U.S. GEOLOGICAL SURVEY CIRCULAR 1026

Late Cenozoic History of the Interior Basins of Alaska and the Yukon

Proceedings of a cooperative workshop between earth scientists from Canada and the United States of America

AVAILABILITY OF BOOKS AND MAPS OF THE U.S. GEOLOGICAL SURVEY

Instructions on ordering publications of the U.S. Geological Survey, along with prices of the last offerings, are given in the current-year issues of the monthly catalog "New Publications of the U.S. Geological Survey." Prices of available U.S. Geological Survey publications released prior to the current year are listed in the most recent annual "Price and Availability List." Publications that are listed in various U.S. Geological Survey catalogs (**see back inside cover**) but not listed in the most recent annual "Price and Availability List" are no longer available.

Prices of reports released to the open files are given in the listing "U.S. Geological Survey Open-File Reports," updated monthly, which is for sale in microfiche from the U.S. Geological Survey, Books and Open-File Reports Section, Federal Center, Box 25425, Denver, CO 80225. Reports released through the NTIS may be obtained by writing to the National Technical Information Service, U.S. Department of Commerce, Springfield, VA 22161; please include NTIS report number with inquiry.

Order U.S. Geological Survey publications **by mail** or **over the counter** from the offices given below.

BY MAIL

Books

Professional Papers, Bulletins, Water-Supply Papers, Techniques of Water-Resources Investigations, Circulars, publications of general interest (such as leaflets, pamphlets, booklets), single copies of Earthquakes & Volcanoes, Preliminary Determination of Epicenters, and some miscellaneous reports, including some of the foregoing series that have gone out of print at the Superintendent of Documents, are obtainable by mail from

U.S. Geological Survey, Books and Open-File Reports
Federal Center, Box 25425
Denver, CO 80225

Subscriptions to periodicals (Earthquakes & Volcanoes and Preliminary Determination of Epicenters) can be obtained ONLY from the

Superintendent of Documents
Government Printing Office
Washington, D.C. 20402

(Check or money order must be payable to Superintendent of Documents.)

Maps

For maps, address mail orders to

U.S. Geological Survey, Map Distribution
Federal Center, Box 25286
Denver, CO 80225

Residents of Alaska may order maps from

Alaska Distribution Section, U.S. Geological Survey,
New Federal Building - Box 12
101 Twelfth Ave., Fairbanks, AK 99701

OVER THE COUNTER

Books

Books of the U.S. Geological Survey are available over the counter at the following Geological Survey Public Inquiries Offices, all of which are authorized agents of the Superintendent of Documents:

- **WASHINGTON, D.C.**--Main Interior Bldg., 2600 corridor, 18th and C Sts., NW.
- **DENVER, Colorado**--Federal Bldg., Rm. 169, 1961 Stout St.
- **LOS ANGELES, California**--Federal Bldg., Rm. 7638, 300 N. Los Angeles St.
- **MENLO PARK, California**--Bldg. 3 (Stop 533), Rm. 3128, 345 Middlefield Rd.
- **RESTON, Virginia**--503 National Center, Rm. 1C402, 12201 Sunrise Valley Dr.
- **SALT LAKE CITY, Utah**--Federal Bldg., Rm. 8105, 125 South State St.
- **SAN FRANCISCO, California**--Customhouse, Rm. 504, 555 Battery St.
- **SPOKANE, Washington**--U.S. Courthouse, Rm. 678, West 920 Riverside Ave..
- **ANCHORAGE, Alaska**--Rm. 101, 4230 University Dr.
- **ANCHORAGE, Alaska**--Federal Bldg, Rm. E-146, 701 C St.

Maps

Maps may be purchased over the counter at the U.S. Geological Survey offices where books are sold (all addresses in above list) and at the following Geological Survey offices:

- **ROLLA, Missouri**--1400 Independence Rd.
- **DENVER, Colorado**--Map Distribution, Bldg. 810, Federal Center
- **FAIRBANKS, Alaska**--New Federal Bldg., 101 Twelfth Ave.

Late Cenozoic History of the Interior Basins of Alaska and the Yukon

L. DAVID CARTER, THOMAS D. HAMILTON,
and JOHN P. GALLOWAY, Editors

Proceedings of a cooperative workshop between earth scientists from Canada and the United States of America

U.S. GEOLOGICAL SURVEY CIRCULAR 1026

DEPARTMENT OF THE INTERIOR

MANUEL LUJAN, JR., Secretary

U.S. GEOLOGICAL SURVEY

Dallas L. Peck, Director

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON : 1989

Free on application to the
Books and Open-File Reports Section
U.S. Geological Survey
Federal Center, Box 25425
Denver, CO 80225

Library of Congress Cataloging-in-Publication Data

Late Cenozoic history of the interior basins of Alaska and the Yukon / L.
David Carter, Thomas D. Hamilton, and John P. Galloway, editors.
p. cm. — (U.S. Geological Survey circular ; 1026)
Proceedings of a joint Canadian-American workshop hosted by the U.S.
Geological Survey at Anchorage, AK, in Feb. 1987.
Bibliography: p.
Supt. of Docs. no.: I 19.4/2:1026
1. Geology, Stratigraphic—Quaternary—Congresses. 2. Geology—Canada—
Beaufort Sea Region—Congresses. 3. Geology—Alaska—Beaufort Sea Region—
Congresses. I. Carter, L. David. II. Hamilton, Thomas D. (Thomas Dudley),
1936– . III. Galloway, John P. IV. Series: Geological Survey circular ;
1026.
QE696.L286 1989 89-600059
551.7'9—dc20 CIP

CONTENTS

INTRODUCTION 1
L. David Carter and Thomas D. Hamilton

WORKSHOP PROGRAM 3

WORKSHOP PARTICIPANTS 4

JOINT CANADIAN-AMERICAN INVESTIGATION OF THE CENOZOIC GEOLOGY OF THE LOWLANDS BORDERING THE BEAUFORT SEA 7
Jean-Serge Vincent, L. David Carter, John V. Matthews, Jr., and David M. Hopkins

PAPERS

Late Pleistocene Spruce (*Picea*) in Northern Interior Basins of Alaska and the Yukon: Evidence from Marine Deposits in Northern Alaska 11
L. David Carter and Thomas A. Ager

Offshore Constraints on the Late Pleistocene Glacial History at the Mouth of the Mackenzie River 15
Steve M. Blasco, Julie Brigham-Grette, and Philip R. Hill

Continental Ice Advances in Northwestern Canada and their Significance to Interior Basins in the Yukon 18
Jean-Serge Vincent

Was the Late Pleistocene Northwest Laurentide Ice Sheet Wet-Based? 22
James Beget

Quaternary Chronology, Yukon and Western District of Mackenzie 25
Owen L. Hughes

The Old Crow and Bluefish Basins, Northern Yukon: Development of the Quaternary History 30
Charles E. Schweger

Climatic Change in Eastern Beringia During Oxygen Isotope Stages 2 and 3: Proposed Thermal Events 34
John V. Matthews, Jr., Charles E. Schweger, and Owen L. Hughes

Paleosols of Northwestern Canada 39
Charles Tarnocai

Upper Cenozoic Deposits, Kanuti Flats and Upper Kobuk Trench, Northern Alaska 45
Thomas D. Hamilton

Quaternary Deposits at Birch Creek, Northeastern Interior Alaska: The Possibility of Climatic Reconstruction 48
Mary E. Edwards and Patricia F. McDowell

Late Quaternary Paleofloods along the Porcupine River, Alaska: Implications for Regional Correlation 51
Robert M. Thorson

Late Cenozoic Sedimentary History along Major Fault Zones, Alaska 55
Warren Yeend

Paleoglaciology of the Selwyn Lobe of the Cordilleran Ice Sheet and Quaternary Stratigraphy of the East-Central Yukon 60
Lionel E. Jackson, Jr.

Late Cenozoic Stratigraphy and Sedimentology of Gravelly Deposits in Central and Southern Yukon 66
Stephen R. Morison

Quaternary Stratigraphy of the Fairbanks Area, Alaska 72
Troy L. Péwé

Pleistocene Glacial Events, Southeastern Copper River Basin, Alaska 78
Donald R. Nichols

A Working Glacial Chronology for the Western Copper River Basin, Alaska 81
John R. Williams

Glacial Lake Atna, Copper River Basin, Alaska 85
Oscar J. Ferrians, Jr.

History of Late Pleistocene and Holocene Vegetation in the Copper River Basin, South-Central Alaska 89
Thomas A. Ager

Pleistocene Vertebrate Localities in the Yukon 93
C. Richard Harington

Arctic Microtine Biochronology—Current Status 99
Charles A. Repenning

Comment on the Present Status of Dating Loess by Thermoluminescence 103
Glenn W. Berger

SUMMARIES OF DISCUSSION SESSIONS 107

Geochronology 107
Julie Brigham-Grette

Neotectonics in Alaska and Northwestern Canada 108
Robert M. Thorson

Late Tertiary Events 109
John V. Matthews, Jr.

Early and Middle Pleistocene Events 110
Thomas D. Hamilton

Late Pleistocene Glacial Advances 111
Jean-Serge Vincent

Late Pleistocene Interstadial Intervals 112
Charles E. Schweger

Canadian-American Collaborative Research 113
L. David Carter

INTRODUCTION

By L. David Carter *and* Thomas D. Hamilton

This volume contains the proceedings of a joint Canadian-American workshop on the late Cenozoic history of the interior basins of Alaska and the Yukon, a meeting that was hosted by the U.S. Geological Survey at Anchorage, Alaska, in February 1987. The principal aim of the workshop was fostering communication between Canadian and American scientists on problems of late Cenozoic history common to both northwestern Canada and Alaska.

The Anchorage workshop was the second in a planned series of meetings that are intended to be held at two- or three-year intervals. The first meeting of this series, which was hosted by the Geological Survey of Canada, was held in Calgary, Alberta, in April 1984. Its topic was the correlation of Quaternary deposits and events around the margin of the Beaufort Sea. A correlation chart that was assembled at the meeting and 16 short papers prepared by individual participants were published by the Geological Survey of Canada (Heginbottom and Vincent, 1986).

During the Calgary meeting, it was agreed that a field trip would be held along the shore of the Beaufort Sea during the summer of 1985 and that a second workshop focusing on long geologic and climatic records of interior basins of Alaska and the Yukon would be held in Anchorage, Alaska, in the spring of 1987.

The field trip was carried out in late July and early August 1985. Participants were Jean-Serge Vincent and John V. Matthews, Jr. (Geological Survey of Canada), David M. Hopkins (University of Alaska), and L. David Carter (U.S. Geological Survey). The group's objective was to study the chronology and geologic relations of deposits along the arctic coastal lowlands in order to clarify sea-level history and to resolve the age of late Tertiary and early to middle Pleistocene glacier advances (see Vincent and others, this volume).

Planning for the subsequent Anchorage workshop was carried out jointly by members of the U.S. Geological Survey and the Geological Survey of Canada. Twenty-three scientists from the Geological Survey of Canada, other Canadian government agencies, Canadian and United States universities, the Alaska Division of Mines and Geology, and the U.S. Geological Survey participated in the meeting. Twenty of the following papers were prepared for the workshop and subsequently revised for publication; two additional papers were solicited after the workshop in order to expand coverage of topics that proved to be of particular interest. Brief synopses of the seven discussion sessions that followed the presentation of papers were prepared by discussion leaders, and these also are included in this volume.

Some of the major rivers, tectonic basins, and faults referred to in the papers and discussion-session reports are shown in figure 1. More detailed maps showing all of the localities and geographic features mentioned in papers dealing with the Yukon can be found in Morison and Smith (1987, figs. 39, 41, and 42).

One central theme that emerged from the workshop is the critical role of volcanic ash layers (tephras) in dating and correlating deposits of late Cenozoic age in Alaska and the Yukon. It is unfortunate that the age of one particularly widespread unit, the Old Crow tephra, is still unsettled as this volume goes to press. Most of the following topical reports and discussion summaries were written when an age of 86±8 ka was generally accepted for the Old Crow tephra (Wintle and Westgate, 1986), but subsequent age measurements indicate that it may be as old as 110±12 ka (Berger, this volume) or 149±13 ka (Westgate, 1988).

In this volume, we follow the convention of using **ka** for "thousand years before present" and **Ma** for "million years before present." Timespans are designated as **k.y.** and **m.y.** for thousands and millions of years, respectively. Geologic-climatic units such as glaciations, interglaciations, stades, and interstades are considered informal stratigraphic units, as are all tephra layers discussed except the formally defined Ester, Dome, Wilber, Jarvis, and White River Ash Beds (Péwé, 1975).

REFERENCES CITED

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 1984: Geological Survey of Canada Open File Report 1237, 60 p.

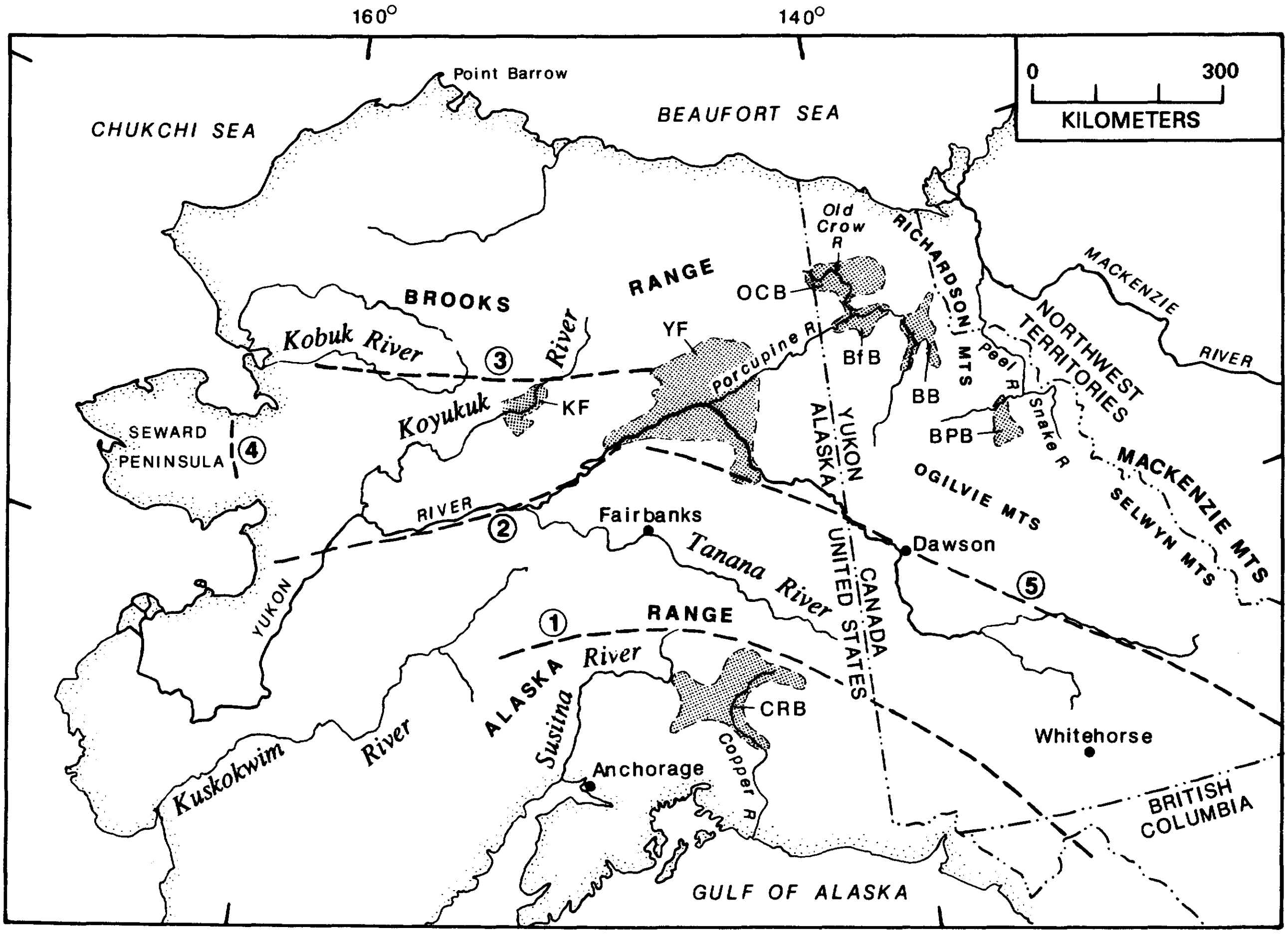
Manuscript approved for publication, January 5, 1989.

Morison, S.R., and Smith, C.A.S., eds., 1987, Guidebook to Quaternary research in Yukon; XIIth INQUA Congress field excursions A20a and A20b: Ottawa, National Research Council of Canada, 110 p.

Péwé, T.L., 1975, Quaternary stratigraphic nomenclature in central Alaska: U.S. Geological Survey Professional Paper 862, 32 p.

Westgate, J.A., 1988, Isothermal plateau fission-track age of the late Pleistocene Old Crow tephra, Alaska: Geophysical Research Letters, v. 15, p. 376–379.

Wintle, A.G., and Westgate, J.A., 1986, Thermoluminescence age of Old Crow tephra in Alaska: Geology, v. 14, p. 594–597.



EXPLANATION

	Basins
BB	Bell Basin
BfB	Bluefish Basin
BPB	Bonnet Plume Basin
CRB	Copper River Basin
KF	Kanuti Flats
OCB	Old Crow Basin
YF	Yukon Flats

— — —	Fault
1	Denali fault
2	Kaltag fault
3	Kobuk fault
4	Kugruk fault
5	Tintina fault

Figure 1. Some major geographic and geologic features of Alaska and the Yukon mentioned in papers and discussion-session reports in this volume.

WORKSHOP PROGRAM

Wednesday, February 11

Welcome and Introduction
L.D. Carter and T.D. Hamilton

Resume of the joint Canadian-American field trip of 1985
Vincent, Jean-Serge, and Carter, L. David

Brief Presentations

I. Northern Alaska, Yukon, and Mackenzie District
Brigham-Grette, Julie
Carter, L. David
Vincent, Jean-Serge
Beget, James E.

II. Interior basins of northwest Canada
Hughes, Owen L.
Morison, Stephen R.
Jackson, Lionel E., Jr.
Tarnocai, Charles
Schweger, Charles E.

III. Basins of central Alaska
Hopkins, David M.
Hamilton, Thomas D.
Williams, John R.
Edwards, Mary E.
Péwé, Troy L.
Williams, John R.
Ferrians, Oscar, J., Jr.

IV. Regional studies
Thorson, Robert M.
Yeend, Warren E.
Repenning, Charles A.
Judge, Alan S.
Harington, C. Richard
Ager, Thomas A.
Matthews, John V., Jr.

Thursday, February 12

Discussion Sessions

1. Geochronology
Discussion Leader: Brigham-Grette, Julie

2. Neotectonics
Discussion Leader: Thorson, Robert M.

3. Late Tertiary events
Discussion Leader: Matthews, John V., Jr.

4. Early and middle Pleistocene events
Discussion Leader: Hamilton, Thomas D.

Friday, February 13

Discussion Sessions

5. Late Pleistocene glacial advances
Discussion Leader: Vincent, Jean-Serge

6. Late Pleistocene interstadial intervals
Discussion Leader: Schweger, Charles E.

7. Canadian-American collaborative research
Discussion Leader: Carter, L. David

WORKSHOP PARTICIPANTS

Thomas A. Ager
Branch of Paleontology and Stratigraphy
U.S. Geological Survey
National Center—MS 970
Reston, Virginia 22092

James E. Beget
Department of Geology and Geophysics
408 Brooks Building
University of Alaska
Fairbanks, Alaska 99775-0760

Julie Brigham-Grette
Department of Geology
University of Alberta
Edmonton, Alberta
Canada T6G 2E3
Present address:
Department of Geology and Geography
University of Massachusetts
Amherst, Massachusetts 01003-0026

L. David Carter
Branch of Alaskan Geology
U.S. Geological Survey
4200 University Drive
Anchorage, Alaska 99508-4667

Mary E. Edwards
Marlboro College
Marlboro, Vermont 05344
Present address:
Department of Geography
University of Oregon
Eugene, Oregon 97403-1218

Oscar J. Ferrians, Jr.
Branch of Alaskan Geology
U.S. Geological Survey
4200 University Drive
Anchorage, Alaska 99508-4667

Thomas D. Hamilton
Branch of Alaskan Geology
U.S. Geological Survey
4200 University Drive
Anchorage, Alaska 99508-4667

C. Richard Harington
National Museum of Natural Sciences
Ottawa, Ontario
Canada K1A ON8

David M. Hopkins
Department of Geology and Geophysics
408 Brooks Building
University of Alaska
Fairbanks, Alaska 99775-0760

Owen L. Hughes
Geological Survey of Canada
3303-33rd Street
Calgary, Alberta
Canada T2L 2A7

Lionel E. Jackson, Jr.
Geological Survey of Canada
100 West Pender Street
Vancouver, British Columbia
Canada V6B 1R8

Alan S. Judge
Geological Survey of Canada
1 Observatory Crescent
Ottawa, Ontario
Canada K1A OY3

John V. Matthews, Jr.
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario
Canada K1A OE8

Stephen R. Morison
Department of Indian and Northern Affairs
200 Range Road
Whitehorse, Yukon
Canada Y1A 3V1

Troy L. Péwé
Department of Geology
Arizona State University
Tempe, Arizona 85287

Richard D. Reger
Alaska Division of Mining and Geology
794 University Ave.
Fairbanks, Alaska 99701

Charles A. Repenning
Branch of Paleontology and Stratigraphy
U.S. Geological Survey—MS 919
Box 25046—Denver Federal Center
Denver, Colorado 80225-0046

WORKSHOP PARTICIPANTS—*Continued*

Charles E. Schweger
Department of Anthropology
University of Alberta
Edmonton, Alberta
Canada T6G 2E3

Charles Tarnocai
Land Resource Research Institute
Central Experimental Farm
Ottawa, Ontario
Canada K1A OC6

Robert M. Thorson
Department of Geology and Geophysics
U–45, University of Connecticut
Storrs, Connecticut 06268

Jean-Serge Vincent
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario
Canada K1A OE8

John R. Williams
Branch of Alaskan Geology—MS 917
U.S. Geological Survey
345 Middlefield Road
Menlo Park, California 94025

Warren E. Yeend
Branch of Alaskan Geology—MS 917
U.S. Geological Survey
345 Middlefield Road
Menlo Park, California 94025

Other Contributors to this Volume

Glenn W. Berger
Department of Geology
Western Washington University
Bellingham, Washington 98225

Steve M. Blasco
Geological Survey of Canada
Atlantic Geoscience Centre
Bedford Institute of Oceanography
P.O. Box 1006
Dartmouth, Nova Scotia
Canada B2Y 4A2

John P. Galloway
Branch of Alaska Geology—MS 917
U.S. Geological Survey
345 Middlefield Road
Menlo Park, California 94025

Philip R. Hill
Geological Survey of Canada
Atlantic Geoscience Centre
Bedford Institute of Oceanography
P.O. Box 1006
Dartmouth, Nova Scotia
Canada B2Y 4A2

Patricia F. McDowell
Department of Geography
University of Oregon
Eugene, Oregon 97403-1218

Donald R. Nichols
Branch of Geologic Risk Assessment
U.S. Geological Survey—MS 966
Box 25046 Denver Federal Center
Denver, Colorado 80225–0046

JOINT CANADIAN-AMERICAN INVESTIGATION OF THE CENOZOIC GEOLOGY OF THE LOWLANDS BORDERING THE BEAUFORT SEA

By Jean-Serge Vincent, L. David Carter, John V. Matthews, Jr., *and* David M. Hopkins

In late July and early August, 1985, we examined the geology of the lowlands along the margin of the Beaufort Sea in the Yukon and Northwest Territories of Canada and in Alaska (fig. 1). This field project was organized to address geologic problems identified during a joint Canadian-American workshop on the correlation of Quaternary deposits and events around the margin of the Beaufort Sea. The workshop was held in Calgary in 1984 and hosted by the Geological Survey of Canada (Heginbottom and Vincent, 1986).

Suites of sediments along the Canadian margin of the Beaufort Sea record up to four advances of glaciers of Laurentide provenance and associated sea-level changes of glacioisostatic origin. Deposits of varied origin also record extensive nonglacial periods extending back at least to the Miocene. After meeting in Inuvik, we flew to Banks Island where the longest and most complete record exists. J.A. Westgate (University of Toronto) and Michel Lamothe (Geological Survey of Canada) accompanied us on this part of the trip in order to collect, respectively, tephra and thermoluminescence (TL) samples. At the Duck Hawk Bluffs, near Sachs Harbour, Upper Cretaceous marine sediments (Kanguk Formation) and Tertiary (Miocene and Pliocene?) fossil-bearing deltaic sand and gravel (Beaufort Formation) are overlain by preglacial mainly eolian and lacustrine sediments of the Worth Point Formation. The Worth Point Formation is overlain by a complex sequence composed of lower Pleistocene glacial and marine deposits associated with the Banks glaciation, interglacial middle Pleistocene lacustrine, fluvial, and paludal sediments of the Morgan Bluffs Formation, marine deposits associated with the late(?) middle Pleistocene Thomsen glaciation, interglacial lacustrine and paludal sediments of the Cape Collinson Formation, glacial and marine deposits of the early Wisconsin M'Clure stade of the Amundsen glaciation, and finally by Holocene deposits of varied origin (Vincent and others, 1983). Deposits of Late Wisconsin age crop out only on the east coast of the island. Our studies showed that (1) Cretaceous tephra reported from the Worth Point Formation had been ice-thrust from the Kanguk Formation over Beaufort Formation sediments during the Banks glaciation; (2) the Worth Point Formation is more complex than previously assumed; and (3) the Beaufort Formation contains a rich macroflora that includes many taxa not previously known from it.

The group then moved to the Polar Continental Shelf Project Base at Tuktoyaktuk, on the mainland, and investigated various localities around Liverpool Bay, the Eskimo Lakes, and the Mackenzie River Delta. Glacial deposits of the middle(?) Pleistocene Mason River glaciation, of the Toker Point stade-Franklin Bay stade (early Wisconsin?), and of the Sitidgi stade (late Wisconsin) were briefly examined, but particular attention was given to pre-Toker Point stade nonglacial sediments (see Vincent, this volume, Hughes, this volume, and Rampton, 1982 for glacial limits and explanation of terminology). These nonglacial sediments include ancestral Mackenzie River sediments, eolian(?) sands of the Kittigazuit Formation, and complex sequences of middle Pleistocene and lower(?) upper late Pleistocene (Sangamon?) lacustrine, paludal, and marine deposits. Also investigated were marine shell-bearing sand and gravel that occurs on islands of the Mackenzie River Delta and postdates the Toker Point stade. Working out of Shingle Point, we examined exposures of the Buckland Till (= Toker Point stade), and the Flaxman Member of the Gubik Formation in Alaska (see Carter and Ager, this volume) was traced from the U.S.-Canada border east to Komakuk Beach, where it has been removed by erosion.

From Shingle Point the group moved to Barter Island in Alaska. Working out of this base, the group examined excellent exposures of the Flaxman Member at Simpson Cove and Pokok Bay, and Pliocene(?) marine deposits of the Colvillian transgression (see Carter and others, 1986, for terminology of marine transgressions) were examined in coastal exposures between Pokok Bay and Pokok Lagoon. Poorly exposed Pliocene(?) marine deposits of the Fishcreekian transgression near the Niguanak River and along Carter Creek also were visited. At Carter Creek we sampled the well-exposed Tertiary Nuwok Member of the Sagavanirktok Formation (Detterman and others, 1975). We visited and extensively sampled fluvial gravelly sand that is exposed on a tributary of the Niguanak River and is older than the Fishcreekian transgression and possibly older than any of the recognized late Cenozoic marine transgressions. This site has yielded an intriguing flora and insect fauna that suggests a forest/tundra environment (unpublished Geological Survey of Canada macrofossil reports available on request from John Matthews). Unlike the present-day forest/tundra zone, it contained larch and five-needle pine in addition to spruce. Both tree and shrub birches were present as well as some southern, nonboreal plants like honeysuckle (*Lonicera*).

The group then moved west to Prudhoe Bay and examined a variety of deposits ranging in age from Eocene to Holocene including (1) Eocene deltaic(?) deposits with intercalated tephras and overlying coarse gravel beds that also contain tephras, (2) gravel pits excavated in lower(?), middle, and upper Pleistocene deposits near the Kuparuk River and near Milne Point, (3) larch-bearing fluvial deposits on the Kikiakrorak River, (4) thick Wisconsin eolian sand at Judy Creek, (5) Pliocene(?) marine deposits of the Fishcreekian transgression at Fish Creek, (6) Pliocene(?) marine deposits of the Bigbendian transgression near Ocean Point on the Colville River, and (7) sedimentary facies and structures in the bed of a drained lake near Fish Creek.

Continuing west toward Barrow we stopped at the Kogru River to see Pelukian(?) and pre-Pelukian marine sediments and overlying marine sediments of the Simpsonian transgression (Flaxman Member of the Gubik Formation) at Cape Simpson. Out of Barrow, we spent two days studying the superposed Colvillian, Bigbendian, Fishcreekian, and Wainwrightian transgressions at Skull Cliff that were described by Brigham (1985).

Some of the most important findings of our trip were the following:

1. Plant macrofossils are widely present in the Beaufort and older parts of the Gubik Formations and will be useful in establishing a Neogene climatic framework of the margins of the Arctic Basin.
2. The Worth Point Formation of Banks Island, which records an interval during the late Pliocene or early Pleistocene when larch, rather than spruce, formed the tree line, may correlate with deposits at various sites in Alaska that apparently also document the same type of vegetation.
3. Similarly, the Banks glaciation, the oldest continental glacial event recognized in northwestern Canada, is probably recorded by glaciomarine deposits of late Pliocene or early Pleistocene age in Alaska.
4. Determining the stratigraphic position at which shield erratics first appear in Gubik marine beds would establish a minimum age for the first Laurentide glacial episode in Canada.
5. The widespread "brown sands" of the Kittigazuit Formation near Tuktoyaktuk, traditionally thought to be deltaic in origin, may instead be remnants of a large dune field.
6. The glaciomarine Flaxman Member extends into Canada on the Yukon Coastal Plain at least to Komakuk Beach, and is either the same age as or younger than the Laurentide Buckland Till laid down during the

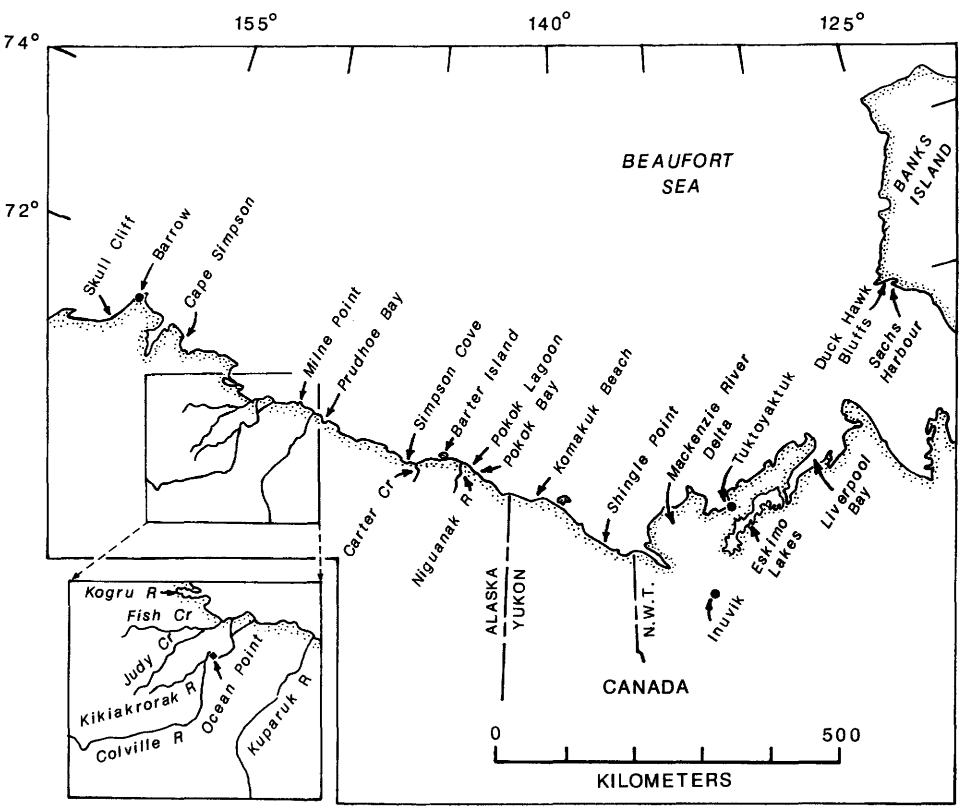


Figure 1. Location of places mentioned in text. N.W.T., Northwest Territories.

Buckland glaciation (=M'Clure stade of the Amundsen glaciation).

7. Although most of the marine deposits in Canada represent high sea levels due to glacioisostatic depressions, some of the high sea-level events were likely eustatic in origin. In particular, driftwood-bearing marine beds exposed along the northeast coast of Liverpool Bay were likely deposited during a eustatically higher sea level and may correlate with the Pelukian transgression of Alaska. In contrast, all Alaskan marine deposits formed during eustatic high sea level events, except perhaps part of the Flaxman Member.

REFERENCES CITED

Brigham, J.K., 1985, Marine stratigraphy and amino-acid geochronology of the Gubik Formation, western Arctic Coastal Plain, Alaska: U.S. Geological Survey Open-File Report 85-381, 218 p.

Carter, L.D., Brigham-Grette, Julie, and Hopkins, D.M., 1986, Late Cenozoic marine transgressions of the Alaskan Arctic Coastal Plain, *in* Heginbottom, J.A., and Vincent, J-S., eds., Correlation of Quaternary deposits and events around the margin of the Beaufort Sea: Geological Survey of Canada Open File Report 1237, p. 21–26.

Detterman, R.L., Reiser, H.N., Brosgé, W.P., and Dutro, J.T., Jr., 1975, Post-Carboniferous stratigraphy, northeastern Alaska: U.S. Geological Survey Professional Paper 886, 46 p.

Heginbottom, J.A., and Vincent, J-S., eds., 1986, Correlation of Quaternary deposits and events around the margin of the Beaufort Sea: Geological Survey of Canada Open File Report 1237, 60 p.

Rampton, V.N., 1982, Quaternary geology of the Yukon Coastal Plain: Geological Survey of Canada Bulletin 317, 49 p.

Vincent, J-S., Occhietli, S., Rutter, N., Lortie, G., Guilbault, J.P., and de Boutray, B., 1983, The late Tertiary-Quaternary stratigraphic record of the Duck Hawk Bluffs, Banks Island Canadian Arctic Archipelago: Canadian Journal of Earth Sciences, v. 20, no. 11, p. 1694–1712.

Carter, L. David, Hamilton, Thomas D., and Galloway, John P., eds., 1989, Late Cenozoic History of the Interior Basins of Alaska and the Yukon: U.S. Geological Survey Circular 1026.

Late Pleistocene Spruce (*Picea*) in Northern Interior Basins of Alaska and the Yukon: Evidence from Marine Deposits in Northern Alaska

By L. David Carter *and* Thomas A. Ager

The record of spruce (*Picea*) in upper Pleistocene sediments older than 20 ka in the northern interior basins of Alaska and the Yukon is poorly dated. The few finite radiocarbon dates on spruce occurrences have large analytical uncertainty (Hopkins and others, 1981b; Shackleton, 1982) and thus are of uncertain validity. Most late Pleistocene sediments that contain spruce macrofossils or abundant spruce pollen are older than the range of radiocarbon dating, and there is debate about whether they represent interstadial or interglacial conditions (Péwé, 1975; Schweger and Matthews, 1985). Marine deposits on the Alaskan Arctic Coastal Plain provide evidence that during the last interglacial interval and during at least one and possibly two Wisconsin interstadials spruce was growing in northern interior basins. We discuss this evidence below and comment on its significance for the interpretation of the climatic and sedimentary record of northern interior basins.

Marine deposits of the Alaskan Arctic Coastal Plain record at least two and possibly three late Pleistocene marine transgressions (Carter and others, 1986; Carter, unpub. data). The oldest of these was the Pelukian transgression, defined by Hopkins (1967) as producing shoreline features and deposits a few meters above present sea level that can be traced discontinuously around the coast of western and northern Alaska. Pelukian deposits on the Arctic Coastal Plain extend from Harrison Bay westward to near Barrow (fig. 1), and they are the youngest marine deposits that contain a fauna indicating more open water and warmer climatic conditions than exist at present (Hopkins and others, 1981a). Shells of the bivalve *Hiatella arctica* collected from Pelukian deposits yield amino-acid ratios that are barely distinguishable from those determined for modern specimens (Brigham and Miller, 1983), which supports Hopkins' (1967) conclusion that the Pelukian transgression occurred during the most recent pre-Holocene interglacial interval.

In deep-sea sediment cores, the last interglacial period is generally equated with oxygen-isotope stage 5e, which peaked about 123 ka, and is the youngest Pleistocene oxygen-isotope stage in which the oxygen isotope compositions of the shells of benthic foraminifers are similar to Holocene isotopic compositions (Martinson and others, 1987). We correlate the Pelukian transgression with stage 5e because oxygen-isotope compositions of shells of the bivalve *Astarte borealis* from Pelukian deposits are about the same as those for modern *A. borealis* shells from the Beaufort Sea (Carter and others, 1985). This correlation is supported by six of eight thermoluminescence (TL) ages on Pelukian sediments. The six analyses are between 119 and 134 ka and average about 124 ka (Carter, unpub. data). Two other TL ages of 89 and 140 ka fall outside the age range of oxygen-isotope stage 5e and may be spurious.

Driftwood logs up to 25 cm in diameter locally are common in Pelukian deposits of the Arctic Coastal Plain, and some of this wood has been identified as spruce (R.C. Koeppen, USDA Forest Products Laboratory, written commun., 1978). The most likely source for spruce driftwood is the Mackenzie River, but for the western part of the coastal plain a source in or north of the Yukon River drainage basin is also possible (Giddings, 1952), and a source on the North Slope of the Brooks Range cannot be ruled out. Spruce was thus almost certainly present in northern Alaska or northwestern Canada during the last interglacial interval. This conclusion was reached previously on the basis of the presence of spruce pollen, cones, and wood in interior basin sediments that were assumed to be of last-interglacial age (Hopkins, 1972; Péwé, 1975). However, the assumption of a last-interglacial age for these sediments was based in part on the presence of spruce pollen and macrofossils, and some of these same deposits have more recently been interpreted as much younger than 125 ka (Hopkins, 1982; Schweger and Matthews, 1985).

The next youngest transgression is informally named the Simpsonian transgression (Carter and others, 1986) and is represented by the Flaxman Member of the Gubik Formation, which occurs along the Beaufort Sea coast and inland to altitudes of about 7 m (fig. 1). The Flaxman Member is a glaciomarine deposit that contains ice-rafted erratics of Canadian provenance (Rodeick, 1979), with one of the source areas identified as Bathurst Inlet, about 1,000 km to the east of figure 1 (J.-S. Vincent, Geological Survey of Canada, personal commun., 1985). Transportation of the erratics to the Beaufort Sea coast by icebergs records the ablation of an ice sheet in the Canadian Arctic. Remains of Pacific marine mammals collected from the Flaxman Member include ribbon seal (*Histriophoca fasciata*) and gray whale (*Eschrichtius* sp.) (Repenning, 1983) and show that, as now, a connection with the Bering Sea existed during the Simpsonian transgression. Marine mollusk shells from the Flaxman Member are enriched in ^{18}O relative to modern specimens from the Beaufort Sea (Carter and others, 1985), indicating that there was more glacial ice during the Simpsonian transgression than there is today. The ^{18}O data also positively differentiate Simpsonian and Pelukian deposits.

Preliminary results from 18 TL analyses of sediments from the Flaxman Member range from 50 to 86 ka and average about 70 ka (Carter, unpub. data). These analyses are supported by a uranium-series age of 75 ka determined on whalebone (J.L. Bischoff, U.S. Geological Survey, written commun., 1984). Radiocarbon ages on organic materials from the Flaxman Member range from about 19.5 ka to indeterminate at more than 48 ka, and in some cases finite and indeterminate ages have been measured on pairs of samples from the same site (Sellmann and Brown, 1973; Stafford and others, 1987; Carter, unpub. data). For example, the uranium-series age cited above was measured on bone from a gray whale that previously had yielded radiocarbon ages of about 22.5 ka and >38 ka (Stafford and others, 1987). We conclude that finite radiocarbon ages on organic materials from the Flaxman Member are probably minimum ages, and we tentatively accept the TL ages as approximating the time of the Simpsonian transgression. We emphasize, however, that analyses of the TL data are incomplete, and at this time we can say only that the data suggest that the Simpsonian transgression coincided with one of the two global high sea-level events dated on New Guinea as 59 and 81 ka (Chappell and Shackleton, 1986).

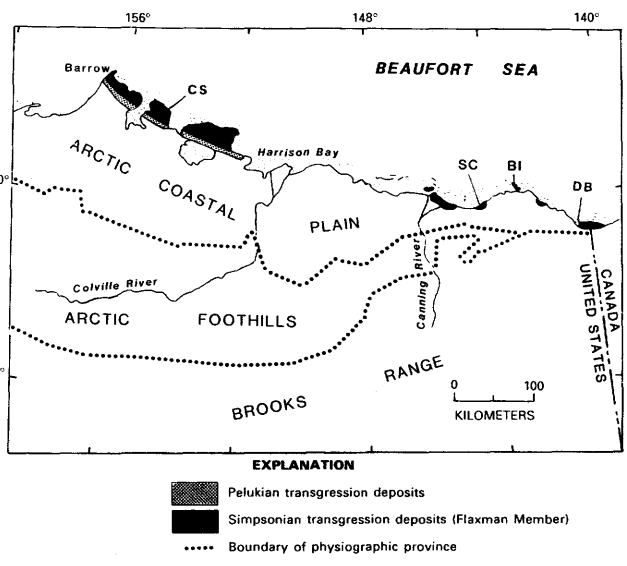


Figure 1. Northern Alaska showing distribution of deposits of Pelukian transgression and Simpsonian transgression (Flaxman Member of Gubik Formation) and locations mentioned in the text (CS, Cape Simpson; SC, Simpson Cove; BI, Barter Island; DB, Demarcation Bay).

Spruce macrofossils (identified by J.T. Quirk, USDA Forest Products Laboratory, written commun., 1983) have been collected from the Flaxman Member at Cape Simpson and near Barter Island (fig. 1), and have been dated by radiocarbon as 35.75±0.68 ka (USGS–1850) and >48 ka (USGS–1855), respectively. As mentioned above, we consider the finite age as a minimum age. Of the several sediment samples from the Flaxman Member examined for pollen by T.A. Ager, only one sample was suitable for quantitative analysis. This sample contained 7.1 percent spruce pollen (table 1), which falls within the range of percentages of spruce pollen in the modern pollen rain (Nelson, 1979). These data suggest that during at least part of the Simpsonian transgression spruce forests may have been as close as the Mackenzie River valley, or valleys within or south of the Brooks Range, but probably not north of the Brooks Range.

Evidence for possibly a third high sea-level event consists of marine sand that disconformably overlies the Flaxman Member at Demarcation Bay and at Cape Simpson. The marine sand could be merely a regressive facies of the Simpsonian transgression, but preliminary results of two TL analyses of the marine sand at Demarcation Bay are 41.5 and 48 ka (Carter, unpub. data), which suggests that it represents a separate sea-level highstand. Fine-grained organic detritus from the marine sand yielded a radiocarbon age of 29.39±0.24 ka (USGS–1856), but this age conflicts with the TL ages, and evidence from the Canadian Beaufort Sea shelf indicates that at about 27 ka relative sea level may have been as much as 140 m below modern sea level (Hill and others, 1985). Furthermore, the marine oxygen-isotope record suggests that the time of minimum ice volume, and presumably highest sea level, during oxygen-isotope stage 3 occurred about 50 ka (Martinson and others, 1987). We thus provisionally accept the preliminary TL data, and adopt as a working hypothesis an age of about 45 or 50 ka for this unnamed and tentatively identified marine transgression.

Small twigs identified as spruce (J.T. Quirk, USDA Forest Products Laboratory, written commun., 1983) occur in the marine sand at Demarcation Bay. They are localized in lenticular, organic-rich, cross-stratified beds that may be deltaic. If these beds are deltaic, then the spruce probably originated in the adjacent Brooks Range drainage. Otherwise, the most likely source for spruce driftwood on this part of the coastal plain is the Mackenzie River valley.

Evidence from marine sediments of northern Alaska thus indicates that spruce was growing in northern interior basins during at least two and possibly three late Pleistocene marine transgressions: the Pelukian transgression of last-interglacial age (125 ka), the Simpsonian transgression of early or middle Wisconsin age, and a tentatively identified and unnamed transgression of middle Wisconsin age. A pollen or macrofossil record of spruce should be present in interior basin sediments that are correlative with these

Table 1. Pollen and spores from Flaxman Member of Gubik Formation at Simpson Cove

[Analysis by T.A. Ager]

Taxon	Number of grains	Percent
Pollen type		
Betula (birch)	49	19.2
Picea (spruce)	18	7.1
Alnus (alder)	33	12.9
Salix (willow)	8	3.1
Myrica (sweet gale)	1	.4
Ericaceae (blueberry/crowberry type)	7	2.7
Rosaceae (rose family)	1	.4
Shepherdia canadensis (soapberry)	2	.8
Caryophyllaceae (pink family)	3	1.2
Compositae (Asteraceae)		
Tubuliflorae type (for example, *Aster*)	1	.4
Artemisia (wormwood)	6	1.7
Cruciferae (Brassicaceae) (mustard family)	12	4.7
Cyperaceae (sedge family)	28	11.0
Gramineae (Poaceae) (grass family)	75	29.4
Polygonum sp. (buckwheat family)	2	.8
Ranunculaceae undiff. (crow foot family)	3	1.2
Saxifragaceae undiff. (saxifrage family)	2	.8
Unknown	3	1.2
TOTAL POLLEN	255	
Spore type		
Monolete type fern spores	3	1.1
Sphagnum moss spores	17	6.2

marine transgressions, and some upper Pleistocene sediments in interior basins do indeed contain spruce macrofossils and relatively high percentages of spruce pollen. Known occurrences of such sediments are in the Old Crow basin in Yukon Territory, and within Alaska on the Seward Peninsula, in the Koyukuk basin, the Chandalar valley, and in the Fairbanks area (Hamilton, 1979; Schweger and Matthews, 1985). These sediments are all stratigraphically above the Old Crow tephra, which has been dated by TL as 86±8 ka (Wintle and Westgate, 1986) and 110±12 ka (Berger, 1987). These ages are concordant at two standard deviations and support earlier suggestions that the spruce-bearing sediments above the tephra formed after oxygen-isotope stage 5e and represent Wisconsin interstadial conditions (Hopkins, 1982; Schweger and Matthews, 1985). Some scientists, however, are reluctant to accept the TL ages for the tephra, because some of the overlying spruce-bearing sediments contain evidence for climatic conditions as warm as or warmer than today, and they consider it unlikely that such climatic conditions have occurred since oxygen-isotope stage 5e (Péwé, this volume; D.M. Hopkins, University of Alaska, oral commun., 1987). If a correlation with stage 5e is correct, however, then there is no unequivocal record of Wisconsin-age spruce in sediments of northern interior basins. We thus prefer to accept the TL data for the age of the Old Crow tephra because of the likelihood that some of the overlying sediments that contain spruce macrofossils or abundant spruce pollen may correlate with Wisconsin-age marine deposits in northern Alaska. If we are correct, then climate in the northern interior basins of Alaska and the Yukon may have been as warm as or warmer than today during at least one Wisconsin interstadial event.

