerr, F.A., (1936): The physiography of the Cordilleran region of northern B.C. and adjacent areas. Transactions, Royal Society of Canada, Section 4, pp 137-154.

Kershaw, K., (1973): Quantitative and Dynamic Plant Ecology. Edward Arnold, London, 37 pp.

Kindle, E.D., (1945): Geological reconnaissance along the Canol Road, from Teslin River to Macmillan Pass, Yukon. Geological Survey of Canada Paper 45-21, (Including Preliminary Map 45-21A).

Kindle, E.D., (1949): Dezadeash map-area, Yukon. Geological Survey of Canada Paper 48-24 (Preliminary Map 48-1019A).

Kindle, E.D., (1953): Dezadeasch map-area, Yukon Territory. Geological Survey of Canada Memoir 268, 59 pp.

King, H.F., (1958): Notes on occurences in highly metamorphosed Precambrian rocks. Proceedings of the Australian Institute of Mining and Metallurgy.

Klassen, R.W., (1978): A unique stratigraphic record of late Tertiary - Quaternary events in southeastern Yukon. Canadian Journal of Earth Sciences, Vol. 15, pp 1884-1886.

Klassen, R.A., (1982): Quaternary Stratigraphy and Glacial History of Bylot Island, Northwest Territories, Canada. Unpublished Ph.D. Thesis, University of Illinois, 161 pp.

Klassen, R.W., (1987): The Tertiary Pleistocene Stratigraphy of the Liard Plain Southeastern Yukon Territory. Geological Survey of Canada Paper 86-17, 16 pp.

Klassen, R.A., and Shilts, W.W., (1977): Glacial dispersion of uranium in the district of Keewatin, Canada. In, G.R. David (ed.), Prospecting in Areas of Glaciated Terrain, Institute of Mining and Metallurgy, pp 80-88.

Kline, J.T. and Bundtzen, T.K., (1986): Two glacial records from west-central Alaska. In, T.D. Hamilton, et al. (eds), Glaciation in Alaska, Alaska Geological Survey, pp 123-150.

Koch, G.S. Jr., and Link, R.F., (1971): The coefficient of variation - A guide to the sampling of ore deposits. Economic Geology, 66, pp 293-301.

Kostaschuk, R.A., and Smith, D.G., (1983): Late Quaternary lacustrine and deltaic sedimentation near Banff, Alberta. Canadian Journal of Earth Sciences, Vol. 20(12), pp 1834-1842.

Krauskopf, K.B., (1967): Introduction to Geochemistry. McGraw-Hill.

Krinsley, D.B., (1965): Pleistocene geology of the southwest Yukon Territory, Canada. Journal of Glaciology, Vol. 5(40), pp 385-397.

Krumbein, W.C, and Graybill, F.A., (1965): An Introduction to Statistical Models in Geology. McGraw-Hill, New York, 475 pp.

Krumbein, W.C., and Pettijohn, F.J., (1938): Manual of Sedimentary Petrology. Appelton-Century-Crofts Inc., New York, 549 pp.

Krumbein, W.C., and Sloss, L.L., (1951): Stratigraphy and Sedimentation. W.H. Freeman and Company, San Francisco, California, 660 pp.

Kukla, G., and Nakagawa, H., (1977): Late Cenozoic magnetostratigraphy comparisons with bioclimate and lithozones. Quaternary Research, 7, pp 283-293.

Lacasse, D., (1987): Glaciation locale durant le Pléistocène supérieur et l'Holocène dans les chaînes Ruby et Dalton, S.O. du Territoire du Yukon. Thèse de maîtrise, Département de Géographie, Université d'Ottawa.

Lanphere, M.A., (1978): Displacement history of the Denali fault system, Alaska and Canada. Canadian Journal of Earth Sciences, Vol. 15, No. 5, pp 817-822.

Learned, R.E., and Boissen, R., (1973): Gold; a useful pathfinder element in the search for porphyry copper deposits in Puerto Rico. Geochemical Exploration, 1972, IMM, London, pp 93-103.

Lematre, S., (1985): L'influence de la circulation atmosphrique sur le regime thermique dans la vallee du ruisseau Gladstone au cours de l'ete de 1982. M.A. Thesis, University of Ottawa, 103 pp.

Lemke, R.W., Laird, W.M., Tipton, J.A., and Lindvall, R.M., (1965): Quaternary geology of northern Great Plains. In, H.E. Wright, Jr. and D.G. Frey (ed.), The Quaternary of the United States, Princeton University Press, Princeton, New Jersey, pp 15-27.