REFERENCES CITED

Berger, G.W., 1987, Thermoluminescence dating of the Pleistocene Old Crow tephra and adjacent loess, near Fairbanks, Alaska: Canadian Journal of Earth Sciences, v. 24, p. 1975–1984.

Brigham, J.K., and Miller, G.H., 1983, Paleotemperature estimates of the Alaskan Arctic Coastal Plain during the last 125,000 years, *in* International Conference on Permafrost, 4th, Proceedings, Fairbanks, Alaska, July 17–22, 1983: Washington, D.C., National Academy Press, p. 80–85.

Carter, L.D., O'Neil, J.R., and Stipp, J.J., 1985, High sea level along the Alaskan arctic coast about 70 or 80 ka: Evidence for an Antarctic ice surge?: Arctic Land-Sea Interactions, 14th Arctic Workshop, Dartmouth, Nova Scotia, November 6–8, 1985, p. 22–24.

Carter, L.D., Brigham-Grette, J., and Hopkins, D.M., 1986, Late Cenozoic marine transgressions of the Alaskan Arctic Coastal Plain, *in* Heginbottom, J.A., and Vincent, J-S., eds., Correlation of Quaternary deposits and events around the Beaufort Sea: Geological Survey of Canada Open File Report 1237, p. 21–26.

Chappell, J., and Shackleton, N.J., 1986, Oxygen isotopes and sea level: Nature, v. 324, p. 137–140.

Giddings, J.L., Jr., 1952, Driftwood and problems of Arctic sea currents: Proceedings of the American Philosophical Society, v. 96, p. 129–142.

Hamilton, T.D., 1979, Quaternary stratigraphic sections with radiocarbon dates, Chandalar quadrangle, Alaska: U.S. Geological Survey Open-File Report 79–751, 17 p.

Hill, P.R., Mudie, P.J., Moran, K., and Blasco, S.M., 1985, A sea-level curve for the Canadian Beaufort shelf: Canadian Journal of Earth Sciences, v. 22, p. 1383–1393.

Hopkins, D.M., 1967, Quaternary marine transgressions in Alaska, *in* Hopkins, D.M., ed., The Bering Land Bridge: Stanford, Calif., Stanford University Press, p. 47–90.

———1972, The paleogeography and climatic history of Beringia during late Cenozoic time: Internord, v. 12, p. 121–150.

———1982, Aspects of the paleogeography of Beringia during the late Pleistocene, *in* Hopkins, D.M., Matthews, J.V., Jr., Schweger, C.E., and Young, S.B., eds., Paleoecology of Beringia: New York, Academic Press, p. 3–28.

Hopkins, D.M., McDougall, K., and Brouwers, E., 1981a, Microfossil studies of Pelukian and Flaxman deposits, Alaska coast of the Beaufort Sea, *in* Smith, P.A., Hartz, R.W., and Hopkins, D.M., eds., Offshore permafrost studies and shoreline history as an aid to predicting offshore permafrost conditions: U.S. National Oceanic and Atmospheric Administration, Environmental Assessment of the Alaskan Continental Shelf, Annual Report, Task D–9, Research Unit 204 and 273, April 1979 to March 1980, Appendix G, p. 64–71.

Hopkins, D.M., Smith, P.A., and Matthews, J.V., Jr., 1981b, Dated wood from Alaska and the Yukon: Implications for forest refugia in Beringia: Quaternary Research, v. 15, p. 217–249.

Martinson, D.G., Pisias, N.G., Hays, J.D., Imbrie, J., Moore T.C., Jr., and Shackleton, N.J., 1987, Age dating and the orbital theory of ice ages: Development of a high-resolution 0 to 300,000-year chronostratigraphy: Quaternary Research, v. 27, p. 1–29.

Nelson, R.E., 1979, Quaternary environments of the Arctic Slope of Alaska: Seattle, University of Washington, M.S. thesis, 141 p.

Péwé, T.L., 1975, Quaternary geology of Alaska: U.S. Geological Survey Professional Paper 835, 145 p.

Repenning, C.A., 1983, New evidence for the age of the Gubik Formation, Alaskan North Slope: Quaternary Research, v. 19, p. 356–372.

Rodeick, C.A., 1979, The origin, distribution, and depositional history of gravel deposits on the Beaufort Sea continental shelf, Alaska: U.S. Geological Survey Open-File Report 79-234, 87 p.

Schweger, C.E., and Matthews, J.V., Jr., 1985, Early and middle Wisconsinan environments of eastern Beringia: Stratigraphic and paleoecological implications of the Old Crow tephra: Géographie Physique et Quaternaire, v. 39, p. 275–290.

Sellmann, P.V., and Brown, J., 1973, Stratigraphy and diagenesis of perennially frozen sediments in the Barrow, Alaska region, *in* The North American Contribution to the Second International Conference on Permafrost, Yakutsk, U.S.S.R., July 13–28, 1973: Washington, D.C., National Academy of Sciences, p. 171–181.

Shackleton, J., 1982, Environmental histories from Whitefish and Imuruk Lakes, Seward Peninsula, Alaska: Ohio State University, Institute of Polar Studies Report No. 76, 49 p.

Stafford, T.W., Jr., Jull, A.J.T., Brendel, K., Duhamel, R.C., and Donahue, D., 1987, Study of bone radiocarbon dating accuracy at the University of Arizona NSF Accelerator Facility for Radioisotope Analysis: Radiocarbon, v. 29, p. 24–44.

Wintle, A.G., and Westgate, J.A., 1986, Thermoluminescence age of the Old Crow tephra in Alaska: Geology, v. 14, p. 594–597.

Carter, L. David, Hamilton, Thomas D., and Galloway, John P., eds., 1989, Late Cenozoic History of the Interior Basins of Alaska and the Yukon: U.S. Geological Survey Circular 1026.

Offshore Constraints on the Late Pleistocene Glacial History at the Mouth of the Mackenzie River

By Steve M. Blasco, Julie Brigham-Grette, *and* Philip R. Hill

The middle to late Pleistocene history of the Mackenzie River Delta, Mackenzie trough, and adjacent continental shelf of the Beaufort Sea has a direct bearing on the interpretation of Laurentide ice limits, which influenced the chronology of events in the upper Porcupine and Peel River valleys of the northern Yukon. The limit of the early Wisconsin Buckland glaciation as mapped by Rampton (1982; and see Vincent, this volume) extends offshore beyond Herschel Island, where Mackay (1959) originally described ice-thrusted sediments in coastal exposures coincident with this former ice limit. To the south, the Buckland limit has been traditionally extended by Hughes (see Hughes, 1986) to be correlative with the ice limits of the Hungry Creek glaciation (Hughes and others, 1981). Ice occupying this position in the north-central Yukon dammed the eastward drainage of the Peel River and impounded glacial lakes in the Old Crow, Bluefish, and Bell River basins just after 36.9 ka (Hughes and others, 1981). Discrepancies in the age and correlation of this maximum ice limit from Hungry Creek to the Beaufort Sea remain unresolved despite fieldwork in the intervening areas by Catto (1986), causing some to speculate that ice may have occupied a similar position more than once during the late Pleistocene. Alternate interpretations bring into question the age of the Buckland limit along the Yukon coast and the age of the Hungry Creek till. This abstract outlines ongoing work offshore that is relevant to determining the sequence and age of glacial events in the middle to late Pleistocene that influenced the stratigraphy of both the northern Yukon and the Beaufort Continental Shelf.

The upper 80 to 100 m of sediment on the central part of the Beaufort Continental Shelf north of Richards Island and the Tuktoyaktuk Peninsula is composed of nearshore marine silt and clay overlain by glaciofluvial outwash sand, which is in turn overlain by a thin transgressive sequence that grades upward into a veneer of recent marine mud discharged from the modern Mackenzie River. This succession formed after 21 ka (Hill and others, 1985). Extensive seabed surveys by industry to locate offshore gravel and drift deposits as a source for construction materials have led to the identification of localized lag and barrier island deposits related to the last transgression, but no glacial tills or ice-contact deposits have been found. If present, such deposits must lie deeper than 80 m beneath the seabed. The glaciofluvial outwash complex of this succession was probably deposited fairly rapidly after 21 ka when an ice front was not far to the south. Seismic and chronologic evidence outlined by Hill and others (1985) suggests continued sea-level rise from at least 27 ka with an inflection around 15 ka, which is interpreted as indicating a slight readvance of the ice. The ice front was probably not far south of the present shoreline from late-middle Wisconsin time on, with a possible northward pulse about 15 ka.

On the western part of the Beaufort (or Yukon) Continental Shelf, northwest of Herschel Island, a discontinuous veneer of reworked sediments related to the last transgression and recent sedimentation unconformably overlies fine-grained sediments deposited before 53 ka (Julie Brigham-Grette, unpub. radiocarbon and thermoluminescence ages). Again, there is no evidence within this shallow sequence of an ice advance on the shelf.

Between the central and western parts of the shelf lies the Mackenzie trough, a northwest-trending linear depression 80 km wide by 150 km long that is partly infilled by more than 300 m of sediment. Seismostratigraphic evidence suggests that the trough was excavated by the repeated action of ice tongues during the Pleistocene rather than forming as a graben, slump, or river valley (Blasco and others, in press). The base of the infilled trough is well defined by a smooth U-shaped angular unconformity 500 m below sea level, which rises to within 100 m of the seafloor on the western shelf and lies about 200 m below the seabed at the edge of the central shelf (fig. 1). The chronostratigraphy of the Yukon shelf indicates that this basal unconformity formed before 53 ka. On the central shelf, extrapolated Holocene sedimentation rates suggest that the base-of-trough unconformity formed before 70 ka.

Four seismostratigraphic units with a composite thickness of 300 m have been identified in the Mackenzie trough (fig. 1). Occupying a valley in the central section of the otherwise broad, flat base are unsampled sediments (unit 4) with a chaotic, hummocky acoustic character and a maximum thickness of 70 m. We believe that these sediments may be a basal till. This unit is overlain by an acoustically massive to poorly stratified sand (unit 3), as much as 200 m thick, which rests directly on the basal unconfirmity over much of the trough. The contract between this sand and the overlying unit 2 is defined by a smooth, weak, trough-wide reflector which appears to terminate against the southwest wall of the trough and to rise into the shallow strata of the central shelf. The planar nature of this unconformity is interrupted along the western axis of the trough by a ridge with approximately 30 m of relief. Unit 2 is an acoustically unstratified, trough-wide clay with a thickness that ranges from 40 to 60 m. The upper surface of this unit is well defined and can be mapped across the trough as a flat-lying to locally hummocky reflector. On the southwest wall, this event intersects the base-of-trough unconformity and rises higher in the section, truncating shallow sediments, to join with the present seafloor of the Yukon shelf. These seafloor sediments (beneath the thin discontinuous surficial veneer discussed above) were deposited before 53 ka.

This reflector is traceable onto the central shelf, where it can be mapped at about 40 m below the seabed. Sediments at this depth are known to be of late Wisconsin age. Overlying this unconformity within the trough is unit 1, a late Wisconsin to Holocene sequence consisting of prograded deltaic bottomset, foreset, and topset beds, truncated at their upper surface by transgressive silt and clay, which is in turn overlain by thick recent marine mud. Unit 1 has a maximum thickness of 125 m. In the east-central part of the trough, 20 m of thick marine mud was deposited since 9.5 ka. Palynological analysis suggests that the base of unit 1 may have formed 12 to 14.6 ka (P.R. Hill, unpub. data) when compared with pollen sequences on land (Ritchie, 1984).

Tentative interpretation of the offshore stratigraphy suggests that the trough was excavated to its present maximum depth by a major ice advance during early Wisconsin time, or earlier, leaving a basal till deposit. The sand and clay (units 3 and 2) represent either a thick, fining-upward transgressive sequence related to the subsequent sea-level rise, or glacial deposits of succeeding advances. These sediments were subsequently overridden and marginally eroded by a limited middle(?) to late Wisconsin ice advance before 14 ka. Loading by ice may have deformed the underlying sand and clay units and destroyed their internal stratigraphy. This scenario correlates well with Beget's (1987) model of thin late Wisconsin ice in the Mackenzie River Delta. Ice retreat was followed by the deposition of a subaqueous deltaic sequence in which no ice-bonding has been observed. Transgression overtook progradation about 12 ka in the central part of the trough. Rising sea levels truncated topset beds, and water depths became great enough for the deposition of marine mud about 9.5 ka. More than 10 m of marine sediments has been deposited in the central part of the trough over the last 6,800 years.

In summary, a major ice advance during early Wisconsin time, or earlier, excavated the Mackenzie trough to its present maximum depth. The trough may have been subsequently reoccupied by ice at least once prior to the last advance. This middle to late Wisconsin ice margin was located not far south of the present shoreline. A thin ice tongue reoccupied the trough at this time but retreated by 14 ka.

The middle to late Pleistocene glacial history of the Beaufort Continental Shelf is very poorly constrained. In addition, evidence to support credible correlations with the onshore stratigraphy is clearly absent, making it difficult at present to contribute to the solution of the conflicting interpretations of the onshore record.

REFERENCES CITED

Beget, James, 1987, Low profile of the northwest Laurentide Ice Sheet: Arctic and Alpine Research, v. 19, p. 81–88.

Blasco, S.M., O'Connor, M.J., Hill, P.R., and Fortin, G., (in press), Late Quaternary history of the Canadian Beaufort continental shelf, *in* Grantz, A., Johnson, L., and Sweeney, J., eds., The geology of the Arctic Ocean region: Boulder, Geological Society of America Decade of North American Geology.

Catto, N.R., 1986, Quaternary sedimentology and stratigraphy, Peel Plateau and Richardson Mountains, Yukon and N.W.T.: Edmonton, University of Alberta, Ph.D. thesis, 358 p.

Hill, P.R., Mudie, P.J., Moran, K., and Blasco, S.M., 1985, A sea level curve for the Canadian Beaufort Shelf: Canadian Jour-

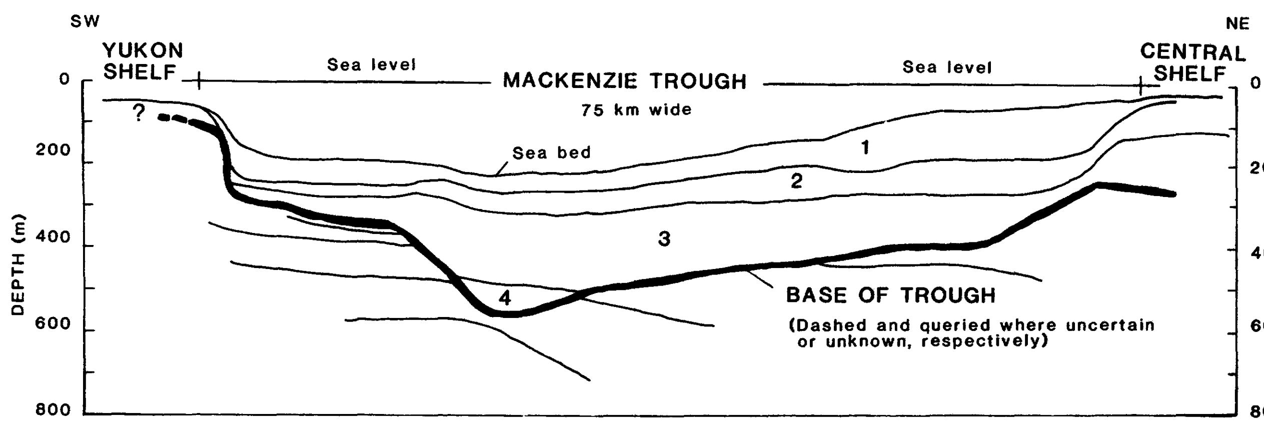


Figure 1. Schematic cross section of Mackenzie trough at approximately 70° north. Seismostratigraphic units 1–4 discussed in text.

nal of Earth Sciences, v. 22, p. 1383–1393.
Hughes, O.L., 1986, Quaternary chronology of Yukon Territory and the western District of Mackenzie: Geological Survey of Canada Open File Report 1237, p. 13–17.
Hughes, O.L., Harrington, C.R., Janssens, J.A., Matthews, J.V., Morlan, R.E., Rutter, N.W., and Schwege, C.E., 1981, Upper Pleistocene stratigraphy, paleoecology, and archaeology of the northern Yukon interior, eastern Beringia, 1. Bonnet Plume Basin: Arctic, v. 34, p. 329–365.
Mackay, J.R., 1959, Glacier ice-thrust features of the Yukon coast: Geographical Bulletin, p. 5–21.
Rampton, V.N., 1982, Quaternary geology of the Yukon Coastal Plain: Geological Survey of Canada Bulletin 317, 49 p.
Ritchie, J.C., 1984, Past and present vegetation of the far northwest of Canada: Toronto, University of Toronto Press, 251 p.

Geological Survey of Canada Contribution 47188

Carter, L. David, Hamilton, Thomas D., and Galloway, John P., eds., 1989, Late Cenozoic History of the Interior Basins of Alaska and the Yukon: U.S. Geological Survey Circular 1026.

Continental Ice Advances in Northwestern Canada and their Significance to Interior Basins in the Yukon

By Jean-Serge Vincent

On the basis of stratigraphic and chronologic evidence in the western District of Mackenzie, the Beaufort Sea coastal plain, and the southwestern Canadian Arctic Archipelago, it has been proposed that continental glaciers advanced westward to the Cordilleran mountain front at least four times during the late Cenozoic (fig. 1; Vincent, 1989). In the westernmost District of Mackenzie, weathered fluvial gravels along the Snake River, Peel River, and Rat River contain stones derived from the Canadian Shield (Catto, 1986). The possibly pre-Wisconsin tills in the Franklin Mountains (Rutter and others, 1973) provide additional evidence for early or middle Pleistocene continental ice advances to the Cordilleran mountain front. However, the westernmost extent of this ice has not been determined.

In the late Pleistocene, two Wisconsin stadial advances to the Cordilleran front are well recorded. The oldest of these is named the Hungry Creek glaciation in the Bonnet Plume Basin (Hughes and others, 1981), the Toker Point stade in the Tuktoyaktuk coastlands (Rampton, 1988), and the Buckland glaciation on the Yukon Coastal Plain (Rampton, 1982). It is this advance that is commonly thought to have been the most extensive ice advance on the mainland during the Pleistocene. However, because ice advances were distinctly more extensive in the southwestern Arctic Archipelago during the early Pleistocene Banks glaciation and the middle Pleistocene Thomsen glaciation than they were in late Pleistocene time (Vincent, 1986), it is possible that ice on the mainland, which came from the same dispersal center, also was most extensive during early and middle Pleistocene advances.

The younger Wisconsin stadial advance is named the Sitidgi stade on the coastal plain (Rampton, 1989), the Russell stade in the southwestern Canadian Arctic Archipelago (Vincent, 1983), and the Tutsieta Lake phase in the District of Mackenzie (Hughes, 1987). The age of this advance is generally accepted as being latest late Wisconsin, but the age of the earlier ice advance is still the subject of some debate, particularly because of the seemingly divergent record of the Bonnet Plume and Rat River basins with that of the Beaufort Sea Coastal Plain (Vincent, 1989; Vincent and Prest, 1987; Blasco and others, this volume). In the Bonnet Plume and Rat River basins, as well as in the interior basins of the Yukon, radiometric ages on materials either predating the older Wisconsin ice advance or closely associated with it argue for a latest middle Wisconsin or earliest late Wisconsin age. In contrast, data from the coastal plain argue for a pre-middle Wisconsin age, perhaps during oxygen isotope stage 4.

Whatever the chronology of Pleistocene continental ice advances, the evidence seems to indicate that ice repeatedly advanced to the Cordilleran front. This being the case, one can try to predict the potential impact of these successive continental ice advances on the interior basins of the Yukon, which were separated from Laurentide ice only by the relatively narrow barrier of the Richardson Mountains. It is then interesting to examine the documented record and see to what extent the predicted impacts have been recognized.

The most obvious impact is certainly that of substantial climate and vegetation change due to the presence of the nearby ice. Other impacts, which I would like to discuss here, are the potential hydrological changes that would result from ice abutting the Cordilleran front. Several scenarios, which could each occur during any ice advance, can be proposed: (1) Drainage of any river flowing eastward across the Richardson Mountains from the interior basins of the Yukon toward the Mackenzie River basin would be impeded, and waters would be retained west of the ice front in lakes which would grow in size until the impounded waters could find an outlet to the north along the ice front or to the west in the unglaciated area toward Alaska. (2) Drainage of rivers on the east flank of the Richardson Mountains or those flowing into the Mackenzie River basin from the Ogilvie, Selwyn, or Mackenzie Mountains, via the Peel River, would be impounded into lakes and eventually diverted into basins in the Yukon if no outlet was present to the north within the Mackenzie River basin along the ice front. (3) Because of tilting resulting from the weight of continental glaciers, one could also envisage that headwaters of streams on the west side of the Richardson Mountains could be diverted into the Mackenzie River basin. There the waters could find their way north along an outlet at the margin of the ice front or be returned eventually to basins in the Yukon when proglacial lakes overflowed to the west. (4) In any of the three scenarios mentioned above, advances or retreats of the ice margin could initiate or interupt drainage of waters into basins in the Yukon thus causing sporadic disruption of the hydrological regime in the interior basins.

Although several cycles of major hydrologic changes could theoretically have occurred, based on the assumption

Figure 1. Proposed limits and tentative ages of Quaternary continental ice in northwestern Canada. Limits queried where uncertain.

that ice extended to the mountain front on several occasions during the Pleistocene, such changes are only well documented for the period spanning latest middle Wisconsin to late Wisconsin time. In the early Pleistocene a (single or multiple) lacustrine episode is well documented (the lower lacustrine unit in the Old Crow and Bluefish basins), but it is thought to be related to tectonic activity such as faulting or basin subsidence rather than glaciation (Schweger, this volume). No record of middle Pleistocene lacustrine events has thus far been documented. In latest middle Wisconsin to late Wisconsin time, ice advanced toward McDougall Pass (between the present headwaters of the Bell and Rat Rivers), caused the damming of the previously east-flowing Porcupine/Bell/Rat Rivers system, created a proglacial lake that occupied the upper Rat River and Bell, Bluefish and Old Crow Basins, and caused water to be diverted west into Alaska via the Ramparts (Catto, 1986; scenario 1 above). At the same time waters dammed in the Peel River basin were diverted north into the Yukon basin, via the Eagle River, cutting into Ford Lake Shale in the process and dispersing palynomorphs of Mississipian age into the basin area (scenario 2; see also Hughes, this volume; and Schweger, this volume). The two late Pleistocene events just mentioned above were likely contemporaneous, and they are the only ones which document a close link between known continental ice advances in the Mackenzie River basin and known major hydrological changes in the interior basins in the Yukon.

The absence of hydrologic evidence in the interior basins in the Yukon for continental glacial events before middle Wisconsin time is somewhat surprising. Why are older events not recorded? Lacustrine sediments that formed prior to latest middle Wisconsin time occur in the Peel River basin (the Deception lacustrine sediments of Hughes and others, 1981), but no evidence of discharge into the Yukon basins has yet been recognized. Perhaps discharge during this event and all other probable early or middle Pleistocene advances was always east of the drainage divide along the mountain front? Are there still possibilities of finding, in the sedimentary record in the Yukon basins, clear evidence of hydrological changes related to the pre-late Wisconsin presence of continental ice? Attempts at answering these questions as well as at investigating the possible role of tilting due to ice loading east of the Richardson Mountains also should be made.

Because of the increasingly detailed chronostratigraphic framework evolving for the interior basins, any direct link that can be made with glacial events east of the Richardson Mountains is critical to the establishment of the chronology of continental ice advances in North America. The interior basins of the northern Yukon, with their thick suite of nonglacial sediments, including marker beds such as widespread tephra layers, offer a unique opportunity for indirectly dating continental glacial events. The chronology derived from the basins for the latter part of late Pleistocene time has significant regional implications. For continental ice to be present in latest middle Wisconsin time at the Cordilleran front, which is far from the Keewatin dispersal center west of Hudson Bay, it must have developed considerably before that time over much of North America. Vincent and Prest (1987) have estimated, in their growth model of the Laurentide ice sheet, that at least 20,000 years would be needed to achieve this position from a Keewatin ice-dispersal center. If the chronology in the basins and the estimate for the growth period are correct, and if the dispersal center was located in Keewatin and not farther west in the District of Mackenzie, then much more ice existed over North America during middle Wisconsin time than is generally accepted. Dredge and Thorleifson (1987), partly on the basis of calculated global ice volumes, have argued for substantial coverage by the Laurentide ice sheet, particularly in northwestern North America, during middle Wisconsin time. These various considerations require a complete reevaluation of our concept of the chronology and understanding of Wisconsin glaciations. The record in the Yukon interior basins should provide much crucial information for this reevaluation.

REFERENCES CITED

Catto, N.R., 1986, Quaternary sedimentology and stratigraphy, Peel Plateau and Richardson Mountains, Yukon and NWT: Edmonton, University of Alberta, Ph.D. thesis, 728 p.

Dredge, L.A., and Thorleifson, H., 1987, The Middle Wisconsinan history of the Laurentide Ice Sheet: Géographie Physique et Quaternaire, v. 41, p. 215–235.

Hughes, O.L., 1987, Late Wisconsinan Laurentide glacial limits of northwestern Canada: The Tutsieta Lake and Kelly Lake phases: Geological Survey of Canada Paper 85–25, 19 p.

Hughes, O.L., Harington, C.R., Janssens, J.A., Matthews, J.V., Jr., Morlan, R.E., Rutter, N.W., and Schweger, C.E., 1981, Upper Pleistocene stratigraphy, paleoecology, and archaeology of the northern Yukon Interior, eastern Beringia, 1. Bonnet Plume Basin: Arctic, v. 34, p. 329–365.

Rampton, V.N., 1982, Quaternary geology of the Yukon coastal plain: Geological Survey of Canada Bulletin 317, 49 p.

———1988, Quaternary geology of the Tuktoyaktuk Coastlands, Northwest Territories: Geological Survey of Canada Memoir 423, 98 p.

Rutter, N.W., Boydell, A.N., Savigny, K.W., and Van Everdingen, R.O., 1973, Terrain evaluation with respect to pipeline construction, Mackenzie Transportation Corridor, southern part, Lat. 60°–64°: Environmental-Social Committee Northern Pipelines, Report Number 73–36, 145 p.

Vincent, J-S., 1983, La géologie du Quaternaire et la géomorphologie de l'ile Banks, Arctique canadien: Commission géologique du Canada Mémoire 405, 118 p.

———1986, Framework of Quaternary events of the northern Interior Plains of Canada, *in* Heginbottom, J.A., and Vincent, J-S., eds., Correlation of Quaternary deposits and events around the margin of the Beaufort Sea: Contributions from a joint Canadian-American Workshop, April 1984: Geological Survey of Canada Open File Report 1237, p. 6–9.

———1989, Quaternary geology of the northern Canadian interior plains, *in* Fulton, R.J., ed., Quaternary geology of Canada and Greenland: Geological Survey of Canada, Geology of Canada Number 1. [Also Geological Society of America, The Geology of North America, v. K–1]

Vincent, J-S., and Prest, V.K., 1987, The Early Wisconsinan history of the Laurentide Ice Sheet: Géographie Physique et Quaternaire, v. 41, no. 2, p. 199–213.

Geological Survey of Canada Contribution 20887

Carter, L. David, Hamilton, Thomas D., and Galloway, John P., eds., 1989, Late Cenozoic History of the Interior Basins of Alaska and the Yukon: U.S. Geological Survey Circular 1026.

Was the Late Pleistocene Northwest Laurentide Ice Sheet Wet-Based?

By James Beget

The Laurentide ice sheet was the largest glacier complex generated in the northern hemisphere. Few data are available on the thickness of much of the Laurentide ice sheet, principally because it originated and dispersed over areas of low relief. However, on both its northwestern and southwestern margins, Laurentide ice limits and glacier thicknesses for both the latest Pleistocene and penultimate glaciations can be reconstructed over reaches of hundreds of kilometers where the ice sheet terminated against the western Cordillera (Mathews, 1974; Beget, 1987). Field evidence in this region suggests that the Laurentide ice sheet was only 10 to 25 percent as thick as typical ice sheet profiles from Antarctica or Greenland. Typical basal shear stresses developed by the thin ice sheet margins were 1 to 30 kPa, much less than the 100 kPa thought typical of ice sheets and generally used in theoretical ice sheet reconstructions (Beget, 1987).

Boulton and Jones (1979) and Fisher and others (1985) suggested that in areas where the Laurentide ice sheet rested on weak, unconsolidated sediments the glacier's thickness would have been strongly influenced by the mechanical strength of the subglacial sediment. Although it is commonly assumed that basal tills are hard, dewatered, and overconsolidated at the time of deposition, sheared and water-saturated tills have much lower shear strength than ice. In situ measurements of till rheology at Breidamerkurjokull, Iceland, by Boulton and Jones (1979) showed that the strength of sheared till was only 20 to 26 kPa. The average cohesion (shear strength at zero normal stress) experimentally determined for poorly sorted clay-sand-silt mixtures similar to till range from only 6 to 28 kPa (Huang, 1983; Selby, 1982). The average shear strength of 10 representative tills when water saturated was only 10±10 kPa (Milligan, 1976). The yield strength of sediments is strongly influenced by the grain size distribution and the amount and types of clay present in the sediment (Johnson, 1970; Hampton, 1975; Rodine and Johnson, 1976), but the presence of water also plays a critical role in reducing shear strength of sediment.

I have suggested that the low shear stress characteristic of the northwest Laurentide ice sheet in the Mackenzie River basin indicates that the glacier in this area was wet-based, and that subglacial tills were water-saturated and at low or zero effective normal stress due to high pore-water pressures (Beget, 1987). This inference is supported by independent, geomorphic evidence for a "warm-melting" subglacial thermal regime beneath the northwest Laurentide ice sheet (Sugden, 1977, 1978).

The low profile of the northwest Laurentide ice sheet probably in part explains why the ice sheet was confined to the broad valley of the MacKenzie River, rather than overtopping the Richardson Mountains and entering the interior basins to the west. The history of lake formation in the Old Crow Basin may also be related to the dynamics of the Laurentide ice sheet. The field evidence seems to require a wet-based glacier, in spite of the extremely cold temperatures which characterized this area during Pleistocene glaciations. Here I examine possible mechanisms to generate water at or introduce water to the base of the northwest Laurentide ice sheet, and make first-order estimates of the amounts of water available.

Possible sources of water beneath the late Pleistocene northwest Laurentide ice sheet

Colder than modern air temperatures during the late Pleistocene probably had little significant effect on the amount of thermal energy introduced to the glacier by geothermal heating or mechanical heating during sliding or sediment deformation beneath the Laurentide ice. It is possible to estimate the magnitude of geothermal meltwater production (M_g) from the equation

$$M_g = Y_g/\rho_i L \qquad (1)$$

where Y_g is the geothermal heat flux, ρ_i is ice density, and L is the latent heat of fusion of ice. M_g in nonvolcanic areas with average heat flow is typically on the order of 0.005 to 0.01 m/yr. Similarly, if the value of Y_s, or heat production due to sliding and (or) creep in basal sediment, were known, it would be possible to estimate meltwater production due to sliding (M_s) from the equation

$$M_s = Y_s/\rho_i L. \qquad (2)$$

The flow velocities of the northwest Laurentide ice sheet are poorly constrained, restricting the application of equation 2. The range of meltwater production due to internal energy dissipation (M_s) in modern glaciers is 0.01 to 0.4 m/yr (Drewry, 1986), and this serves as a good first approximation for the Pleistocene northwest Laurentide ice sheet.

Additional possible sources of water include rain during the ablation season. While it is generally agreed that the Alaskan and Canadian Arctic during Pleistocene glacial episodes was somewhat more arid than present, it is still likely that some rain fell directly on the marginal parts of the Laurentide ice sheet during summers. Current summer precipitation in northwest Canada is on the order of 10 cm/yr. Perhaps as much as several centimeters of rain may have fallen on the glacier each summer, some portion of which may have traveled to the base of the glacier via moulins or crevasses. Similarly, some input of water to the glacier system may have occurred from groundwater, or from runoff from surrounding nonglaciated areas to ice-marginal streams.

I have previously suggested that during the period of late Pleistocene deglaciation, if summer temperatures and atmospheric conditions were at all similar to those of today, as much as several meters of meltwater may have been produced annually during the ablation season from the surface of the northwest Laurentide ice sheet (Beget, 1987). It is more difficult to approximate earlier Pleistocene conditions when the glacier was growing or in equilibrium, but ablation rates were almost certainly lower. If, as a first approximation, late Pleistocene ablation rates at latitude 69° N. are assumed to have been similar to those of the northernmost glaciers known today at 80°–82° N., then ablation rates of 1.0 to 0.5 m/yr might be expected (Budd and Allison, 1975). Such ablation rates would be several times to an order of magnitude greater than the possible water production estimated above from all other sources.

It is possible that surface meltwater reached the base of the Laurentide ice sheet via moulins, as has been suggested for the White Glacier on Axel Heiberg Island at 80° N. (Iken, 1972; Muller and Iken, 1973). The White Glacier is wet-based despite mean annual air temperatures well below freezing. Significant velocity increases were observed at this high-latitude glacier following periods of increased surface ablation, suggesting that surface meltwater was efficiently transmitted through the 200- to-300-m-thick glacier and distributed across the glacier base.

Although most meltwater on the surface of the northwest Laurentide ice sheet probably refroze or ran off in surface drainage, the contribution of even a small percentage of the water available from surface melt to the glacier base would probably exceed the amount of water available from all other sources. The high heat capacity and latent heat of fusion of water would also introduce thermal energy to the glacier base.

If water was present at the base of the northwest Laurentide ice sheet due to geothermal heating, frictional heating, surface meltwater influx, or some combination of these factors, then the rate of introduction of water to the glacier base, calculated by summing the individual estimates above, might have been as much as 1.5 m/yr. Groundwater charging and influx from marginal drainages might contribute still more. While much of any water at the glacier base undoubtedly exited the system as streamflow, this analysis suggests that sufficient sources of water existed during the Pleistocene for the northwest Laurentide ice sheet to have been wet-based.

Summary

Recent mapping indicates profiles of Pleistocene glacier lobes on the northwest perimeter of the Laurentide ice sheet were unusually thin. If the low profile of the northwest Laurentide ice sheet is due to subglacial deformation of low strength sediments in the MacKenzie River Delta area, then tills and subglacial sediments contained water and the glacier was wet-based. Possible sources for water at the glacier base might include geothermal or frictional melting of the glacier base, groundwater recharge, influx of ice-marginal drainage, or introduction of surface melt via moulins. The latter is potentially the most significant, and has been shown to be an important process at modern high latitude glaciers.

I thank Professors D.M. Hopkins and M. Keskinen for reading early versions of the manuscript.

REFERENCES CITED

Beget, J.E., 1987, Low profile of the northwest Laurentide Ice Sheet: Journal of Arctic and Alpine Research, v. 19, p. 81–88.

Boulton, G.S., and Jones, A.S., 1979, Stability of temperate ice caps and ice sheets resting on beds of deformable sediment: Journal of Glaciology, v. 24, p. 29–43.

Budd, W.F., and Allison, I.F., 1975, An empirical scheme for estimating the dynamics of unmeasured glaciers: International Association for Scientific Hydrology Publication No. 104, p. 246–256.

Drewry, D., 1986, Glacial geologic processes: London, Edward Arnold, 276 p.

Fisher, D.A., Reeh, N., and Langley, K., 1985, Objective reconstructions of the late Wisconsin Laurentide ice sheet and the significance of deformable beds: Géographie Physique et Quaternaire, v. 39, p. 229–238.

Hampton, M.A., 1975, Competence of fine-grained debris flows: Journal of Sedimentary Petrology, v. 45, p. 834–844.

Huang, Y.H., 1983, Stability analysis of earth slopes: New York, Van Nostrand Reinhold, 305 p.

Iken, A., 1972, Measurements of water pressure in moulins as part of a movement study of the White Glacier, Axel Heiberg Island, Northwest Territories, Canada: Journal of Glaciology, v. 11, p. 53–58.

Johnson, A.M., 1970, Physical processes in geology: San Francisco, Freeman, Cooper, 577 p.

Mathews, W.H., 1974, Surface profiles of the Laurentide Ice Sheet in its marginal areas: Journal of Glaciology, v. 13, p. 37–43.

Milligan, V., 1976, Geotechnical aspects of glacial tills, *in* Legget, R.F., ed., Glacial till: Ottawa, Royal Society of Canada, p. 269–291.

Muller, F., and Iken, A., 1973, Velocity fluctuations and water regime of arctic valley glaciers: International Association of Scientific Hydrology, v. 95, p. 165–182.

Rodine, J.D., and Johnson, A.M., 1976, The ability of debris, heavily freighted with coarse clastic materials, to flow on gentle slopes: Sedimentology, v. 23, p. 213–234.

Selby, M.J., 1982, Hillslope materials and processes: Oxford, Oxford University Press, 264 p.

Sugden, D.E., 1977, Reconstruction of the morphology, dynamics, and thermal characteristics of the Laurentide ice sheet at its maximum: Arctic and Alpine Research, v. 9, p. 21–47.

———1978, Glacial erosion by the Laurentide ice sheet: Journal of Glaciology, v. 20, p. 367–391.

Carter, L. David, Hamilton, Thomas D., and Galloway, John P., eds., 1989, Late Cenozoic History of the Interior Basins of Alaska and the Yukon: U.S. Geological Survey Circular 1026.

Quaternary Chronology, Yukon and Western District of Mackenzie

By Owen L. Hughes

Local Quaternary chronologies for the Yukon and western District of Mackenzie have been evolving over a period of more than 30 years, beginning with a proposed chronology for the central Yukon by Bostock (1966; loc. 1, fig. 1). This was followed by chronologies for the Ogilvie Mountains (Vernon and Hughes, 1966; loc. 2, fig. 1) the northeastern St. Elias Mountains of the southwestern Yukon (Denton and Stuiver, 1967; loc. 3, fig. 1), and the Snag-Klutlan area of the western Yukon (Rampton, 1971; loc. 4, fig. 1). A chronology of late Quaternary fluctuations of the western margin of the Laurentide ice sheet (loc. 5, fig. 1) and consequent major drainage changes in the area west of the ice sheet was offered by Hughes (1972) and modified by Hughes and others (1981). A chronology for the Yukon Coastal Plain was developed by Rampton (1982; loc. 6, fig. 1).

The first attempt to correlate all of these local chronologies and to relate them to certain of the local chronologies of Alaska and more easterly areas of northern Canada arose from a workshop meeting held in Calgary in 1984 (Hughes, 1986, table 6–1, p. 16; Heginbottom and Vincent, 1986, correlation chart, p. 58). Those correlations are mostly unchanged (fig. 2), but substantial new data of diverse kinds serve to support parts of the correlation that were then speculative.

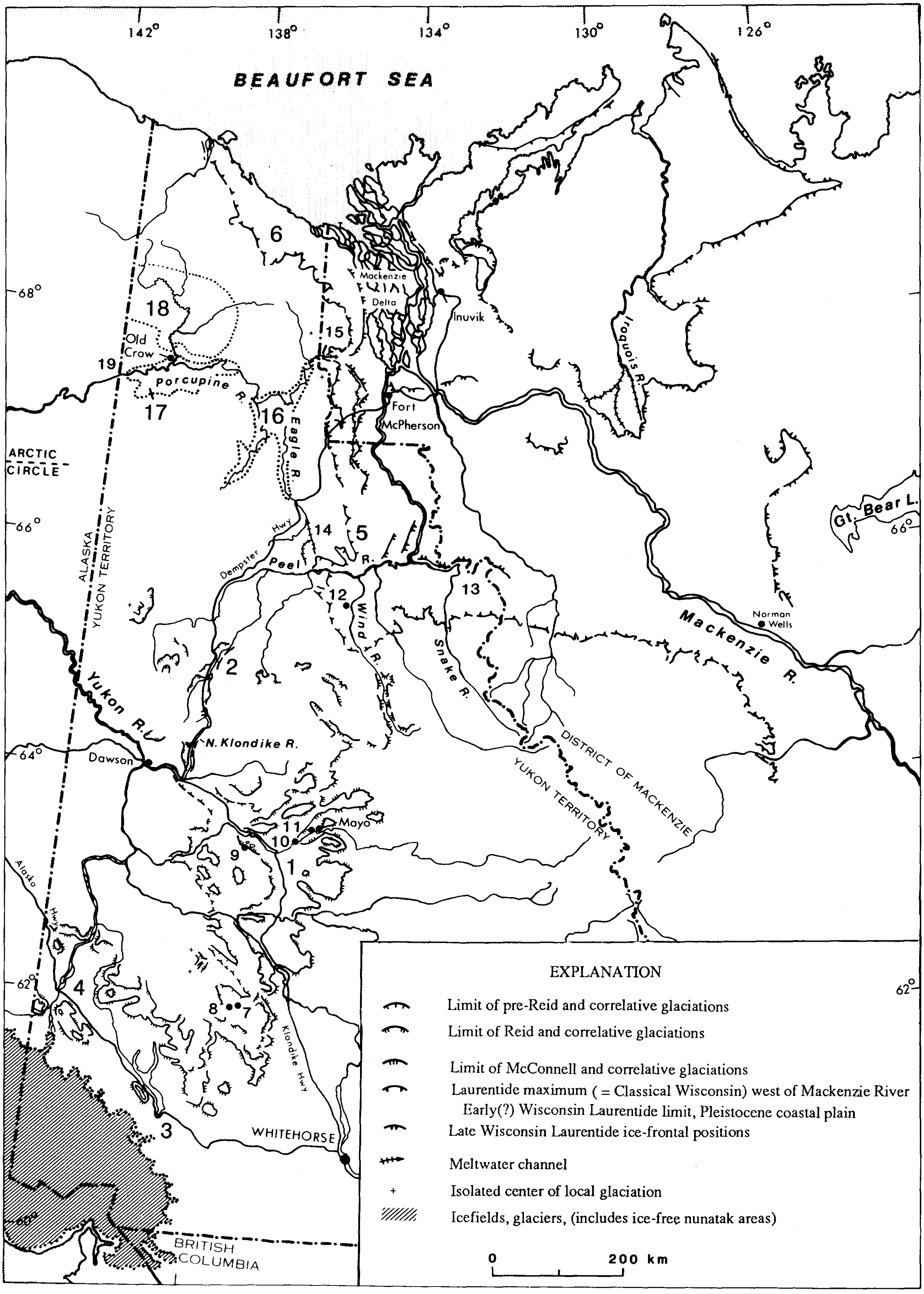
Central and southern Yukon

Of the fourfold glacial sequence of the Cordilleran ice sheet of the central Yukon (Bostock, 1966), the oldest event, the Nansen glaciation, was inferred from till at Nansen Creek (loc. 7, fig. 1). The next younger event, the Klaza glaciation, was inferred from a major glacial diversion of Klaza River (loc. 8, fig. 1). The succeeding Reid and McConnell glaciations were based on well-defined terminal moraines in the valley of the Stewart River (locs. 9 and 10, fig. 1). The Mirror Creek and McCauley glaciations of the western Yukon (Rampton, 1971) are based on terminal moraine systems, as are the glaciations of the Ogilvie Mountains (Vernon and Hughes, 1966). Only the sequence for the southwestern Yukon (Denton and Stuiver, 1967) is based entirely on rock-stratigraphic units. The use of morphostratigraphic units in some areas, rock-stratigraphic units in another, and mixtures of the two in one other area greatly complicates correlations between areas. In addition, within any one area there will likely always be some residual uncertainty as to correlations between glacial deposits seen in section and morphologically defined glaciations, or between stratigraphically defined glaciations and related moraine systems. With this cautionary note, recent developments in the chronology for the central Yukon and correlations with bordering areas are discussed below.

Correlation of the Reid glaciation with the Delta glaciation of the Alaska Range is now strongly supported by recognition that tephra occurring above outwash of both Reid and Delta age is the Sheep Creek tephra (Hamilton and Bischoff, 1984; tephra identified by J.A. Westgate, University of Toronto, oral commun., 1986). Bones from immediately below the tephra at Canyon Creek, Alaska, have yielded uranium-series ages of about 75 to 80 ka (Hamilton and Bischoff, 1984), providing a minimum age for the Delta and Reid glaciations. This age, if valid, admits the possibility that the Reid and Delta glaciations are of early Wisconsin age (oxygen-isotope stage 5b or 5d). The intensity of soil development in the Diversion Creek paleosol developed on Reid drift (Tarnocai, this volume) suggests, however, that the soil is of Sangamon age and the drift of Illinoian age.

Soil developed on moraines of "intermediate" age in the Ogilvie Mountains is essentially the same as the Diversion Creek paleosol developed on drift of Reid age (Tarnocai and others, 1985; Smith and others, 1986). This supports a correlation made originally on the basis of comparable preservation of glacial landforms. The pre-Illinoian Wounded Moose paleosols (Tarnocai, this volume) similarly support broad correlation between pre-Reid deposits of the Cordilleran ice sheet and deposits of "older glaciations" in the Ogilvie Mountains.

Former ambiguity in the correlation between the Wisconsin history of the Cordilleran ice sheet and that of the piedmont glaciers of southwestern Yukon has been resolved by a finite accelerator radiocarbon date from beneath till of McConnell age at the Mayo Indian Village section (loc. 11, fig. 1). In the absence of a finite date it was possible to assume that the Cordilleran ice sheet had advanced to near its maximum position perhaps 40 to 50 ka and remained there until retreat began in late Wisconsin time, whereas piedmont glaciers of southwestern Yukon attained maximum positions after 29 ka, following an interstade that began before 30 ka. The new date from beneath


BEAUFORT SEA
142°
138°
134°
130°
126°
68°
66°
64°
62°
ARCTIC CIRCLE
ALASKA
YUKON TERRITORY
Mackenzie Delta
Inuvik
Old Crow
Porcupine R.
Eagle R.
Fort McPherson
Iroquois R.
Gt. Bear L.
Dempster Hwy
Peel R.
Wind R.
Snake R.
Mackenzie R.
Norman Wells
Yukon R.
N. Klondike R.
Dawson
DISTRICT OF MACKENZIE
YUKON TERRITORY
Mayo
Alaska Hwy
Klondike Hwy
WHITEHORSE
BRITISH COLUMBIA
1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
16
17
18
19
EXPLANATION
Limit of pre-Reid and correlative glaciations
Limit of Reid and correlative glaciations
Limit of McConnell and correlative glaciations
Laurentide maximum (= Classical Wisconsin) west of Mackenzie River
Early(?) Wisconsin Laurentide limit, Pleistocene coastal plain
Late Wisconsin Laurentide ice-frontal positions
Meltwater channel
Isolated center of local glaciation
Icefields, glaciers, (includes ice-free nunatak areas)
0
200 km

till of McConnell age (29,640±260, TO 292) suggests contemporaneity of advances in the two regions.

The Laurentide ice sheet and northern Yukon basins

Two Laurentide glaciations, the Deception glaciation of probable early Wisconsin age and the late Wisconsin Hungry Creek glaciation, have been recognized in the western District of Mackenzie from deposits at the Hungry Creek section (Hughes and others, 1981; loc. 12, fig. 1), and along Snake River (Catto, 1986; loc. 13, fig. 1). Although the Deception glaciation could be as old as Illinoian and therefore correlative with the Reid glaciation, it may be only a stade of the Hungry Creek glaciation. During the Hungry Creek glaciation, the most extensive glaciation in the northwestern sector of the Laurentide ice sheet, the Peel River was diverted northward into the Porcupine River drainage basin via the Eagle River discharge channel (loc. 14, fig. 1). The Porcupine River, which formerly drained eastward through the Richardson Mountains via McDougall Pass (loc. 15, fig. 1), was blocked by a tongue of Laurentide ice that extended into the pass. The Bell, Bluefish, and Old Crow Basins (locs. 16 to 18, fig. 1) were inundated beneath interconnected glacial lakes that found a westward outlet to the Yukon River via the present-day Ramparts of the Porcupine River (loc. 19, fig. 1). Cyclic sediments in the Bell Basin suggest that there may have been many jokulhlaup-caused lowerings of the interconnected lakes through McDougall Pass before westward drainage was permanently established.

The linkage between the Hungry Creek glaciation, diversion of the Peel River, reversal of the Porcupine River, and deposition of glacial lake sediments in the interconnected basins has recently been elucidated in two ways. First, the Eagle River discharge channel by which the Peel River was diverted northward is deeply incised into the Ford Lake Shale of Visean age. Anomalously large quantities of Visean palynomorphs appear abruptly in pollen profiles from all three basins, apparently providing a marker horizon that is correlative with initiation of the channel (Walde, 1985; C.E. Schweger, University of Alberta, unpub. data; J. Utting, Geological Survey of Canada, unpub. data). Second, radiocarbon dates at hand by 1984 from beneath Hungry Creek Till and from beneath glaciolacustrine sediments of Bluefish and Old Crow Basins were in agreement in placing the culmination of the Hungry Creek glaciation and the initiation of the glaciolacustrine stage at about 30 ka. However, dates from beneath glaciolacustrine sediments in the Bell Basin were all "greater than," suggesting an anomalous history for the Bell Basin. Recently acquired finite dates from beneath and within glaciolacustrine sediments of the southeast arm of the Bell Basin indicate a late Wisconsin history similar to that of the Bluefish and Old Crow Basins. Unresolved dating anomalies remain for the southwestern arm of the Bell Basin.

Retreat following the Hungry Creek glaciation had begun by 16.4 ka (GSC–2690; Ritchie, 1982, table 15; Hughes and others, 1981, p. 358). On the Yukon Coastal Plain, where the last glaciation has been termed the Buckland glaciation by Rampton, there had been considerable retreat by 22 ka (Rampton, 1982, p. 25). Three late Wisconsin readvances, or perhaps significant stillstands during general retreat, have been inferred from moraines and other ice marginal features: the Sabine phase (Rampton, 1982, p. 45) and the Tutsieta Lake and Kelly Lake phases (Hughes, 1987). The Sabine phase is younger than 22 ka but is otherwise undated. The Tutsieta phase culminated 13 to 14 ka and the Kelly lake phase before 10.6 ka.

S.R. Morison (Department of Indian and Northern Affairs), J.V. Matthews, Jr. (Geological Survey of Canada), C.E. Schweger (University of Alberta), C. Tarnocai (Land Resources Research Institute), and C.A.S. Smith (Land Resources Research Institute), have contributed much valuable discussion of the Quaternary stratigraphic history. A. Duk-Rodkin (Geological Survey of Canada) assisted in the preparation of figure 1 and S.R. Morison contributed additions and changes to figure 2.

Figure 1. Glacial limits of Cordilleran and Laurentide ice sheets, Yukon and Northwest Territories. General localities cited in the text are indicated by large numerals, and specific sites and features by smaller numerals. Dotted lines around localities 16–18 indicate Bell, Bluefish, and Old Crow Basins, respectively.

REFERENCES CITED

Bostock, H.S., 1966, Notes on glaciation in central Yukon Territory: Geological Survey of Canada Paper 65–36.

Catto, N.R., 1986, Quaternary sedimentology and stratigraphy, Peel Plateau and Richardson Mountains, Yukon and N.W.T.: Edmonton, University of Alberta, Ph.D. dissertation.

Denton, G.H., and Stuiver, M., 1967, Late Pleistocene glacial stratigraphy and chronology, northeastern St. Elias Mountains, Canada: Geological Society of America Bulletin, v. 78, p. 485–510.

Hamilton, T.D., and Bischoff, J.L., 1984, Uranium series dating of fossil bones from the Canyon Creek locality in central Alaska: U.S. Geological Survey Circular 939, p. 26–29.

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, Calgary, April 1984: Geological Survey of Canada Open File Report 1237, 60 p.

Hughes, O.L., 1972, Surficial geology of northern Yukon Territory and northeastern District of Mackenzie, N.W.T.: Geological Survey of Canada Paper 64–36.

———1986, Quaternary chronology of Yukon Territory and western District of Mackenzie, *in* 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, Calgary, April 1984: Geological Survey of Canada Open File Report 1237, p. 13–17.

———1987, Late Wisconsinan Laurentide glacial limits of northern Canada: The Tutsieta Lake and Kelly Lake phases: Geological Survey of Canada Paper 85–25.

Hughes, O.L., Harington, C.R., Janssens, J.A., Matthews, J.V., Jr., Morlan, R.E., Rutter, N.W., and Schweger, C.E., 1981, Upper Pleistocene stratigraphy, paleoecology, and archeology of the northern Yukon interior, eastern Beringia, I. Bonnet Plume Basin: Arctic, v. 34, p. 329–365.

Rampton, V.N., 1971, Late Pleistocene glaciations of the Snag Klutlan area, Yukon Territory, Canada: Arctic, v. 24, p. 277–300.

———1982, Quaternary geology of the Yukon Coastal Plain: Geological Survey of Canada Bulletin 317, 49 p.

Ritchie, J.C., 1982, The modern and Late-Quaternary vegetation of the Doll Creek area, northern Yukon, Canada: The New Phytologist, v. 90, p. 563–603.

Smith, C.A.S., Tarnocai, C., and Hughes, O.L., 1986, Pedological

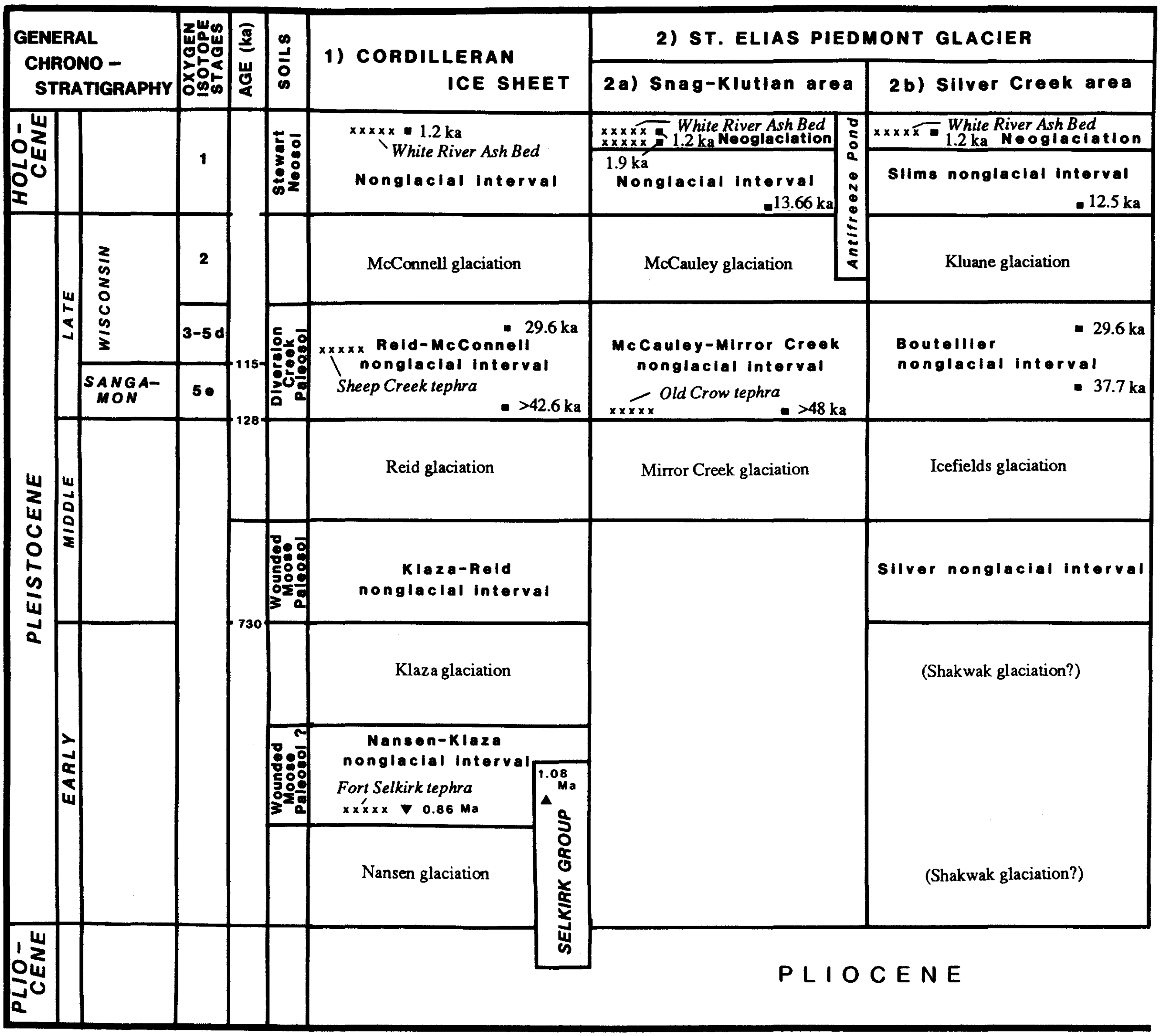


Figure 2. Correlation chart for Yukon and western District of Mackenzie. Modified from Heginbottom and Vincent (1986, correlation chart, p. 58; and Hughes, 1986, table 6–1, p. 16). Squares, radiocarbon ages; diamond, uranium-series age; triangle, fission-track age; inverted triangle, K-Ar age.

investigations of Pleistocene glacial drift surfaces in the central Yukon: Géographie Physique et Quaternaire, v. 40, p. 29–37.

Tarnocai, C., Smith, C.A.S., and Hughes, O.L., 1985, Soil development on Quaternary deposits of various ages in the central Yukon Territory: Geological Survey of Canada Paper 85–1A, p. 229–238.

Vernon, P., and Hughes, O.L., 1966, Surficial geology of Dawson, Larsen Creek and Nash Creek map areas, Yukon Territory: Geological Survey of Canada Bulletin 136, 25 p.

Walde, K., 1985, Pollen analysis and taphonomy of locality 15 alluvial sediments, Old Crow Basin, Yukon: Edmonton, University of Alberta, M.Sc. thesis.

Geological Survey of Canada Contribution 40888

Carter, L. David, Hamilton, Thomas D., and Galloway, John P., eds., 1989, Late Cenozoic History of the Interior Basins of Alaska and the Yukon: U.S. Geological Survey Circular 1026.

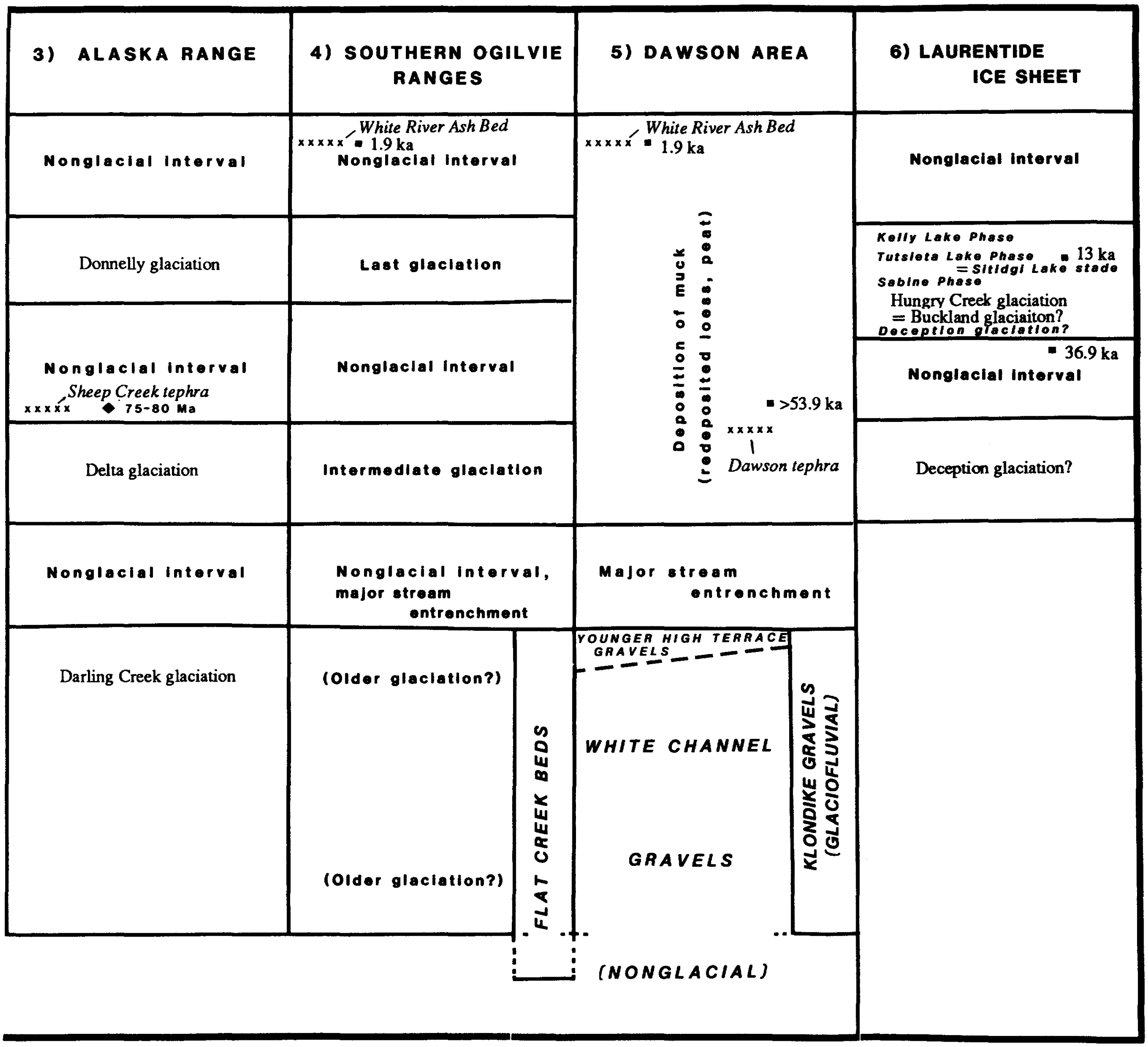


Figure 2. Continued.

The Old Crow and Bluefish Basins, Northern Yukon: Development of the Quaternary History

By Charles E. Schweger

The Old Crow and Bluefish Basins of northern Yukon contain thick sequences of unconsolidated fossil-rich sediments. Fieldwork carried out during the 1960's by O.L. Hughes (Geological Survey of Canada) led to the recognition of a long, "Late Tertiary into Quaternary time" record exposed along river bluffs in these basins. The accelerated research activity since that time is reviewed here by highlighting key developments that have led to our present understanding of the Quaternary history and chronology of this area. Much of these data are unpublished or in preparation, and some conclusions should be more properly called hypotheses.

Hughes (1969, 1972) described, for the Bluefish Basin locality HH228, a sequence of two glaciolacustrine units separated by alluvial sand and silt (fig. 1). The upper lake unit was dated as forming between 12.5 and 32.4 ka, assuring correlation with the late Wisconsin glaciation when Keewatin ice dammed both the Peel River drainage and McDougall Pass, which is the former course of the Porcupine River through the Richardson Mountains. It was assumed that a similar glacial event during or before early Wisconsin time resulted in deposition of the lower lake unit.

The discovery in 1966 of permineralized bone artifacts in the Old Crow Basin resulted in field surveys during 1967 and 1970 by T.D. Hamilton (University of Alaska), who mapped shorelines, reported on stratigraphy, and evaluated features and processes of archeological importance in a series of unpublished maps and reports. Radiocarbon dating placed the ages of the bone artifacts between 27 and 35 ka, thus bringing worldwide attention to the archeological potential of the region (Irving and Harington, 1973). Two multidisciplinary research groups took to the field during the summer of 1975. The Yukon Refugium Project, organized by R.E. Morlan (Archaeological Survey of Canada), has hypothesized a human presence in the Old Crow Basin to 35 ka (Morlan, 1986; but also see Nelson and others, 1986), whereas the Northern Yukon Refugium program proposed a human presence to at least 150 ka (Irving and others, 1986). From the outset both groups relied heavily on the pioneering paleontological research of C.R. Harington, who had systematically searched and collected in the Old Crow and Bluefish Basins since 1966 (Harington, 1978). The Yukon Refugium Project, the work of which is stressed here, placed emphasis on detailed stratigraphy, recovery of in-place faunal remains, nested subsampling for paleoecological reconstruction, and geochronology.

The Old Crow tephra (Westgate and others, 1981) was discovered in 1968 by Hughes at locality HH228 on the Porcupine River in the Bluefish Basin, and was collected and described by J.A. Westgate in 1976 (Westgate, 1982). Also in 1968, paleomagnetic studies indicated reversed geomagnetic polarity of most of the lower lake sediment, unit 3, at HH228 (Pearce and others, 1982). Because of subsequent dating of the Old Crow tephra and the prominent unconformity between units 3 and 4, unit 3 is thought to have been deposited before 730 ka, which is the time-boundary between the Brunhes Normal-Polarity and Matuyama Reversed-Polarity Chrons. Reversed geomagnetic polarity for unit 3 has since been verified at other locations. Unit 3 is known to be rich in pine as well as spruce pollen (Lichti-Federovich, 1974), unlike the modern pollen rain and certainly unlike the pollen found in unit 5, the upper glaciolacustrine unit. This has led to the hypothesis that the lower lake unit is not of glacial origin but probably related to tectonic activity along faults, a conclusion supported by the detailed bedrock mapping of Norris (1981). Subsequent fieldwork in the Bluefish Basin indicated that unit 3 may in fact represent three separate lake events.

Fieldwork during 1977 at Old Crow River locality CRH 15 (fig. 1) led to the recognition there of the Old Crow tephra, disconformity A, and a definition of the Old Crow Basin composite stratigraphy (Morlan and Matthews, 1978). Over the next four years Morlan, together with colleagues and graduate students, concentrated fieldwork along the lower Old Crow River. Excavations at a number of river bluffs resulted in a large collection of vertebrate remains with well-documented stratigraphic context. These provided for the development of a taphonomic approach to Old Crow paleontology and archeology (Morlan, 1980; Morlan and Matthews, 1983). Thesis work on a wide variety of topics was carried out by University of Alberta students: Janssen, fossil bryophytes (1981); McCourt, palynology (1982); Walde, palynology (1985); Bombin, phytoliths (1984); and Hedlund, paleomagnetics (1986). Screening of large-volume sediment samples was undertaken in order to develop a paleoecological record based on plant and insect remains and augmented by fossil pollen studies. In 1980, fieldwork was expanded to include upper Porcupine River localities in the Bell Basin as well as sites of stratigraphic importance along the Porcupine River in Alaska (Thorson and Dixon, 1983).

Tephra studies have come to play a dominant role in the developing geochronology. In 1981 the Little Timber tephra was discovered at locality CRH 94, Old Crow River. It has been characterized by Westgate and dated as forming before 1.2 Ma, depending upon the annealing factor. The Old Crow tephra was dated by fission track as less than 120 ka (Naesser and others, 1982).

An Old Crow River survey in 1983 by J.V. Matthews, Jr. (Geological Survey of Canada) and the author resulted in discovery of the Surprise Creek tephra at locality CRH 47 (fig. 1), while the same year paleobotanist, Ruth Stockey (University of Alberta) initiated her research on the HH228, unit 1, cone flora. Extant species *Pinus monticola*, *Pinus contorta*, and *Picea glauca* have been identified, and the extinct forms *Larix minuta* and *Picea* sp. suggest a Pliocene age. This age placement is supported by paleosol

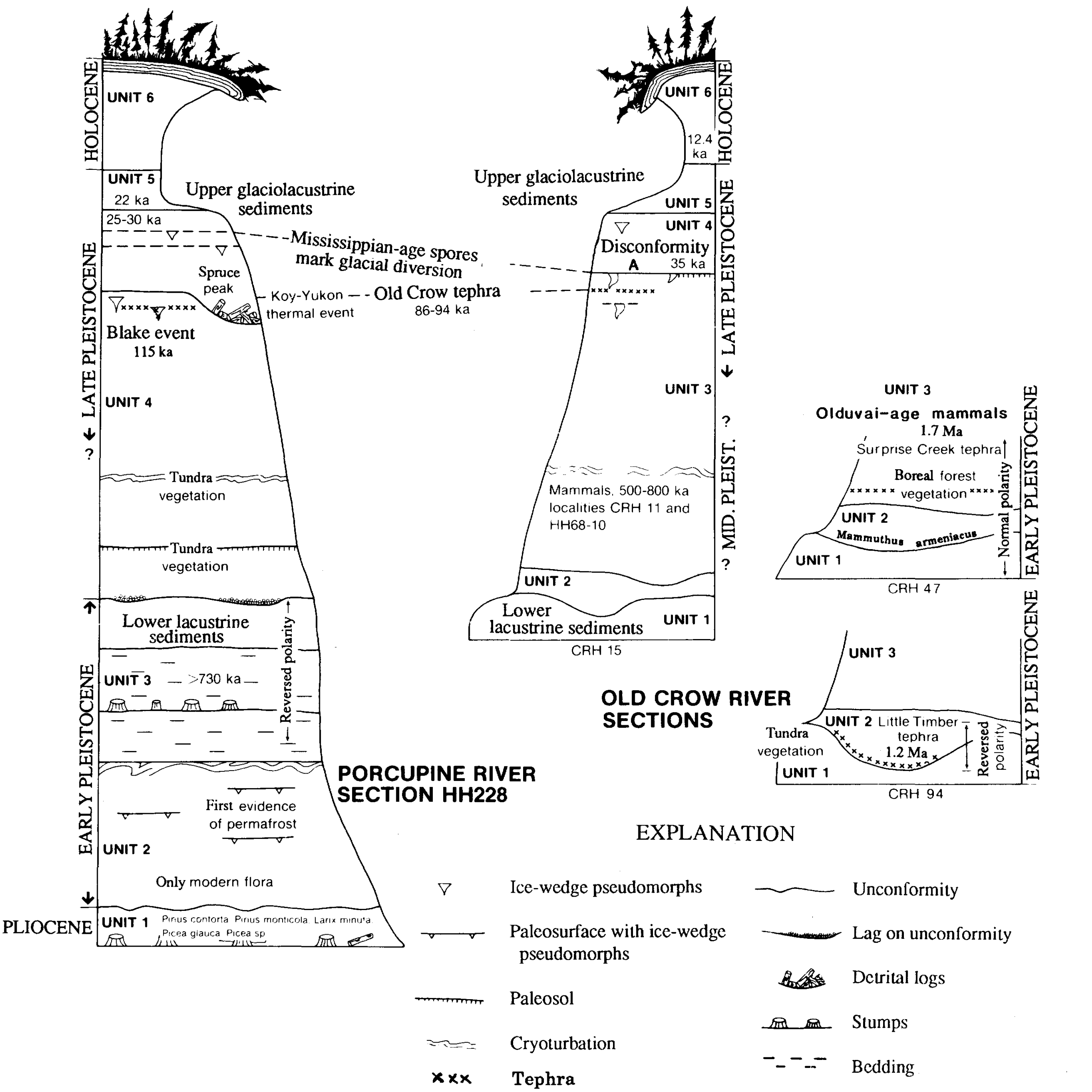


Figure 1. Summary of Quaternary and uppermost Tertiary stratigraphy and geochronology for Old Crow and Bluefish Basins, northern Yukon. Old Crow tephra is an informal unit of Westgate and others (1981) and Westgate (1982). "Blake event" is the Blake Reversed-Polarity Subchron of the Brunhes Normal-Polarity Chron; "Olduvai" is the Olduvai Normal-Polarity Subchron of the Matuyama Reversed-Polarity Chron.

studies (Tarnocai, this volume) and pollen evidence (Schweger) from Bluefish River and Burnt Hill Bluffs (see Tarnocai, this volume).

Walde (1985) demonstrated the paleohydrological significance of reworked palynomorphs from the Upper Mississippian Ford Lake Shale (Norris, 1984). These spores were dispersed throughout the northern Yukon Basins when ice blocked the Peel River and diverted water through the upper Eagle River valley, eroding overflow channels in the shale. Mississippian-age spores are found immediately above disconformity A, locality CRH 15, which has been recently dated at about 35 ka (R.E. Morlan, Archaeological Survey of Canada, oral commun., 1987). This new date for disconformity A plus other recent accelerator mass spectrometer radiocarbon dates in the Bell Basin have considerable regional significance, because they imply that Keewatin ice had approached its maximum western extent earlier than previously thought (Hughes and others, 1981). Glaciolacustrine unit 5 was deposited in a deep-water lake after 25 ka (^{14}C age on tusk, Cadzow Bluff), through 21 ka (^{14}C age on fine organics, Bluefish River section), and until just before 12.5 ka (^{14}C age on bison bone, locality CRH 11) by which time the lake had drained through the Ramparts of the Porcupine River to the west. The interval between the first blockage of the eastward drainage of Porcupine River and development of a deep glacial lake may have been characterized by lowerings of lake level in the basins via jökulhlaups through McDougall Pass.

Stratigraphy, paleontology, and geochronology of locality HH228 has received considerable attention during and since 1983. Westgate and others (1985) documented the 115-ka reversal event marking the Blake Reversed-Polarity Subchron (of the Brunhes Normal-Polarity Chron) below the Old Crow tephra, which has recently been dated by thermoluminescence as forming 86±8 ka (Wintle and Westgate, 1986). The stratigraphy and paleontology of upper unit 4 provides evidence of the Koy-Yukon thermal event, a period 50–60 ka in which climate in eastern Beringia was warmer than the present climate (Schweger and Matthews, 1985).

Fieldwork during 1985 was centered on Old Crow River localities CRH 94 and CRH 47 (fig. 1), which are Little Timber and Surprise Creek tephra localities, respectively. Nearly five tons of sediment from near the Surprise Creek tephra at locality CRH 47 was screened for small mammals. C.A. Repenning (U.S. Geological Survey, oral commun., 1987) places this fauna near the base of the Pleistocene, within the Olduvai Normal-Polarity Subchron (of the Matuyama Reversed-Polarity Chron), which is dated at 1.7 Ma. This is corroborated by the fact that Surprise Creek tephra fell at a time of normal geomagnetic polarity when boreal forest was present in the Old Crow Basin. The stratigraphically lower Little Timber tephra was deposited on tundra vegetation at a time of reversed polarity. Considerable work remains to be done on these two important localities.

At the present time, research continues on several fronts: small mammals, pollen, seeds, cones, insects, radiocarbon chronology, stratigraphy, paleosols, and tephrochronology. Over a decade after the Yukon Refugium Project was initiated, it has moved into a stage of synthesis, where large amounts of diverse information are being integrated into a comprehensive history of the region. The history of the Old Crow and Bluefish Basins is of regional significance because it links events in the District of Mackenzie to the east with those in interior Alaska to the west. These basins are now known to have the most complete Pliocene and early Pleistocene record in northwestern North America. The middle Pleistocene is poorly known, although R.E. Morlan's small-mammal record from Old Crow River locality CRH 11 has been placed between 500 and 800 ka by C.A. Repenning (U.S. Geological Survey, oral commun., 1987) and sediments of similar age may be exposed at other sections (such as HH68–10). The history of the late Pleistocene is well established and dated and points up the fact that at no time during the Pleistocene did Keewatin ice extend as far westward as during the interval 35–12.5 ka. The research accomplished to date clearly points up the importance of joint multidisciplinary research with a high degree of information sharing and a long-term commitment to the region.

REFERENCES CITED

Bombin, M., 1984, On phytoliths, late Quaternary ecology of Beringia, and information evolutionary theory: Edmonton, University of Alberta, Ph.D. dissertation.

Harington, C.R., 1978, Quaternary vertebrate faunas of Canada and Alaska and their suggested chronological sequence: Ottawa, National Museums of Canada Syllogeus 15, 105 p.

Hedlund, M.A.H., 1986, A Pleistocene high-latitude geomagnetic study: Edmonton, University of Alberta, M.A. thesis.

Hughes, O.L., 1969, Pleistocene stratigraphy, Porcupine and Old Crow Rivers, Yukon Territory: Geological Survey of Canada Paper 69–1, p. 209–211.

———1972, Surficial geology of northern Yukon Territory and northwestern District of Mackenzie, Northwest Territories: Geological Survey of Canada Paper 69–36.

Hughes, O.L., Harrington, C.R., Janssens, J.A., Matthews, J.V., Jr., Morlan, R.E., Rutter, N.W., and Schweger, C.E., 1981, Upper Pleistocene stratigraphy, paleoecology, and archaeology of the northern Yukon Interior, eastern Beringia, 1, Bonnet Plume Basin: Arctic, v. 34, p. 329–365.

Irving, W.N., and Harington, C.R., 1973, Upper Pleistocene radiocarbon-dated artifacts from the northern Yukon: Science, v. 179, p. 335–340.

Irving, W.N., Jopling, A.V., and Beebe, B.F., 1986, Indicators of Pre-Sangamon humans near Old Crow, Yukon, Canada, *in* Bryan, A.L., ed., New evidence for the peopling of the Americas: Orono, University of Maine, p. 49–63.

Janssen, J.A.P., 1981, Subfossil bryophytes in eastern Beringia: Their paleoenvironmental and phytogeographic significance: Edmonton, University of Alberta, Ph.D. dissertation.

Lichti-Federovich, S., 1974, Palynology of two sections of Late Quaternary sediments from the Porcupine River, Yukon Territory: Geological Survey of Canada Paper 74–23, 6 p.

McCourt, G.H., 1982, Quaternary palynology of the Bluefish

Basin, northern Yukon Territory: Edmonton, University of Alberta, M.S. thesis.

Morlan, R.E., 1980, Taphonomy and archaeology in the upper Pleistocene of the northern Yukon Territory: A glimpse of the peopling of the New World: National Museum of Man, Mercury Series, Archaeological Survey of Canada, Paper 94, 380 p.

———1986, Pleistocene archaeology in Old Crow Basin: A critical reappraisal, *in* Bryan, A.L., ed., New evidence for the peopling of the Americas: Orono, University of Maine, p. 27–48.

Morlan, R.E., and Matthews, J.V., 1978, New dates for early man: Geos, winter 1978, p. 2–5.

———1983, Taphonomy and paleoecology of fossil insect assemblages from Old Crow River (CRH–15) northern Yukon Territory, Canada: Géographie Physique et Quaternaire, v. 37, p. 147–157.

Naeser, N.D., Westgate, J.A., Hughes, O.L., and Péwé, T.L, 1982, Fission-track ages of late Cenozoic distal tephra beds in the Yukon Territory and Alaska: Canadian Journal of Earth Sciences, v. 19, p. 2167–2178.

Nelson, D.E., Morlan, R.E., Vogel, J.S., Southon, J.R., Harington, C.R., 1986, New dates on northern Yukon artifacts: Holocene not upper Pleistocene: Science, v. 232, p. 749–751.

Norris, D.K., 1981, Geology; Old Crow, Yukon Territory: Geological Survey of Canada Map 1518A.

———1984, Geology of the northern Yukon and northwestern District of Mackenzie: Geological Survey of Canada, Map 1581A, scale 1:500,000.

Pearce, G.W., Westgate, J.A., and Robertson, S., 1982, Magnetic reversal history of Pleistocene sediments at Old Crow, northwestern Yukon Territory: Canadian Journal of Earth Sciences, v. 19, p. 919–929.

Schweger, C.E., and Matthews, J.V., 1985, Early and middle Wisconsian environments of eastern Beringia: Stratigraphic and paleoecological implications of the Old Crow tephra: Géographie Physique et Quaternaire, v. 39, p. 275–290.

Thorson, R.M., and Dixon, E.J., Jr., 1983, Alluvial history of the Porcupine River, Alaska: Role of glacial-lake overflow from northwest Canada: Geological Society of America Bulletin, v. 94, p. 576–589.

Walde, K., 1985, Pollen analysis and taphonomy of locality 15 alluvial sediments, Old Crow Basin, Yukon: Edmonton, University of Alberta, M.A. thesis.

Westgate, J.A., 1982, Discovery of a large-magnitude, late Pleistocene volcanic eruption in Alaska: Science, v. 218, p. 789–790.

Westgate, J.A., Matthews, J.V., Jr., and Hamilton, T.D., 1981, Old Crow tephra: A new late Pleistocene stratigraphic marker across Alaska and the Yukon Territory: Geological Society of America Abstracts with Programs, v. 13, p. 579.

Westgate, J.A., Walter, R.C., Pearce, G.W., and Gorton, M.P., 1985, Distribution, stratigraphy, petrochemistry, and paleomagnetism of the late Pleistocene Old Crow tephra in Alaska and the Yukon: Canadian Journal of Earth Sciences, v. 22, p. 893–906.

Wintle, A.G., and Westgate, J.A., 1986, Thermoluminescence age of Old Crow tephra in Alaska: Geology, v. 14, p. 594–597.

Carter, L. David, Hamilton, Thomas D., and Galloway, John P., eds., 1989, Late Cenozoic History of the Interior Basins of Alaska and the Yukon: U.S. Geological Survey Circular 1026.

Climatic Change in Eastern Beringia During Oxygen Isotope Stages 2 and 3: Proposed Thermal Events

By John V. Matthews, Jr., Charles E. Schweger, *and* Owen L. Hughes

Pollen and macrofossil data associated with recently acquired Accelerator Mass Spectrometer (AMS) and standard ^{14}C dates in addition to recent thermoluminescence (TL) dates on the Old Crow and Sheep Creek tephras from Alaska and the Yukon allow us to discern new detail on climate during oxygen-isotope stages 2 and 3 in eastern Beringia (Alaska and the Yukon). At least three thermal "events" appear to have occurred. One of these, the 60 ka Koy-Yukon thermal event, has been discussed previously (Schweger and Matthews, 1985). The other two, the 25- to 27-ka Sixtymile and the 18- to 22-ka Hanging Lake thermal events, are proposed here for the first time. The designation of these two new events conforms to the North American Stratigraphic Code (North American Commission on Stratigraphic Nomenclature, 1983). Like the earlier named Koy-Yukon thermal event (Schweger and Matthews, 1985), they have only regional importance and no formal chronostratigraphic significance.

The Koy-Yukon thermal event

Schweger and Matthews (1985) proposed the name Koy-Yukon thermal event for an interval of warmer-than-present summer climate which occurred after the deposition of both Old Crow tephra and Sheep Creek tephra, but prior to 59 ka. They estimated its age as about 60 ka, the start of isotope stage 3, when the marine record registers significant interstadial warming and when insolation curves indicate approximately 20 percent greater summer insolation than at present in the northern part of the northern hemisphere. The 59 ka minimum age is based on a minimum limiting age on wood from the Eva Formation (Fairbanks area, Péwé, 1975), deposition of which was penecontemporaneous with the interval of deep thawing in the Fairbanks region that Schweger and Matthews (1985) assigned to the Koy-Yukon thermal event. A TL date on sediments associated with the Old Crow tephra implies that it could be as young as 86±8 ka (Wintle and Westgate, 1986). The age of the Sheep Creek tephra is still poorly known, but Hamilton and Bischoff (1984) estimated it to be slightly younger than 80 ka, whereas a recent TL date suggests it is 75±7 ka (T.L. Péwé, Arizona State Univ., oral commun., 1987). If these dates are correct, the Sheep Creek tephra was deposited during isotope stage 4, a conclusion supported by the pollen data from the Ash Bend site in the northern Yukon (C.E. Schweger, unpub. data). The Sheep Creek tephra occurs in the Gold Hill Loess at Fairbanks, which means that it predates the Eva Formation and the interval of thawing associated with the Koy-Yukon thermal event.

As indicated above, several lines of evidence show that the Koy-Yukon thermal event postdates the last interglaciation (stage 5e) and stage 4, yet its fauna and flora clearly suggest that summer climate was warmer than at present. This is best shown by fossils from unit 4c at the Twelvemile Bluff exposure in the northern Yukon. The fluvial and lacustrine deposits within that unit contain plant macroremains of species like *Carex sychnocephala* and *Chenopodium gigantospermum*, neither of which extend as far north today. A similar pattern is exhibited by some of the insect taxa (such as *Bradycellus lecontei* and *Bembidion quadrimaculatum*), and the presence of traces of *Corylus* and *Typha* in the pollen spectra. Furthermore, it is now clear that the extremely high frequencies of spruce shown in Lichti-Federovich's (1973) pollen diagram for the Twelvemile Bluff section occur in unit 4c. Unit 4c cuts out another unit containing a series of large ice-wedge pseudomorphs which probably reflect regional thawing. Higher in the sequence, above the beds indicative of the warming of the Koy-Yukon thermal event, is another series of ice-wedge pseudomorphs, which formed before 37 ka and may also mark regional thawing.

Sixtymile thermal event

Pollen and macrofossils (fig. 1) from an alluvial unit below late Wisconsin (McConnell glaciation) till at the Mayo Indian Village section near the town of Mayo in the central Yukon suggest a period of dry, arctic climate during which trees were absent or existed only as isolated groves in the lowlands. Today the region is well forested. Fossil seeds of *Corispermum hyssopifolium* (bugseed), a plant very likely to have been growing on the floodplain at the time of deposition, are dated by the AMS radiocarbon method at 29.6±0.3 ka (TO–292). Some of the more important macrofossils from the dated level are listed in figure 1. The insect assemblage implies at least a cold, dry local floodplain climate and probably an arctic regional climate. Like some others from Beringia, the insect assemblage includes at least one beetle, *Harpalus amputatus*, which is normally an occupant of grassland areas. Another species, the weevil *Connatichela artemisiae* Anderson (fig.

1) feeds on sage (*Artemisia*), a plant which is well represented in the pollen diagram and probably grew on the flood plain at 29.6 ka.

Spruce percentages rise slightly above the dated level, suggesting slight climatic amelioration. The best evidence for such a trend comes from a site in the Sixtymile District near Dawson City. A spruce stump in growth position from that region has been dated at 26.08±0.3 ka (B–13870) (C.R. Harington, Canada National Museum of Natural Sciences, oral commun., 1986). It shows that climate had warmed enough for growth of spruce in the valley bottoms of the central Yukon, unlike the situation at 29.6 ka.

We believe that these data may signify a period of slightly warmer climate—the Sixtymile thermal event—which occurred between about 25 and 27 ka. As indicated in figure 2 this conclusion appears to be supported by (1) an AMS age on mosses from the northern Yukon Bell Basin that implies that meltwater levels in the Bell Basin were fluctuating between 25 and 26 ka, probably due to changes of the Laurentide ice margin on the east side of the Richardson Mountains; (2) pollen spectra indicative of interstadial conditions at Harding Lake in central Alaska (Nakao and Ager, 1985); and (3) ages in the 25- to 27-ka range on soils and wood in the Alaska Range (Hamilton and Thorson, 1983). The Sixtymile thermal event was followed at about 24 ka by climatic deterioration and renewed glaciation in both Alaska and the Yukon (Hamilton and Thorson, 1983; Hamilton, 1986; Klassen, 1986; fig. 2).

Hanging Lake thermal event

Several sites in eastern Beringia yield circumstantial evidence for an interval of warmer climate between 18 and 22 ka. Cwynar (1982) first recognized this event in his discussion of the pollen sequence at Hanging Lake in the northern Yukon (fig. 3), hence the designation "Hanging Lake thermal event." Climatic warming evidently occurred at about the same time in western Alaska, judging from the pollen sequence at Kaiyak Lake in northwestern Alaska (Anderson, 1985; fig. 3). Other probable indicators of contemporaneous warming include (1) dates and macrofossils from the Rat Pass area near the Yukon-Northwest Territories border that suggest spruce may have been growing near there at about 21.3 ka (Matthews and Hughes, unpub. data); (2) an abrupt peak of Ericaceae pollen and slightly higher percentages of *Picea*, *Alnus*, and *Betula* at a level now dated at 20.8±0.2 ka (GSC–3946) in the pollen record from the northern Bluefish Basin (HH75–24) section

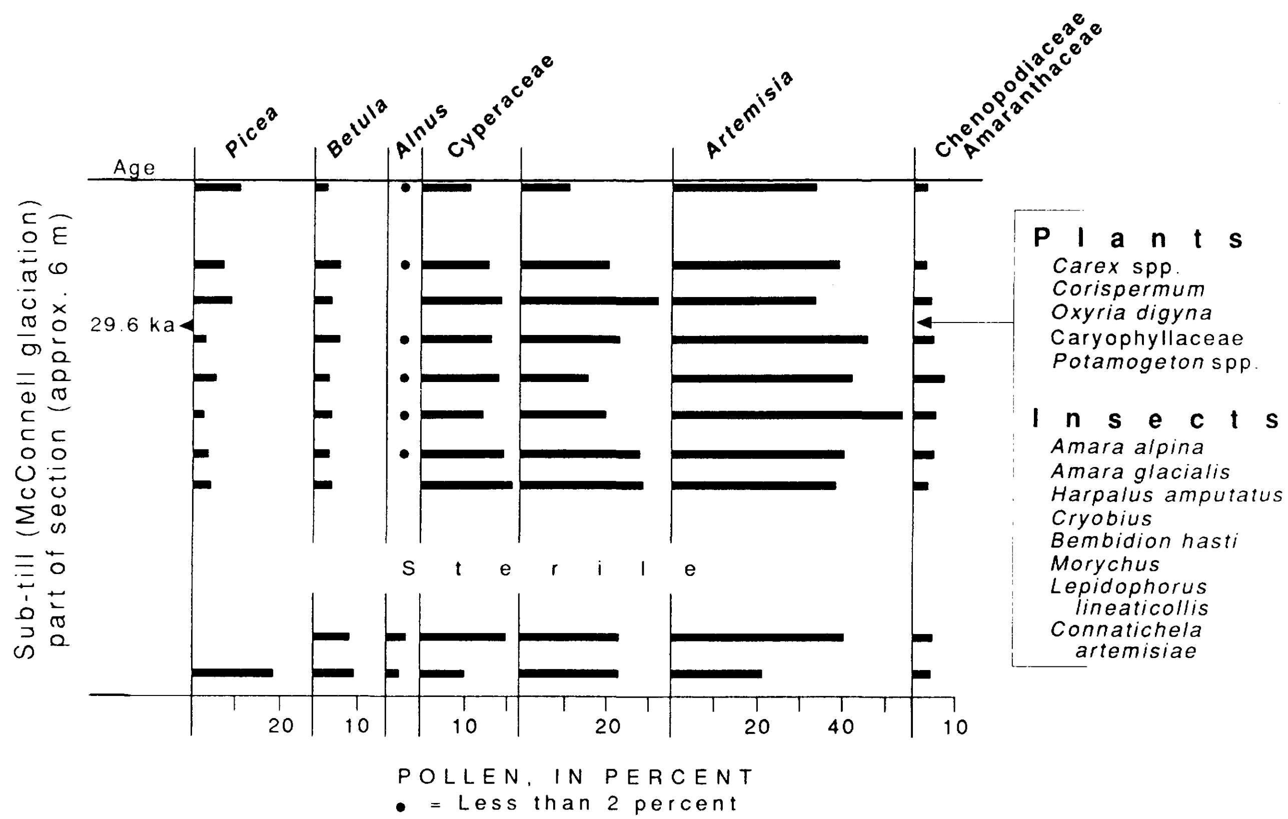


Figure 1. Summary of pollen from base of the Mayo Indian Village section, including insect and plant macrofossils identified from the same level as the 29.6-ka Accelerator Mass Spectrometer ^{14}C date on *Corispermum hyssopifolium* seeds. Identifications: pollen, C.E. Schweger; macrofossils, J.V. Matthews, Jr.

(summarized in fig. 4, which is based on data in McCourt, 1982); (3) presence of paleosols and indicators of cessation of alluviation and stabilized periglacial features in the Wind River and Koyukuk regions of Alaska (fig. 2) (Hamilton, 1986). However, we note that there is also contradictory evidence of warming; for example, T.A. Ager (in Weber and Ager, 1984) found only traces of spruce pollen in a sample from interior Alaska that is associated with the Mount Harper ash, which was deposited about 21 ka.