Lerbekmo, J.F. and Campbell, F.A., (1969): Distribution, composition and source of the White River Ash, Yukon Territory. Canadian Journal of Earth Sciences, Vol. 6, pp 109-116.

Leverett, F., (1899): The Illinois glacier lobe. United States Geological Survey, Monogram 38, 817 pp.

Levinson, A.A., (1980): Introduction to Exploration Geochemistry. Applied Publishing Ltd., Chicago, 612 pp.

Linden, A., (1975): Till petrographical studies in an Archean bedrock area in southern central Sweden. Striae 1, 57 pp.

Linton, D.L., (1950): Unglaciated enclaves in glaciated regions. Journal of Glaciology, 1, pp 451-453.

Linton, D.L., (1955): The problems of tors. Geographical Journal, 121, pp 470-487.

Loughnan, F.C., (1969): Chemical Weathering of the Silicate Minerals. Elsevier.

Lowe, J.J. and Walker, M.J.C., (1984): Reconstructing Quaternary Environments. Longman, New York.

Luckman, B.H., and Osborne, G.O., (1979): Holocene glacier fluctuations in the middle Canadian Rocky Mountains. Quaternary Research, Vol. 11, pp 52-77.

Luepke, G. (ed.), (1985): Economic Analysis of Heavy Minerals in Sediments. Benchmark Papers in Geology Series Vol. 86, Van Nostrand Reinhold Company, New York, 321 pp.

Lyell, C., (1839): Nouveaux Elments de Gologie Paris.

Mackay, J.R., and Mathews, W.H., (1973): Geomorphology and Quaternary history of the Mackenzie River Valley near Fort Good Hope, Northwest Territories, Canada. Canadian Journal of Earth Sciences, Vol. 10(1), pp 26-41.

Mann, D.H., (1986): Wisconsin and Holocene glaciations of southeast Alaska. In, T.D. Hamilton et al. (eds), Glaciation in Alaska, Alaska Geological Survey, pp 237-265.

Martini, E., and Worsley, T., (1970): Standard Neogene calcarious nanno-plankton zonation. Nature, 235, pp 289-290.

Mathews, W.H., (1989): Development of Cordilleran landscapes during the Quaternary. In, R.J. Fulton (ed.), Geology of Canada and Greenland, Geology of Canada No. 1, pp 32-34.

Mathewson, C.C., (1981): Engineering Geology. Charles E. Merrill Pub. Co., Toronto, 450 pp.

McAnerney, J.M., (1967): Early gold mining in frozen ground. Polar Notes (7), pp 37-44.

McCammon, R.B., (1974): The statistical treatment of geochemical data. In, A.A. Levinson (ed.), Introduction to Exploration Geochemistry, Applied Pub. Ltd., 612 pp.

McConnell, R.G., (1905): Report on the Klondkie Coldfields. Geological Survey of Canada Annual Report Part B, pp 1-7.

McConnell, R.G., (1907): Report on gold values in the Klondike high level gravels. Geological Survey of Canada Publication 979.

McDonald, B.C., and Kelly, R.G., (1968): Procedures used in sedimentology laboratory. Geological Survey of Canada.

Mcgimsey, R. Richter, D.H., Dubois, G.D., and Miller, T.P., (1992): A postulated new source for the White River Ash, Alaska. United States Geological Survey, Anchorage, Alaska, pp 212-218.

McIntyre, M.P., (1980): Physical Geography. Third Edition, John Wiley and Sons, New York, 507 pp.

Miller, D.J., (1953): Late Cenozoic marine glacial sediments and marine terraces of Middleton Island, Alaska. Journal of Geology, Vol. 61, pp 17-40.

Miller, M.M., (1964): Morphogenetic classification of Pleistocene glaciations in the Alaska-Canada Boundary Range. Proceedings of the American Philosophical Society, Vol. 108(3), (June), pp 247-256.

Miller, R.L., and Kahn, J.S., (1962): Statistical Analysis in the Geological Sciences. John Wiley and Sons.

Milner, H.B., (1940): Sedimentary Petrography. Third Edition, Thomas Murby and Company, London.

Milner, H.B., (1962): Sedimentary Petrography, Vol. 2. The Macmillanc Company, New York, 715 pp.

Mitchell, R.L., (1964): Trace elements in soils. In, F.E. Bear (ed.), Chemistry of the Soil, Second Edition, American Chemistry Society, Monogram Series 160, Reinhold, pp 320-368.