Discussion

We have attempted to point out several facts that seem to indicate minor climatic oscillations during oxygen-isotope stages 2 and 3. We acknowledge that the evidence is at best circumstantial, but believe that by proposing names for the presumed climatic fluctuations we will stimulate the research and discussion necessary to test the validity of our proposal.

Paleoecologists working in east Beringia have often expressed regret that there was no unequivocal interglacial fauna and flora with which to compare middle Wisconsin faunas and floras. As indicated above, the deposits assigned to the Koy-Yukon thermal event include the expected interglacial biotic elements, yet appear to be much younger than stage 5e. Work now under way at the Birch Creek exposure by Mary Edwards and colleagues (Edwards, this volume) may resolve this dilemma. In our view, the Birch Creek section is the key to clarifying the record of climatic events during both stages 3 and 5 be-

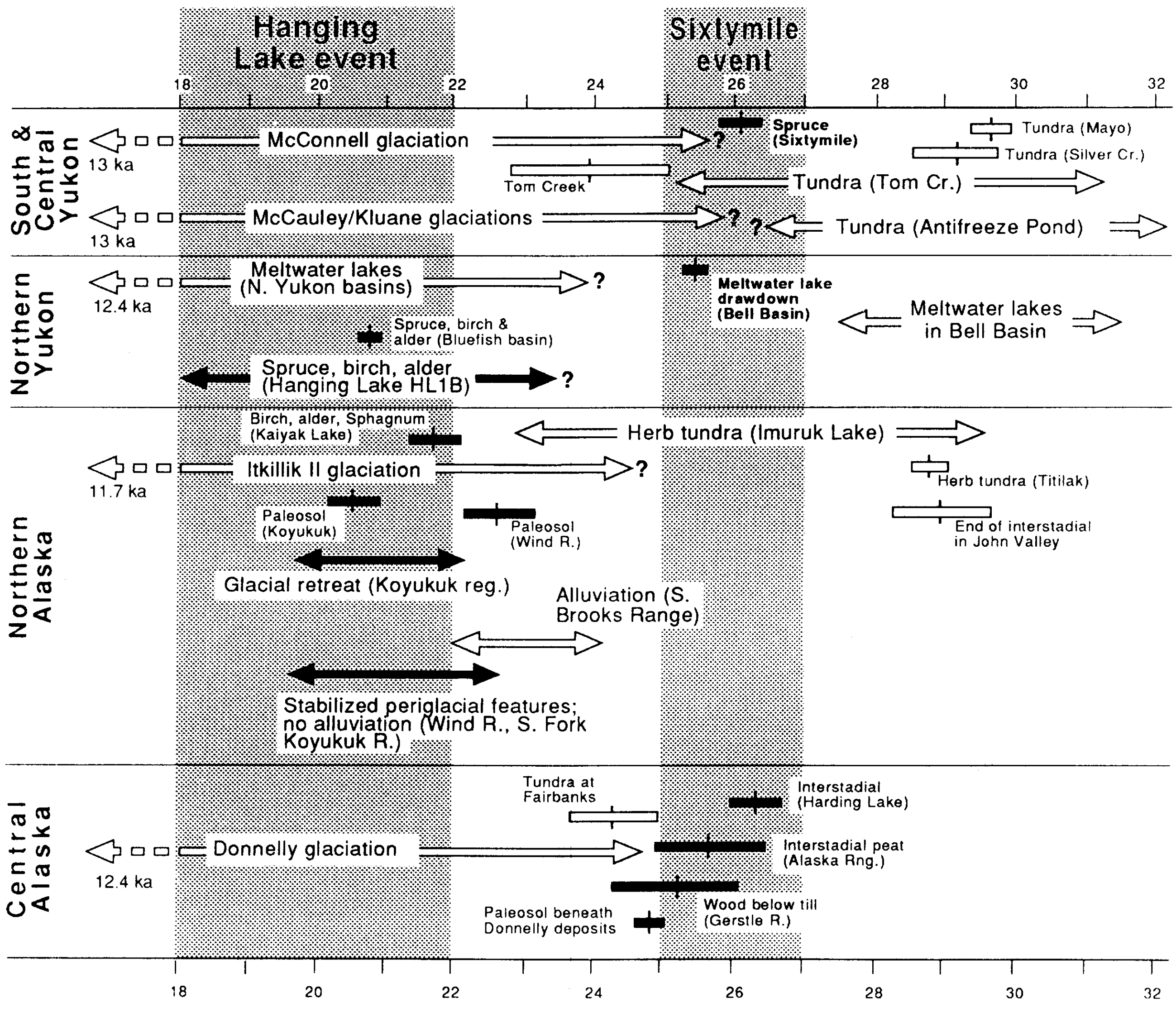


Figure 2. Plot of ^{14}C dates on phenomena thought to imply existence of Sixtymile and Hanging Lake thermal events in eastern Beringia (from various sources). Bars with vertical hash mark represent ^{14}C dates and 1 or 2 σ error values. Arrows signify duration of glaciations or other climatically related phenomena; arrows dashed where approximate and queried where uncertain. Black ^{14}C bars and black arrows signify dates or events that suggest climatic warming.

cause it contains sediments bracketing those stages as well as several paleosols and another undated tephra older than the Old Crow tephra.

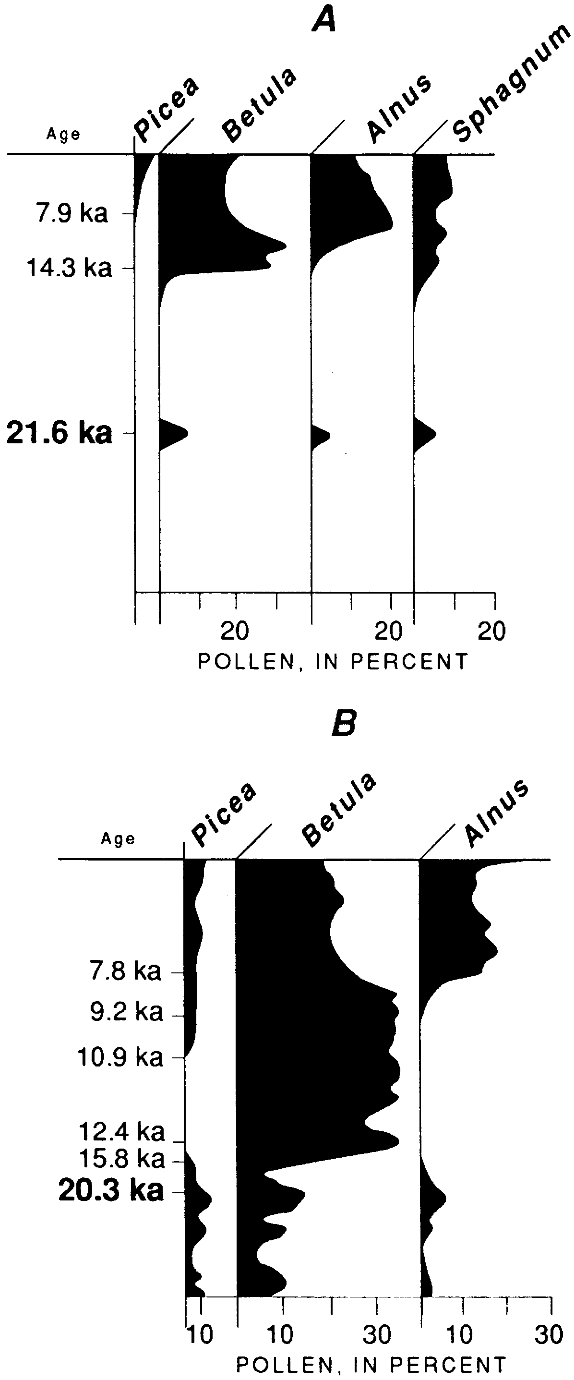


Figure 3. Summary of pollen from (*A*) Kaiyak Lake, northwestern Alaska (modified from Anderson, 1985), and (*B*) Hanging Lake, northern Yukon (modified from Cwynar, 1982), showing pollen fluctuations at around 20 ka that are thought to reflect warming associated with the Hanging Lake thermal event.

REFERENCES CITED

Anderson, P.M., 1985, Late Quaternary vegetational change in the Kotzebue Sound area, Northwestern Alaska: Quaternary Research, v. 24, p. 307–321.

Cwynar, L.C., 1982, A late-Quaternary vegetation history from Hanging Lake, northern Yukon: Ecological Monographs, v. 52, p. 1–24

Hamilton, T.D., 1986, Late Cenozoic glaciation of the Central Brooks Range, *in* Hamilton, T.D., Reed, K.M., and Thorson, R.M., eds., Glaciation in Alaska: The geologic record: Anchorage, Alaska Geological Society, p. 9–49.

Hamilton, T.D., and Bischoff, J.L., 1984, Uranium series dating of fossil bones from the Canyon Creek locality in central Alaska: U.S. Geological Survey Circular 939, p. 26–29.

Hamilton, T.D., and Thorson, R.M., 1983, The Cordilleran ice sheet in Alaska, *in* Porter, S.C., and Wright, H.W., Jr., eds., Late-Quaternary environments of the United States; v. 1, The late Pleistocene: University of Minnesota Press, p. 38–52.

Klassen, R.W., 1986, Tertiary-Pleistocene stratigraphy of the Liard Plain, southeastern Yukon Territory: Geological Sur-

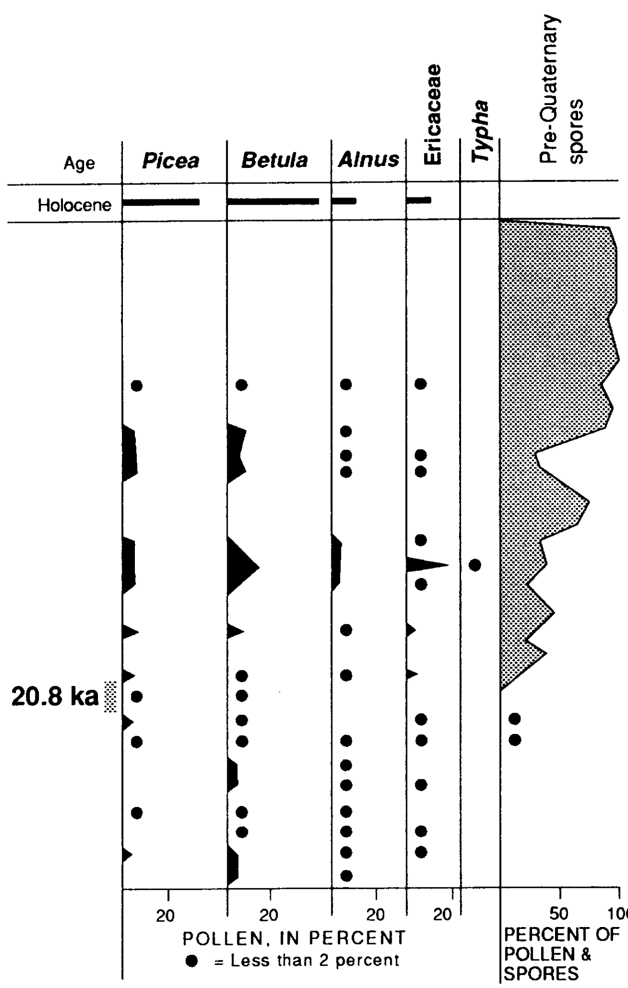


Figure 4. Summary of pollen percentages in upper part of Bluefish Basin (loc. HH75–24) that may signify slightly warmer climate shortly after 20.8 ka—the Hanging Lake thermal event. From McCourt (1982) and J.V. Matthews, Jr. (unpub. data).

vey of Canada Paper 86–17, 16 p.
Lichti-Federovich, S., 1973, Palynology of six sections of late Quaternary sediments from the Old Crow River Yukon Territory: Canadian Journal of Biology, v. 51, p. 553–564.
McCourt, G.H., 1982, Quaternary palynology of the Bluefish Basin, northern Yukon Territory: Edmonton, University of Alberta, M.S. thesis, 165 p.
Nakao, K., and Ager, T.A., 1985, Climatic changes and vegetational history of the Interior Alaska inferred from drilled cores at Harding Lake, *in* Nakao, K., ed., Hydrological regime and climatic changes in the Arctic Circle; Report of the Alaskan Paleohydrology Research Project—1982/83: Japan, Hokkaido University, Laboratory of Hydrology, Department of Geophysics, Faculty of Science, p. 43–52.
Péwé, T.L., 1975, Quaternary stratigraphic nomenclature in unglaciated central Alaska: U.S. Geological Survey Professional Paper 862, 32 p.
Schweger, C.E., and Matthews, J.V., Jr., 1985, Early and Middle Wisconsinan environments of eastern Beringia; Stratigraphic and paleoecological implications of the Old Crow tephra: Géographie Physique et Quaternaire, v. 39, no. 3, p. 275–290.
Weber, F.R., and Ager, T.A., 1984, Glacial-lake deposits in the Mt. Harper area, Yukon-Tanana Upland, *in* Coonrad, W.L., and Elliott, eds., The United States Gelogical Survey in Alaska; Accomplishments during 1981: U.S. Geological Survey Circular 868, p. 68–70.
Wintle, A.G., and Westgate, J.A., 1986, Thermoluminescence age of Old Crow tephra in Alaska: Geology, v. 14, p. 594–597.

Geological Survey of Canada Contribution 20687

Carter, L. David, Hamilton, Thomas D., and Galloway, John P., eds., 1989, Late Cenozoic History of the Interior Basins of Alaska and the Yukon: U.S. Geological Survey Circular 1026.

Paleosols of Northwestern Canada

By Charles Tarnocai

Paleosols were investigated in three areas in northwestern Canada: the central Yukon, the Old Crow area, and the Mackenzie Mountains. In the central Yukon, the work focused on paleosols developed on old glacial materials. In the unglaciated Old Crow area, paleosols developed on a variety of parent materials were studied in both the Old Crow Basin and the Bluefish River uplands. In the Mackenzie Mountains, paleosols were studied in the Little Bear River section, which is composed of Cordilleran and Laurentide glacial deposits of Pleistocene age. The brief discussion of these paleosols given below includes information concerning the relationship of their development to past climates.

Paleosols of the central Yukon

Paleosols in the central Yukon formed during the Pleistocene (table 1). Paleosols that formed during the early Pleistocene are called Wounded Moose paleosols, whereas those that developed during the late Pleistocene are called Diversion Creek paleosols (Smith and others, 1986). The morphological and analytical data presented in tables 2 and 3 include not only the data for these paleosols, but for comparison also include data for the Stewart soils, which developed during the Holocene on materials derived from the late Wisconsin McConnell glaciation (Smith and others, 1986).

Early Pleistocene paleosols

The Wounded Moose paleosols developed during pre-Illinoian time on both till and glaciofluvial materials deposited during the pre-Reid glaciations (table 1).

A maximum solum thickness of 2 m was reported for those soils developed on pre-Reid outwash (Tarnocai and others, 1985). The colors of the upper paleo-argillic (Bt) horizons of these paleosols are 5YR for outwash and 7.5YR for till (table 2). These paleosols developed strongly weathered Bt horizons with thick clay skins. Clay skins occur as coatings on ped and pebble surfaces and form bridges between sand grains. As a result of cryoturbation, the clay skins in the upper B horizon are usually fragmented and dispersed. The Wounded Moose paleosols are strongly leached. The uppermost paleo Bt horizon has average pH values ranging between 4.7 and 4.8. This horizon also has a somewhat higher concentration of exchangeable cations than the uppermost paleo B horizons in the Diversion Creek paleosols developed on corresponding materials. Carbon and nitrogen values are low in this horizon, and the average sodium pyrophosphate extractable Fe + Al values are 0.10 percent (table 3).

According to Foscolos and others (1977), Wounded Moose paleosols contain kaolinite, illite, and montmorillonite-kaolinite mixed-layer clay minerals. Chlorite and vermiculite are absent. Although chloritic intergrades are present, they are not as conspicuous as in the Diversion Creek paleosols. These chloritic intergrades decrease with depth, and no such intergrade was found in the IIC horizon. The montmorillonite-kaolinite mixed-layer clay minerals were identified below a depth of 190 cm in these paleosols (Foscolos and others, 1977).

Most Wounded Moose paleosols display strong cryoturbation in the form of disrupted and displaced soil horizons and oriented stones. Sand wedges and sand involutions of various sizes are also common. The sand wedges are much wider and more common in these paleosols than in the Diversion Creek paleosols. Ventifacts are commonly found at the paleosol surface. During the early Pleistocene, when development of Wounded Moose paleosols began, the climate was probably warm and subhumid and grassland and shrub vegetation predominated (Foscolos and others, 1977; Rutter and others, 1978). These conditions, in contrast to modern soil-forming conditions, promoted strong weathering, leaching, and soil development and the formation of montmorillonite. This was followed by a more temperate and humid climate that induced the degradation of montmorillonite to kaolinite through an intermediate step of mixed-layer montmorillonite-kaolinite. This latter type of climate was probably responsible for the red colors and very thick Bt horizons that characterize these paleosols. These two climatic regimes produced the rubification, deep sola, eluviated clay, and thick clay skins of the Wounded Moose paleosols. During subsequent glacial periods, Wounded Moose paleosols were affected by strong erosional and cryogenic processes as a result of the cold arctic climate. Thermal cracking caused the development of sand wedges and sand involutions, and cryoturbation produced discontinuous and displaced soil horizons and oriented features.

Late Pleistocene paleosols

The Diversion Creek paleosols developed on deposits formed during the Reid glaciation, with soil formation beginning during the Sangamonian interglacial period

Table 1. Paleosols of northwestern Canada

[Time scale used in this table is that of Heginbottom and Vincent (1986)]

General chronostratigraphy				Ice sheets (glaciations)	Paleosols			Major soil-forming processes and soil features
Time (Ma)	Period	Epoch	Stage		Central Yukon	Old Crow area	Little Bear River section	
		HOLOCENE						Weak Brunisolic development in the central Yukon, Cryosolic development in the Old Crow area.
0.010								
	QUATERNARY	LATE PLEISTOCENE	WISCONSINAN: LATE	McConnell and Buckland				Moderate cryogenic processes with the development of periglacial features in a permafrost environment. Severe erosion and the formation of ventifacts.
0.023								
			WISCONSINAN: MIDDLE			Old Crow paleosols		Old Crow paleosols: Cryosolic soil development with formation of patterned ground and permafrost.
0.064								
			WISCONSINAN: EARLY					
0.115								
			SANGAMONIAN		Diversion Creek		Paleosol 5	Diversion Creek paleosols: Moderate Brunisolic and weak Luvisolic soil development, leaching of carbonates, moderate Bm and weak Bt horizon development, occasional thin clay skins, moderate weathering. Paleosol 5: Moderate Brunisolic soil development with moderately thick Bm horizons, very little leaching, slight rubification in the Bm horizon.
0.128								
		MIDDLE PLEISTOCENE	ILLINOIAN	Reid				Strong cryogenic processes with the development of periglacial features in a permafrost environment. Severe erosion and formation of ventifacts.
0.320								
							Paleosol 4	INTERGLACIAL PERIODS: Wounded Moose paleosols: Strong Luvisolic soil development, leaching, development of Bt horizons with red colors (2.5YR and 5YR) and clay skins, strong weathering, deep sola.
0.730							Paleosol 3	Paleosols 1 to 4: Strong Brunisolic and weak Luvisolic soil development, some leaching and development of moderately thick Bm horizons with slightly reddish colors in Paleosol 4 and moderate rubification in paleosols 1 to 3, some argillans in paleosol 3.
		EARLY PLEISTOCENE	PRE-ILLINOIAN	Pre-Reid glacial and interglacial periods	Wounded Moose paleosols		Paleosol 2	
1.10							Paleosol 1	GLACIAL PERIODS: Strong cryogenic processes with the development of periglacial features in a permafrost environment. Severe erosion and formation of ventifacts on the Wounded Moose paleosols.
1.70								
	TERTIARY	PLIOCENE				Burnt Hill and Bluefish River paleosols		Burnt Hill and Bluefish River paleosols: Luvisolic and Podzolic soil development, leaching of carbonates, development of strong colors (5YR) and clay accumulation in the B horizon.
5.00								

(table 1). They are found on both till and glaciofluvial materials.

These paleosols are associated with a moderately well developed B horizon, and the dominant color of the uppermost B horizon is 7.5YR (table 2). The Bt horizons, which are found only in Diversion Creek paleosols developed on till, are associated with thin clay skins on ped surfaces and clay bridges between sand grains. These Diversion Creek paleosols have leached sola. The uppermost paleo B horizons have average pH values ranging between 4.6 and 5.3 and a low concentration of exchangeable cations. The carbon and nitrogen values are low in these horizons and the sodium pyrophosphate extractable Fe + Al values are generally less than 0.11 percent (table 3). Sand wedges are present in these paleosols but not common. Ventifacts are regularly observed on the surfaces of Diversion Creek paleosols. Cryoturbated features such as oriented stones, silt cappings, and distorted horizons are more common on

Table 2. Summary of soil morphologies of the central Yukon soils

[For definition of soil horizons see Agriculture Canada Expert Committee on Soil Survey (1987); solum thickness is mean thickness with range of values in brackets; equal sign means equal proportions]

Soil and parent material	Solum thickness (cm)	Dominant color hue in Munsell soil notation		Dominant primary structure		Soils with IIBt horizons (pct)	IIB horizons with clay skins (pct)	Soils with sand wedges and involutions (pct)	Soils with strong cryoturbation (pct)	Coarse fragments		
		Uppermost IIB	Lower IIB	Type	Grade					Degree of weathering	Frost orientation	Ventifacts
Wounded Moose (outwash)	109 (58-205)	5YR	7.5YR	Blocky	Weak to moderate	100	100	34	50	Strong chemical alteration	Common	Common
Wounded Moose (till)	91 (50-123)	7.5YR	7.5YR	Blocky = granular	Moderate to strong	100	100	75	100	Strong chemical alteration	Common	Common
Diversion Creek (outwash)	45 (9-90)	7.5YR	7.5YR	Blocky = platy	Weak to moderate	20	80	20	0	Moderate chemical alteration; frost shattering	Occasional	Abundant
Diversion Creek (till)	56 (14-106)	10YR	10YR	Blocky = granular	Weak to moderate	33	83	0	20	Moderate chemical alteration; frost shattering	Common	Common
Stewart (drift)	21 (0-40)	10YR	--	Variable	Weak	None	None	None	None	Weak chemical alteration; some frost shattering	None	None

Table 3. Average values of various parameters for the uppermost B horizons and the C horizons in the central Yukon soils

Soil and parent material	Soil horizon	Number of samples	$CaCO_3$ equiv. (pct)	pH	C (pct)	N (pct)	Sodium pyrophosphate (Fe+Al) (pct)	Exchangeable cations (meq/100 g)				Total		
								Ca	Mg	K	Al	Sand (pct)	Silt (pct)	Clay (pct)
Wounded Moose (outwash)	IIBt	14	0.0	4.8	0.22	0.01	0.10	7.19	3.85	0.09	0.34	57.2	23.0	19.8
	IIC	6	.0	5.3	.12	.01	.06	2.46	.84	.05	.16	90.4	5.0	4.6
Wounded Moose (till)	IIBt	6	.0	4.7	.23	.02	.10	10.70	3.85	.09	.25	50.9	30.3	18.7
	IIC	3	.0	5.0	.09	.00	.06	6.80	1.73	.10	.00	81.5	11.8	6.7
Diversion Creek (outwash)	IIBm	6	.23	5.3	.26	.01	.08	3.30	1.10	.09	.07	74.2	19.2	6.5
	IIC	2	.00	5.8	.13	.00	.05	1.70	.45	.03	.00	91.0	6.7	2.3
Diversion Creek (till)	IIBm	9	.00	4.6	.30	.02	.11	2.84	.85	.07	.15	63.5	31.1	5.3
	IIBt	6	.00	4.7	.30	.03	.11	7.02	1.90	.18	.13	53.7	32.5	13.7
	IIC	10	.00	5.3	.57	.03	.07	3.80	.25	.14	.02	66.1	26.0	7.9
Stewart (outwash)	IIBm	3	.00	4.3	.53	.04	.12	1.63	.40	.08	.98	56.6	38.9	4.5
	IIC	4	.00	4.8	.22	.20	.06	1.37	.40	.07	.03	84.0	14.9	1.1
Stewart (till)	IIBm	8	.94	5.3	.73	.04	.28	1.92	.43	.06	.30	62.5	3.9	6.6
	IIC	8	1.68	6.3	.33	.01	.08	2.23	.42	.02	.04	17.2	27.8	6.4

the Diversion Creek paleosols developed on till than on those developed on outwash. Weathering of coarse fragments in these paleosols is moderate and frost-shattered coarse fragments are common.

Foscolos and others (1977) and Rutter and others (1978) suggested that a cool and humid climate during this period was responsible for the development of Brunisols (paleo) or weak Luvisols (paleo). This cooler climate resulted in a weaker soil development than is found in the Wounded Moose paleosols. The Diversion Creek paleosols were exposed to only one glacial period, the McConnell glaciation, and thus cryogenic features are also weaker and fewer than those found in the Wounded Moose paleosols.

Paleosols of the Old Crow area

Late Pliocene and Pleistocene paleosols

Paleosols examined in the Old Crow area, a portion of the Yukon that was never glaciated, include examples of possible late Pliocene age from the Bluefish River valley, the Bluefish River uplands, and the Burnt Hill section along the Old Crow River. Other paleosols of early to middle Wisconsin age were examined in sections in the Old Crow Basin (table 1). Although most of these paleosols were buried by thick fluvial deposits, some of them are covered only by a thin loess cap.

A paleosol examined at the Burnt Hill exposure along the Old Crow River had a well-developed reddish (5YR 3/3) B horizon and a leached Ae horizon. Analysis of pollen from this soil indicated that the area had been dominated by spruce, pine, alder, birch, *Corylus* sp., ericaceous shrubs, and moss species. The pine pollens were identified as Haploxylon types (white pine), *Pinus monticola*, and Diploxylon types. Small amounts of *Tsuga* and *Abies* pollen were also present. The pollen assemblage found in this paleosol indicates that the vegetation was a mixed, closed-canopy forest with a dense alder layer. The understory was dominated by *Corylus* sp., ericaceous shrubs, and moss species. The presence of white pine, *Corylus*, and *Tsuga* point to a climate much warmer than at present. Similar pollen assemblages, dated as late Pliocene (Nelson and Carter, 1985), were found in the Gubik Formation, at the Ocean Point site in Alaska, although no *Corylus* was found at that site and *Salix* pollen was much less common than in the Burnt Hill paleosol. The similarity of the pollen assemblage indicates that the Burnt Hill paleosol also developed in the late Pliocene (Schweger, in press).

Paleosols in the Bluefish River area developed from calcareous materials derived from limestone bedrock. They have a well-developed B horizon with strong colors, dominantly 5YR 4/5 moist. The B horizon contains a significant clay accumulation (table 4). The Bluefish River paleosols have average pH values of 7.4 in the uppermost paleo Bt horizon and 7.5 in the Ck horizon, which would normally indicate that very little leaching has occurred. The difference in the concentration of calcium carbonate in these two horizons, however, indicates that a significant amount of leaching has taken place. The average calcium carbonate concentration in the Ck horizon is 63.5 percent while in the Bt horizon it is 17.0 percent. These paleosols contain high amounts of total carbon. Since the concentration of carbon is generally higher in the parent material than in the overlying Bt horizons, this carbon is most likely derived from the bedrock (table 4). The Burnt Hill paleosol (table 5), on the other hand, has developed on noncalcareous fluvial materials. Its morphology reflects the effects of advanced podzolization processes. Although the sodium-pyrophosphate-extractable Fe and Al values are lower than in the contemporary (Holocene) Podzols (Agriculture Canada Expert Committee on Soil Survey, 1987), the ammonium-oxalate-extractable Fe and Al values are very similar. This would indicate that the use of the sodium-pyrophosphate-extractable Fe and Al criteria developed for recent Podzols is questionable when attempting to identify podzolization in paleosols. It is possible that the sodium-pyrophosphate-extractable Fe and Al values change with time. In modern Podzols occurring in southern Canada, the sodium-pyrophosphate-extractable Fe is approximately 50 percent of the ammonium-oxalate-extractable Fe (Wang and McKeague, 1982; McKeague and others, 1983). In the the Burnt Hill paleosol this ratio is much less, possibly because the sodium-pyrophosphate-extractable Fe, which is bound in an iron-organic complex (McKeague and others, 1971), has oxidized into an amorphous hydrous iron oxide. The concentration of this form of amorphous iron is determined by the ammonium oxalate test.

Most of these paleosols were buried under various sediments in the Bluefish River valley. In one case, however, a strongly truncated paleosol was found on the Bluefish River uplands. Soil material from the surface horizon of a buried paleosol was analyzed for pollen. It had essentially the same pollen assemblage as did the paleosol at the Burnt Hill exposure. The development of paleosols found in the Bluefish River uplands, the Bluefish River valley, and the Burnt Hill exposure are very similar. The soil development and, especially, the pollen data suggest that the Bluefish River paleosols are also of late Pliocene age.

The most recent paleosols in the Old Crow area were found buried within a dominantly fluvial succession overlain by late Wisconsin lake sediments in the Old Crow Basin. These Cryosols (paleo), which had developed in a permafrost environment, were associated with gleyed horizons, mottles, cryoturbated features, patterned ground, and ice wedges. A bone buried in this soil yielded a radiocarbon age of 42±1.2 ka (R.E. Morlan, Archaeological Survey of Canada, oral commun., 1987).

Paleosols of the Little Bear River section

The Little Bear River lies near the boundary between the Mackenzie Plain and the Mackenzie Mountains, in a zone of overlap between the former extent of local montane glaciers originating in the Canyon Ranges of the Mackenzie Mountains and the maximum extent of the Laurentide ice sheet. The Little Bear River section is located at latitude 64° 27.5' N. and longitude 126° 42' W. at an elevation of approximately 900 m. Tills of montane origin overlain by the Laurentide boulder gravel can be differentiated in this section. These tills and boulder gravel, which are underlain by bedrock of the Franklin Mountain Formation, have an aggregate thickness of approximately 40 to 50 m.

Five paleosols were identified in association with five different montane tills (table 1). In addition, a modern soil developed on boulder gravel of Laurentide origin occurs on the surface. Paleosols developed on the top of each till indicate prolonged intervals of soil formation during the interglacial periods (Hughes and others, in press).

Spruce logs found in a buried peat deposit on the top of paleosol 4 gave a radiocarbon age greater than 47 ka (Hughes and others, in press). Pollen data obtained from

Table 4. Average values of various parameters for the uppermost B horizons and the C horizons in the Bluefish River paleosols

Soil and parent material	Soil horizon	Number of samples	$CaCO_3$ equiv (pct)	pH	C (pct)	N (pct)	Dithionite citrate (Fe + Al) (pct)	Ammonium oxalate (Fe + Al) (pct)	Sodium pyrophosphate (Fe + Al) (pct)	Total Sand (pct)	Total Silt (pct)	Total Clay (pct)
Bluefish River	IIBt	3	17.0	7.4	2.13	0.06	1.76	0.27	0.05	51.5	30.8	18.0
(residual)	IICk	3	63.5	7.5	7.68	.24	.79	.11	.03	43.9	44.8	11.3

Table 5. Analytical data for the Burnt Hill paleosol

[*, calculated]

Soil and parent material	Soil horizon	Number of samples	$CaCO_3$ equiv. (pct)	pH	C (pct)	N (pct)	Dithionite citrate Fe (pct)	Dithionite citrate Al (pct)	Ammonium oxalate Fe (pct)	Ammonium oxalate Al (pct)	Sodium pyrophosphate Fe (pct)	Sodium pyrophosphate Al (pct)	C.E.C.* meq/100 g	Total Sand (pct)	Total Silt (pct)	Total Clay (pct)
Burnt Hill	Ae	1	0.0	4.6	0.06	0.02	0.43	0.04	0.31	0.03	0.05	0.01	2.80	92.9	5.5	1.6
(fluvial)	Bf	1	.0	6.7	.31	.02	3.94	.04	3.54	.03	.02	.00	3.90	91.7	5.9	2.3
	BC	1	.0	6.7	.23	.02	.81	.04	.97	.03	.05	.00	3.93	92.5	5.6	2.0
	C	1	.0	6.4	.23	.01	.75	.04	.92	.03	.05	.00	2.87	89.7	8.5	1.7

this same deposit indicate that the area was covered by a closed-canopy coniferous forest at that time (Hughes and others, in press). The area is now alpine tundra, just above treeline. This information and the degree of soil development indicate that paleosols 1 to 4 probably developed during pre-Illinoian stages whereas paleosol 5 likely developed during the Sangamonian Stage (table 1).

The uniqueness of the Little Bear River section is that it contains, in one location, five distinctly different paleosols developed on similar parent materials. The greater solum and B horizon thicknesses, the increase of redness in the B horizons, and the occurrence of some argillans in the older paleosols indicate stronger soil development with age (table 6). This suggests that, since the formation of the oldest paleosols, the climate has steadily cooled. The lack of evidence for cryoturbation, especially in Gleysols (paleo), indicates that these paleosols developed in a permafrost-free environment.

All of these paleosols developed in a relatively dry environment in which the precipitation was probably similar to that occurring today, but because of the higher temperatures and thus correspondingly higher rate of evapotranspiration, the environment was drier than at the present time. Paleosols 1 and 2 developed under a slightly warmer climate than did paleosols 3, 4, and 5, which developed in a boreal environment dominated by closed-canopy coniferous forest vegetation. The modern soil has developed in a permafrost environment, the coldest of all the climates under which these soils have developed.

Summary

1. The older the paleosol, the stronger the soil development. Paleosols developed during the late Pliocene and the pre-Illinoian stages of the Pleistocene have stronger soil development than those developed in the Sangamonian and early part of the Wisconsinan Stages of the Pleistocene.
2. This stronger soil development in the late Pliocene and the pre-Illinoian stages of the Pleistocene is primarily due to the warmer and moister climate and, to a lesser degree, to the great length of time during which weathering took place.
3. All pre-Wisconsin paleosols began developing in a permafrost-free environment.
4. Cryogenic features found in all pre-Wisconsin paleosols, except those which were buried, result from the cold climate which existed during the subsequent glacial periods.
5. Early Wisconsin paleosols in the Old Crow Basin began developing in a permafrost environment.

REFERENCES CITED

Agriculture Canada Expert Committee on Soil Survey, 1987, The Canadian system of soil classification: Research Branch, Agriculture Canada, Publication No. 164, second edition, 164 p.

Foscolos, A.E., Rutter, N.W., and Hughes, O.L., 1977, The use of pedological studies in integrating the Quaternary history of

Table 6. Average values of the uppermost B horizons in the Little Bear River paleosols

[Redness rating = [(10–H)xCl]/V, where H is the YR hue, C is the chroma and V is the value (Torrent and others, 1980, 1983]

	Number of pedons	Total B horizon thickness (cm)	Color	Redness rating	pH	$CaCO_3$ (pct)	N (pct)	C (pct)	Dithionite citrate (Fe+Al) (pct)	Ammonium oxalate (Fe+Al) (pct)	Sodium pyroposphate (Fe+Al) (pct)	Clay (pct)
Paleosol 1	0	--	--	--	--	--	--	--	--	--	--	--
Paleosol 2	2	68	5YR 4/4 7.5YR 4/2	3.1[a]	7.2	1.3	0.10	0.16	1.29	0.13	0.03	10.9
Paleosol 3	1	95	7.5YR 4/2	1.2	7.4	1.4	.01	.28	1.92	.19	.06	8.3
Paleosol 4	1	68	10YR 3/2	.0	7.4	18.2	.05	2.32	1.92	.46	.06	26.8
Paleosol 5	3	81	7.5YR 4/6 10YR 4/6 10YR 4/6	1.2[a]	7.3	.6	.03	.29	1.90	.25	.06	20.6
Modern soil	1	42	10YR 4/6	.0	6.4	.0	.16	2.63	2.33	1.15	.56	18.7

[a]Mean value.

central Yukon: Geological Survey of Canada, Energy, Mines and Resources Canada, Bulletin 271, 48 p.

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 1984: Geological Survey of Canada Open File Report 1237, 60 p.

Hughes, O.L., Tarnocai, C., and Schweger, C.E., in press, Pleistocene stratigraphy, paleopedology and paleoecology, Little Bear River section, western District of Mackenzie: Géographie Physique et Quaternaire.

McKeague, J.A., Brydon, J.E., and Miles, N.M., 1971, Differentiation of forms of extractable iron and aluminum in soils: Soil Science Society of America, Proceedings, v. 35, p. 33–38.

McKeague, J.A., Wang, C., Coen, G.M., DeKimpe, C.R., Laverdiere, M.R., Evans, L.J., Kloosterman, B., and Green, A.J., 1983, Testing chemical criteria for spodic horizons on podzolic soils in Canada: Soil Science Society of America Journal, v. 47, p. 1052–1053.

Nelson, R.E., and Carter, L.D., 1985, Pollen analysis of a late Pliocene and early Pleistocene section from the Gubik Formation of arctic Alaska: Quaternary Research, v. 24, p. 295–306.

Rutter, N.W., Foscolos, A.E., and Hughes, O.L., 1978, Climatic trends during the Quaternary in central Yukon based upon pedological and geomorphological evidence, *in* York Quaternary Symposium on Quaternary Soils, 2d, Proceedings: Norwich, England, Geo Abstracts, p. 309–359.

Schweger, C.E., in press, Late Tertiary-early Pleistocene palynology, Old Crow-Bluefish Basins, northern Yukon, Canada: Proceedings of a Workshop on Neogene Environments, Denver, October 1987.

Smith, C.A.S., Tarnocai, Charles, and Hughes, O.L., 1986, Pedological investigations of Pleistocene glacial drift surfaces in the central Yukon: Géographie Physique et Quaternaire, v. 15, p. 29–37.

Tarnocai, C., Smith, S., and Hughes, O.L., 1985, Soil development on Quaternary deposits of various ages in the central Yukon Territory: Geological Survey of Canada Paper 85–1A, Current Research, Part A, p. 229–238.

Torrent, J., Schwertmann, U., and Schulze, D.G., 1980, Iron oxide mineralogy of some soils of two river terrace sequences in Spain: Geoderma, v. 23, p. 91–208.

Torrent, J., Schwertmann, U., and Alferez, F., 1983, Quantitative relationships between soil color and hematite content: Soil Science, v. 136, p. 354–358.

Wang, C., and McKeague, J.A., 1982, Illuviated clay in sandy podzolic soils of New Brunswick: Canadian Journal of Soil Science, v. 62, p. 79–89.

Carter, L. David, Hamilton, Thomas D., and Galloway, John P., eds., 1989, Late Cenozoic History of the Interior Basins of Alaska and the Yukon: U.S. Geological Survey Circular 1026.

Upper Cenozoic Deposits, Kanuti Flats and Upper Kobuk Trench, Northern Alaska

By Thomas D. Hamilton

Tectonic deformation has caused the Kanuti Flats and upper Kobuk trench (fig. 1) to be continuing foci of late Cenozoic deposition. The Kanuti Flats, a broad lake-dotted plain within the Koyukuk River valley (Wahrhaftig, 1965, p. 26), has been subjected to late Cenozoic subsidence which continues to the present day (T.D. Hamilton, unpub. field mapping, 1985 and 1987). Its geologic record spans about two million years and includes glacial and glaciolacustrine deposits correlated with all of the major glacial events defined for the central Brooks Range (Hamilton, 1986a, 1986b). From oldest to youngest, these are the Gunsight Mountain, Anaktuvuk River, Sagavanirktok River, and Itkillik glacial episodes. The upper Kobuk trench is part of a structural feature that extends west to the Bering Sea (Grantz, 1966); activity along its bounding faults has continued into the Holocene (Hamilton, 1984). The record from the upper Kobuk trench extends into the Matuyama Reversed-Polarity Chron and includes deposits of Anaktuvuk River through Itkillik ages.

Drift of presumed Tertiary age correlated with Gunsight Mountain glacial advances forms a subdued, arcuate, probable end moraine about 4 to 5 km wide that is traceable for nearly 100 km along the south margin of the Kanuti Flats (fig. 2). The moraine is thickly covered by loess and lacustrine silt of Quaternary age, but concentrations of erratic boulders along river channels that cut the moraine attest to its glacial origin.

Glacier advances of Anaktuvuk River age occurred prior to about 0.8 Ma throughout the central Brooks Range on ancient landscapes 50 m or more above levels of modern streams (Hamilton, 1986b, 1986c). Glaciers that flowed south from the Brooks Range coalesced into broad piedmont lobes in lowlands around the present-day Kobuk and Koyukuk Rivers (fig. 2). Glacial lobes terminated in a large proglacial lake in the Kanuti Flats, where subdued arcuate end moraines generally are covered by eolian and lacustrine silt more than 15 m thick. Knolls and ridges along moraine crests contain wave-washed gravel mixed with larger glacial clasts. River bluffs expose strongly jointed, compact till and stony glaciolacustrine silt that are strongly oxidized along joint planes. Glaciers in the Kobuk valley coalesced into an extensive ice body that extended west to Kotzebue Sound and probably was contemporaneous with the early Pleistocene Nome River advance of the Seward Peninsula (Kaufman and Hopkins, 1986).

Glaciation of Anaktuvuk River age was followed by a long nonglacial interval during which drainage systems within the Brooks Range became adjusted to levels close to those of modern flood plains and structural troughs developed along the south flank of the range (Hamilton, 1986a). Bluffs and knolls within the Koyukuk-Kobuk region expose rounded, quartzose, pebble-small-cobble gravel that filled east-west-trending structural troughs to measured thicknesses as great as 55 m during a long interval of reduced glaciation in mountain valleys. Paleomagnetic determinations indicate that deposition of nonglacial fine gravel in the upper Kobuk trench began during the Matuyama Reversed-Polarity Chron and continued into the Brunhes Normal-Polarity Chron (Hamilton, 1986a, 1986c).

Glacier flow patterns of Sagavanirktok River age were similar to those of the Anaktuvuk River glacial interval, but glaciers were markedly smaller. Glaciers filled structural troughs along the south flank of the Brooks Range (fig. 2) and extended south beyond the troughs along major valleys, within which they commonly terminated in lakes. Drift is not as broadly eroded as is drift of Anaktuvuk River age, and it occurs close to modern valley floors. Multiple end moraines and stratigraphic relations suggest two major ice advances in some valleys. The Sagavanirktok River glaciation is much younger than the early Pleistocene Anaktuvuk River glaciation and is separated by a major

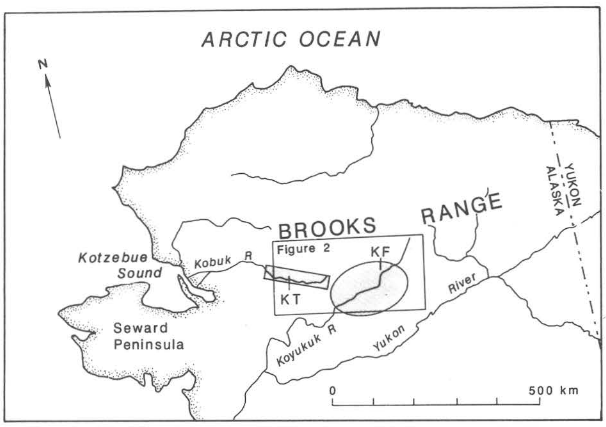


Figure 1. Northern Alaska, showing Kanuti Flats (KF), upper Kobuk trench (KT), and area of detail map (fig. 2).

interglacial period from succeeding glacial advances of Itkillik age. Because of evidence for multiple advances in some valleys, "Sagavanirktok River" is considered a generic term for multiple ice advances of middle Pleistocene age.

All glacial deposits of late Pleistocene age are assigned to the Itkillik glaciation, which is divisible into two or three major phases. The outer border of Itkillik drift is the most striking discontinuity in the entire Brooks Range glacial succession; it separates subdued older glacial deposits from steeper, stonier, and more irregular drift with drainage derangements and discontinuous vegetation.

During Itkillik I time, glaciers extended through southern valleys of the Brooks Range and terminated 25–45 km beyond the range along the north margin of the Kanuti Flats. Glaciers farther west were smaller and generally did not extend into the upper Kobuk trench. Moraines have smooth, somewhat flattened, hummocky crests that support open vegetation; moist depressions are covered by dense muskeg. All ice advances of Itkillik I age are older than the effective range of radiocarbon dating, and stratigraphic relations described in Westgate and others (1983) suggest that the maximum advance may have taken place shortly before deposition of the Old Crow tephra about 86±8 ka (Wintle and Westgate, 1986). A younger major advance occured sometime between about 86 and 55 ka. Drift-mantled glacier ice persisted for long intervals in many valleys, and relict ice of Itkillik I age is still locally present (Hamilton, 1982).

Drift of Itkillik II age formed during late Wisconsin time between about 24 and 11.5 ka. Rugged end moraines commonly are located at the south flank of the Brooks Range, where they enclose elongate lakes or lacustrine plains. Younger end moraines formed within mountain valleys during a major readvance of glaciers between 13 ka and 11.5 ka.

Many glaciers of Sagavanirktok River and Itkillik age either terminated in standing water or extended to within a few kilometers of large lakes in the upper Kobuk trench and the Kanuti Flats (Hamilton, 1986a). Some moraines are anomalously broad and flat, and they consist either of fine-grained glaciolacustrine sediment with dropstones or of deltaic sand with steeply dipping foreset beds. Outwash trains in other cases extend southward and become deltas at concordant altitudes that reflect former lake levels. Because the lakes were not dammed by glaciers or by any

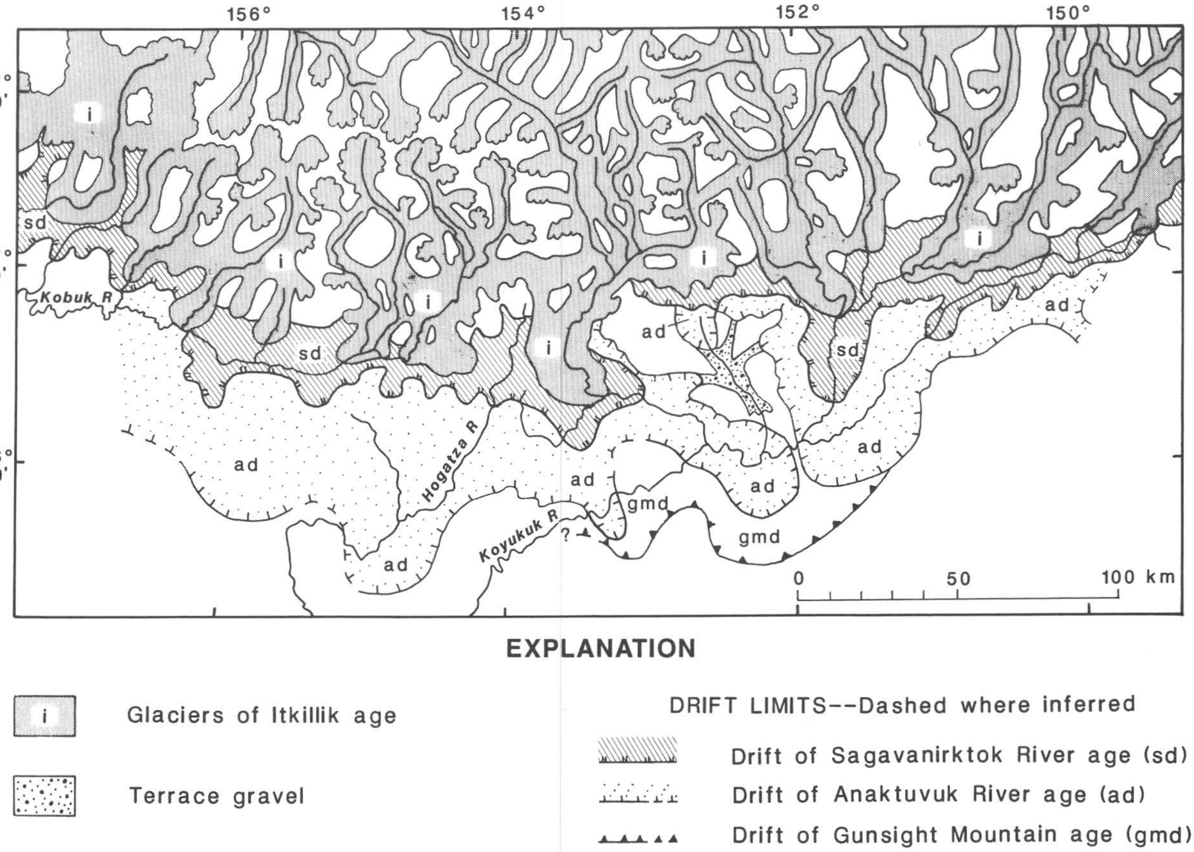


Figure 2. Glacier limits and drift deposits, Kanuti Flats and upper Kobuk trench, northern Alaska.

other known physical agents, and because their reconstructed surfaces rise northward toward the Brooks Range, their origin must be the result of isostatic deflection and drainage reversals within the low-gradient Kobuk and Koyukuk drainage systems.

REFERENCES CITED

Grantz, Arthur, 1966, Strike-slip faults in Alaska: U.S. Geological Survey Open-File Report 267, 82 p.

Hamilton, T.D., 1982, Relict Pleistocene glacier ice in northern Alaska, *in* Arctic workshop, 11th, March 1982, Institute of Arctic and Alpine Research: Boulder, University of Colorado, p. 25–26.

———1984, Late Quaternary offsets along the Kobuk and related fault zones, northwestern Alaska: Geological Society of America Abstracts with Programs, v. 16, p. 288.

———1986a, Late Cenozoic glaciation of the central Brooks Range, *in* Hamilton, T.D., Reed, K.M., and Thorson, R.M., eds., Glaciation in Alaska—The geologic record: Anchorage, Alaska Geological Society, p. 9–49.

———1986b, Glaciation of the Brooks Range, *in* Heginbottom, J.A., and Vincent, J-S., eds., Correlation of Quaternary deposits and events around the margin of the Beaufort Sea: Geological Survey of Canada Open File Report 1237, p. 27–29.

———1986c, Correlation of Quaternary glacial deposits in Alaska, *in* Richmond, G.M., and Fullerton, D.S., eds., Quaternary glaciations in the United States of America: Quaternary Science Reviews, v. 5 (Quaternary Glaciations in the Northern Hemisphere), p. 171–180.

Kaufman, D.S., and Hopkins, D.M., 1986, Glacial history of the Seward Peninsula, *in* Hamilton, T.D., Reed, K.M., and Thorson, R.M., eds., Glaciation in Alaska—The geologic record: Anchorage, Alaska Geological Society, p. 51–77.

Wahrhaftig, Clyde, 1965, Physiographic divisions of Alaska: U.S. Geological Survey Professional Paper 482, 52 p.

Westgate, J.A., Hamilton, T.D., and Gorton, M.P., 1983, Old Crow tephra—A new late Pleistocene stratigraphic marker across north-central Alaska and western Yukon Territory: Quaternary Research, v. 19, p. 38–54.

Wintle, A.G., and Westgate, J.A., 1986, Thermoluminescence age of Old Crow tephra in Alaska: Geology, v. 14, p. 594–597.

Carter, L. David, Hamilton, Thomas D., and Galloway, John P., eds., 1989, Late Cenozoic History of the Interior Basins of Alaska and the Yukon: U.S. Geological Survey Circular 1026.

Quaternary Deposits at Birch Creek, Northeastern Interior Alaska: The Possibility of Climatic Reconstruction

By Mary E. Edwards *and* Patricia F. McDowell

Extensive sections of Quaternary sediments, exposed by rivers or mining activities, offer excellent opportunities to study Quaternary paleoenvironments in unglaciated interior Alaska. Thick accumulations of loess containing paleosol complexes occur in the Fairbanks area and the Yukon Flats (Péwé, 1975; McDowell and Edwards, 1984). Woody and organic-rich units ("forest beds") and lacustrine sediments are also present. Tephra layers provide a means of correlation and dating (Westgate and others, 1983).

We suggest that a sequence of climatic oscillations is recorded in such sections. If a chronology can be established and records correlated, the inferred climatic sequence may extend the terrestrial proxy climate record through marine oxygen-isotope stage 5 and beyond, and provide independent data for the testing of paleoclimatic models. Ritchie and others (1983) have demonstrated that the paleoecological record in northwest Canada reflects a climatic warming about 10 ka, attributable to increased summer insolation at high latitudes (Milankovitch, 1941). Other insolation maxima controlled by orbital elements occurred about 50, 80 to 90, and 130 ka (Berger, 1978). The known values of important climate boundary conditions such as ice volume, sea-surface temperature, and sea level suggest that warm intervals could have been associated with these maxima (P.J. Bartlein, University of Oregon, oral commun., 1987), but few data are available to test this hypothesis.

This report describes a 45-m-high river-cut section of Quaternary deposits on Birch Creek, near Circle, Alaska (fig. 1). We briefly discuss the stratigraphy of this site, problems associated with dating the sediments, and the application of our data to paleoclimatic reconstruction.

Study area

The Yukon Flats tectonic basin (altitude 150 to 250 m) contains a marginal upland surrounding a central lowland, through which flows the Yukon River (Williams, 1962; fig. 1). The southern marginal upland is mantled by loess as thick as 35 m, which overlies gravel or bedrock. Williams (1962) suggested that the loess is of Pleistocene age, derived from the flood plain of the Yukon and its tributaries in glacial periods. The present surface of the southern marginal upland is characterized by numerous lakes.

The modern climate is cold continental (Fort Yukon July mean temperature is 17 °C; January mean is –28 °C; annual precipitation is <20 cm). Discontinuous permafrost underlies the region. Below 700 m altitude the vegetation is boreal forest dominated by spruce (*Picea*), birch (*Betula*), alder (*Alnus*), poplar (*Populus*), and willow (*Salix*).

The present boreal forest developed during the early Holocene, according to late Quaternary pollen diagrams for the Yukon Flats (Edwards and Brubaker, 1984) and the Tanana lowland (Ager, 1975). Full-glacial pollen assemblages are characterized by herbaceous taxa such as Poaceae, Cyperaceae, *Artemisia*, and Chenopodiaceae-Amaranthaceae and suggest tundra vegetation.

Section BC–1

Section BC–1 (fig. 2) faces east and south and is exposed for about 400 m along Birch Creek. We sampled a 45-m-high part of the section that is free from large ice masses, and where slumping is confined to the lowermost 2.5 m of the section. The basal sediments, units A and B, from 2.5 to 14 m above creek level, are fluvial deposits of silt, sand, and gravel that include organic lenses and pieces of detrital wood. The presence of pollen of *Picea*, *Alnus*, and *Betula* and the absence of palynomorphs characteristic of Tertiary sediments suggest that the deposits are of Pleistocene age. Unit A contains a white ash layer that is 1 to 3 cm thick. Its composition suggests a source in the Wrangell Mountains, but the ash has not been dated and is not yet chemically or petrographically characterized (J.A. Westgate, University of Toronto, written commun., 1987).

Between 14 m and 31 m above river level are three units of silt. The lowermost (unit C) probably consists of retransported loess; units D and E are massive silts, probably undisturbed or slightly retransported loess. The uppermost part of units B, C, and D contain relatively large amounts of disseminated organic matter and macrofossils (such as rootlets and pieces of wood), and are either leached of carbonates, and(or) display roots, bioturbation, and zones resembling soil horizons, suggesting that they are paleosols. A 10-cm-thick volcanic ash layer 25 m above river level has been identified as the Old Crow tephra (Westgate and others, 1985).

Above unit E is an unconformity overlain by 12 m of lacustrine sediment and peat (unit F). The lowermost part of unit F consists of a layer of detrital, beaver-chewed *Picea*, *Salix*, and *Populus* wood. Approximately 9 m of lacustrine sediments lies above the woody layer and is in turn overlain by about 3 m of silty peat. The section is capped by several meters of loess (unit G).

Pollen has been extracted from units E and F. Unit E contains Chenopodiaceae-Amaranthaceae, Poaceae, other herbs, and high frequencies of indeterminable pollen (>100-grain counts from loess). Abundant pollen of *Picea*, *Betula*, and *Alnus*, with minor quantities of *Populus*, Poaceae, Cyperaceae, and *Typha* characterizes unit F (>300-grain counts from lacustrine silt).

Chronology

Three radiocarbon ages (QL–1897, QL–1898, QL–1899) on unit F are older than 50 ka. A maximum age for unit E is provided by the Old Crow tephra, which occurs widely in eastern Beringia (Westgate and others, 1985). Fission-track data indicate a maximum age of 120 ka (Naeser and others, 1982; Westgate and others, 1985). Thermoluminescence dating of bracketing loess has given an age of about 86 ka (Wintle and Westgate, 1986), and 110 ka (Berger, 1987). This would suggest that unit E loess deposition and subsequent lake development occurred after 120 ka and before 50 ka. Holocene lake sedimentation rates of 1 m/2,000 yr to 1 m/6,000 yr determined from modern lakes in the study area (M. Edwards, unpub. data) suggest that the lake or pond represented by unit F was present for a considerable interval of time.

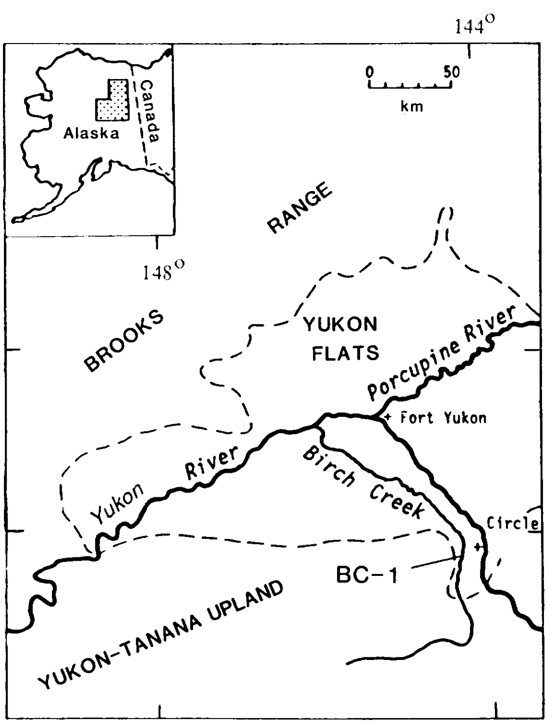


Figure 1. Central Alaska, showing location of Birch Creek section. Dashed line is boundary of Yukon Flats.

Implications for paleoclimatic reconstruction

In a survey of early- and mid-Wisconsin paleoenvironmental data from eastern Beringia, Schweger and Matthews (1985) showed that a phase of cold, treeless conditions followed the deposition of the Old Crow tephra. A subsequent warm period is marked variously at sites across eastern Beringia by increased pollen frequencies of boreal forest taxa, deep thawing, and macrofossil evidence of forested conditions. The BC–1 stratigraphy shows a similar pattern, with loess characterized by a herbaceous pollen assemblage overlying the Old Crow tephra, and a subsequent warm period represented by unit F.

The boreal forest vegetation indicated by the unit F pollen spectra, and the presence of *Typha*, which today reaches its northern limits in the study region, imply a very warm interval, probably of interglacial character. If the dating of the Old Crow tephra is correct, this interval occurred in early Wisconsin time, and perhaps reflects an insolation maximum at 80 to 90 ka or about 50 ka (approximately oxygen-isotope stages 5a+5c or 3). However, we suggest that the age of the Old Crow tephra must be further

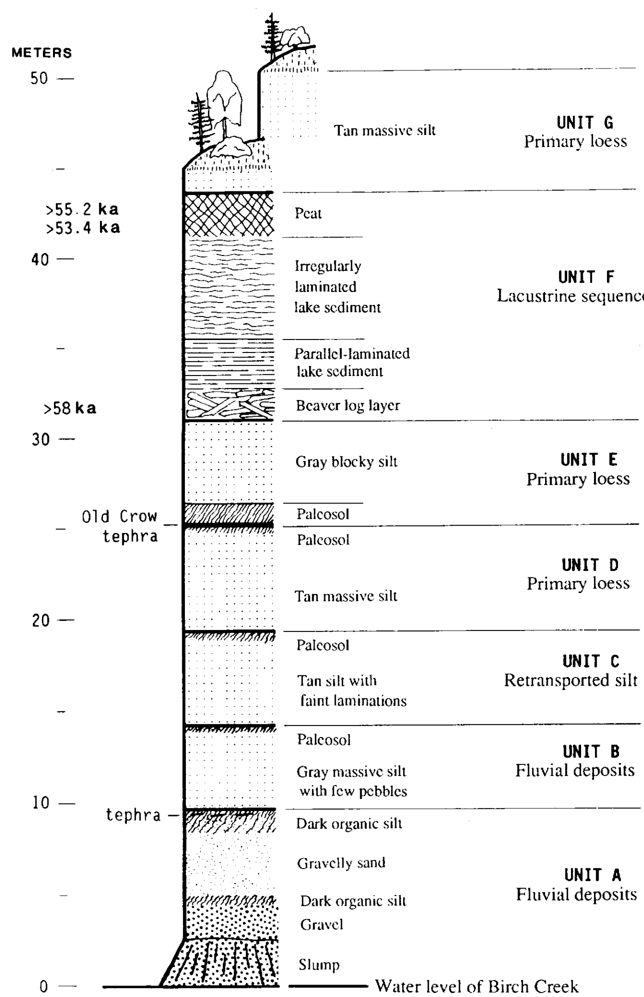


Figure 2. Summary stratigraphic diagram of section BC–1, including Old Crow tephra of Westgate and others (1985).

studied. The fission-track method is at the lower limit of its range, and dating therefore rests heavily on two thermoluminescence results. It is critical that the age of the Old Crow tephra be more firmly established. If it is actually much older, the very warm interval represented by unit F would probably correspond to the last interglacial (130 ka; oxygen-isotope stage 5e).

At BC–1, determining the age of the basal tephra in unit A is crucial for dating the lower part of the section; work is in progress to date the tephra and to characterize it chemically and petrographically. Further plans for work at BC–1 involve a detailed analysis of the paleosols and the extraction of pollen from the lower units in an attempt to describe environments of earlier parts of the Pleistocene.

This work was funded by National Science Foundation Grant DPP 8303734. We thank P.J. Bartlein (University of Oregon), D.M. Hopkins (University of Alaska), J.V. Matthews, Jr. (Geological Survey of Canada), and C.E. Schweger (University of Alberta) for helpful discussions and advice. Robert Anderson (University of Washington) and P.W. Dunwiddie (University of Washington) provided field assistance, P.W. Dunwiddie made the fossil wood identifications, L.B. Brubaker (University of Washington) provided laboratory facilities, and Mulu Malmo (University of Washington) provided technical help.

REFERENCES CITED

Ager, T.A., 1975, Late Quaternary environmental history of the Tanana River valley, Alaska: Columbus, Ohio State University, Institute of Polar Studies Report No. 54, 117 p.

Berger, A.L., 1978, Long-term variations of caloric insolation resulting from the Earth's orbital elements: Quaternary Research, v. 9, p. 139–167.

Berger, G.W., 1987, Thermoluminescence dating of the Pleistocene Old Crow tephra and adjacent loess, near Fairbanks, Alaska: Canadian Journal of Earth Sciences, v. 24, p. 1975–1984.

Edwards, M.E., and Brubaker, L.B., 1984, A 23,000-year pollen record from northern interior Alaska: Boulder, Colorado, American Quaternary Association, VIII Biennial Meeting Abstracts, p. 35.

McDowell, P.F., and Edwards, M.E., 1984, Loess stratigraphy and landscape development, Yukon Flats area, interior Alaska: Boulder, Colorado, American Quaternary Association, VIII Biennial Meeting Abstracts, p. 81.

Milankovitch, Milutin, 1941, Canon of insolation and the ice-age problem: Belgrade, Koniglich Serbische Akademie Geogr. Spec. Publ. 132, 484 pp. (Translated by Israel Program for Scientific Publications, Jerusalem, 1969.)

Naeser, N.D., Westgate, J.A., Hughes, O.L., and Péwé, T.L., 1982, Fission-track ages of late Cenozoic distal tephra beds in the Yukon Territory and Alaska: Canadian Journal of Earth Sciences, v. 29, p. 2167–2178.

Péwé, T.L., 1975, Quaternary geology of Alaska: U.S. Geological Survey Professional Paper 835, 145 p.

Ritchie, J.C., Cwynar, L.C., and Spear, R.W., 1983, Evidence from northwest Canada for an early Holocene Milankovitch thermal maximum: Nature, v. 305, p. 126–128.

Schweger, C.E., and Matthews, J.V., Jr., 1985, Early and middle Wisconsinan environments of eastern Beringia: Stratigraphic and paleoecological implications of the Old Crow tephra: Géographie Physique et Quaternaire, v. 39, p. 275–290.

Westgate, J.A., Hamilton, T.D., and Gorton, M.P., 1983, Old Crow Tephra: A new late Pleistocene stratigraphic marker across north-central Alaska and western Yukon Territory: Quaternary Research, v. 19, p. 38–54.

Westgate, J.A., Walter, R.C., Pearce, G.W., and Gorton, M.P., 1985, Distribution, stratigraphy, petrochemistry, and paleomagnetism of the late Pleistocene Old Crow Tephra in Alaska and the Yukon: Canadian Journal of Earth Sciences, v. 22, p. 893–906.

Williams, J.R., 1962, A geologic reconnaissance, Yukon Flats district, Alaska: U.S. Geological Survey Bulletin 1111–H, p. 290–331.

Wintle, A.G., and Westgate, J.A., 1986, Thermoluminescence age of the Old Crow tephra in Alaska: Geology, v. 14, p. 594–597.

Carter, L. David, Hamilton, Thomas D., and Galloway, John P., eds., 1989, Late Cenozoic History of the Interior Basins of Alaska and the Yukon: U.S. Geological Survey Circular 1026.

Late Quaternary Paleofloods along the Porcupine River, Alaska: Implications for Regional Correlation

By Robert M. Thorson

A preliminary overview of the alluvial history of the Porcupine River valley was presented by Thorson and Dixon (1983). In that paper, we described erosional and depositional evidence for exceptionally high discharges through the Porcupine canyon that resulted from intermittent drainage of lakes that were impounded in basins of Yukon Territory by the Keewatin sector of the Laurentide ice sheet. We concluded that at least two paleoflood episodes, separated by an interval of local aggradation, occurred within approximately the last 35,000 radiocarbon years.

This report presents additional stratigraphic and erosional evidence for the Porcupine River paleofloods and provides preliminary paleohydrologic estimates for peak discharge, mean channel velocity, and flood duration. I compare features along the Porcupine River to those reported for the Channeled Scabland of Washington (Bretz, 1969; Baker and Nummendal, 1978), the Snake River canyon (Malde, 1968), and torrential flood deposits in mountain valleys (Costa, 1983). Nearly every feature described in these areas is present along the Porcupine River upstream from the Yukon Flats (fig. 1). High-water marks are recognized on aerial photographs by either pronounced erosional discontinuities or by massive eddy-bars along reentrants (figs. 2, 3). Constricted reaches are characterized by steep waterline gradients, trough-shaped cross sections, and hanging tributaries (fig. 4). Expanded reaches exhibit low gradients and broad longitudinal gravel bars that were subsequently segmented by narrow canyons.

Specific features interpreted as having been formed by Porcupine River paleofloods include the following:

1. Hexagonally jointed basalt boulders to 3 m in intermediate diameter have been transported more than 1 km from the nearest outcrop.
2. A thin sheet of gray pelitic pebble-cobble gravel was spread uniformly across uneroded loess-mantled fine-grained alluvium in Fishhook Bend (fig. 1), which is the broadest of the expanded reaches.
3. An anastomosing plexus of steep bedrock channels (coulees) of at least 14 m average depth characterizes the lowest 30 km of the canyon.
4. A 700-m-wide amphitheatre-shaped vertical headwall at the mouth of Rapid River in the upper Ramparts (fig. 1) is interpreted as a plunge-pool cataract created by westward drainage (fig. 2).
5. A 20-m-high gravel bar 1 km west of the Canadian border, whose full height is dominated by steeply dipping coarse foreset beds, is similar to the pendant bars described for both the Snake River and Missoula flood tracts.
6. A scour channel 32 m deep, 1 km south of Rapid River that is alternately filled with open-work coarse gravel and matrix-supported boulders may have been emplaced during alternating episodes of scour and deposition from suspension, respectively. Large (about 1 m in diameter) chaotically oriented rip-up clasts of Tertiary clay and lignite fill the lower 15 m of the channel.

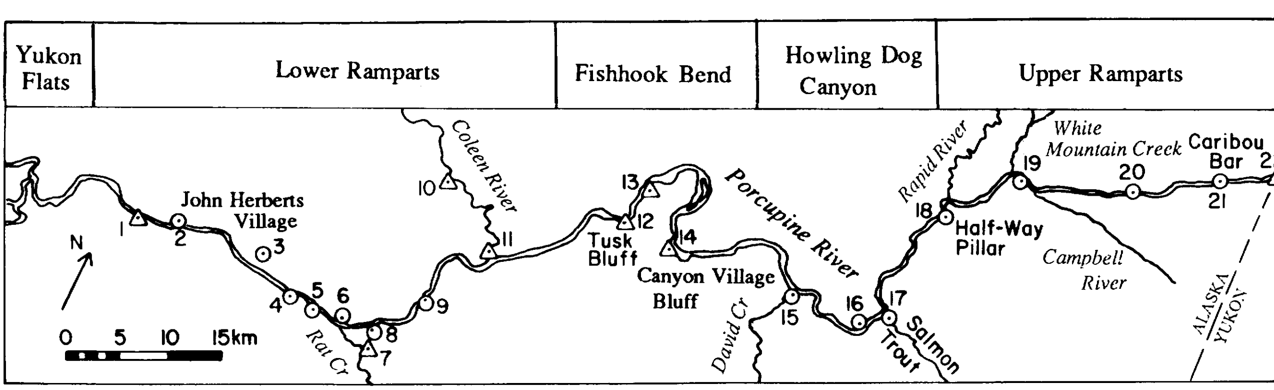


Figure 1. Northeastern Alaska and northern Yukon showing location of study area and geographic features mentioned in text (modified from Thorson and Dixon, 1983). Site locations (circles and triangles) are not mentioned in the text, but are discussed in Thorson and Dixon (1983).

7. Scalloped irregular topography (scabland) characterized by scour depressions occupied by lakes and by eminences up to 7 m in height occurs on many bedrock surfaces below the high-water mark (whaleback eminences at the mouth of Salmon Trout River (fig. 1) are streamlined parallel to the paleoflow direction, transverse to lithologic trends).
8. Evidence for negligible slopes of the high-water mark upstream from tightly constricted reaches may indicate hydraulic impoundment.
9. Stratified sediments characterized by rhythmically interbedded silt and ripple-drift cross-laminated sand occur near the base of Tusk Bluff in Fishhook Bend (fig. 1) and near the mouths of several tributaries. These resemble slack-water sediments reported from other paleofloods.

Preliminary estimates for paleohydrologic parameters were obtained using a simplified version of the slope-area method outlined by Baker (1973, p. 17). A constricted reach at White Mountain Creek (fig. 4) was used to infer channel geometry. Maximum and minimum flow depths were obtained from the vertical interval defined by the cataract and scour channel at Rapid River, respectively. The energy gradient was estimated from the slope of the high-water mark for 10 km down valley from the measured constricted reach. Using conservative (minimizing) estimates for Manning roughness (0.040), energy slope (0.0026), hydraulic radius (26.3 m), and channel area (15,360 m^2), peak discharge could have exceeded 134,000 m^3 s^{-1}. Using the average channel depth of 26 m, and assuming uniform flow, mean velocity should have exceeded 8.7 m s^{-1}. These minimum estimates for discharge and average velocity are comparable to those determined by Malde (1968) for the Bonneville Flood and to the highest peak flows of the Amazon River (Oltman and others, 1964), but are substantially lower than the most conservative estimates for Missoula outburst flooding (Beget, 1986). Use of the DuBoys equation for boundary shear (Baker, 1973, p. 23) and estimates based on unit stream power (Costa, 1983) confirm that strong flows of this magnitude are required to transport the large boulders found in flood alluvium along the Porcupine River.

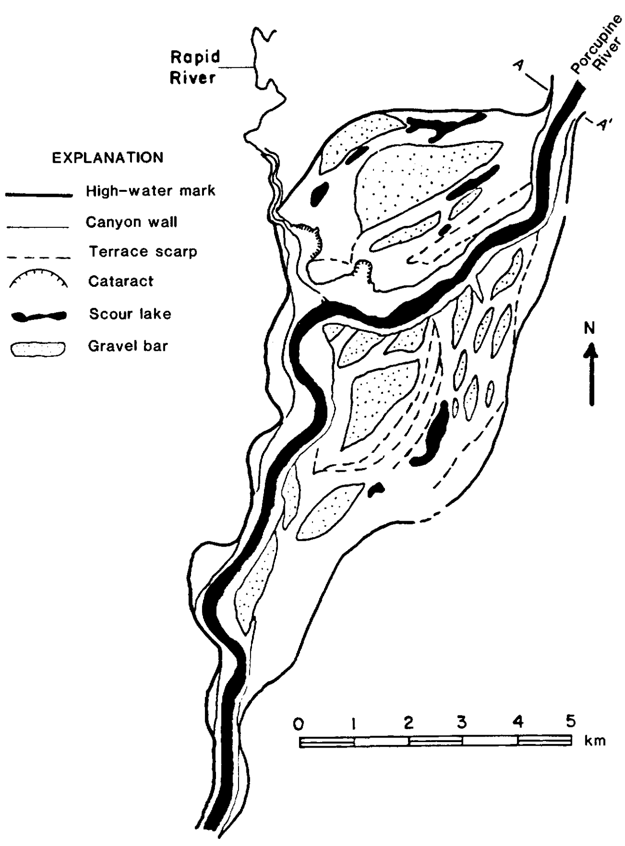


Figure 2. Geomorphic map showing features associated with paleofloods along the Porcupine River near Rapid River. A–A′ shows location of cross-valley profile (fig. 4).

Thorson and Dixon (1983) inferred that the drainage of glacial lakes in the Yukon Territory was required to explain their observations. Determination of the duration of sustained flow from estimates of the lake volume can be used to test this hypothesis. Using the combined area of Old Crow and Bluefish Basins (5,700 km^2) and an average depth of only 30.5 m (100 ft), flow at peak discharges could have been sustained for 15 days. Direct drainage of Keewatin meltwater across the northern Yukon without impoundment is a highly improbable alternative explanation because the flood discharge required greatly exceeds the probable steady-state meltwater flux. For example, assuming a mean ablation rate of 0.5 m/yr (water equivalent of ice), an ablation area larger than all of Canada would have been required to supply the needed discharge.

Flood discharges along the Porcupine River should have had a dramatic effect on downstream areas. The large fan with braided surface morphology below John Herberts Village in the Yukon Flats is at least partially a result of flood deposition, and may have indirectly caused aggradation of rivers draining the southern Brooks Range. Hydraulic ponding at the Ramparts of the Yukon River and at other constrictions may have initiated temporary lacustrine slackwater sedimentation in such alluvial basins. The flux and character of sediment load to the lower Yukon River the continental shelf, and the Bering Sea may also have been impacted. For example, Griggs and others (1970) reported that fluvial gravel resulting from Missoula flooding reached abyssal depths more than 600 km from the outburst flood source. First appearances of retransported palynomorphs (such as the Mississippian-age spores marking disconformity A in the Old Crow Basin (Schweger, this volume)) could be used as marker horizons to supplement lithologic, geomorphic, and isotopic correlations of paleoflood horizons.

Outburst flooding of glacially impounded lakes in the northern Yukon Territory might provide a unique opportunity to link previously disjointed stratigraphic records across the whole of eastern Beringia. Although speculative, possible correlations could be made for several paleoflood episodes linking (1) the westward-advancing Keewatin sector of the Laurentide ice sheet, (2) lacustrine sediments and erosional channels within and between Yukon lakes, (3) the terrace stratigraphy within the middle Porcupine River valley, (4) the Yukon Flats and its south-draining tributaries, (5) altered sedimentation regimes of the continental shelf, and (6) abyssal sedimentation in the Bering Sea. More detailed analysis of Porcupine River outburst flooding is in progress. Many questions remain, particularly those dealing with the number and timing of flood events and with changing channel characteristics.

REFERENCES CITED

Baker, V.R., 1973, Paleohydrology and sedimentology of Lake Missoula flooding in eastern Washington: Geological Society of America Special Paper 144, 79 p.

Baker, V.R., and Nummendal, Dag, 1978, The Channeled Scabland: Washington, D.C., National Aeronautics and Space Administration Planetary Geology Program, 186 p.

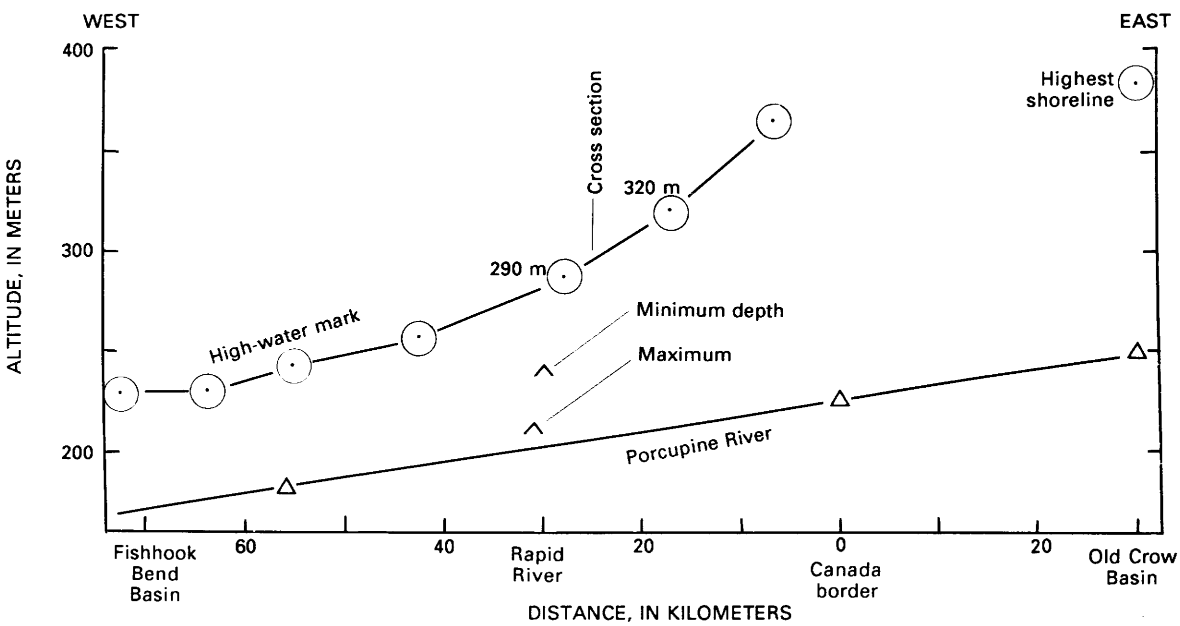


Figure 3. Longitudinal profile of the Porcupine River (lower line) and the high-water mark (upper line) associated with the highest paleoflood. Location of channel cross section A–A′ (fig. 4) and points used to estimate maximum and minimum flow depths shown.

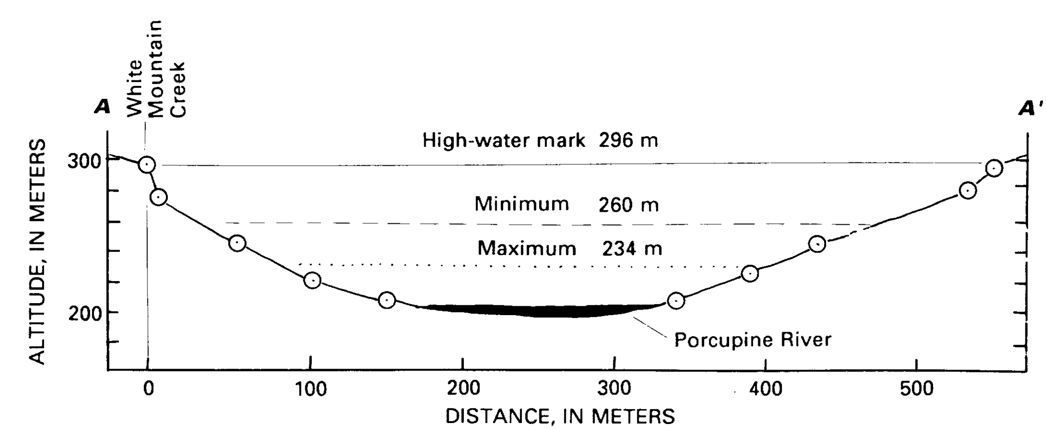


Figure 4. Cross-valley profile of the Porcupine River canyon at White Mountain Creek showing high-water mark and minimum and maximum depths of flood flows. No vertical exaggeration. Location of section shown on figures 1 and 2.

Beget, J.E., 1986, Comment on "Outburst Floods from Glacial Lake Missoula" by Clarke, G.K.C., Mathews, W.H., and Pack, R.T. (letter to the editor): Quaternary Research, v. 25, p. 136–138.

Bretz, J H., 1969, The Lake Missoula floods and the Channeled Scabland: American Journal of Science, v. 77, p. 505–543.

Costa, J.E., 1983, Paleohydraulic reconstruction of flash flood peaks from boulder deposits in the Colorado Front Range: Geological Society of America Bulletin, v. 94, p. 986–1004.

Griggs, B.B., Kulm, L.D., Waters, A.C., and Fowler, G.A., 1970, Deep-sea gravel from Cascadia Channel: Journal of Geology, v. 78, p. 611–619.

Malde, H.E., 1968, The catastrophic late Pleistocene Bonneville flood in the Snake River Plain, Idaho: U.S. Geological Survey Professional Paper 596, 52 p.

Oltman, R.E., Sternberg, H. O'R., Ames, F.C., and Davis, L.C., Jr., 1964, Amazon River investigations—Reconnaissance measurements of July 1963: U.S. Geological Survey Circular 486, 15 p.

Thorson, R.M., and Dixon, E.J., 1983, Alluvial history of the Porcupine River, Alaska: Role of glacial-lake overflow from northwest Canada: Geological Society of America Bulletin, v. 94, p. 576–589.

Carter, L. David, Hamilton, Thomas D., and Galloway, John P., eds., 1989, Late Cenozoic History of the Interior Basins of Alaska and the Yukon: U.S. Geological Survey Circular 1026.

Late Cenozoic Sedimentary History Along Major Fault Zones, Alaska

By Warren Yeend

The presence of small, isolated, commonly fault-bounded outcrops of clastic rocks of known or suspected Tertiary age throughout much of interior Alaska is puzzling. Were these rocks once part of a large blanket deposit covering an extensive area, which subsequently was faulted and eroded to such an extent that only small remnants are left of the original deposit? Or were they deposited only in small basins or trenches such as might occur along faults, and which could explain their common spatial association with major fault zones today? The age of the clastic rocks is rarely well documented, but plant fragments and pollen of Tertiary age are found in fine-grained beds associated with some clastic units. Many of the clastic rocks are gold-bearing and, although of low unit value (less than \$0.10/yd^3 at \$400/oz), appear to have been a source of gold for a few Pleistocene and Holocene placers. Four specific locations of such clastic rocks (fig. 1) are discussed below, and one—the Tintina fault zone in the Circle quadrangle—is suggested as a modern depositional analog for the others.

The Denali fault zone roughly parallels the Alaska Range in the Mount Hayes quadrangle (fig. 2). Tertiary sandstone and conglomerate form scattered outcrops south of the main strand of the Denali fault; these units are gold-bearing and are referred to, at least in the Slate Creek area, as "Round Wash" (Yeend, 1981). The Tertiary clastic rocks are extensively fragmented at this locality. Many have beds that dip up to 60°, are bounded by steeply dipping faults with vertical offsets of more than 400 m, and in places are thrust beneath older rocks. The sandstones are light-colored and are interbedded with conglomerate, gray siltstone, shale, mudstone, sparse white rhyodacite ash, and thin coal layers. Abundant crossbedding, ripple marks, and pebble imbrication imply a high-energy alluvial environment; and the geometry of the deposits suggests that they probably formed as alluvial fans. A potassium-argon hornblende age of 5.4 Ma for interbedded white rhyodacite ash (Bond, 1976; Turner and others, 1980) and plant leaves of *Sequoia* from beds immediately above and below rhyodacitic ash indicate a Pliocene or latest Miocene age (J.A. Wolfe, U.S. Geological Survey, unpub. data cited in Turner and others, 1980) for at least some of these deposits. The conglomerates, which are stratigraphically below the sandstones, have clasts up to 30 cm diameter of both local and foreign rock types. Thin beds of coal are present locally in intercalated sandstone layers. Plant fossils resembling *Alnus evidens* suggest a late Oligocene age (Stout, 1976). Minimum thickness of both the sandstone and conglomerates is on the order of 300 to 500 m (Nokleberg and others, 1982).

Outcrops of Tertiary rocks associated with placer gold deposits are present in the vicinity of the Kaltag fault in the eastern part of the Tanana quadrangle (fig. 3). The Kaltag fault here has probably exerted a strong control on the present position of the Yukon River. Rocks of Tertiary and suspected Tertiary age crop out at several localities on both sides of the fault, although they may be even more prevalent beneath the alluvium-covered lowland next to the river. The high-level gravel associated with the fault is chiefly pebble to small-cobble size with some sand and silt. Clasts are subrounded to well rounded, semi-consolidated, and consist of local rock types. South of the Yukon River the gravel is in part auriferous (Mertie, 1937) and is believed to be of late Tertiary and perhaps early Quaternary age (Chapman and others, 1982). The rocks believed to be Tertiary are poorly to well consolidated, interbedded polymictic pebble conglomerate, boulder conglomerate, grit, and sandstone, with some siltstone, shale, and lignite. Plant fossils and pollen suggest an early Tertiary and

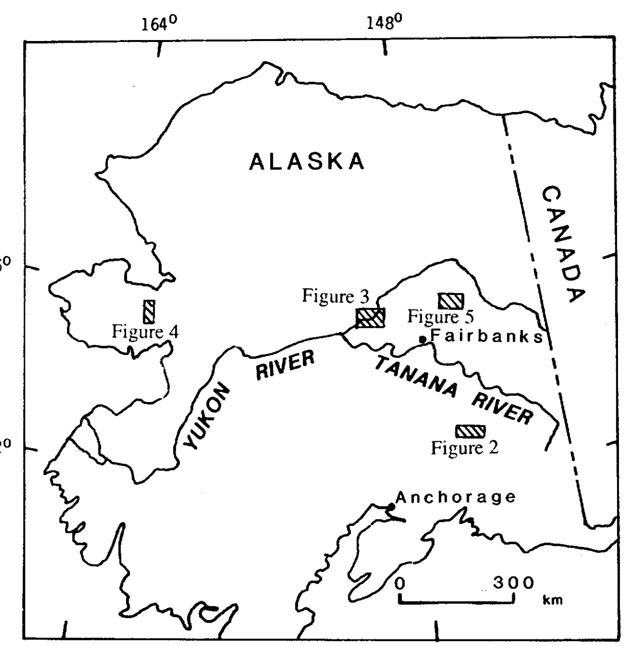


Figure 1. Index map of Alaska showing location of figures 2–5.

Miocene age, respectively (Chapman and others, 1982). Thickness of the rocks in the western part of the map area is possibly as much as 1,500 m (Paige, 1959). These isolated exposures of gently folded Tertiary rocks have definitely been offset by movement along the Kaltag fault.

On the eastern side of the Seward Peninsula, the Kugruk fault zone generally trends north-south and is 10 to 22 km wide (fig. 4). Only a few outcrops of Tertiary and suspected Tertiary aged sediments have been identified within the fault zone, but these rocks probably are more prevalent beneath the Pleistocene and Holocene deposits that cover most of the fault zone. The Tertiary(?) rocks consist of both fine-grained clastics and conglomerate composed of carbonate clasts. The fine-grained clastics that occur in the north half of the fault zone are tan to light-gray siltstone, sandstone, and pebbly sandstone, friable to

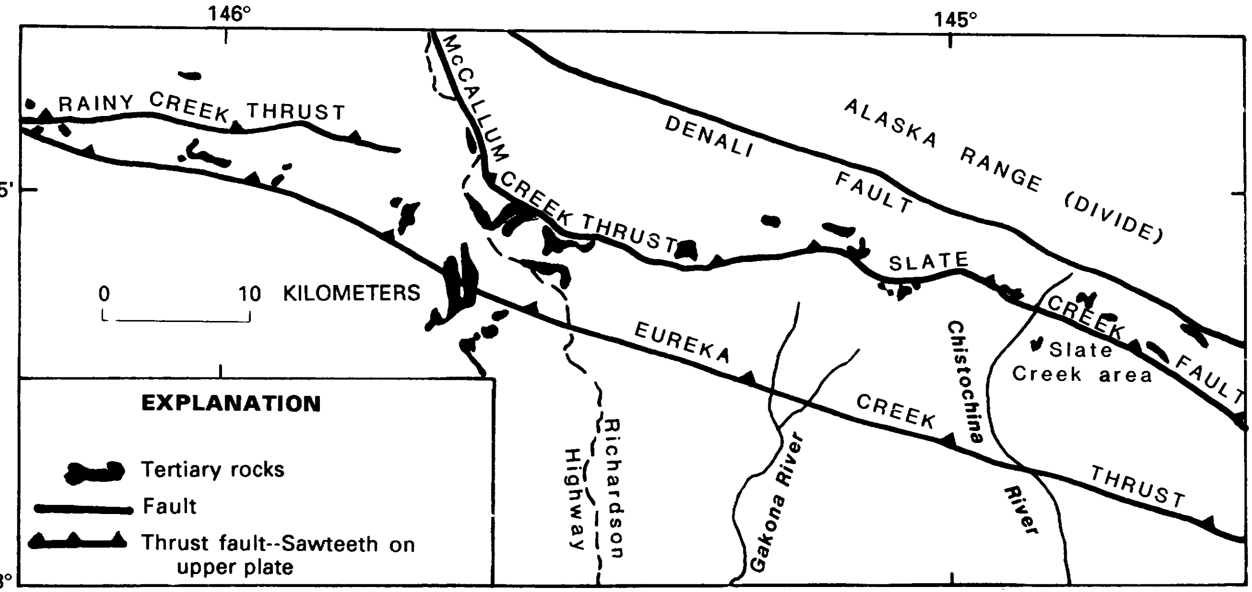


Figure 2. Outcrops of Tertiary rocks along the Denali fault zone, Mount Hayes quadrangle.

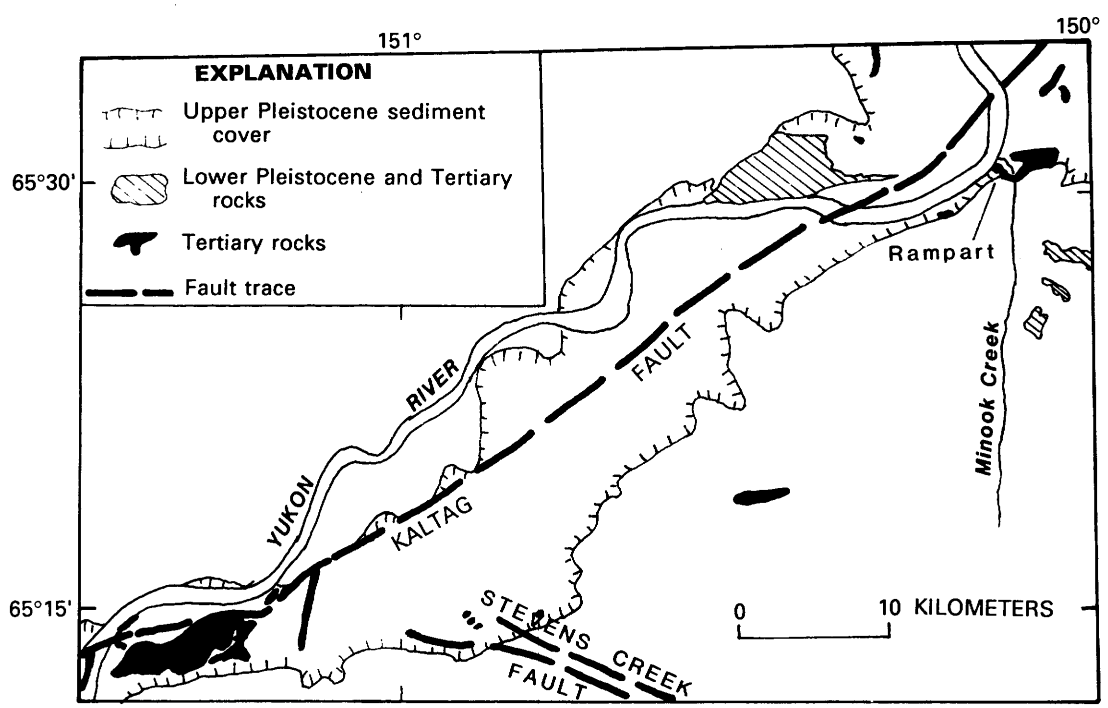


Figure 3. Outcrops of Tertiary and lower Pleistocene rocks along the Kaltag fault zone, Tanana quadrangle.

well-indurated with calcite cement. Beds are 5 to 60 cm thick and commonly are graded; sedimentary structures include channels, small-scale ripples, and crossbeds. Siltstone layers are locally rich in carbonaceous plant debris, and coal seams are present. Clasts are derived from local rocks. Pollen of both Cretaceous and Tertiary (Eocene to early Miocene?) age has been identified from this unit (Till and others, 1986). The conglomerate is poorly sorted, with rounded and subrounded cobbles of carbonate rocks up to 72 cm in diameter. Bedding is rarely discernible, but local sandy interbeds and crude bedding are present. No age-diagnostic fossils have been obtained, although conodonts from the clasts have been assigned an age range of Ordovician to Devonian. Deposition may have been as a series of small alluvial fans (Till and others, 1986). Both of the clastic units are strongly deformed, with dips locally vertical.