Mitchell, G.F., (1972): The Pleistocene history of the Irish Sea, second approximation. Scientific Proceedings of the Royal Dublin Society, A 4(13), pp 181-199.

Molnia, B.F., (1986): Glacial history of the northeastern Gulf of Alaska - A Synthesis. In, T.O. Hamilton, et al. (eds), Glaciation in Alaska, Alaska Geological Survey, pp 219-236.

Monger, J.W. and Berg, H.C., (1984): Lithotectonic terrane map of western Canada and southeastern Alaska. In, Silberling, N.J. and Jones, D.L. (eds.), Lithotectonic Terrane Maps of the North American Cordillera, United States Geological Survey Open File 84-523, B1-B31.

Monger, J.W.H. and Berg, H.C., (1987): Lithotectonic Terrane Map of Western Canada and Southeastern Alaska. United States Geological Survey, Map MF-1874-B.

Monger, J.W.H., Price, R.A., and Tempelman-kluit, D., (1982): Tectonic accretion and the two metamorphic and plutonic welts in the Canadian Cordillera. Geology, Vol. 10, pp 70-75.

Moore, R.C., (1949): The Pliocene/Pleistocene boundary. American Association of Petroleum Geologists Bulletin, Vol. 33, pp 1276-1280.

Morison, S.R., (1987): Stop 26, White Channel placer deposits in the Klondike area. In, S.R. Morison and C.A.S. Smith (eds.), Quaternary Research in Yukon, International Union for Quaternary Research, 12th International Congress (Ottawa, Ontario, Guidebook Field Exersion A-20a and A20b), pp 68-71.

Morison, S.R., (1989): Placer deposits in Canada. In, R.J. Fulton (ed.), Chapter 11, Quaternary Geology of Canada and Greenland, Geology of Canada, No. 1.

Morner, N.A.,(1976): The Pleistocene/Holocene boundary: A proposed boundary stratotype in Gothenburg, Sweden. Boreas, 5, pp 193-275.

Muller, J.E., (1967): Kluane Lake map-area, Yukon Territory. Geological Survey of Canada Memoir 340, 137 pp.

Naeser, N.D., Westgate, J.A. Hughes, O.L., and Pewe, T.L., (1982): Fission-track ages of late Cenozoic distal tephra beds in the Yukon Territory and Alaska. Canadian Journal of Earth Sciences, Vol. 19, pp 2167-2178.

Nichol, I., (1986): Geochemical exploration for gold deposits in areas of glaciated overburden: problems and new developments. In, Prospecting in Areas of Glaciated Terrain 1986; Institution of Mining and Metallurgy, London, pp 7-16.

Norris, D.K., (1971): Tectonic complex, northern Yukon Territory and western District of Mackenzie, Canada. Geological Society of America Abstracts with Programs, Vol. 3, No. 6, pp 398-399.

Olausson, E., (1980): The Pleistocene-Holocene boundary stratotype section: A progress report of the working group. INQUA Newsletter, (2), pp 16-19.

Ovchinnikov, L.N., and Grigoryan, S.V., (1971): Primary halos in prospecting for sulphide deposits. Geochemical Exploration, CIM Special, Vol. 11, pp 375-380.

Palmer, J., and Neilson, R.A., (1962): The origin of granite tors on Dartmoor, Devonshire. Proceedings, Yorkshire Geological Society, 33, pp 315-340.

Paré, D., (1985): Application of Heavy Mineral Analysis to Problems of Till Provenance along a Transect from Longlac, Ontario to Somerset Island. Unpublished M.A. Thesis, Carleton University, Ottawa, 76 pp.

Parfenoff, A., Pomerol, C., and Tourenq, J., (1970): Les minéraux en grains, Méthodesd'étude et de détermination. Masson, Paris, 574 pp.

Parrish, R.R., (1981): Uplift rates of Mt. Logan, Yukon Territory and British Columbia's Coast Mountains using fission track dating methods (abstract), Vol. 62, pp 59-60.

Peltier, L., (1962): Area sampling for terrain analysis. Professional Geographer, Vol. 14(2), pp 24-28.

Penck, A., (1882): Die Vergletscherung der deutschen Alpen. Leipzig, J.A. Barth.

Perttunen, M., (1977): The lithological relation between till and bedrock in the region of Hameenlinna, Southern Finland. Geological Survey of Finland, Bulletin 291, 68 pp.

Pettijohn, F.J., (1957): Sedimentary Rocks. Second Edition, Harper and Bros., New York, 718 pp.

Pettijohn, F.J., (1975): Sedimentary Rocks. Third Edition, Harper & Row, New York.

Peuraniemi, V., (1990): Heavy minerals in glacial material. In, R. Kujansuu, and M. Saarnista (eds.), Glacial Indicator Tracing, A.A. Balkena, Rotterdam, pp 1-14.

Péwé, L.T., (1975): Quaternary Geology of Alaska. United States Geological Survey Professional Paper 835, 145 pp.

Péwé, L.T. et al., (1953): Multiple glaciation in Alaska. United States Geological Survey Circular 289, pp 12-13.

Pitulko, V.M., (1968): Features of geochemical searches for rare-metal deposits in permafrost areas. International Geological Review, Vol. 11, pp 1239-1246.

Pitulko, V.M., (1969): Features of geochemical searches for rare metal deposits in Permafrost areas. Transactions of the International Geological Revue, Vol 11(11), pp 1239-1246.

Pollard, W., (1983): A study of seasonnal frost mounds, North Fork Pass, Northern interior, Yukon Territory. Ph.D. Thesis, Department of Geography, University of Ottawa, 243 pp.

Porter, S.C., Pierce, K.L., and Hamilton, T.D., (1983): Late Wiscosin mountain glaciation in the western United States. In, Porter S.C. (ed.), Late-Quaternary Environments of the United States, Volume 1, The Late Pleistocene, University of Minnesota Press, Minneapolis, pp 71-111.

Power, M.A., (1988): Microearthquake seismicity on the Duke River, Denali Fault System. In, G. Abbott (ed.), Yukon Geology, Volume 2, pp 61-68.

Prest, V.K., (1969): Retreat of Wisconsin and recent ice in North America. Geological Survey of Canada, Map 1257 A.

Prest, V.K., (1983): The late Wisconsinian glacier complex. In, R.J. Fulton (ed.), Quaternary Stratigraphy of Canada, Geological Survey of Canada Paper 84-10, pp 21-36.

Prest, V.K., et al., (1967): Glacial maps of Canada. Geological Survey of Canada, Map 1253A, Scale 1:5,000,000.

Prest, V.K., Grant, D.R., and Rampton, V.N., (1967): Glacial map of Canada, Geological Survey of Canada, Map 1253A.

Price, R.J., (1971): Glacial and Fluvioglacial Landforms. Oliver and Boyd, Edinburgh, 242 pp.

Price, L.W., (1972): The Periglacial Environment, Permafrost and Man. Association of American Geographers, Resource Paper # 14, Commission on College Geography, Washington, 88 pp.

Price, L.W., (1973): Rates of mass wasting in the Ruby Range, Yukon Territory. In, North American Contribution, Permafrost, 2nd International Conference on Permafrost, Yakutsk, Siberia, Printing and Publishing Office, National Academy of Science, Washington, D.C., pp 235-245.

Price, R.A. and Douglas, R.J.W., (1972): Variations in Tectonic Styles in Canada. Geological Association of Canada Special Paper Number 11, 81 pp.

Pullan, R.A., (1959): Tors. Scottish Geographical Magazine, Vol. 75, pp 51-55.

Rampton, V.N., (1969): Pleistocene Geology of the Snag- Klutan area, Southwestern Yukon Territory, Canada. Unpublished Ph.D. Dissertation, Minnesota University, Minneapolis, 237 pp.

Rampton, V.N., (1971): Late Quaternary vegetational and climatic history of the Snag-Klutlan area, Southwestern Yukon Territory, Canada. Geological Society of America Bulletin, 82, pp 959-978.

Rampton, V.N., (1971): Late Pleistocene glaciations of the Snag-Klutlan area, Yukon Territory. Arctic, 24, pp 277-300.

Rampton, V.N., (1972): Surficial deposits; Alaska Highway, Whitehorse, to Alaska-Yukon boundary. In, Report of Activities, Part B, November 1971 to March 1972, Geological Survey of Canada, Paper 72-1B, pp 118-122.

Rampton, V.N., (1979): Surficial geology Malloch Hill, Aklavik, Mackenzie Delta and Stanton map-areas. Geological Survey of Canada, Maps 30-1979, 31-1979, 32-1979 and 33-1979.

Rampton, V.N., (1981): Surficial materials and landforms of Kluane National Park, Yukon Territory. Geological Survey of Canada Paper 79-24.

Rampton, V.N., (1982): Quaternary geology of the Yukon Coastal Plain. Geological Survey of Canada Bulletin 317, 49 pp.