The Tintina fault zone is a topographic low about 10 km wide that trends northwest-southeast across the Circle quadrangle (fig. 5). Rocks of suspected Tertiary age crop out rarely; however, it is believed that much of the fault trench is underlain by Tertiary gravel and conglomerate. Recent placer mining within the fault trench in the vicinity of Crooked Creek has exposed orange-colored, deeply decomposed, gold-bearing Tertiary(?) gravel. The gravel is truncated by the Hot Springs fault at the one locality where it is exposed at the fault contact. Elsewhere, Tertiary(?) rocks consist of both conglomerate and sandstone. The conglomerate is well to poorly consolidated, poorly stratified, and consists of well-rounded clasts up to 45 cm in diameter. The sandstone is fine to coarse grained, locally argillaceous, and commonly has a "salt and pepper" appearance; its minimum thickness is 90 m. Poor to moderately well preserved plant material and impressions are present, including *Metasequoia* and broadleaf types. Lignite and coal are found in float associated with the conglomerate and sandstone. Beds dip up to 55 degrees (Foster and others, 1983). The age is uncertain but presumed to be Tertiary (Mertie, 1937).

Tertiary(?), Pleistocene, and Holocene age alluvial fans have been deposited in the Tintina fault trench, in places obscured by overlying loess, alluvium, and undifferentiated silt (fig. 6), and this topographic and geologic environment may be a modern analog of conditions that have existed at some time along many other major fault zones in Alaska. The situation along this portion of the Tintina fault zone—that of a topographic low bounded by topographic high areas—is ideal for the generation of alluvial fan debris which eventually fills the fault trench. In time this topographic situation will be altered through changing fault activity, erosion will occur, and the area will no longer be suitable for receiving sediment. It then might resemble some of the older fault zones where the Tertiary fan detritus has been fragmented and only isolated exposures are apparent. The Denali fault zone where it extends through the Mount Hayes quadrangle is near the divide of the Alaska Range, and is, obviously, not similar topographically to the Tintina fault zone in the Circle quadrangle. Therefore, it is no longer an environment of fan deposition, except very locally. The Kaltag fault zone in the Tanana

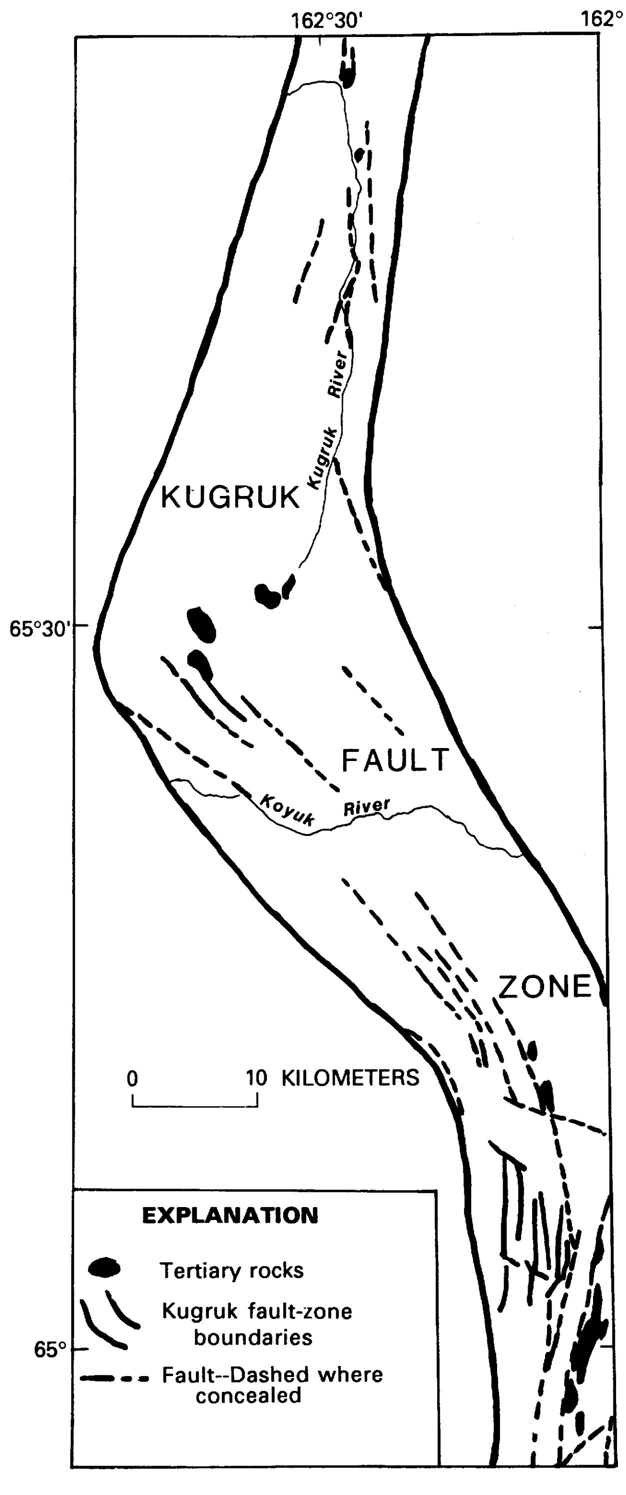


Figure 4. Outcrops of Tertiary rocks within the Kugruk fault zone, Seward Peninsula.

quadrangle is better situated for receiving contemporary fan deposits, as the Yukon River flows in a topographic low that is bounded by highlands. The Kugruk fault zone does not seem to exert much control over topography where it crosses the Seward Peninsula. Drainages cut across the zone as well as run parallel to it, and it is not receiving much sediment except very locally.

REFERENCES CITED

Bond, G.C., 1976, Geology of the Rainbow Mountains-Gulkana Glacier area, eastern Alaska Range, with emphasis on upper Paleozoic strata: Alaska Division of Geological and Geophysical Surveys Geologic Report 45, 47 p.

Chapman, R.M., Yeend, Warren, Brosgé, W.P., and Reiser, H.N., 1982, Reconnaissance geologic map of the Tanana quad-

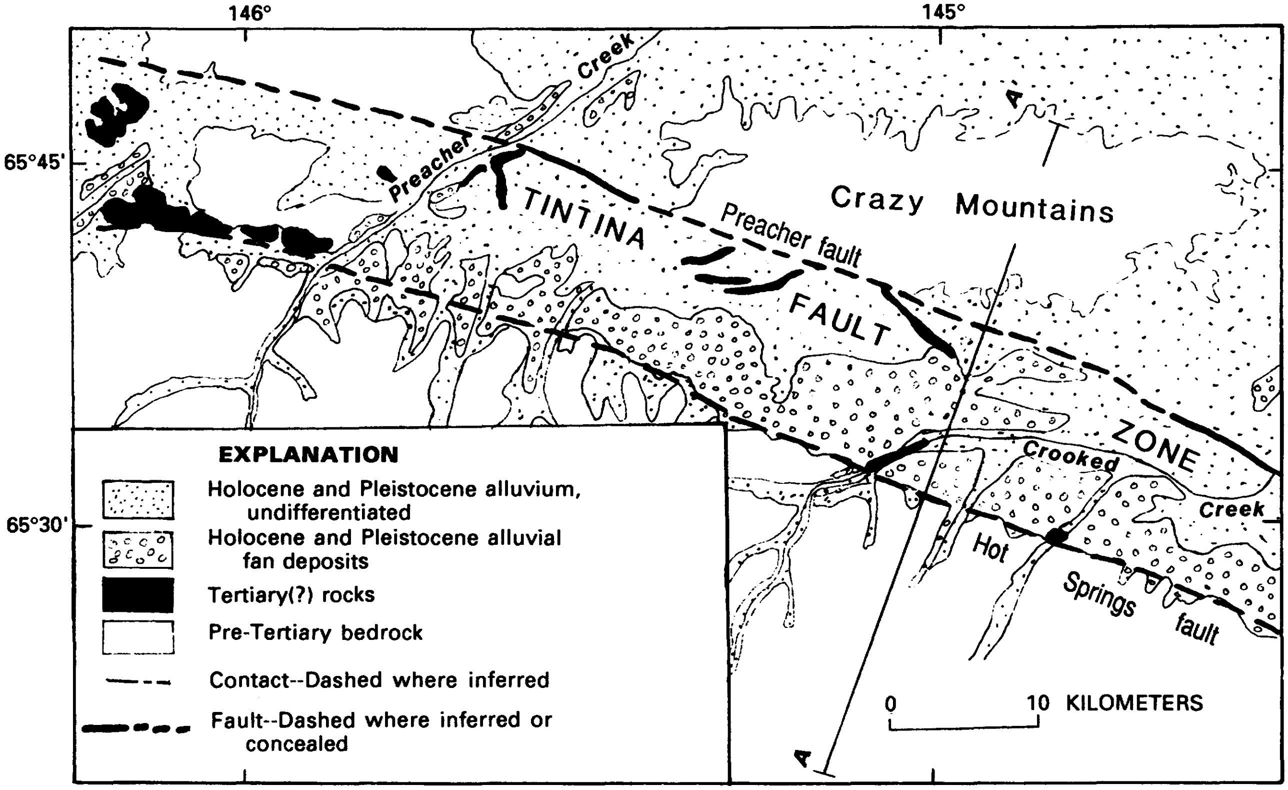


Figure 5. Simplified geologic map of a part of the Tintina fault zone, Circle quadrangle (modified from Foster and others, 1983). A–A', line of cross section in figure 6.

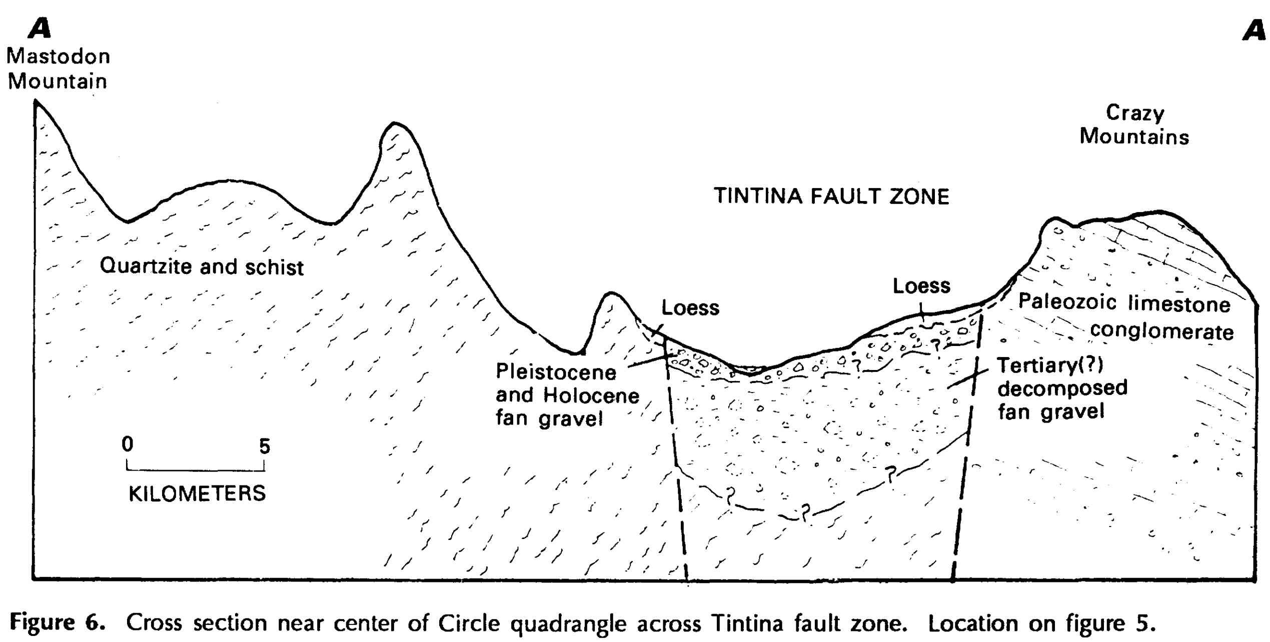


Figure 6. Cross section near center of Circle quadrangle across Tintina fault zone. Location on figure 5.

rangle, Alaska: U.S. Geological Survey Open-File Report 82–734.

Foster, H.L., Laird, Jo, Keith, T.E.C., Cushing, G.W., and Menzie, W.D., 1983, Preliminary geologic map of the Circle quadrangle, Alaska: U.S. Geological Survey Open-File Report 83–170–A, scale 1:250,000.

Mertie, J.B., Jr., 1937, The Yukon-Tanana region, Alaska: U.S. Geological Survey Bulletin 872, 276 p.

Nokleberg, W.J., and others, 1982, Geologic map of the southern Mount Hayes quadrangle, Alaska: U.S. Geological Survey Open-File Report 82–52.

Paige, R.A., 1959, Tertiary geology of the Cheyenne Creek area, Alaska: Seattle, University of Washington, M.S. thesis, 66 p.

Stout, J.H., 1976, Geology of the Eureka Creek area, east-central Alaska Range: Alaska Division of Geological and Geophysical Surveys Geologic Report 46, 32 p.

Till, A.B., Dumoulin, J.A., Gamble, B.M., Kaufman, D.S., and Carroll, P.I., 1986, Preliminary geologic map and fossil data, Solomon, Bendeleben, and southern Kotzebue quadrangles, Seward Peninsula, Alaska: U.S. Geological Survey Open-File Report 86–276, scale 1:250,000.

Turner, D.L., Triplehorn, D.M., Naeser, C.W., and Wolfe, J.A., 1980, Radiometric dating of ash partings in Alaskan coal beds and upper Tertiary paleobotanical stages: Geology, v. 8, p. 92–96.

Yeend, Warren, 1981, Placer gold deposits, Mount Hayes quadrangle, Alaska, *in* Silberman, Miles, and others, eds., Proceedings of the symposium of mineral deposits of the Pacific Northwest: U.S. Geological Survey Open-File Report 81–355, p. 74–83.

Carter, L. David, Hamilton, Thomas D., and Galloway, John P., eds., 1989, Late Cenozoic History of the Interior Basins of Alaska and the Yukon: U.S. Geological Survey Circular 1026.

Paleoglaciology of the Selwyn Lobe of the Cordilleran Ice Sheet and Quaternary Stratigraphy of the East-Central Yukon

By Lionel E. Jackson, Jr.

This paper reports the results of surficial geologic mapping and investigation of the Quaternary stratigraphy in east-central Yukon carried out between 1980 and 1986 by Jackson (1982, 1986a, 1986b, 1987a, 1987b), Jackson and others (1984, 1985), and Duk-Rodkin and others (1986), and by previous workers in the same or contiguous areas (Campbell, 1967; Klassen, 1978, 1982; Klassen and Morison, 1981; Dyke, 1981). The region discussed includes an area which was covered by the Selwyn lobe of the Cordilleran ice sheet during the McConnell glaciation (Hughes and others, 1969). The morphology, modes of flow, and deglaciation of the ice sheet are discussed along with regional Quaternary stratigraphy and chronology.

Setting

Figure 1 shows the context of the Selwyn lobe within the limits of the McConnell glaciation and older glaciations in southern Yukon Territory, and figure 2 outlines the portion of the region formerly covered by the Selwyn lobe for which data exist to permit its detailed reconstruction. This region includes the Glenlyon Range, the MacMillan River basin sector of the Yukon Plateau, parts of the Tintina Trench and Pelly Mountains, and the Selous Mountains, which are part of the Selwyn Mountains.

The surface of the Yukon Plateau within the MacMillan River basin is incised by two sets of valleys. Broad southwest-trending valleys were incised during preglacial times, and are occupied predominantly by underfit streams. Narrower northwest-trending valleys were created by glacial and glaciofluvial erosion during the Quaternary and contain major streams. The result is a pattern of anastomosing valleys surrounding island-like table lands that are surmounted by small mountain ranges. Valley bottoms range in elevation from 640 to 760 m, the plateau surface from 1,070 to 1,370 m. Mountain summits may have an additional 600 m of relief with maximum elevations up to about 2,130 m.

Evidence of past glaciation is ubiquitous. Plateau areas and the sides of larger valleys are marked by whaleback ridges and crag-and-tails which allow past ice-flow directions to be easily determined. Valley bottoms contain complexes of glaciofluvial deposits or thick fills of glaciolacustrine sediments. The gravelly floors of channels that were walled by ice form flights of terraces that commonly extend from mountain summits to valley bottoms.

The valleys are extensively mantled by boreal forest whereas much of the plateau surfaces lie above tree line and are vegetated by thick brush. Alpine areas are vegetated by tundra or are barren. The entire region is within the zone of discontinuous permafrost.

Paleoglaciology of the Selwyn lobe

The Selwyn lobe is the name given by Hughes and others (1969) to a sector of the Cordilleran ice sheet that flowed west to northwest across 5 degrees of longitude and between latitudes 61° and 64° N. in east to east-central Yukon Territory (fig. 1). It was separated from the Cassiar lobe on the south by the Pelly Mountains. The two lobes interacted near their termini. The Selwyn lobe shared divides with glaciers flowing east and south into the Mackenzie River Valley. The largest of these flowed into the Liard River basin and is named the Liard lobe.

During the McConnell glaciation, isolated mountain summits stood above the Selwyn lobe as nunataks (Campbell, 1967; Duk-Rodkin and others, 1986). Glacial deposits surrounding these former nunataks have a fresh morphology characteristic of deposits of McConnell age elsewhere in the territory, in contrast to the subdued morphology typical of deposits of Reid and older glaciations (Bostock, 1966; Hughes and others, 1969). The maximum McConnell terminus (fig. 2) is well defined by moraines and incised ice-marginal channels.

Duk-Rodkin and others (1986) constructed a generalized profile of the distal part of the Selwyn lobe (fig. 3), based upon geomorphic features indicative of ice limits around former nunataks. Profile slopes were 0°30' over the outer 50 km, 0°19' over the next 53 km, and 0°03' farther east. The profile of the Selwyn lobe was flatter than profiles of the contemporary Greenland and Antarctic ice sheets (Paterson, 1981, p. 153–167). The best numerical approximation to the measured profile is the curvilinear profile of a valley glacier modeled by Shilling and Hollin (1981) after Nye (1952a, 1952b). This finding is not surprising, because flow in the Selwyn lobe was similar to flow in a valley glacier; shear stress occurred not only along the sole of the ice sheet but along the sides of high-relief roughness elements (valley and mountain sides). Flow within the ice sheet, based upon ice flow indicators such as whalebacks and crag-and-tails, occurred as anastomosing ice streams that followed major east-west-trending valleys such as the Tintina trench and the MacMillan River valley (fig. 2). Ice flow was entirely channeled by underlying topography rather than predominantly independent of it, as is largely the case in the contemporary ice sheets of Greenland (Flint, 1971, p. 51–54) and Antarctica (Denton and others, 1971).

Deglaciation of the Selwyn lobe

The Selwyn lobe disappeared rapidly at the close of the McConnell glaciation through a combination of down-wasting and stagnation. Many cirques in the Selwyn and Pelly Mountains have ice-stagnation landforms on their floors which are continuous with similar features in adjacent valleys. It can be concluded from these relationships that ice stagnated in these cirques during or before stagnation in adjacent valleys. These stagnation features occur up to altitudes of 1,830 m. Consequently, it appears that early in the deglaciation of the Selwyn lobe, the equilibrium line rose significantly above an altitude of 1,830 m and remained above it until the present day (Jackson, 1986b). This rise in the equilibrium line above the ice sheet resulted in the wholesale starvation of the Selwyn lobe. The resulting thinning of the ice sheet is documented along many valley sides by flights of terraces that once were the floors of ice-walled channels. Ice in valleys oriented transverse to

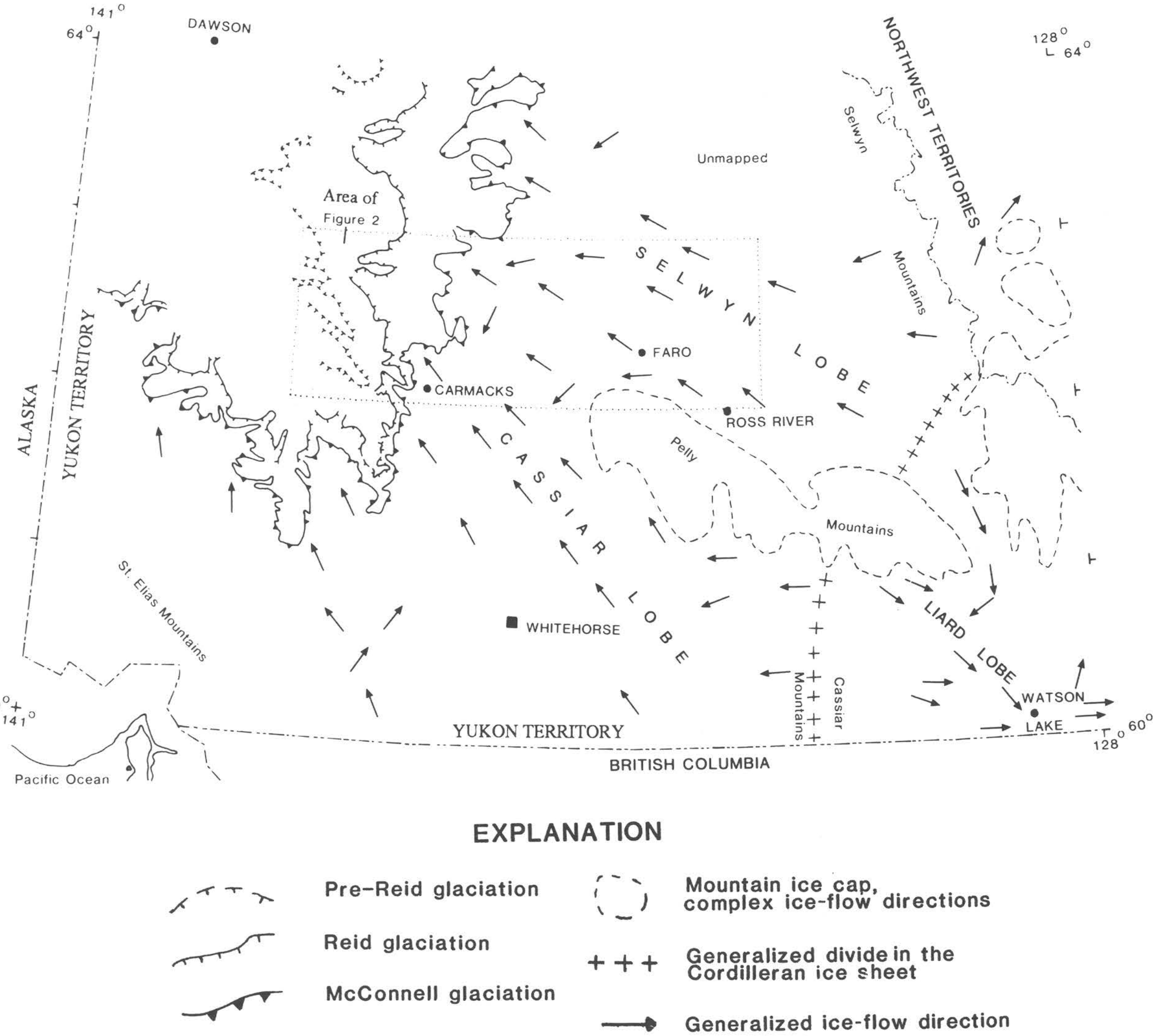


Figure 1. Glacial limits and ice divides in southern Yukon Territory.

the general direction of glacial flow stagnated between ice streams and gave rise to extensive complexes of stagnation landforms.

A succession of progressively lower and younger lakes was ponded within the Pelly Mountains and mountain ranges to the north during the period of downwasting of the Selwyn and Cassiar lobes (figs. 4A and 4B).

Quaternary stratigraphy and chronology

The deposition of Quaternary sediments within the region formerly occupied by the Selwyn lobe was controlled chiefly by topography. Lodgement and meltout tills are the dominant sediments on plateau remnants; ice-stagnation gravels are extensive in valleys oriented transverse to former ice flow, and glaciolacustrine sediments occur in valleys which were fortuitously blocked by ice during deglaciation. Only the Tay and Pelly River valleys (Tintina Trench) were sufficiently wide and deep enough to accumulate and preserve sediments representing more than one glacial interval. Figure 5 summarizes the stratigraphy of the most informative sections within the region covered by the Selwyn lobe. Of these, sections along the Pelly River and Lapie Canyon (14686 S–1 and 7686 S–2, respectively, on fig. 4A), provide the most complete record of depositional and erosional events of two glacial intervals. The age of the lowest till (PM) is unknown. The recession following deposition of the lowest till was marked by glaciolacustrine sedimentation in both the Tay River and Pelly River valleys (sections 23685 C–3 and 14686 S–1, fig. 4B). In the Pelly River valley, glaciolacustrine sediments are unconformably overlain by horizontally stratified gravel deposited during a subsequent period of fluvial aggradation and prior to the McConnell glaciation. Gravel beds in an

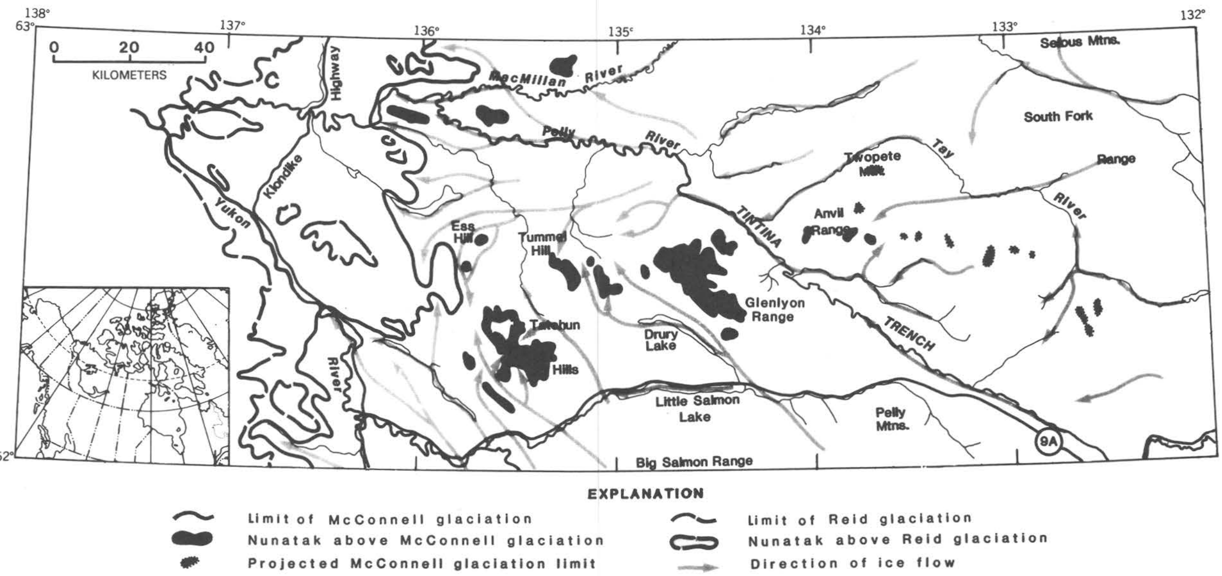


Figure 2. Generalized glacial limits, Carmacks (115 I), Glenlyon (105 L), and Tay River (105 K) map areas, Yukon Territory.

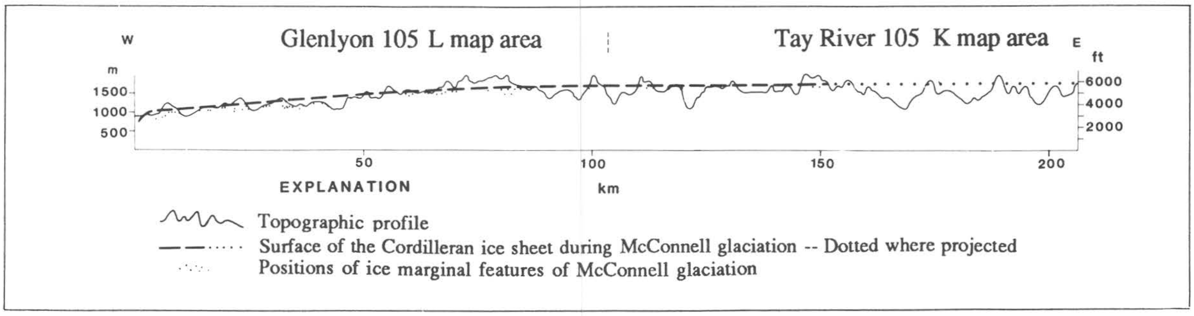
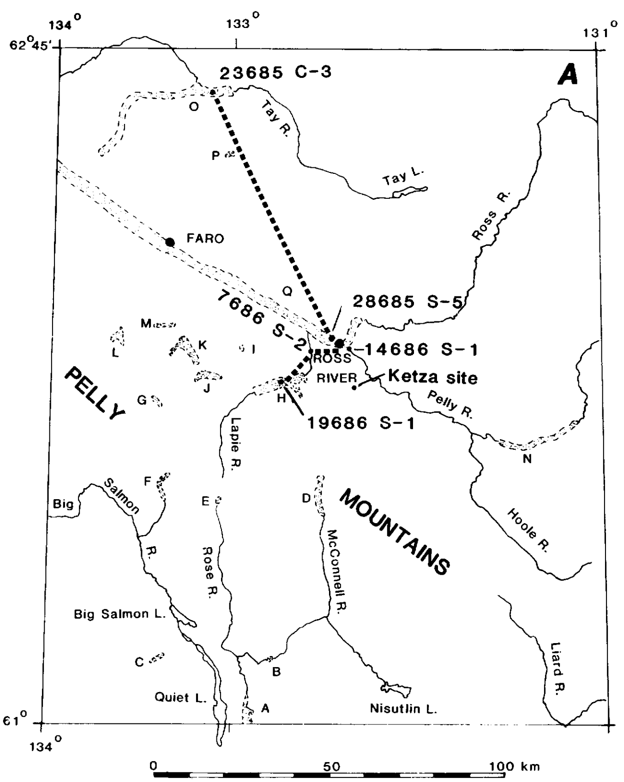


Figure 3. Generalized topographic profile oriented east-west across area of figure 2 showing reconstructed upper limit of the McConnell glaciation (from Duk-Rodkin and others, 1986). Profile constructed by projecting eight separate east-west profiles onto single plane. Vertical exaggeration x 6.

analogous stratigraphic position below McConnell-age till at the Ketza site (fig. 4A) yielded bone fragments from *Bison* sp. (perhaps *B. priscus*) and a cheek tooth from *Mammuthus primigenius* (identified by C.R. Harington, Canada National Museum of Natural Sciences, written commun., 1986). A radiocarbon date on a *B. priscus* radius fragment from this site yielded an age of 26.35±0.28 ka (TO–393). This provides a maximum age for the overlying McConnell till (M in fig. 5).

The succession above the McConnell-age till is essentially a repeat of stratigraphy seen above the lower till: glaciolacustrine sedimentation in the Pelly and Tay River valleys and massive to horizontally bedded gravelly outwash in adjacent areas. Radiocarbon dates from the continental divide in the Selwyn Mountains indicate that ice had disappeared from the divide area before about 9 ka (MacDonald, 1983; Jonasson and others, 1983).

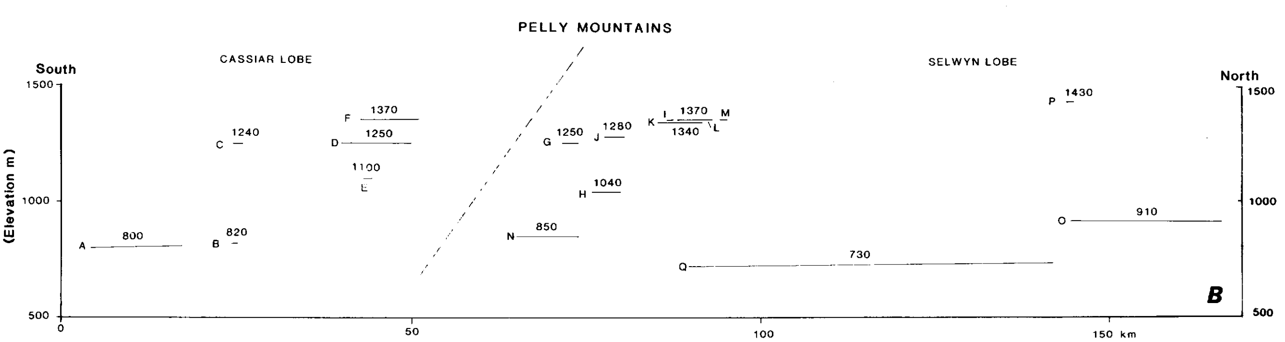


Holocene events

As has been demonstrated elsewhere in glaciated regions of the Cordillera, geomorphic events during the Holocene have reflected the readjustment of slopes and streams from a glacial to a nonglacial regime (Ryder, 1971a, 1971b; Church and Ryder, 1972; Jackson and others, 1982; Ritter and Ten Brink, 1986). Alluvial fans built predominantly of debris-flow diamictons are ubiquitous Holocene landforms. These were graded to braided flood plains. As glacially deposited sediments were progressively eroded from stream basins, sediment loads decreased and debris-flow activity waned. Paraglacial fans became inactive and were incised from head to toe as braided stream courses were incised. Incision of outwash plains in the Pelly River valley began before 8 ka (Jackson, unpub. data). Near the town of Ross River, the Pelly River had incised to 14 m above its present flood plain, by 5.92±0.7 ka (TO–196). The White River Ash Bed (Péwé, 1975, p. 79–81), which formed about 1.4 ka, is within the overbank sediment of the present floodplain. At some point between about 6 and 1 ka, the Pelly River changed from a braided planform to a meandering one. This change in channel planform in the post middle-Holocene has been documented elsewhere in the Cordillera (Jackson and others, 1982).

The initiation and spread of blanket bog throughout the region is second only to paraglacial sedimentation in its modification of the land surface during the Holocene. Bog bottom dates from Selwyn Mountains indicate that blanket-bog expansion was well under way by 8.8 ka (Jackson, 1987b).

Figure 4. Distribution and surface elevations of glaciolacustrine sediments deposited in glacial lakes during deglaciation of Selwyn and Cassiar lobes in areas adjacent to Pelly Mountains. *A*, Distribution of sediments (patterned areas). Heavy dashed line and numbered localities show plane of stratigraphic section in figure 5. Single letters designate glaciolacustrine surface elevations shown in *B*. *B*, Surface elevations of glaciolacustrine sediments shown in *A* (letters) projected onto north-south plane at longitude 133° W. Dashed line separates lake basins under influence of Selwyn lobe from those under influence of Cassiar lobe.



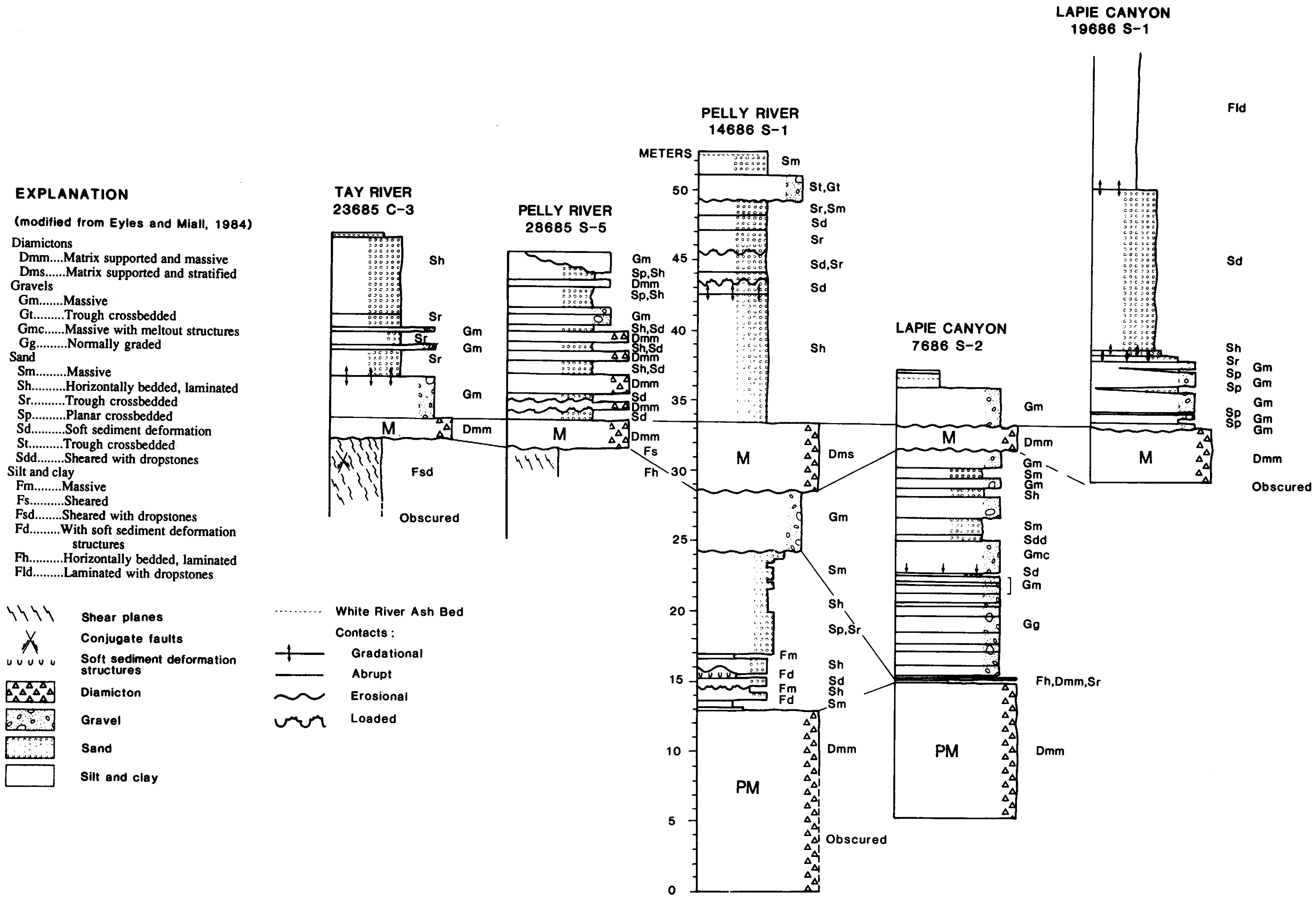


Figure 5. Lithostratigraphic cross section from Tay River valley to Lapie River valley (line of section shown in fig. 4*A*). Section constructed using top of till of McConnell glaciation (M) as datum. PM denotes till that predates pre-McConnell glaciation.

REFERENCES CITED

Bostock, H.S., 1966, Notes on glaciation in central Yukon Territory: Geological Survey of Canada Paper 65-56, 18 p.

Campbell, R.B., 1967, Reconnaissance geology of Glenlyon map area, Yukon Territory: Geological Survey of Canada Memoir 352, 92 p.

Church, M., and Ryder, J.M., 1972, Paraglacial sedimentation: A consideration of fluvial processes conditioned by glaciation: Geological Society of America Bulletin, v. 83, p. 3072–3095.

Denton, G.H., Armstrong, R.L., and Stuiver, M., 1971, The late Cenozoic glacial history of Antarctica, *in* Turekian, K.K., ed., The late Cenozoic glacial ages: New Haven, Yale University Press, p. 267–306.

Duk-Rodkin, A., Jackson, L.E., Jr., and Rodkin, O., 1986, A composite profile of the Cordilleran ice sheet during McConnell Glaciation, Glenlyon and Tay River map areas, Yukon Territory: Geological Survey of Canada Paper 86–1B, Current Research, Part B, p. 257–262.

Dyke, A.S., 1981, Surficial geology of Francis Lake map area (105 H): Geological Survey of Canada Open File Report 898, scale 1:125,000.

Eyles, N., and Miall, A.D., 1984, *in* Walker, R.G., ed., Facies models, Geoscience Canada reprint series 1: Toronto, Geological Association of Canada, Glacial facies, p. 15–38.

Flint, R.F., 1971, Glacial and Quaternary geology: New York, Wiley, 892 p.

Hughes, O.L., Campbell, R.B., Muller, J.E., and Wheeler, J.O., 1969, Glacial limits and flow patterns, Yukon Territory south of 65 degrees north latitude: Geological Survey of Canada Paper 68–34, 9 p.

Jackson, L.E., Jr., 1982, Terrain conditions, Nahanni (105 I): Geological Survey of Canada Open File Report 886, scale 1:125,000.

———1986a, Terrain inventory Finlayson Lake map area (105 G), Yukon Territory: Geological Survey of Canada Open File Report 1379, scale 1:125,000.

———1986b, Terrain inventory, Tay River map area (105 K): Geological Survey of Canada Open File Report 1380, scale 1:125,000.

———1987a, Terrain inventory, Quiet Lake map area (105 F): Geological Survey of Canada Open File Report 1536, scale 1:125,000.

———1987b, Terrain inventory and Quaternary history of the Nahanni map area, Yukon and Northwest Territories: Geological Survey of Canada Paper 86–18, 23 p.

Jackson, L.E., Jr., MacDonald, G.M., Foscolos, A.E., and Clarke, A.H., 1985, An occurrence of pre-McConnell nonglacial sediments, Selwyn Mountains, Northwest Territories: Geological Survey of Canada Paper 85–1B Current Research, Part B, p. 169–175.

Jackson, L.E., Jr., MacDonald, G.M., and Wilson, M.C., 1982, Paraglacial origin for terraced river sediments in Bow Valley, Alberta: Canadian Journal of Earth Sciences, v. 19, p. 2219–2231.

Jackson, L.E., Jr., Morison, S.R., McKenna, K., and Duk, A., 1984, Terrain inventory, Sheldon Lake map area (105 J): Geological Survey of Canada Open File Report 1033.

Jonasson, I.R., Jackson, L.E., and Sangster, D.F., 1983, A Holocene zinc orebody formed by the supergene replacement of mosses: Journal of Geochemical Exploration, v. 18, p. 189–194.

Klassen, R.W., 1978, A unique stratigraphic record of late Tertiary-Quaternary events in southeastern Yukon: Canadian Journal of Earth Sciences, v. 15, p. 1884–1886.

———1982, Surficial geology Wolf Lake map area (105 A): Geological Survey of Canada Map 14–1982, scale 1:250,000.

Klassen, R.W., and Morison, S.R., 1981, Surficial geology Watson Lake (105 B): Geological Survey of Canada Map 21–1981.

MacDonald, G.M., 1983, Holocene vegetation history of the upper Natla River area, Northwest Territories, Canada: Arctic and Alpine Research, v. 15, p. 169–180.

Nye, J.F., 1952a, The mechanics of glacier flow: Journal of Glaciology, v. 2, p. 103–107.

———1952b, A comparison between the theoretical and the measured long profile of the Unteraar Glacier: Journal of Glaciology, v. 2, p. 103–107.

Paterson, W.S.B., 1981, The physics of glaciers (2d ed.): New York, Pergamon Press, 380 p.

Péwé, T.L., 1975, Quaternary geology of Alaska: U.S. Geological Survey Professional Paper 835, 145 p.

Ritter, D.F., and Ten Brink, N.W., 1986, Alluvial fan development and the glacial-glaciofluvial cycle, Nenana Valley, Alaska: Journal of Geology, v. 94, p. 613–625.

Ryder, J.M., 1971a, The stratigraphy and morphology of paraglacial alluvial fans in south central British Columbia: Canadian Journal of Earth Sciences, v. 8, p 279–298.

———1971b, Some aspects of morphology of paraglacial alluvial fans in south central British Columbia: Canadian Journal of Earth Sciences, v. 8, p. 1252–1264.

Shilling, D.H., and Hollin, J.T., 1981, Numerical reconstructions of valley glaciers and small ice caps, *in* Denton, G.H., and Hughes, T.J., eds., The last great ice sheets: New York, Wiley-Interscience, p. 207–220.

Geological Survey of Canada Contribution 17487

Carter, L. David, Hamilton, Thomas D., and Galloway, John P., eds., 1989, Late Cenozoic History of the Interior Basins of Alaska and the Yukon: U.S. Geological Survey Circular 1026.

Late Cenozoic Stratigraphy and Sedimentology of Gravelly Deposits in Central and Southern Yukon

By Stephen R. Morison

Placer deposit studies in west-central and southern Yukon have shown that discrete periods of sedimentation of gold-bearing gravel occurred in five generalized settings (table 1). These settings occurred in both unglaciated terrain and in glaciated terrain during interglacial and glacial intervals, and thus form important components of the late Cenozoic history of the Yukon. This paper represents a first attempt at relating types of fluvial gravelly sedimentation to the late Cenozoic history of the Yukon. It therefore is very general and forms only a working framework.

Pliocene to early Pleistocene

Pliocene to lower Pleistocene gravelly sequences are preserved as high-level terraces in unglaciated terrain and areas within the limits of pre-Reid glaciations. Examples include terraces proximal to the upper limit of glaciation in the Clear Creek drainage basin and the high-level terraces in the Klondike placer district (fig. 1).

In the Clear Creek drainage basin, preglacial fluvial gravel is found stratigraphically below till units that were

Table 1. Stratigraphy and general characteristics of placer gold deposits in Canada (from Morison, in press)

AGE	TERTIARY	QUATERNARY			
	PLIOCENE	PLEISTOCENE			HOLOCENE
ENVIRONMENT AND GEOMORPHIC LOCATION OF PLACERS	Buried alluvial sediments in benches above valley floors	Preglacial or nonglacial buried alluvial sediments in benches above valley floors; valley fill alluvial sediments; alluvial terraces	Interglacial valley fill alluvial sediments; alluvial terraces	Glacial benches of proglacial and ice contact deposits; terminal valley moraines and alpine drift	Valley bottom alluvial plains and terraces; colluvium and slope deposits
GENERAL SEDIMENT CHARACTERISTICS	Mature sediments; well-sorted alluvium with a diverse assemblage of sediment types	Locally derived gravel lithology; moderately to well sorted alluvium which is crudely to distinctly stratified	Mixed gravel lithology; moderately to well sorted alluvium, crudely to distinctly stratified	Regionally derived gravel lithology; variable sorting and stratification depending upon type of glacial drift	Mixed gravel lithology; moderately to well sorted alluvium, crudely to distinctly stratified; poorly sorted, massive slope deposits
GOLD DISTRIBUTION	Greater concentration with depth	Discrete concentrations throughout to pay streaks at base of alluvium	Discrete concentrations throughout to pay streaks at base of alluvium	Dispersed throughout	Discrete concentrations throughout to pay streaks at base of alluvium; pay streaks follow slope morphology
MINING PROBLEMS	Thick overburden	Thick overburden; variable grade	Variable grade	Low grade and larger volume of material	Variable grade and small volume of auriferous sediment
EXAMPLES	"White Channel Gravel" of the Klondike area Yukon Territory	Preglacial fluvial gravels, Clear Creek drainage basin and unglaciated terrain and Sixtymile River area, Yukon Territory	Interglacial stream gravels in Atlin and Cariboo areas, British Columbia	Glaciofluvial gravel in Clear Creek drainage basin, Yukon Territory	Valley bottom creek and gulch placers in Clear Creek drainage basin, Yukon Territory

deposited prior to the Reid glaciation (fig. 2). The stratigraphy and sedimentology of this sequence suggests that gold-bearing fluvial gravel was buried by melt-out till, possibly basal, which was subsequently covered by redeposited ice-marginal sediments (Morison, 1985a). Glacial sediments in this sequence are not auriferous, which demonstrates minimal subglacial erosion of underlying fluvial gravel and supports deposition through meltout of the overlying drift.

High-level terraces in the Klondike placer area are composed of White Channel placer gravel deposits that are overlain by non-auriferous sequences of either the Klon-

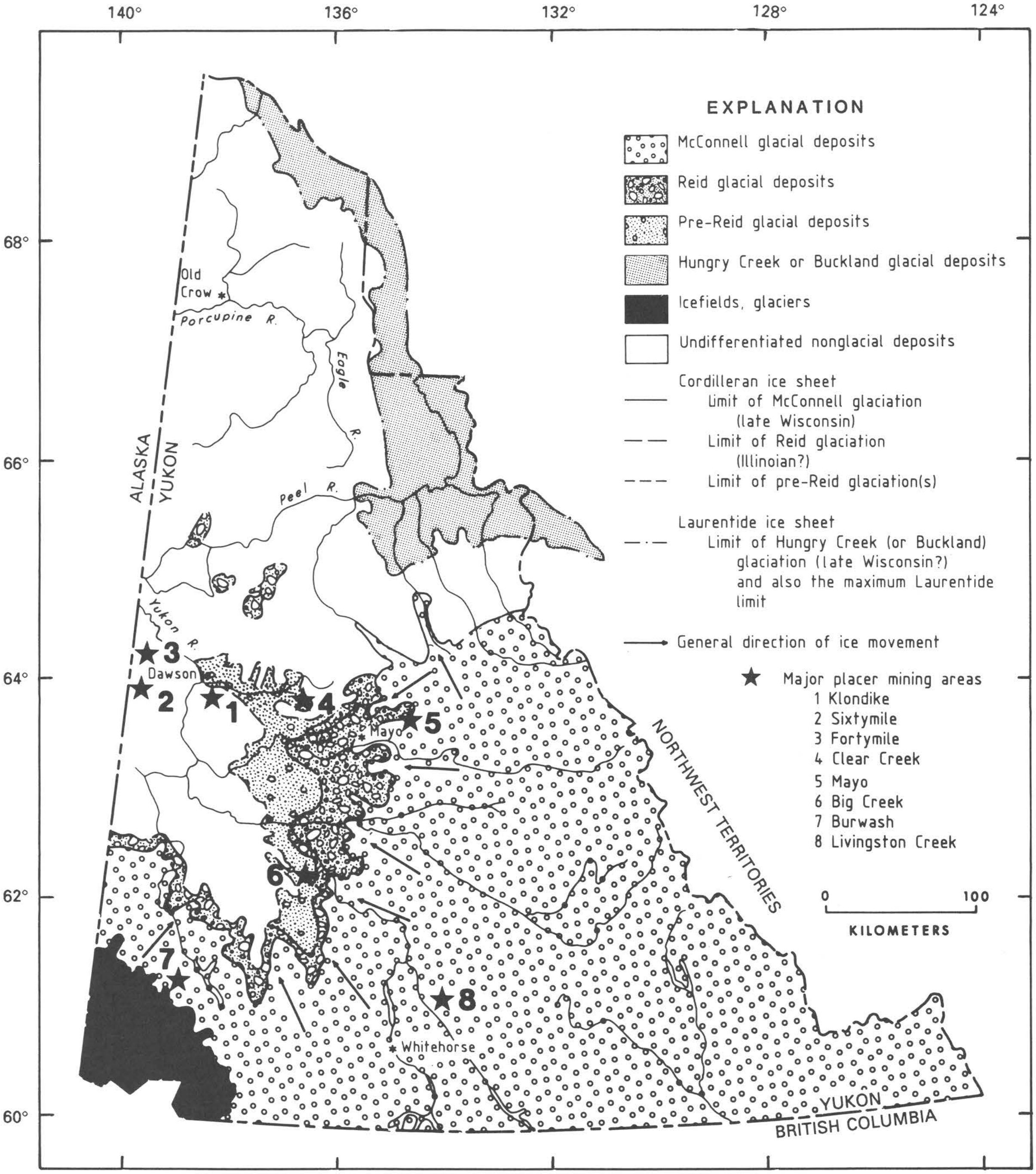


Figure 1. Glacial limits in Yukon Territory (modified from Hughes and others, 1983, and Tarnocai, 1987).

dike Gravel (McConnell, 1907), which is glaciofluvial in origin (Hughes and others, 1972), or younger Pleistocene fluvial gravel such as the iron-stained sequence in the lower part of the Hunker Creek drainage basin (Dufresne and Morison, 1985; Morison, 1985b). White Channel alluvium is characterized by 14 lithofacies types which range from laminated silt and clay to massive and disorganized boulder gravel (Morison, 1985b). Cryogenic features such as ice-wedge casts are found in the upper part of the White Channel sequence and indicate climatic cooling, which may coincide with the Pliocene-Pleistocene boundary. White Channel alluvium was deposited in a braided river environment (fig. 3) that included the following main components (Morison and Hein, 1987): (1) a main valley proximal setting which was dominated by high discharge and flood channels in a crudely braided environment; (2) main valley medial to distal settings which formed well-defined braided sequences of channels and low-relief bars; and (3) debris flows from valley walls in association with either tributary alluvial fan sedimentation or as independent gravity flows.

Pleistocene

Gravel deposits in unglaciated terrain

In unglaciated terrain, placer gravel deposits are found as valley-bottom fill and low to high terraces. An example is the Miller Creek area in the Sixtymile River placer area (fig. 1), where four distinct stages in drainage-basin evolution have been identified (Hughes, 1986; Hughes and others, 1987). These stages (fig. 4A–D) consist of: (1) downcutting and tributary-valley widening; (2) main valley braidplain sedimentation and alluvial fan aggradation, which was completed before 40 ka (GSC–3934 and GSC–4032); (3) debris-flow sedimentation, which began approximately 26 ka (Beta–13870), and a change in the main valley channel pattern from braided to meandering;

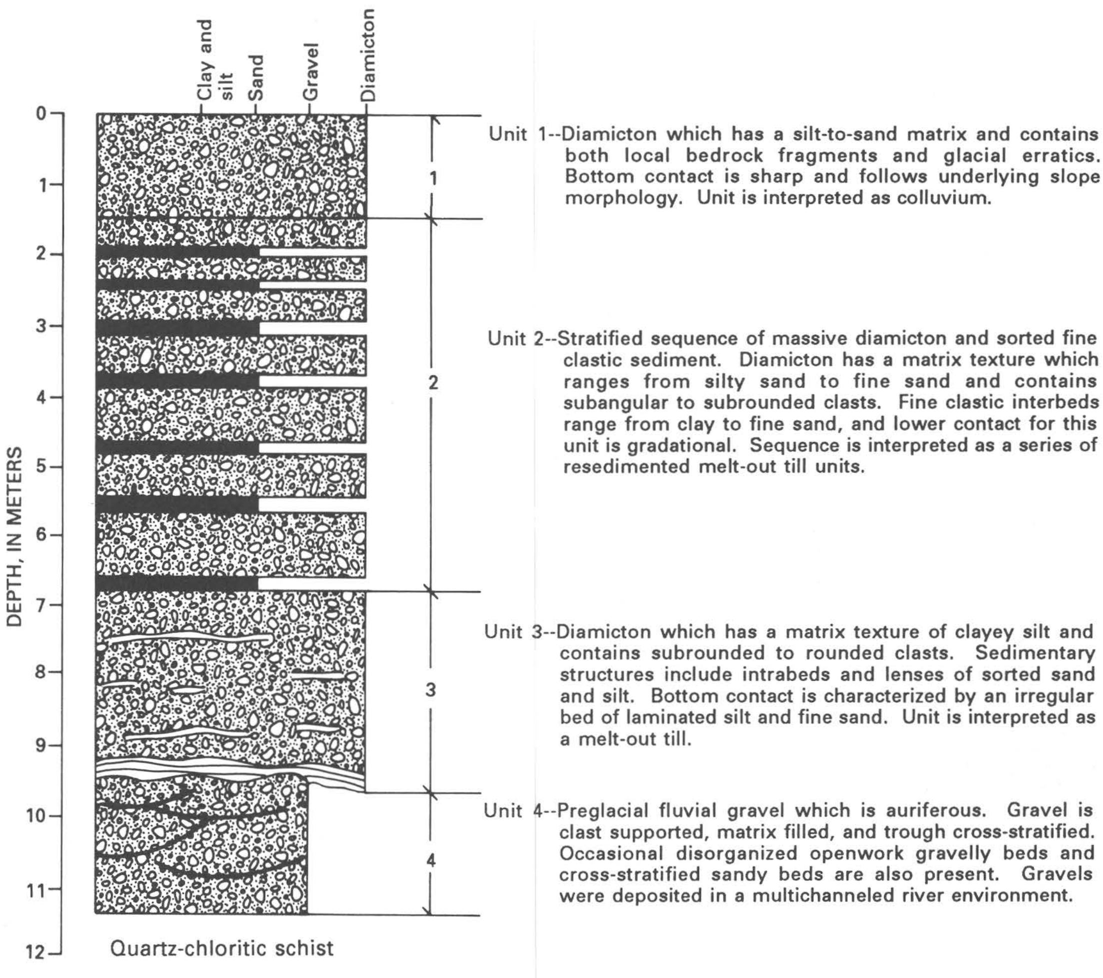


Figure 2. Stratigraphy and sedimentology of buried placer deposit at mouth of Left Clear Creek, Yukon. Sediment texture shown diagrammatically by relative horizontal extent of units.

(4) loess and colluvial sedimentation, and establishment of the modern Sixtymile River flood plain. Although it is difficult to estimate the age of incision and the development of gravelly terraces, major changes in stream channel patterns are thought to indicate rejuvenated downcutting through gravelly deposits.

Interglacial gravel deposits

In glaciated terrain, drainage basins that were not eroded by subsequent Cordilleran glacial advances contain placer gold deposits that were concentrated through resedimentation by interglacial streams. Examples are not well documented; however, in the Livingstone Creek placer area (fig. 1), gold-bearing gravelly alluvium is found below drift of the McConnell glaciation (Bostock and Lees, 1960). Also, valley-bottom gravelly fill was deposited and preserved in areas of pre-Reid glaciation, such as the Big Creek drainage basin (placer area 6, fig. 1), which were not occupied by ice of the subsequent Reid and McConnell glaciations.

Significant stream entrenchment occurred in the Klondike placer area (fig. 1) between the Old (pre-Reid?) and Intermediate (Reid?) glaciations of the southern Ogilvie Range (Hughes and others, 1972, 1983; and Vernon and Hughes, 1966). Following deposition of drift from the Old

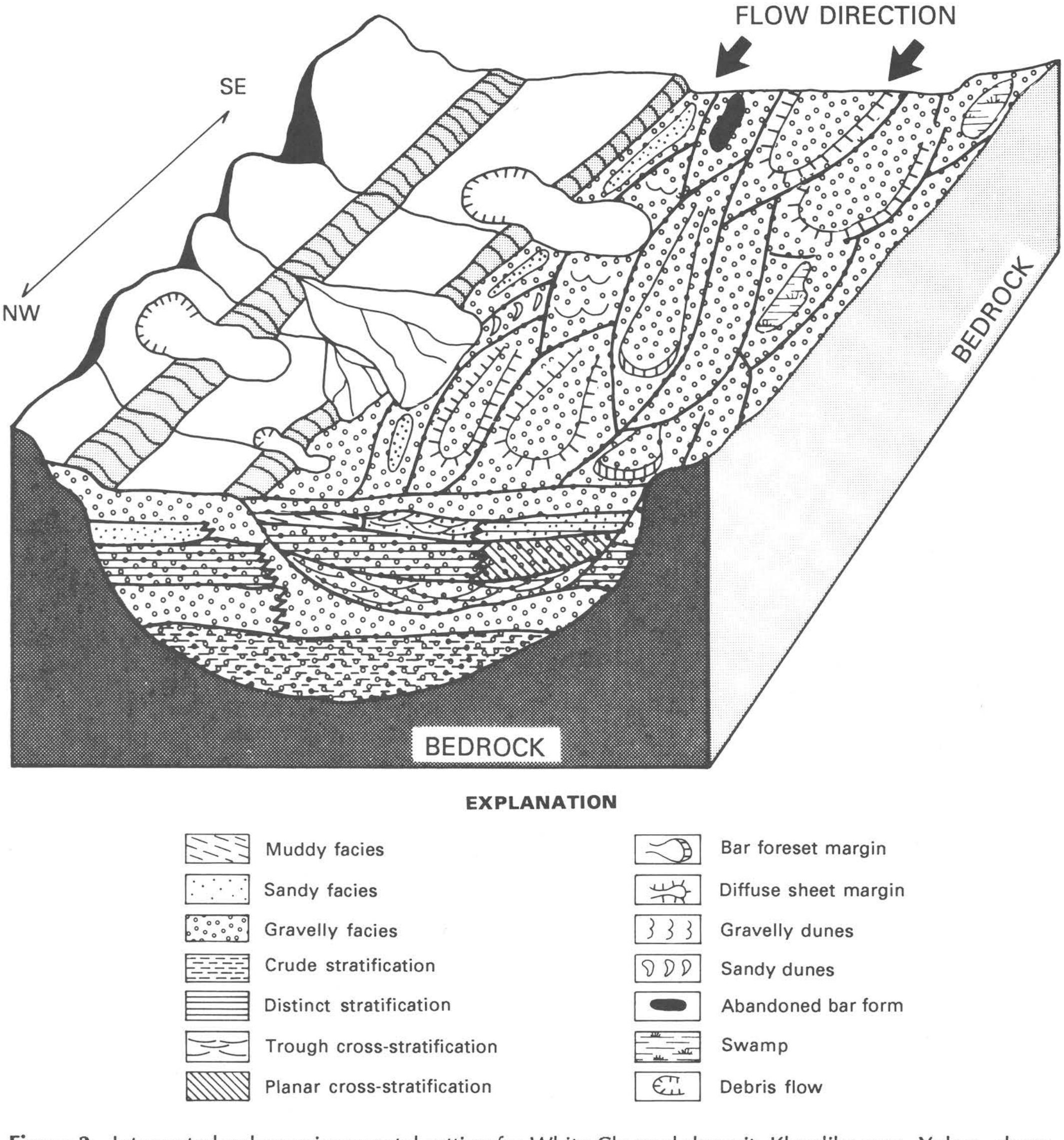
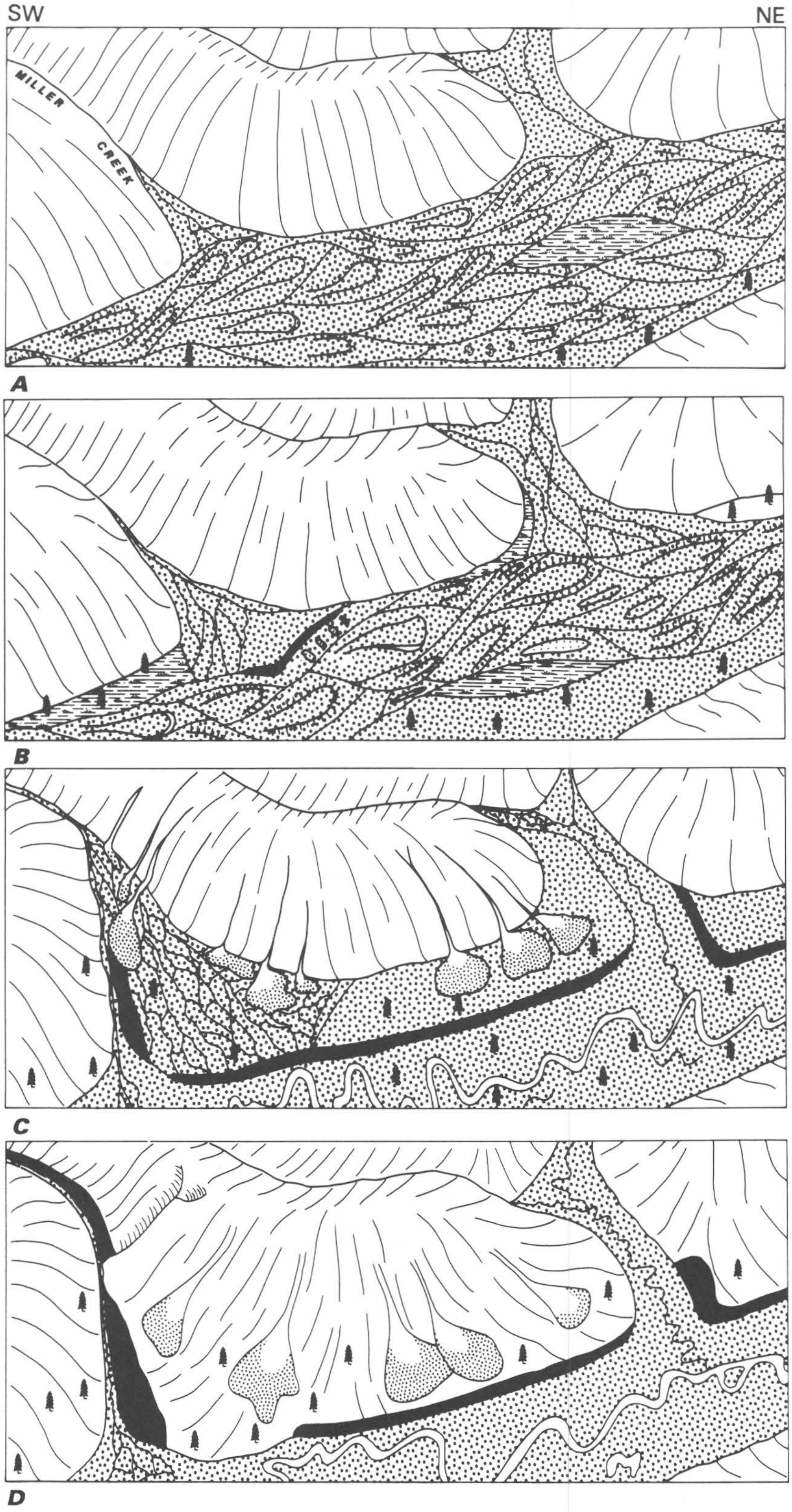


Figure 3. Interpreted paleoenvironmental setting for White Channel deposit, Klondike area, Yukon, showing fully developed gravelly braided river flanked by alluvial fans and debris flows from colluvial slopes.



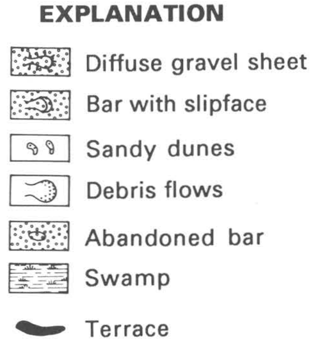


Figure 4. Sketches showing Pleistocene evolution of unglaciated terrain. *A*, Fluvial environments of Sixtymile River and Miller Creek prior to fan aggradation. Note initiation of tributary valley widening and braidplain environment of main Sixtymile River valley. Stream flow is left to right. *B*, Aggradation of Miller Creek alluvial fan, which was complete before 40 ka. Note erosion of distal fan sediments by Sixtymile River. *C*, Debris-flow sedimentation and terrace development 26 ka. Miller Creek is being diverted to southwest side of valley, and Sixtymile River has changed channel pattern from braided to meandering. *D*, Post-terrace development with active debris-flow sedimentation, valley widening at mouth of Miller Creek, and establishment of modern Sixtymile River flood plain.

glaciation (for example, Flat Creek beds and Klondike Gravel), the Klondike River and North Klondike River were incised to approximately present-day stream levels before the Intermediate glaciation (Hughes and others, 1983). This represents downcutting of approximately 200 m below the drift surface of the Old glaciation (Hughes and others, 1983).

Glacial gravel deposits

During glacial intervals, ice-marginal and proglacial processes deposited gravelly sequences which incorporated placer gold from regional bedrock or paleoplacer sources. Examples are found in areas glaciated during pre-Reid glaciations such as the Clear Creek drainage basin (placer area 4, fig. 1), where placer gold occurs in glaciofluvial sediments (Morison, 1985a). Drift of the Reid and McConnell glaciations does not usually contain placer gold because bedrock or paleoplacer sources were largely eroded and dispersed or covered by drift during the pre-Reid glacial intervals.

Holocene

Holocene gravelly deposits with concentrated placer gold include colluvium (for example, Dublin Gulch, near the Mayo placer area), valley-bottom alluvial blankets in gulches and streams (for example, the Clear Creek drainage basin), and bar deposits in major rivers (fig. 1). Gulch and stream placer deposits are generally thin (less than 5 m), and gravelly sediments are poorly sorted, angular, and locally derived.

REFERENCES CITED

Bostock, H.S., and Lees, E.J., 1960, Laberge map-area: Geological Survey of Canada Memoir 217, 32 p.

Dufresne, M.B., and Morison, S.R., 1985, Stratigraphy and alteration of the White Channel Gravel at Dago Hill, A progress report, Klondike area, Yukon; Yukon Exploration and Geology 1983: Exploration and Geological Services Division, Northern Affairs Program, Yukon, p. 55–59.

Hughes, O.L., Rampton, V.N., and Rutter, N.W., 1972, Quaternary geology and geomorphology, southern and central Yukon (Northern Canada): International Geological Congress, 24th Excursion, Guidebook All, 59 p.

Hughes, O.L., Van Everdingen, R.O., and Tarnocai, C., 1983, Regional setting-Physiography and geology, *in* French, H.M., and Heginbottom, J.A., eds., Guidebook to permafrost and related features of the northern Yukon Territory and Mackenzie Delta, Canada: Guidebook 3, International Conference on Permafrost, 4th, p. 5–12.

Hughes, R.L., Morison, S.R., and Hein, F.J., 1987, Placer gravels of Miller Creek, Sixtymile area, 116B–116C: Yukon Geology, v. 1, Exploration and Geological Services Division, Northern Affairs Program, Yukon, p. 50–55.

Hughes, R.L., 1986, Sedimentology of the Sixtymile River placer gravels, Yukon Territory: Edmonton, University of Alberta, M.S. thesis, 210 p.

McConnell, R.G., 1907, Report on gold values in the Klondike high level gravels: Geological Survey of Canada Publication no. 884, p. 1–71.

Morison, S.R., 1985a, Placer deposits of Clear Creek drainage basin 115P, central Yukon; Yukon Exploration and Geology 1983: Exploration and Geological Services Division, Northern Affairs Program, Yukon, P. 88–93.

———1985b, Sedimentology of White Channel placer deposits, Klondike area, wests-central Yukon; Edmonton, University of Alberta, M.S. thesis 149 p.

———in press, Placer deposits in Canada, *in* Geology and economic minerals of Canada: DNAG Quaternary Volume, Geological Survey of Canada Publication.

Morison, S.R., and Hein, F.J., 1987, Sedimentology of the White Channel Gravels, Klondike area, Yukon Territory: Fluvial deposits of a confined valley, *in* Ethridge, F.G., Flores, R.M., and Harvey, M.D., eds., Recent developments in fluvial sedimentology: Society of Economic Paleontologists and Mineralogists Special Publication no. 39, p. 205–216.

Tarnocai, C., 1987, Quaternary soils of Yukon Territory: Guidebook to Quaternary Research in Yukon, Field Excursions A20a and A20b, XII INQUA Congress, Ottawa, July–August 1987, 110 p.

Vernon, P., and Hughes, O.L., 1966, Surficial geology of Dawson, Larsen Creek and North Creek map areas, Yukon Territory: Canadian Geological Survey Bulletin 136, 25 p. (includes Maps 1170A, 1171A, 1172A, scale 1:250,000).

Carter, L. David, Hamilton, Thomas D., and Galloway, John P., eds., 1989, Late Cenozoic History of the Interior Basins of Alaska and the Yukon: U.S. Geological Survey Circular 1026.

Quaternary Stratigraphy of the Fairbanks Area, Alaska

By Troy L. Péwé

A geologic record of the past 2 million years or more is preserved in the nonglacial sediments of the Fairbanks area in central Alaska (fig. 1). Perennially frozen gravel, rich in placer gold, and loess and retransported loess with several interbedded tephra layers preserve a rich floral and faunal history. This paper summarizes an earlier synthesis of the stratigraphy, age, and paleoclimatic significance of these sediments (Péwé, 1975a), especially the Gold Hill Loess, and then discusses more recent tephrochronology and thermoluminescence (TL) data and their bearing on the previous interpretations. These data include preliminary results, and the interpretations presented represent my opinions as of March 1987. The tephrochronology studies were initiated in 1971, when J.A. Westgate (University of Toronto) began studying many tephra samples that I had previously collected from the loess. In 1983, Westgate, G.W. Berger (Western Washington University), Ann Wintle (Cambridge University), and I began joint field investigations to refine the age and stratigraphic interpretations of the loess by expanding detailed tephrochronology studies and initiating TL and paleomagnetic studies. The reader is referred to Péwé (1975a) and references therein for detailed descriptions of individual stratigraphic units and their contained fossils.

Stratigraphy

The oldest unconsolidated deposit in the Fairbanks vicinity is the Cripple Gravel (Péwé, 1975a); a gold-rich, brownish, coarse sandy gravel with angular clasts that is preserved on buried bedrock benches (fig. 2). The Cripple Gravel is thought to be late Pliocene and (or) early Pleistocene in age, but no fossils have been reported from it. It is interpreted as composed of solifluction debris that was produced in a periglacial climate and partly reworked by streams. The Cripple Gravel formed when drainage directions were different from those of today.

After modern drainageways had been established by rejuvenated streams, a gold-bearing, poorly stratified gravel named the Fox Gravel (Péwé, 1975a) accumulated in valley bottoms. The Fox Gravel is believed to be laterally equivalent to the Tanana Formation, which consists of unsorted and angular fractured and weathered bedrock fragments in a silty-sandy matrix and is interpreted as a solifluction deposit. Both the Fox Gravel and the Tanana Formation are thus thought to have formed during an interval of rigorous periglacial climate. The Fox Gravel contains large bones of mammoth and bison that are thought to be of early Pleistocene age. At various locations, the Fox Gravel is overlain by the Dawson Cut Formation, Gold Hill Loess, or Goldstream Formation (fig. 2).

The Dawson Cut Formation (Péwé, 1975a) is a silt unit 1 to 3 m thick that contains peat lenses, logs, and forest beds. Logs are as much as 30 cm in diameter and are weathered, flattened, and iron stained. Spruce stumps as much as 20 cm in diameter are rooted in peaty soil. The Dawson Cut Formation is believed to represent a pre-Illinoian interglacial interval.

The next younger formation is the Gold Hill Loess (Péwé, 1975a), which is as much as 55 m thick and includes conspicuous folds, contortions, and fault blocks that are 2 to 35 m across. Because of the uniformity in texture and composition of the silt, even prominent structural features are seen only as displacements of tephra layers, organic-rich beds, iron-stained layers, or paleosols. Several conglomerate layers composed of silt clasts in a silt matrix occur at topographic breaks in cliff faces developed in the Gold Hill Loess, especially in the middle of the formation at Gold Hill mining cut, Sheep Creek mining cut, and "Ester Island" (Péwé, 1952, and unpub. data, 1956), and these may record significant interruptions in the record. These topographic breaks have become prominent in the last 15 to 20 years as the silt cliffs thawed and retreated due to weathering and erosion. One of these silt-conglomerate layers occurs just above the Ester Ash Bed, which is described below.

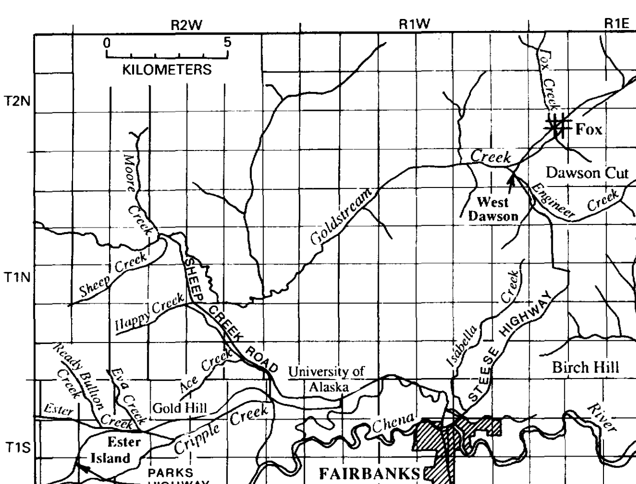


Figure 1. Map of Fairbanks area showing localities mentioned in text.

Remains of small arctic and alpine mammals occur throughout the Gold Hill Loess, and tree pollen is rare. Bones of large mammals such as mammoth, horse, and bison are common but not abundant, and no mammal carcasses have been found. Retransported loess with abundant mammal fossils probably accumulated in valley bottoms during or shortly after deposition of the Gold Hill Loess. If so, then most of the valley-bottom accumulations must have been removed by the major valley streams prior to deposition of the late Pleistocene Goldstream Formation, except in rare downfaulted areas such as the one near Ester island (fig. 1) known as Cripple Sump, where retransported loess correlative with the Gold Hill Loess is preserved. The retransported loess at Cripple Sump is rich in fossils, including remains of *Praeovibos*, the only Pleistocene large mammal that may have been restricted to pre-Wisconsin time in Alaska.

A prominent tephra layer, the Ester Ash Bed (Péwé, 1955), occurs 2.8 m above the base of the Gold Hill Loess at several localities. A persistent tephra layer named the Dome Ash Bed (Péwé, 1975a) also occurs near the top of the Gold Hill Loess in many exposures near Fairbanks and as far away as 200 km. This bed is commonly faulted. More than one ash bed is present in some exposures, and recent studies have shown that some of these are distinct from both the Ester Ash Bed and the Dome Ash Bed and provide new and important marker horizons.

Based on sedimentological and paleontological studies, I proposed that the Gold Hill Loess represented a long interval of rigorous periglacial climate when treeline was 450 to 600 m lower than today (Péwé, 1975a). Because sedimentological breaks were not thought to represent significant time intervals, and no major environmental changes were recognized in the floral and faunal record, the Gold Hill Loess was believed to have formed during a single major glaciation, which was interpreted as Illinoian in age.

The long sedimentary record represented by the Gold Hill Loess is rarely preserved in its entirety at any one exposure. On upper slopes the loess is more easily eroded and only the upper part of the record is preserved. In valley bottoms most of the loess was removed by erosion during the last major warm period (interglacial interval). Only on low hills near the Tanana River (7 km to the south), or on middle slopes, is the most complete record available. Exposures at Gold Hill, Sheep Creek, and Ester island exhibit the thickest deposits and the longest record.

After deposition of most or all of the Gold Hill Loess, large blocks of this silt with tephra layers were locally tilted into the valley by slumping, presumably in response to erosion of the loess. Prior to and during erosion, the iron coating on minerals in the upper part of the deposit was reduced from the ferric to the ferrous state, probably by percolating ground water. The thickness of the silt with ferrous iron is about 10 to 20 m. These relations, plus the absence of ice wedges and mammal carcasses, indicate a major interval of erosion and thawing of permafrost during a climate warmer than now. Frozen carcasses and ice wedges of Wisconsin age exist in the silt of the Fairbanks area under the modern climate.

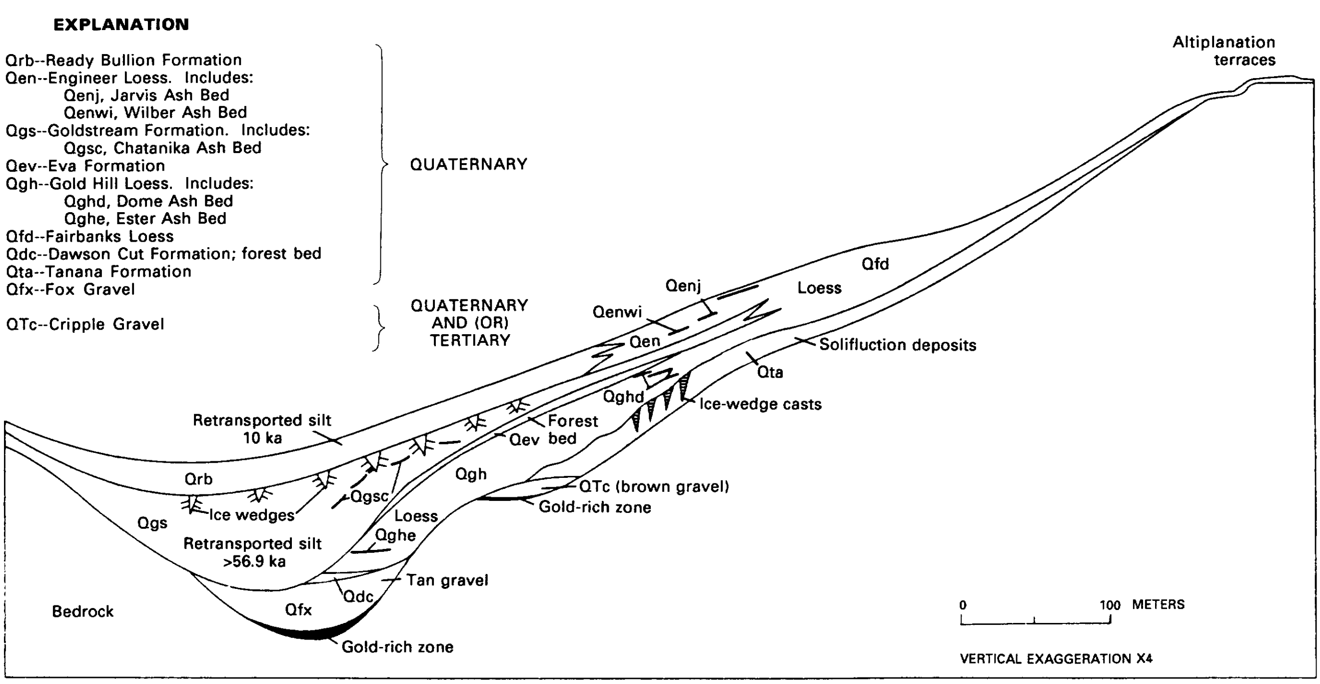


Figure 2. Schematic composite cross section of creek valley near Fairbanks illustrating stratigraphic relations of Quaternary deposits (from Péwé, 1975a).

Unconformably overlying the Gold Hill Loess is the Eva Formation (Péwé, 1975a), a forest bed 1 m thick that contains rooted stumps and prostrate logs of white spruce and birch up to 25 cm in diameter and 3 m long. These remains appear to indicate a climate at least as warm as now. The forest bed probably formed during the latter part of the warm interval that caused the erosion and slumping of the Gold Hill Loess. Radiocarbon ages show that this warm interval occurred before 57 ka, and it is presumed to be of Sangamon (last interglacial) age (Péwé, 1975a, 1975b), which is now equated with oxygen-isotope substage 5e (Richmond and Fullerton, 1986; Edwards and others, 1987).

Overlying the Eva Formation is the Goldstream Formation (Péwé, 1975a), a widespread deposit of perennially frozen, poorly bedded, organic-rich, gray to black retransported loess that is 10 to 35 m thick. Vertebrate remains including mammoth, horse, and bison are common in this deposit, and rare frozen carcasses also are present. Palynological analyses and studies of small-mammal fossils indicated that trees were rare on the landscape during deposition of the Goldstream Formation. All of these characteristics, plus the presence of large ice wedges, are interpreted as indicating a rigorous periglacial climate. Large, flat-topped ice wedges in the upper middle part of the unit probably formed during a short warming interval. The Chatanika Ash Bed (Péwé, 1975a), dated as forming about 14 ka, occurs near the top of the Goldstream Formation and is deformed by large ice wedges. Numerous radiocarbon ages have been obtained on organic materials from this deposit; they range from about 10 ka near the top of the unit to more than 38.5 ka near the middle, and so the Goldstream Formation was thought to be Wisconsin in age.

Unconformably overlying the Goldstream Formation is the Engineer Loess (on hillslopes) and the Ready Bullion Formation (in valley bottoms). Both of these units are of Holocene age and have basal radiocarbon ages of about 10 ka (Péwé, 1975a). The Engineer Loess contains at least four tephra layers: the Wilber, Jarvis, and White River Ash Beds (Péwé, 1975a) and an unnamed ash bed. The Ready Bullion Formation is a poorly to well-stratified, perennially frozen, organic-rich, retransported loess. Prior to deposition of the Ready Bullion Formation, ice wedges at the top of the Goldstream Formation thawed downward 0.5 to 1 m, and faulting and slumping occurred over and adjacent to them. These relations were interpreted to indicate that a brief interval of thawing and erosion occurred about 10 ka. Later, the permafrost table rose into the Ready Bullion Formation and now is about 0.25 to 0.5 m below the surface. Active ice wedges locally are present (Hamilton and others, 1983).

The Fairbanks Loess (Péwé, 1958) is the most widespread Quaternary deposit in the Fairbanks area. It is defined to consist of all the loess on hilltops and upper slopes that cannot be identified as either Gold Hill Loess or Engineer Loess, and it is in part correlative with the retransported silt of the Goldstream Formation.

Tephrochronology studies

Naeser and others (1982) reported a fission-track age of greater than 450 ka for the Ester Ash Bed. Detailed work on (1) tephra layers at the Gold Hill, Dawson, and Cripple Sump mining cuts that formerly were interpreted as correlative with the Ester Ash Bed (Péwé, 1975a), and (2) a tephra bed at Eva Creek also thought to correlate with the Ester Ash Bed, revealed that each tephra layer is chemically and petrologically distinct and that all are older than the Ester Ash Bed (J.A. Westgate, University of Toronto, written commun., 1984). Two of the tephra layers occur at the Cripple Sump (south side of Ester Island, fig. 1) and Dawson mining cuts (West Dawson and Dawson cut, fig. 1). The lower of these two beds is here informally termed the Cripple Sump tephra, and the upper bed is informally termed the Dawson tephra. The tephra layer near the base of the Fairbanks Loess at Gold Hill (fig. 1; Péwé, 1975a) is here informally termed the Canal tephra, and the one at Eva Creek (fig. 1) is informally termed the Weigh Scale tephra. The stratigraphic positions of the Canal tephra and the Weigh Scale tephra relative to one another and to the Cripple Sump and Dawson tephras are not clear. According to the recent application of the term "Illinoian" to the time period from 302 to 132 ka (Richmond and Fullerton, 1986), all five tephra layers and the lower part of the Gold Hill Loess are pre-Illinoian in age. I suggest that the loess record in the Fairbanks area may extend back 1 m.y.

Petrographic and chemical examination of tephras thought to correlate with the Dome Ash Bed have revealed that they may comprise as many as seven different tephra layers (J.A. Westgate, University of Toronto, written commun., 1985) (fig. 3). Thus, contrary to previous interpretations (Péwé, 1975a), the Dome Ash Bed is not present in all exposures of the upper part of the Gold Hill Loess. However, preliminary examination of tephra samples from Eva Creek, Dawson Cut, and at West Dawson (fig. 1) suggests that the Dome Ash Bed is present at these localities (J.A. Westgate, University of Toronto, written commun., 1980).

A tephra layer present at Sheep Creek (fig. 1) is petrographically and chemically distinct from those mentioned above and has been informally termed the Sheep Creek tephra (Westgate and others, 1982, 1985). This tephra bed also has been identified at Dawson Cut and Eva Creek (fig. 1). At Canyon Creek, about 85 km southeast of Fairbanks, the Sheep Creek tephra occurs as angular retransported fragments (Weber and others, 1981), which are associated with mammal bones dated by the uranium-series method as about 80 ka (Hamilton and Bischoff, 1984). The Sheep Creek tephra also has been identified in Yukon Territory (Hughes, this volume).

The thickest and perhaps most widespread of the group of tephras in the middle and upper parts of the Gold Hill Loess is the informally termed Old Crow tephra (Westgate and others, 1981; Westgate, 1982). It occurs in several places in Alaska and the Yukon (Westgate and others, 1985), and some of the thickest known occurrences are near Fairbanks, where it is 10 to 30 cm thick. It has been identified by petrographic and chemical analyses at Eva Creek, Gold Hill, and Sheep Creek near Fairbanks (fig. 1), at the Halfway House locality on the Parks Highway about 35 km west of Fairbanks, and in an exposure on the Richardson Highway about 100 km southeast of Fairbanks (Westgate and others, 1985).

At the Halfway House locality, a tephra layer stratigraphically above the Old Crow tephra is here informally termed the Halfway House tephra. The Halfway House tephra also has been identified at West Dawson (J.A. Westgate, University of Toronto, written commun., 1985) (fig. 1).

In addition to the tephra beds described above, at least three other tephra layers occur in the middle and upper parts of the Gold Hill Loess. Thus there are at least seven tephra beds in the middle and upper parts of the Gold Hill Loess. However, except for the relationship of the Old Crow tephra and the Halfway House tephra, the stratigraphic position of any one tephra bed relative to the others is unknown (fig. 3).

Thermoluminescence studies

To refine age interpretations for the upper part of the Gold Hill Loess and its enclosed tephras in the Fairbanks area, a TL sampling program was begun in 1983. Loess immediately above the Sheep Creek tephra yielded a TL

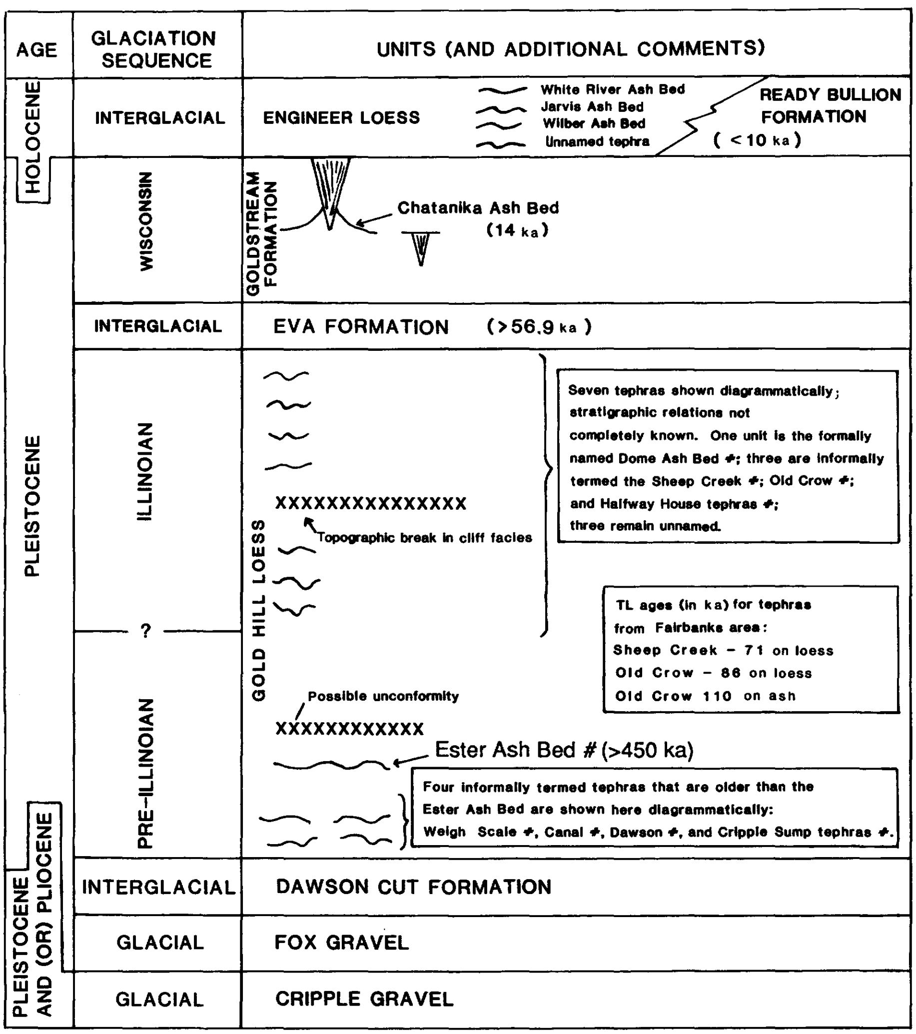


Figure 3. Generalized stratigraphic section of Quaternary deposits in Fairbanks area. Symbol (#) indicates tephra layers that have been petrographically and chemically characterized.

age of 71±3ka (Ann Wintle, Cambridge University, written commun., 1987). Lower and upper age limits of 130 and 220 ka and a probable age of 150 ka were determined by TL analyses of glass from the Old Crow tephra at the Halfway House locality (Berger and Huntley, 1985), but techniques used in these analyses are now considered invalid, and more recent TL analyses produced an age of 110±12 ka (Berger, 1987; and this volume). Loess immediately above the tephra has been dated by TL as 108±16 ka (Berger, 1987; and this volume), and, using a different technique, loess above and below the tephra yield TL ages of about 86±8 ka (Wintle and Westgate, 1986).