Rampton, V.N., (1989): Surficial Geology and Geomorphology, Burwash Creek (115G-C). Geological Survey of Canada.

Rampton, V.N., Dugal, J.B., (1974): Quaternary stratigraphy and geomorphic processes on the Arctic Coastal Plain and adjacent areas, Demarcation Point, Yukon Territory to Malloch Hill, District of Mackenzie. In, Report of Activities, Part A, April to October 1973, Geological Survey of Canada Paper 74-1A, pp 283.

Rankama, K., and Sahama, T.G., (1950): Geochemistry. University of Chicago Press.

Rice, H.M.A., (1936): Glacial phenomena near Cranbrook, British Columbia. Journal of Geology, Vol. 44, pp 68-73.

Ricker, K.E., (1968): Quaternary Geology of the Southern Ogilvie Ranges, Yukon Territory, and an investigation of morphological, perglacial, pedological, and botanical criteria for possible use in the chronology of morainal sequences. M.A. Thesis, University of British Columbia.

Riddihough, R.P., (1982): Contemporary movements and tectonics on Canada's west coast: a discussion. Tectonophysics, Vol. 86, pp 319-341.

Ridler, R.H., and Shilts, W.W., (1974): Exploration for Archean Polymetallic Sulphide Deposits in Permafrost Terrain: An Integrated Geological/Geochemical Technique; Kaminak Lake Area, District of Keewatin. Geological Survey of Canada Paper 73-34, 33 pp.

Roed, M.A., (1975): Cordilleran and Laurentide multiple glaciation, West Central Alberta. Canadian Journal of Earth Sciences, Vol. 12(9), pp 1493-1515.

Ruddiman, W.F., (1971): Pleistocene sedimentation in the equatorial Atlantic; stratigraphy and faunal paleoclimatology. Geological Society of America Bulletin, Vol. 82, pp 283-302.

Rutter, N.W., (1980): Late Pleistocene history of the Western Canadian ice-free corridor. Journal of Anthropology, Vol. 1(1), pp 1-8.

Rutter, N.W., (1981): Relationship between late Pleistocene Laurentide and Cordilleran glaciations, Canada. In, V. Sibrava and F.M. Shotton (eds.), Quaternary Glaciations in the Northern Hemisphere, Geological Survey of Czechoslovakia, IGCP Project 73/1/24, Report (6), pp 205-218.

Rutter, N.W., (1984): Pleistocene history of the Western Canadian ice-free corridor. In, R.J. Fulton (ed.), Quaternary Stratigraphy of Canada - A Canadian Contribution to IGCP Project 24, Geological Survey of Canada Paper 84-10.

Rutter, N.W., et al., (1985): Dating methods of Pleistocene deposits and their problems, Geoscience Canada, Reprint Series 2. Geological Association of Canada, 87 pp.

Ryder, J.M. and Clague, J.J., (1989): British Columbia Quaternary stratigraphy and history, Cordilleran Ice-Sheet. In, R.J. Fulton (ed.), Geology of Canada and Greenland, Geology of Canada No. 1, pp 48-58.

Saarnisto, M., (1990): An outline of glacial indicator tracing. In, R. Kujansuu, and M. Saarnisto (eds.), Glacial Indicator Tracing, A.A. Balkema, Rotterdam, pp 1-14.

Sanders, T., Ward, R., Loftis, J., Steele, T., Adrian, D., and Yevjevich, V., (1983): Design of networks for monitoring water quality, Littleton CO. Water Resources Publications, 328 pp.

Saxby, J.D., (1969): Metal-organic chemistry of the geochemical cycle. Revue of Pure Applied Chemistry, 19, pp 131-150.

Schmoll, H.R. and Yehle, L.A., (1986): Pleistocene glaciation of the upper Cook Inlet basin. In, T.D. Hamilton et al. (eds), Glaciation in ALaska, Alaska Geological Survey, pp 193-218.

Schwarcz, H.P., (1978): Dating methods of Pleistocene and their problems: 2) Uranium series disequilibrium dating. Geoscience Canada, Vol. 5, pp 184-188.

Schwartzburg, R.E.C., (1986): Physical Limnology of a Small Sub-Arctic Alpine Lake, Yukon Territory. M.A. Thesis, Department of Geography, University of Ottawa, 101 pp.

Sharp, R.P., (1943): Geology of the Wolf Creek area, St. Elias Range, Yukon Territory, Canada. Geological Society of America Bulletin, Vol. 54, pp 625-650.

Shaw, J., (1972): Pleistocene chronology and geomorphology of the Rocky Mountains in south and central Alberta. In, H.O. Slaymaker and H.J. McPherson (eds.), Mountain Geomorphology, Tantalus, Vancouver, pp 34-46.

Shetsen, I., (1984): Applications of till pebble lithology to the differentiation of glacial lobes in Southern Alberta. Canadian Journal of Earth Sciences, 21, pp 920-933.

Shilts, W.W., (1971): Till studies and their applications to regional drift prospecting. Canadian Mining Journal, (April), pp 45-51.

Shilts, W.W., (1973): Drift prospecting; geochemistry of eskers and till in permanently frozen terrain: District of Keewatin; Northwest Territories. Geological Survey of Canada Paper 72-45.

Shilts, W.W., (1976): Till in mineral exploration, in R.F. Legget (ed.), Glacial Till: An Interdisciplinary Study. Royal Society of Canada, Special Publication # 12, pp 205-224.

Shilts, W.W., (1978): Nature and genesis of mudboils, central Keewatin, Canada. Canadian Journal of Earth Sciences, 15, pp 1053-1068.

Shilts, W.W., (1980): Flow patterns in the central North American ice sheet. Nature, Vol. 286 (5770), pp 213-218.

Shilts, W.W., (1982): Quaternary evolution of the the Hudson - James Bay Region. Naturaliste Canadien, Vol. 109, pp 309-332.

Shilts, W.W., (1984): Quaternary events - Hudson Bay Lowland and southern District of Keewatin. In, R.J. Fulton (ed.), Quaternary Stratigraphy of Canada - A Canadian Contribution to IGCP Project 24, Geological Survey of Canada Paper 84-10.

Shilts, W.W., Cunninghan, C.M., and Kaszycki, C.A., (1979): Keewatin Ice Sheet re-evaluation of the traditional concept of the Laurentide Ice Sheet. Geology, Vol. 7, pp 537-541.

Shumilov, V.Y., (1981): An experimental study of cryogenic factors affecting geological processes in placer formations. In, Permafrost, Fourth International Conference, Proceedings, National Academy Press, Washington, U.S.A., pp 1157-1159.

Shvartsev, S.L., (1972): Geochemical prospecting methods in regions of permanently frozen ground. XXIVth International Geological Congress, Montreal, Section 10, pp 380-383.

Slaymaker, H.O. and Mcpherson, H.J., (1977): An overview of geomorphic processes in the Canadian Cordillera. Zeitschrift for Geomorphologie, Vol. 21, pp 169-186.

Smith, F.M., (1971): Geological and Geochemical Report on the Alaskite Project Claims ED and ADD Groups Claim Sheet 115G8 Rockslide and Alaskite Creeks Area, Yukon Territory. Phelps Dodge Corporation of Canada Ltd., 15 pp.

Smith, F.M., (1971): Geological and geochemical report on the Talbot Project claims "A", "B",and "K" groups, Claim Sheet 115G8, Talbot and Rockslide Creek Area, Yukon Territory. Phelps Dodge Corporation of Canada Limited.

Smith, R.L., (1985): An Investigation of the Effects of Geomorphology on Stream Hydrochemistry. M.A. Thesis, Department of Geography, University of Ottawa, 163 pp.

Smithson, F., (1930): The Reliability of Frequency Estimations of Heavy Mineral Suites. Geological Magazine, 67, pp 134-136.

Souther, J.G., Stanciu, C., (1975): Operation Saint Elias, Yukon Territory: Tertiary volcanic rocks. In, Report of Activities, Part A: April to October, 1974, Geological Survey of Canada Paper 75-1A. pp 63-70

Speidel, D.H. and Agnew, A.F., (1982): The Natural Geochemistry of Our Environment. Westview Press, Boulder, Colorado, 214 pp.

St. Onge, D., (1981): Theories, paradigms, mapping and geomorphology. Canadian Geographer, Vol. 25(4), pp 307-315.

Stalker, A., and McPherson, H.J., (1969): A probable late Pinedale terminal moraine in Castle River Valley, Alberta. Geological Society of America Bulletin, (80), pp 2115-2122.

Sugden, D.E., (1968): The selectivity of glacial erosion in the Cairngorm Mountains, Scotland. Transactions of the Institute of British Geographers, 45, pp 79-92.

Sugden, D.E. and Watts, S.H., (1977): Tors, Felsenmeer, and glaciation in northern Cumberland Peninsula, Baffin Island. Canadian Journal of Earth Sciences, Vol. 14, pp 2817-2823.

Sutherland Brown, A., (1969): Aiyarsh lava flow, British Columbia. Canadian Journal of Earth Sciences, Vol. 6, pp 1460-1468.

Taipale, K., Nevalainen, R., and Saamisto, O., (1986): Silicate analyses and normative composition fo the fine fraction of till: Examples from eastern Finland. Journal of Sedimentary Petrology 56, pp 370-378.

Taliaferro, N.L., (1932): Geology of the Yakataga, Katalla, and Nichawak districts, Alaska. Geological Society of America Bulletin, Vol. 43, pp 749-782.

Tarr, R.S. and Martin, L., (1914): Alaskan Glacier Studies, Washington, D.C. National Geographic Society.

Taylor, P.J., (1975): Distance Decay Models in Spatial Interactions, CATMOG. Institute of British Geographers, 35 pp.

Taylor, S.R., (1969): Trace Element Chemistry of Andesites and Associated Calc-Alkaline Rccks, Proceedings of the Andesite Conference. Oregon Department of Geology and Mines Resources Bulletin, 65, pp 43-63.

Tempelman-Kluit, D.J., (1974): Reconnaissance geology of Aishihik Lake, Snag and part of Stewart River map- areas, west-central Yukon. Geological Survey of Canada Paper 73-41, 97 pp.

Tempelman-Kluit, D.J., (1979): Five occurrences of transported synorogenic clastic rocks in Yukon Territory. Geological Survey of Canada Paper 79-1A, pp 1-12.

Tempelman-Kluit, D.J., (1980): Evolution of physiography and drainage in southern Yukon. Canadian Journal of Earth Sciences, Vol. 17, No. 9, pp 1189-1203.

Tempelman-Kluit, D.J., (1980): Highlights of field work in Laberge and Carmacks map-areas, Yukon Territory. Geological Survey of Canada Paper 80-1A, pp 357-362.

Terasmae, J., (1967): Notes on Quaternary paleoecological problems in the Yukon Territory, and adjacent regions. Geological Survey of Canada Paper 64-46, 11 pp.

Thorson, R.M., (1986): Late Cenozoic glaciation of the northern Nenara River valley. In, T.D. Hamilton et al. (eds), Glaciation in Alaska, Alaska Geological Survey, pp 99-122.

Tipper, H.W., (1963): Geology, Taseko Lakes, British Columbia. Geological Survey of Canada, Map 29, 1963.

Tipper, H.W., (1971): Glacial geomorphology and Pleistocene history of central British Columbia. Geological Survey of Canada Bulletin 196, 89 pp.

Tipper, H.W., (1971): Multiple glaciation in central British Columbia. Canadian Journal of Earth Sciences, Vol. 8, pp 743-752.

Twenhofel, W.H., and Tyler, S.A., (1941): Methods of Study of Sediments. McGraw-Hill Book Company Inc., New York and London.

Tyrrell, T., (1898): Report on the Doobaunt, Kazan, and Ferguson River, and the Northwest coast of Hudson Bay; and on two overland routes from Hudson Bay to Lake Winnipeg. Geological Survey of Canada Annual Report, Series N, Vol. 9, pp 1-218.

van Eysinga, F.W.B., (1978): Geological Time Table. Third Edition, Elsevier Scientific Publication Company, Amsterdam.

Van Tassel, R.E., (1969): Exploration by over-burden drilling of Keno Hill Mines Limited, Second International Geochemical Exploration Symposium. Colorado School of Mines, Vol 64(1), pp 457-478.

Van Der Plas, L., (1962): Preliminary note on the granulometric analysis of sedimentary rocks. Sedimentology, Vol. 1, pp 145-157.

Vanicek, P. and Nagy, D., (1981): On the compilation of the map of contemporary vertical crustal movements in Canada. Tectonophysics, #71, pp 79-86.

Veillette, J.J., and Nixon, F.M., (1980): Portable drilling equipment for shallow permafrost sampling. Geological Survey of Canada Paper 79-21, pp 1-35.

Vernon, P., and Hughes, O.L., (1966): Surficial glaciation, Dawson, Larsen Creek, and Nash Creek map areas, Yukon Territory. Geological Society of Canada Bulletin 136, 25 pp.

Vincent, J.S., (1989): Quaternary geology of the northern Interior Plains of Canada. In, R.J. Fulton (ed.), Chapter 2, Quaternary Geology of Canada and Greenland, Geology of Canada, No. 1.

Virkkala, K., (1971): On the lithology and provenance of the till of a gabbro area in Finland. VIII Congrès INQUA, Paris 1969. Etudes sur le Quaternaire dans le Monde 2: pp 711-714.

Wagner, W.P., (1966): Correlation of Rocky Mountain and Laurentide glacial chronologies in southwestern Alberta. Canada, Ph.D. Thesis, University of Michigan.

Wahrhaltig, C., (1958): Quaternary geology of the Nenana River valley and adjacent parts of the Alaska Range. United States Geological Survey Professional Paper 293-A, pp 1-68.

Wardle, P., (1973): Variations of the glaciers of Westland National Park and the Hooker Range, New Zealand. New Zealand Journal of Botany, Vol. 11, pp 349-388.

Wardle, P., (1978): Further radiocarbon dates from Westland National Park and the Hooker Range, New Zealand. New Zealand Journal of Botany, Vol. 16, pp 147-152.

Warren, H.V., and Delavault, R.E., (1953): Geochemical prospecting finds widespread application in British Columbia. Mining Engineering, 5, pp 980-981.

Washburn, A.L., (1969): Weathering, frost action, and patterned ground in the Mesters Vig District, northeast Greenland. Medd. om Gronland, Vol. 176(4), pp 303.

Washburn, A.L., (1979): Geocryology: A Survey of Periglacial Processes and Environments. Second Edition, Edward Arnold, London, 406 pp.

Watson, R.A., and Wright, H.E., (1980): The end of the Pleistocene: A general critique of chronostratigraphic classification. Boreas, 9, pp 153-163.

Watterson, J. and Theobald, P., (1979): Geochemical exploration, 1978. Proceedings of the 7th International Geochemical Exploration Symposium, Toronto, Canada, Association of Exploration Geochemists, 504 pp.

Wennervirta, H., (1973): Sampling of the bedrock hill interface in geochemical exploration. In, M.J. Jones (ed.), Prospecting in Areas of Glaciated Terrain, Institute of Mining and Metallurgy, London, pp 67-71.

Wentworth, C.K., (1922): A scale of grade and class terms for clastic sediments. Journal of Geology, Vol. 30, pp 377-392.

West, R.J., (1968): Pleistocene geology and biology, with especial reference to the British Isles. Longmans Green and Co., Ltd., London, 377 pp.

West, R.G., (1977): Pleistocene Geology and Biology. Longman, London.

Westgate, J.A., et al., (1977): Wasca Creek Ash (Middle Pleistocene) in southern Saskatchewan: characterization, source, fission track age, paleomagnetism, and stratigraphic significance. Canadian Journal of Earth Sciences, Vol. 4, pp 357-374.

Westgate, J.A., and Briggs, N.D., (1985): Dating methods of Pleistocene and their problems: Tephrochronology and fission-tract dating. Geoscience Canada, Vol. 7, pp 3-10.

Westgate, J.A., (1987): Compositional comparison of Old Crow Tephra, Sheep Creek Tephra, and White River Ash and its significance for the provenance of other widespread Beringian tephras. International Union for Quaternary Research, XIIth International Congress, Ottawa.

Wheeler, J.O., (1961): Whitehorse map area, Yukon Territory. Geological Survey of Canada Memoir 312.

Wheeler, J.O., (1963): Geological map of Kaskawalsh half of Mount St.Elias map sheet, Yukon Territory. Canada Geological Survey Map 1134A.

Wheeler, J.O. and Mcfeely, P., (1987): Tectonic Assemblage Map of the Canadian Cordillera, Open FIle 1565, Geological Survey of Canada.

Wheeler, J.O., Brookfield, A.J., Gabriedge, H. Monger, J.W.H., Tipper, H.W., and Woodsworth, G.J., (1988): Terrane Map of the Canadian Cordillera, Open File 1894, Geological Survey of Canada.

Whitten, D.G.A., (1982): Geology. Penguin Books, New York.

Wickham, J.T., (1979): Pre-Illinoian Stratigraphy of Quincy, Illinois area. ISGS Guidebook 14, pp 69-90.

Wilson, J.T., Falconer, G., Mathews, W.B., and Prest, V.K. (compilers) (1976): Glacial Map of Canada; Geological Association of Canada. Toronto, Ontario.

Yarg, L.A., (1974): Weathering-Induced Changes of Physico - Mechanical Properties of Rock. Nedra, Moscow, 126 pp.

Youngman, P.M., (1975): Mammals of the Yukon Territory, National Museum of Natural Sciences. Publications in Zoology, 10, 192 pp.