The TL ages support recent interpretations of an early Wisconsin age for the Old Crow tephra, the upper part of the Gold Hill Loess, and the overlying Eva Formation (for example, Hopkins, 1982). However, some scientists now question the applicability of TL dating to sediments older than 100 ka (Debenham, 1985; Wintle, 1987), and some geologists have concluded that the TL method severely underestimates the age of loess in the central United States that is beyond the range of radiocarbon dating (Norton and Bradford, 1985; McKay, 1986; Canfield, 1985; Canfield and Mickelson, in press). Therefore, I still believe that a Sangamon age for the Eva Formation and an Illinoian age for the upper part of the Gold Hill Loess is most logical. The Eva Formation and the deep thawing of permafrost that I believe was concurrent with its deposition represent the warmest climatic interval of middle and late Pleistocene time, and it seems most likely that this warm interval is of Sangamon age.

I deeply appreciate comments by the following persons on critical points of earlier drafts of the manuscript: G.W. Berger (Western Washington University), O.J. Ferrians, Jr. (U.S. Geological Survey), D.M. Hopkins (University of Alaska), J.V. Matthews, Jr. (Geological Survey of Canada), R.D. Reger (Alaska Division of Geological and Geophysical Surveys), J.A. Westgate (University of Toronto), and A.G. Wintle (Cambridge University). Support for my work in central Alaska is presently provided by the U.S. Geological Survey through the Arctic Environmental Studies Project and by the National Science Foundation.

REFERENCES CITED

Berger, G.W., 1987, Thermoluminescence dating of the Pleistocene Old Crow tephra and adjacent loess, near Fairbanks, Alaska: Canadian Journal of Earth Sciences, v. 24, p. 1975–1984.

Berger, G.W., and Huntley, D.J., 1985, A thermoluminescence date for the Old Crow tephra: Geological Society of America Abstracts with Programs, v. 17, no. 6, p. 341.

Canfield, H.E., 1985, Thermoluminescence dating and chronology of loess deposition in the central United States: Madison, University of Wisconsin, M.S. thesis, 159 p.

Canfield, H.E., and Mickelson, D.M., in press, Thermoluminescence dating of loess in the central United States: Quaternary Research.

Debenham, N.C., 1985, Use of UV emissions in TL dating of sediments: Nuclear Tracks, v. 10, p. 717–724.

Edwards, R.L., Chen, J.H., Ku, T.L., and Wasserburg, C.J., 1987, Precise timing of the last interglacial period from mass spectrometer determination of thorium-230 in corals: Science, v. 236, p. 1547–1553.

Hamilton, T.D., Ager, T.A., and Robinson, S.W., 1983, Late Holocene ice wedges near Fairbanks, Alaska, U.S.A.: Environmental setting and history of growth: Arctic and Alpine Research, v. 15, p. 157–168.

Hamilton, T.D., and Bischoff, J.L., 1984, Uranium-series dating of fossil bones from the Canyon Creek vertebrate locality in central Alaska: U.S. Geological Survey Circular 939, p. 26–29.

Hopkins, D.M., 1982, Aspects of the paleogeography of Beringia during the late Pleistocene, *in* Hopkins, D.M., Matthews, J.V., Jr., Schweger, C.E., and Young, S.B., eds., Paleoecology of Beringia: New York, Academic Press, p. 3–28.

McKay, E.D., 1986, Illinoian and older loesses and tills at the Maryville Section: Illinois State Geological Survey AMQUA Guidebook, p. 21–30.

Naeser, N.D., Westgate, J.A., Hughes, O.L., and Péwé, T.L., 1982, Fission-track ages of late Cenozoic distal tephra beds in the Yukon Territory and Alaska: Canadian Journal of Earth Sciences, v. 19, p. 2167–2178.

Norton, L.D., and Bradford, J.M., 1985, Thermoluminescence dating of loess from western Iowa: Soil Science Society of America Journal, v. 49, p. 708–712.

Péwé, T.L., 1952, Geomorphology of the Fairbanks area, Alaska: Stanford, California, Stanford University, Ph.D. Thesis, 220 p.

———1955, Origin of the upland silt near Fairbanks, Alaska: Geological Society of America Bulletin, v.66, p. 699–724.

———1958, Geology of the Fairbanks D–2 quadrangle, Alaska: U.S. Geological Survey Geologic Quadrangle Map GQ–110, scale 1:63,360.

———1975a, Quaternary stratigraphic nomenclature in central Alaska: U.S. Geological Survey Professional Paper 862, 32 p.

———1975b, Quaternary geology of Alaska: U.S. Geological Survey Professional Paper 835, 145 p.

Péwé, T.L., and Reger, R.D., 1983, Guidebook to permafrost and Quaternary geology along the Richardson and Glenn Highways between Fairbanks and Anchorage, Alaska: Alaska Division of Geological and Geophysical Surveys, Guidebook 1, International Conference on Permafrost, 4th, 263 p.

Richmond, G.M., and Fullerton, D.S., 1986, Introduction to Quaternary glaciation in the United States of America, *in* Sibrava, Vladimir, Bowen D.Q., and Richmond, G.M., eds., Quaternary glaciations in the Northern Hemisphere: Quaternary Science Reviews, v. 5, p. 3–10.

Weber, F.R., Hamilton, T.D., Hopkins, D.M., Repenning, C.A., and Haas, Herbert, 1981, Canyon Creek: A late Pleistocene vertebrate locality in interior Alaska: Quaternary Research, v. 16, p. 167–180.

Westgate, J.A., 1982, Discovery of a large-magnitude, late Pleistocene volcanic eruption in Alaska: Science, v. 218, p. 789–790.

Westgate, J.A., Matthews, J.V., Jr., and Hamilton, T.D., 1981, Old Crow tephra—A new late Pleistocene stratigraphic marker across Alaska and the Yukon Territory: Geological Society of America Abstracts with Programs, v. 13, p. 579.

Westgate, J.A., Péwé, T.L., and Gorton, M.P., 1982, Tephrachronology of the Gold Hill Loess in central Alaska: Geological Society of America Abstracts with Programs, v. 14,

no. 7, p. 645–646.
Westgate, J.A., Walter, R.C., Pearce, G.W., and Gorton, M.P., 1985, Distribution, stratigraphy, petrochemistry, and paleomagnetism of the late Pleistocene Old Crow tephra in Alaska and the Yukon: Canadian Journal of Earth Sciences, v. 22, p. 893–906.
Wintle, A.G., 1987, Thermoluminescence dating of loess: Catena Supplement 9 (in press).
Wintle, A.G., and Westgate, J.A., 1986, Thermoluminescence age of Old Crow tephra in Alaska: Geology, v. 14, p. 594–597.

Carter, L. David, Hamilton, Thomas D., and Galloway, John P., eds., 1989, Late Cenozoic History of the Interior Basins of Alaska and the Yukon: U.S. Geological Survey Circular 1026.

Pleistocene Glacial Events, Southeastern Copper River Basin, Alaska

By Donald R. Nichols

The deeply incised Copper River and its principal tributaries form bluffs that expose thick sequences of glacial, glaciofluvial, glaciolacustrine, fluvial, colluvial, eolian, and volcanic deposits. These deposits record a long Quaternary history probably extending back to early Pleistocene time. Pre-Quaternary rocks (Jurassic and older) are rarely exposed (Winkler and others, 1981) and, where they are observed, they usually are covered only by relatively young Quaternary deposits; presumably deposits of intermediate age have been eroded.

The onset of glacial conditions during each major Pleistocene glaciation resulted in the advance of valley glaciers that blocked the Copper River, which flows in a deep trench through the Chugach Mountains (fig. 1). For example, near the head of the Copper River Delta, the Miles, Childs, and Allen Glaciers currently front in the Copper River, and ice advances of less than 10 km would likely dam the Copper River, at least temporarily. As glaciers filled the Copper River trench during each glaciation, they created an ice dam, which formed a lake that backed up into the Copper River Basin. Glaciers draining the large snow and ice fields of the Chugach and Wrangell Mountains, and even the Alaska Range, flowed down their respective valleys and terminated in the lake. The Chitina River valley, fed by one of the largest ice fields in Alaska, was occupied by a huge trunk glacier that flowed down the valley to the town of Chitina where it was joined by ice filling the Copper River trench in the Chugach Mountains. This enlarged trunk glacier then flowed up the Copper River valley; its terminus probably floated at times in the proglacial lake filling the basin.

Five apparently stratigraphically distinct drift sheets are exposed in the Pleistocene sediments of the southeastern part of the basin (Nichols, 1965). All appear to be separated by fluvial deposits indicative of nonglacial or nonlacustrine environments that suggest ice positions were no farther advanced than at present in the Chugach Mountains. The three youngest drift sheets are expressed both geomorphically and stratigraphically, but the older two are believed to be represented only stratigraphically. Glaciolacustrine sediments are associated with all but the oldest drift sheet. All radiocarbon samples collected from beneath the youngest drift sheet have yielded ages beyond the maximum radiocarbon dating range, generally older than 40 ka. However, two K-Ar age determinations on plagioclase in a near-surface lava flow, which was scoured by ice of at least the last major glaciation, indicate an age of no older than about 200 ka (Yehle and Nichols, 1980). The oldest near-surface finite dates in the southeastern Copper River Basin are no greater than about 9.4 ka (Rubin and Alexander, 1960, p. 171–172, W–714; Ferrians, 1963, p. C121). Thus this youngest drift is considered equivalent to late Wisconsin, or possibly even the entire Wisconsin time as defined by Richmond and Fullerton (1986).

No direct evidence has been found to indicate the extent of the ice sheet that deposited the oldest drift sheet. Based on relative intensities of European and mid-continent United States glaciations, where older ice sheets generally equaled or extended beyond the limits of younger drift (Flint, 1957, p. 327), it is speculated that ice of the oldest drift sheet probably filled the Copper River Basin and then flowed outward through low divides and drainages in the surrounding mountains—Tahneta Pass, Susitna River Canyon, Delta River, Mentasta Pass, and the Copper River trench. The apparent lack of widespread lacustrine conditions associated with this drift also suggests that the glaciers may have formed a massive ice cap centered in the basin and that deglaciation probably began along the periphery of the basin with ice persisting in the center until a late stage of deglaciation.

The next-to-oldest preserved drift is commonly made up of two diamicton units, distinguishable by color and texture (Nichols, unpub. data), and separated by a persistent bedded zone or layer of pebbles or cobbles. The dual nature of the drift and the association with lacustrine deposits above and below the drift, and locally interbedded within it, is interpreted as representing an advance of ice into the basin, a partial retreat, but not removal of the ice dam, and readvance into the basin. This scenario may indicate a restricted distribution of the ice, which probably did not completely fill the basin even though it may have covered the basin floor.

I arbitrarily correlate the third youngest, stratigraphically exposed drift sheet with the oldest drift having geomorphic expression. Geomorphically, it is represented by erratics and highly modified morainal remnants, which lie on high, relatively flat or gently sloping surfaces of the northern slopes of the Chugach Mountains and above the limits of the two youngest drift sheets. Stratigraphically, till of this ice advance appears more uniform in texture and composition than that of other drifts. Where exposed, it forms steep, prominent ledges and pinnacles, commonly about midway in bluff faces along the Copper River and in tributary valleys. The till generally is underlain by lacustrine deposits; thus, ice probably advanced into a lake-filled basin. However, the presence of coarse fluvial deposits immediately above the till in most exposures indicates that melt water from the ice probably drained off subaerially, rather than subglacially into a proglacial lake. The high altitude of erratics and erosional evidence of south-flowing ice over the tops of 1,525-m-high mountains also suggest that ice filled the basin and flowed outward from a central ice cap. Consequently, deglaciation is inferred to have begun in the mountains peripheral to the basin and to have progressed toward the center of the basin, as was suggested for retreat of the oldest ice advance.

The next-to-youngest drift sheet is represented by the widespread distribution of dual diamicton units and associated lacustrine deposits that are arbitrarily correlated with the higher of two sets of prominent moraines blocking small tributary valleys in the northern Chugach Mountains. The higher, slightly dissected and modified set of moraines decreases in altitude from between 1,056 and 1,160 m in altitude between Tonsina and Tonsina Lake to 975 to 1,056 m north of the Klutina River. The strong similarity in the character of this drift to that of the next-to-oldest drift suggests that this younger ice advance was also somewhat limited in extent and was characterized by a major recession followed by a significant readvance. Although the ice may have covered most, if not all, of the basin floor, it may never have been very thick near the center.

Ice that produced the fifth and youngest drift sheet flowed northward up the Copper River about as far as the mouth of the Gulkana River and fronted in Glacial Lake Atna (Nichols, 1965), the youngest of the succession of proglacial lakes formed by repeated invasions of glaciers

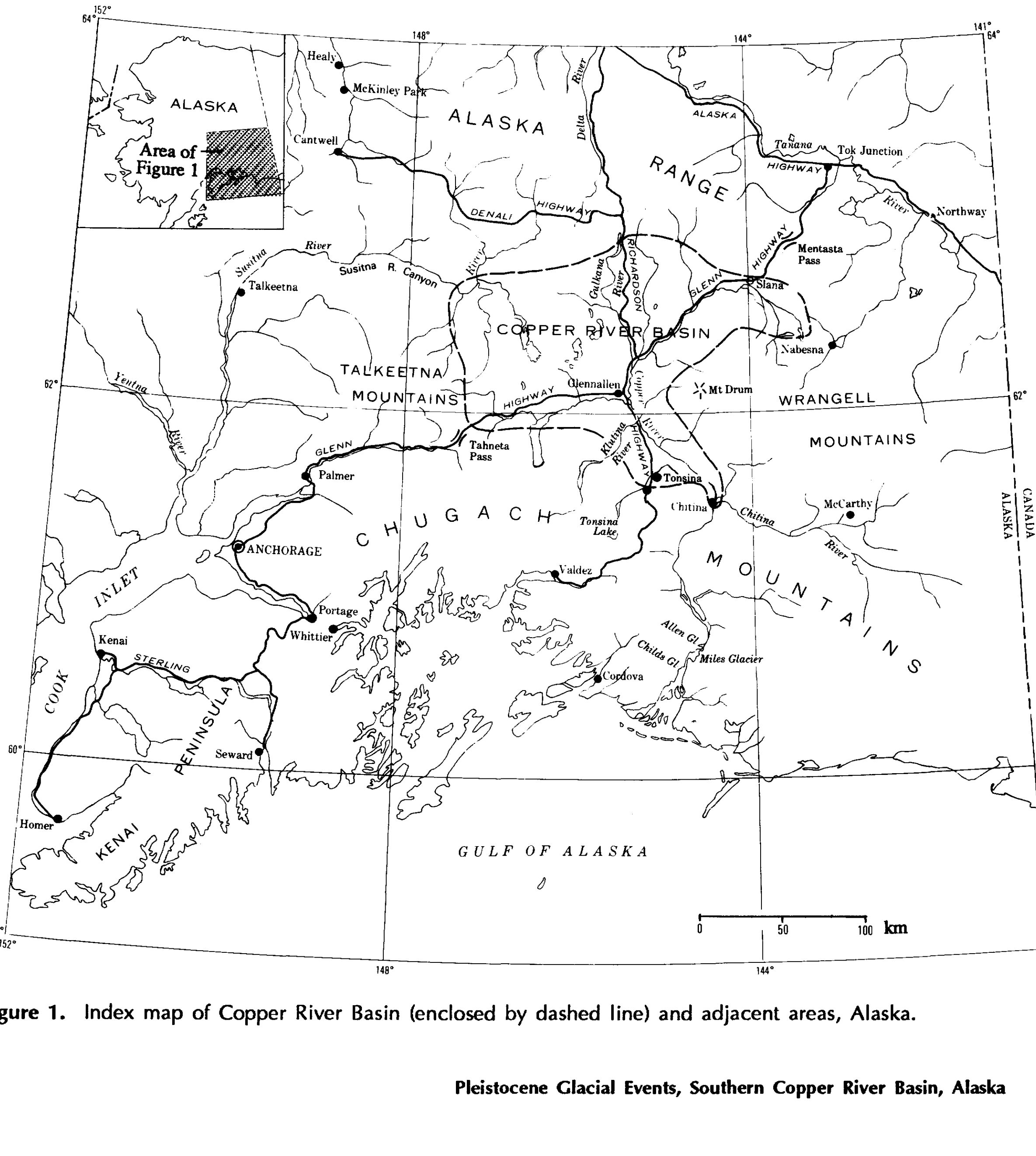


Figure 1. Index map of Copper River Basin (enclosed by dashed line) and adjacent areas, Alaska.

into the basin or along its margins. Ferrians (1984) presented evidence that the lake level was below an altitude of 655 m between 31.3 and 28.3 ka, and subsequently rose above that altitude in the central Copper River Basin. This youngest glaciation is represented by two major, little-modified morainal systems along the western flank of the Wrangell Mountains. The higher, slightly older moraines and related features decrease in altitude from about 1,220 m along the southwest side of Mount Drum to about 975 m along the northwest side. The lower moraine decreases in altitude from about 1,070 m to about 700 m over the same distance. Similar, but slightly lower, north-sloping moraines are present along the northern flank of the Chugach Mountains.

REFERENCES CITED

Ferrians, O.J., Jr., 1963, Glaciolacustrine diamicton deposits in the Copper River Basin, Alaska, *in* Geological Survey Research 1963: U.S. Geological Survey Professional Paper 475–C, p. C120–C125.

———1984, Pleistocene glacial history of the northeastern Copper River basin, Alaska: Geological Society of America Abstracts with Programs, v. 16, no. 5, p. 282.

Flint, R.F., 1957, Glacial and Pleistocene geology: New York, Wiley, 553 p.

Nichols, D.R., 1965, Glacial history of the Copper River Basin, Alaska [abs.]: International Association for Quaternary Research, 7th Congress, Fairbanks, Alaska, 1965, Proceedings, Lincoln, Nebraska Academy of Sciences, p. 360.

Richmond, G.M., and Fullerton, D.S., 1986, Introduction to Quaternary glaciations in the United States of America, *in* Sibrava, V., Bowen, D.Q., and Richmond, G.M., eds., Quaternary glaciations in the northern hemisphere: International Union of Geological Sciences and UNESCO, Report of the International Geological Correlation Programme, Project 24, Pergamon Press, 513 p.

Rubin, Meyer, and Alexander, Corrine, 1960, U.S. Geological Survey radiocarbon dates V, *in* Flint, R.F., and Deevey, E.S., Jr., eds., Radiocarbon supplement: American Journal of Science, v. 2, p. 129–185.

Winkler, G.R., Silberman, M.L., Grantz, Arthur, Miller, R.J., and MacKevett, E.M., Jr., 1981, Geologic map and summary geochronology of the Valdez quadrangle, southern Alaska: U.S. Geological Survey Open-File Report 80–892–A, 2 sheets.

Yehle, L.A., and Nichols, D.R., 1980, Reconnaissance map and description of the Chetaslina volcanic debris flow (new name), southeastern Copper River Basin and adjacent areas, south-central Alaska: U.S. Geological Survey Miscellaneous Field Studies Map MF–1209, scale 1:250,000.

Carter, L. David, Hamilton, Thomas D., and Galloway, John P., eds., 1989, Late Cenozoic History of the Interior Basins of Alaska and the Yukon: U.S. Geological Survey Circular 1026.

A Working Glacial Chronology for the Western Copper River Basin, Alaska

By John R. Williams

Pleistocene piedmont glaciers advanced into the western Copper River Basin from the Chitina-Copper River valley, and also from the surrounding Chugach and Talkeetna Mountains and from the Alaska Range (fig. 1). They filled the basin, perhaps to overflowing, during earlier advances that are marked by erratics, formless drift, and the Daisy Creek moraine system along the eastern flank of the Talkeetna Mountains (fig. 2). During the last major glaciation, equivalent to most if not all of Wisconsin time, glaciers advanced from the Talkeetna Mountains, the Chugach Mountains, and the Alaska Range. They coalesced in the northwestern part of the Copper River Basin during the early phase of glacier expansion, but advances during a later phase were less extensive by about 40 km. Each glacial phase was associated with extensive glacial lakes, termed glacial Lakes Susitna and Atna, respectively (fig. 1). The western Copper River Basin has been studied intermittently by the U.S. Geological Survey since 1952, and this summary of the glacial chronology is based on the distribution, stratigraphy, and radiocarbon dating of the Quaternary deposits (Williams and Galloway, 1986).

Older glaciations of the western Copper River Basin are recognized by erratic clasts and patches of drift, lacking morainal form, that lie above and beyond the Daisy Creek (oldest) moraine system in areas farthest from modern glaciers in the eastern Talkeetna Mountains. Although erratics are difficult to identify in many areas of the eastern Talkeetna Mountains that are underlain by disaggregated Jurassic and Tertiary conglomerate, their distribution west of the Oshetna River on Twin Hills (TH on fig. 2) to more than 1,730 m above sea level is 400 m above the oldest moraine nearby. East of the Oshetna River (near TC, fig. 2) drift and erratics seem to be absent above 1,296 m along the eastern front of the Talkeetna Mountains, suggesting a possible driftless area. Thick, stratified gravel (OG, fig. 2) farther south lies above 1,219 m altitude; the gravel is slightly lithified and contains clasts derived from the Chugach Mountains, which suggests (Grantz, 1960 and Grantz, U.S. Geological Survey, written commun., April 12, 1957) that it is outwash, rather than Tertiary conglomerate.

The Daisy Creek moraine (fig. 2) is the oldest moraine along the eastern front of the Talkeetna Mountains; it is a lateral moraine formed by Chugach glaciers that stands beyond and 122 m higher than Tyone Creek, which is the marginal channel bordering the Curtis Lake moraine of the early phase of the last major glaciation. Exposures on the west bank of Tyone Creek (TC, fig. 2) are of organic-rich fluvial overbank deposits at river level, overlain by gravel that is capped by glaciolacustrine diamicton of glacial Lake Susitna. Daisy Creek drift is assumed to lie stratigraphically below the organic-rich deposits. Dating the organic material by radiocarbon methods provided an age greater than 35 ka (W–357). Ager (this volume) found pollen of spruce, birch, alder, and other plants in the organic beds; the flora was similar to the present lowland spruce forest and also to the interglacial Goose Bay peat, near Anchorage. Daisy Creek drift is pre-interglacial (Illinoian?), or less likely early Wisconsin(?) in age, and formed immediately preceding a very warm climatic interval.

During the *early phase of the last major glaciation*, glaciers from the Talkeetna Mountains moved northward to form the Oshetna moraine system, glaciers from the Alaska Range advanced southward to form the Tysus moraines, and Chugach glaciers moved northward to form the Tolson Creek and Curtis Lake moraines and to merge with the Alaska Range glaciers to form the Heartland Ridge interlobate moraine (HR) against the northward moving Chitina-Copper piedmont glacier (fig. 2). Based on radiocarbon ages from Thaw and Tyone Bluffs (TB, fig. 2) (Thorson and others, 1981), the drift of the Tysus moraines, although not exposed, is believed to lie below river level and to be older than the sediments of glacial Lake Susitna(?), which are at least as old as 32±2.73 ka (Beta–1820). Glacial Lake Susitna (fig. 2) expanded during retreat of the ice of the early phase of the last major glaciation. The lake was blocked by ice in the lower Susitna River canyon (R&M Consultants, 1981) and, during retreat of the glaciers, the lake extended up the Susitna River an unknown distance beyond the Hatchet Lake moraine, northeastward up the Maclaren River, eastward to where dammed by piedmont ice east of Heartland Ridge, and south to and perhaps beyond the Lake Louise sandur (LL) and Old Man moraine of the late phase of the last glaciation. Its level, 914 to 975 m, was controlled by spillways at the southwestern end of Fog Lakes lowland, such as Chunilna Creek (CS, fig. 2) near the site of the ice dam in Susitna canyon. Glacial Lake Susitna had drained, perhaps completely, but at least to some unknown level below 705 m by 29.45±0.61 ka (Beta–1820), which is the radiocarbon age of a mammoth bone embedded in subaerial gravel above the lake sediments at that altitude.

During the *late (Old Man) phase of the last major glaciation* glaciers from the Chugach Mountains advanced northward in two lateral lobes, the western or Little Nelchina lobe up the Little Nelchina River, and the eastern, or Old Man lobe, up Mendeltna Creek and beyond Old Man Lake to the Lake Louise sandur (LL, on fig. 2); a trunk glacier also moved eastward down the Tazlina River valley. Meltwater drainage from the Little Nelchina lobe cut through deposits of glacial Lake Susitna and into the underlying Curtis Lake moraine (fig. 2). This outwash stream formed a gravel kame terrace at about 792 m altitude against ice along the western edge of the Lake Louise sandur of the Old Man lobe. The kame terrace was deposited when lake level was lower than 792 m but higher than 777 m, because the lake deposits do not cover the kame terrace but do cover the sandur and recessional ice-contact features and feeder eskers to the south. This lake, associated with the maximum of the late phase of the last major glaciation, was apparently held between 777 to 792 m by a short-lived Tyone spillway (TSp on fig. 2).

As the ice retreated from the outer Old Man moraine of the Little Nelchina lobe and from the Lake Louise sandur to the Old Man moraine of the Old Man lobe, successive marginal drainage channels from the interlobate area were graded to delta-like features at 774 m, 762 m, and 747 m altitude; the minimum age of the outwash leading to the 762-m delta is 13.28±0.4 ka (W–583). By the time the lake level had fallen to 747 m, the lake was receiving outwash from the Old Man moraine (fig. 2), and the lake was draining across the present Susitna-Copper divide through a single outlet, connecting the valley of the West Fork Gulkana River to the Tyone River. The 747-m shoreline was more stable or longer lived than any of the others, and it is widespread and locally prominent on both sides of the Copper River Basin. The lake thus formed has been called glacial Lake Atna (see Ferrians, this volume). In the northeastern end of Lake Atna (fig. 1) wood exposed in lake bottom sediments was dated at 17.6±0.4 ka (W–1184); that this age may closely approximate the age of the highest stand of glacial Lake Atna in that area at 747 m was among the possible correlations listed by Schmoll (1984). This correlation would permit assignment of an 18 ka age to the 747-m lake level, related outwash, and the Old Man moraine.

In the Susitna River valley the ice advanced southward to the Hatchet Lake moraine, which is similar morphologi-

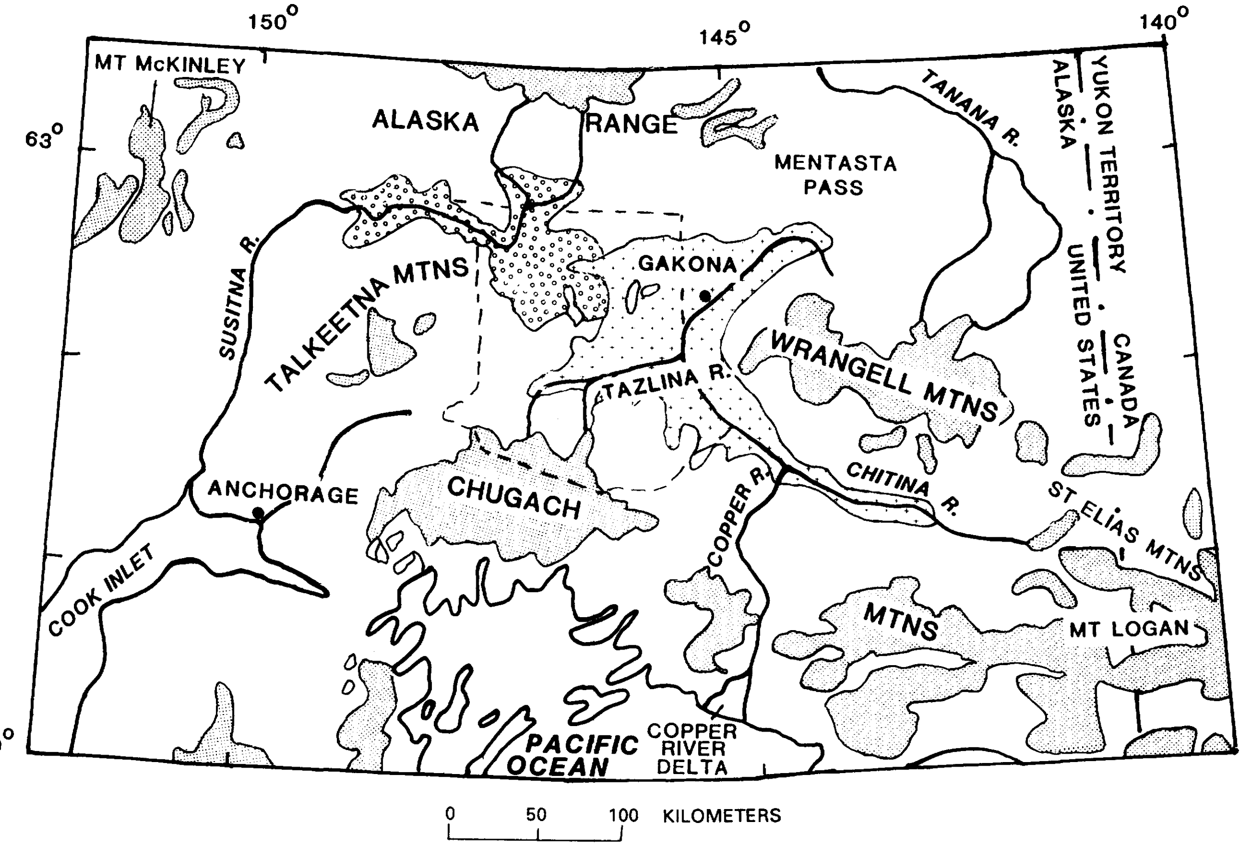


Figure 1. South-central Alaska showing boundary of western Copper River Basin (dashed lines), existing glaciers (dot pattern), glacial Lake Susitna (circle pattern), and glacial Lake Atna (cross pattern).

cally to the Old Man moraines and like them in being about 40 km within the morainal boundaries of the older phase of the last major glaciation. The Hatchet Lake moraine, described by Kachadoorian and others (1955), lies above lacustrine deposits, presumably of glacial Lake Susitna (Reuben Kachadoorian, U.S. Geological Survey, written commun., 1959). The moraine, however, has not been covered by these or younger lake deposits. The upper stratigraphic units at Tyone and Thaw Bluffs (TB), dated between 11.53±0.14 ka (Beta–1821) and 21.73±0.39 ka (DIC–1861) by Thorson and others (1981), are reinterpreted here as glaciolacustrine deposits; they are at elevations of 709 to 722 m, which is above the 29.45 ka gravel that separates the younger lake deposits from those of glacial Lake Susitna. These younger lake sediments also seem equivalent in age to the Hatchet Lake moraine. Exactly how these deposits can be linked to the history of lakes and receding Chugach glaciers, given above, is not known; however, it is possible that a lake at this elevation existed separately or in connection with that above the Tyone spillway.

Both the late and early phases of the last major glaciation are represented by a single drift unit exposed in sections along the Nelchina River near the front of the Chugach Mountains (N, fig. 2). This drift lies locally on a discontinuous organic layer that has alpine tundra vegetation (Ager, this volume) and is older than 38 ka (W–842). Gravel underlying the organic layer can be traced for 11 km at a gradient and direction of flow comparable to the present Nelchina River, down to an altitude of 615 m near Tazlina Lake. The gravel, apparently an advance outwash, lies on lacustrine sand and silt that also are older than 38

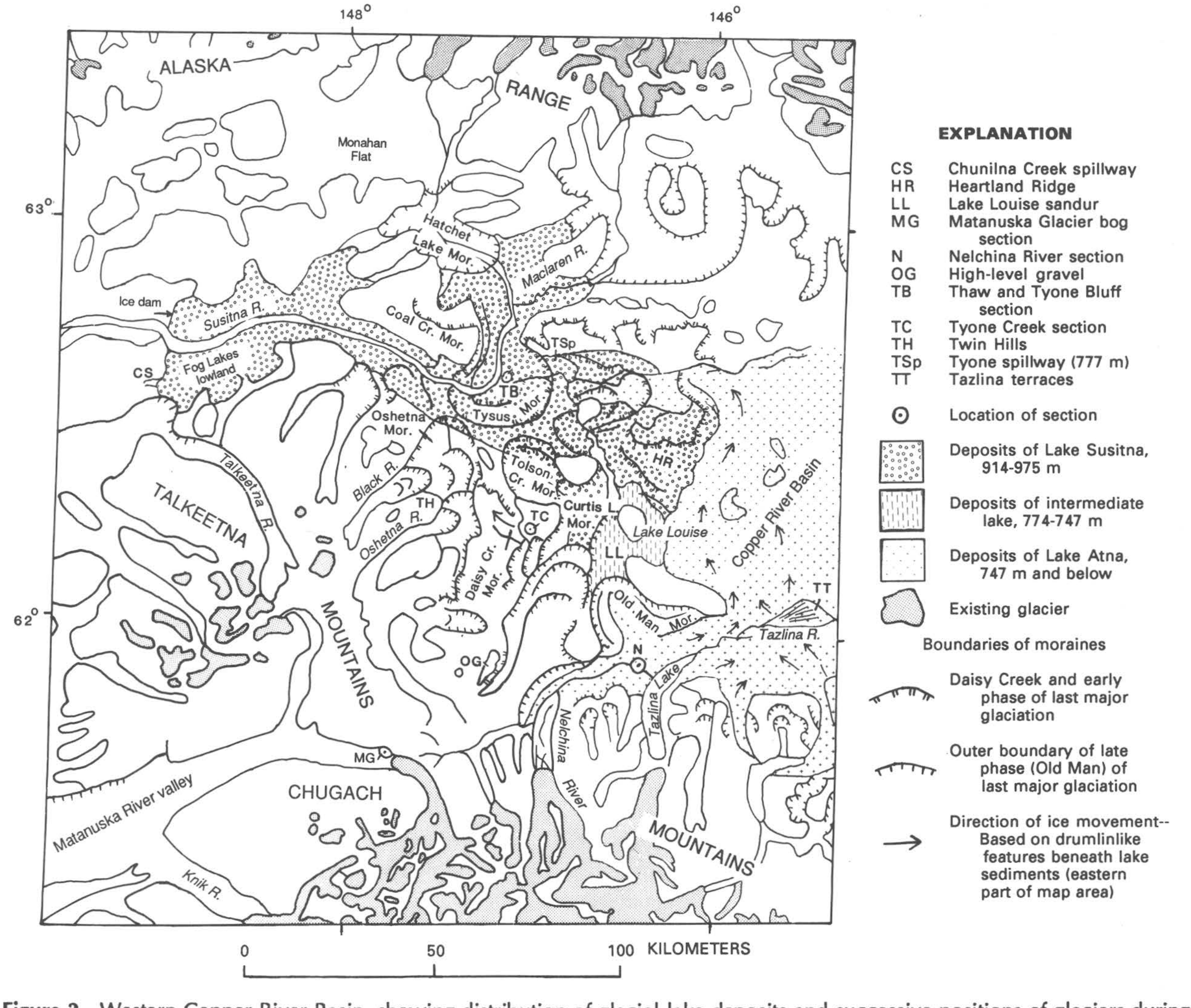


Figure 2. Western Copper River Basin, showing distribution of glacial-lake deposits and successive positions of glaciers during last major glaciation (hachures). Mor., moraine.

ka (W–295). The lacustrine sediments contain a sparse pollen assemblage characteristric of spruce forest, suggesting a warm interstadial or interglacial climate (T.A. Ager, U.S. Geological Survey, written commun., March 1985).

Retreat of the ice from the Old Man moraine first involved bifurcation of the glacier into discrete Tazlina and Nelchina glaciers, the uncovering of the Nelchina-Matanuska divide, advance of the Matanuska and Southeast Fork Glacier eastward up Matanuska River valley toward the divide, and formation of moraines at the mountain front in valleys tributary to the divide area (fig. 2). Drainage from Nelchina Glacier across the divide was ponded east of the extended Matanuska Glacier. A local readvance or halt of Nelchina Glacier about 20 km from its present terminus was later covered by lake waters at least as high as 700 m altitude. Lake Atna, therefore, persisted at very high levels until the retreating glaciers were within 17 km of their present positions. The Matanuska Glacier and others in the coastal mountains (Williams, 1986) had been no more than a few kilometers beyond their present terminal positions in the last 10 to 13 ka. The level of the ice in the Copper River canyon near Tiekel River had been lowered to about 500 m altitude before 14 ka (Sirkin and Tuthill, 1987), based apparently on the age of the basal peat in a nearby bog. These data strongly suggest that Lake Atna in the western Copper River Basin was lowered significantly or drained through the Copper River canyon, the only outlet available lower than about 708 m, slightly before 14 ka or soon thereafter.

Drainage of Lake Atna created an independent lake basin at Tazlina Lake. The lake's outlet, the modern Tazlina River, cut a deep canyon through the lip of the hanging valley and deposited alluvium now preserved as terraces bordering the river downstream. Shorelines of Tazlina Lake were initially 20 m above the present lake. In the upper Matanuska River valley the Matanuska Glacier advanced 2 to 8 km downvalley west of its present terminus (MG, fig. 2) (Williams and Ferrians, 1961) before 13 ka (Williams, 1986). This glacial advance blocked streams from the east and north to create local lakes.

Holocene advances of Matanuska Glacier were 1.6 km beyond the present terminus at 4 ka and a later advance about 0.4 km from the present terminus in the past few hundred years (Williams and Ferrians, 1961). The Holocene deposits are similar to deposits of the Tustumena and Tunnel advances of the Kenai Peninsula (Karlstrom, 1964).

In summary, the ages of two phases of the last major glaciation are 11.53 to about 21.73 ka for the younger and more than 32 ka for the older. Both phases are represented by a single drift sheet near the mountains. This drift is probably equivalent to the deposits of the last major glaciation in the northeastern Copper River Basin, which at Gakona represents the last of three depositional cycles (Ferrians, 1963 and this volume) and began before 55 ka. Drift of the last major glaciation is similar to the uppermost drift of the southeastern Copper River Basin, where basal dates are also greater than 38 ka; this drift postdates the lava flows, volcaniclastic debris flows, mud flows, ash flows and tephra emanating from the volcanic Wrangell Mountains to the east (Nichols, 1984, Nichols and Yehle, 1985).

REFERENCES CITED

Ferrians, O.J., Jr., 1963, Glaciolacustrine diamicton deposits in the Copper River Basin, Alaska: U.S. Geological Survey Professional Paper 475–C, p. C120–C125.

Grantz, Arthur, 1960, Geologic map of the Talkeetna Mountains (A–1) quadrangle and the south third of the Talkeetna Mountains (B–1) quadrangle, Alaska: U.S. Geological Survey Miscellaneous Geologic Investigations Map I–314, scale 1:48,000.

Kachadoorian, Reuben, Hopkins, D.M., and Nichols, D.R., 1955, A preliminary report of geologic factors affecting highway construction in the area between the Susitna and Maclaren Rivers, Alaska: U.S. Geological Survey Open-File Report 92, 74 p., scale 1:40,000.

Karlstrom, T.N.V., 1964, Quaternary geology of the Kenai Lowland and glacial record of the Cook Inlet region, Alaska: U.S. Geological Survey Professional Paper 443, 69 p.

Nichols, D.R., 1984, Quaternary events in the southeastern Copper River Basin, Alaska: Geological Society of America Abstracts with Programs, v. 16, no. 5, p. 325.

Nichols, D.R., and Yehle, L.A., 1985, Volcanic debris flows, Copper River Basin, Alaska: International Conference and Field Workshop on Landslides, 4th, Proceedings, Tokyo, p. 365–372.

R&M Consultants, Inc., 1981, Terrain unit maps, photointerpretation, subtask 5.02: Acres Inc., Report on Susitna Hydroelectric Project, Alaska Power Authority, 1980–81, Geotechnical Report, v. 2, Appendix J, Final draft, 21 sheets.

Schmoll, H.R., 1984, Late Pleistocene morainal and glaciolacustrine geology in the upper Copper River-Mentasta Pass area, Alaska: Geological Society of America Abstracts with Programs, v. 16, no. 5, p. 332.

Sirkin, L.A., and Tuthill, S.J., 1987, Late Pleistocene and Holocene deglaciation and environments of the southern Chugach Mountains, Alaska: Geological Society of America Bulletin, v. 99, p. 376–384.

Thorson, R.M., Dixon, E.J., Jr., Smith, G.S., and Batten, A.R., 1981, Interstadial Proboscidean from south-central Alaska: Implications for biogeography, geology, and archaeology: Quaternary Research, v. 16, no. 3, p. 404–417.

Williams, J.R., 1986, New radiocarbon dates from the Matanuska Glacier bog section, *in* Bartsch-Winkler, Susan, and Reed, K.M., eds., Geologic studies in Alaska by the U.S. Geological Survey during 1985: U.S. Geological Survey Circular 978, p. 85–88.

Williams, J.R., and Ferrians, O.J., Jr., 1961, Late Wisconsin and Recent history of the Matanuska Glacier, Alaska: Arctic, v. 14, no. 2, p. 83–90.

Williams, J.R., and Galloway, J.P., 1986, Map of western Copper River Basin, Alaska, showing lake sediments and shorelines, glacial moraines, and location of stratigraphic sections and radiocarbon-dated samples: U.S. Geological Survey Open-File Report 86–390, 30 p., scale 1:250,000.

Carter, L. David, Hamilton, Thomas D., and Galloway, John P., eds., 1989, Late Cenozoic History of the Interior Basins of Alaska and the Yukon: U.S. Geological Survey Circular 1026.

Glacial Lake Atna, Copper River Basin, Alaska

By Oscar J. Ferrians, Jr.

The Copper River Basin, located in south-central Alaska, is a large intermontaine basin surrounded by high glacier-clad mountains—the Alaska Range to the north, the Talkeetna Mountains to the west, the Chugach Mountains to the south, and the Wrangell Mountains to the east (fig. 1). The basin is approximately 13,000 km^2 in areal extent, and most of it is drained by the Copper River, which heads in the northern Wrangell Mountains, flows in a semicircular course around the west side of the mountains to their south side, then cuts through the Chugach Mountains in a deep canyon and flows into the Gulf of Alaska. Within the Chugach Mountains, the Copper River is bordered on both sides by ice fields. Several glaciers, which include the Allen, Miles, and Childs Glaciers, emanate from the ice fields and terminate along the margin of the river. This relation suggests that during any major glaciation the Copper River would be dammed by advancing glaciers, causing the formation of an extensive proglacial lake in the basin. Stratigraphic evidence from high river bluff exposures in the basin confirms this, indicating that lakes did form during each major glaciation, and that the lakes drained when the glaciers retreated during each major interglaciation (Ferrians and Nichols, 1965; Ferrians and others, 1983).

Previous investigations

In 1898, Schrader (1900) made a reconnaissance study in the Copper River Basin and concluded that the unconsolidated Pleistocene sediments in the central part of the basin were much finer than those at the edges, and that this relation suggested that the fine-grained sediments were deposited in a large body of standing water (either an "inland lake" or an "arm of the sea") with encroaching glaciers depositing gravel and till around the edges. However, Schrader and Spencer (1901, p. 58), after additional studies in the southeastern part of the Copper River Basin, decided that the fine-grained deposits probably are present only in limited areas and are surrounded by coarser deposits. Mendenhall (1905, p. 62–72), who studied the Pleistocene deposits of the central Copper River region, concurred with Schrader and Spencer in concluding that fine-grained sediments were deposited only in several isolated lakes, similar to the present-day Tazlina and Tonsina Lakes. Moffit (1954, p. 159) also concluded that the lake sediments in the Copper River Basin were deposited in several small lakes; however, he stated that "the topographic setting for a large lake seems favorable, but evidence for it has not been recognized." Subsequent studies demonstrated that a large proglacial lake formed in the Copper River Basin during the last major glaciation. These studies form the basis for the following discussion.

Glacial Lake Atna

Extent and strand lines

The lake that formed in the Copper River Basin during the last major glaciation has been named glacial Lake Atna (Nichols, 1965). Evidence in the northeastern Copper River Basin indicates that this lake covered more than 5,200 km^2 of the basin floor, and that numerous glaciers and sediment-laden, glacier-fed streams debouched into it (Ferrians, 1984). Locally, prominent strandlines occur at altitudes of approximately 800, 745, 700, and 640 m, and less prominent ones occur at intermediate and lower levels. These strandlines indicate periods of relatively stable lake level, separated by periods of lowering of lake level.

Several maps show the extent of the lake or the distribution of lake sediments throughout or in various parts of the Copper River Basin (Karlstrom and others, 1964; Coulter and others, 1965; Nichols and Yehle, 1969; Ferrians, 1971a, 1971b; Richter and others, 1979; Williams and Johnson, 1980; Yehle, 1980, 1981; Williams, 1985; Emery and others, 1985; and Williams and Galloway, 1986).

Isostatic effects

Obviously, the tremendous additional weight of the waters of glacial Lake Atna and of the glaciers that advanced into the basin depressed the basin isostatically, and conversely, when the glaciers retreated and the lake drained, the loss of weight resulted in rebound.

Because the shoreline features are not preserved continuously, and because precise measurements of the altitudes of the features have not been made, it is difficult to determine whether or not there has been tilting of the basin floor. The apparent general agreement of the altitudes of shoreline features in many different areas of the basin suggests that the rebound was somewhat uniform and that tilting was not great within any one area.

One line of evidence that suggests that some tilting occurred in the northeastern Copper River Basin is as follows. The Indian Creek delta gravel, 65 km northeast of Gakona, represents the highest recognizable glacial Lake Atna deposit in this local area, and it is at an altitude of 745 m. The Meiers Lake delta gravel, 50 km west of the Indian

Creek deposit, represents the highest Glacial Lake Atna deposit in this local area, and it is at an altitude of 800 m. If one concludes that these two features are coeval, then the 55 m difference in altitude would be a measure of the difference in isostatic rebound between these two sites. If the tilt were uniform, it would be only a little more than 1 m per kilometer of horizontal distance, which would be very difficult to detect in routine geologic field mapping. Possibly the apparent anomalous situation of the highest shoreline features formed during the last major glaciation occurring at different altitudes in different parts of the basin might be explained, in part, by this kind of differential isostatic rebound. For example, lacustrine deposits or shoreline features occur up to altitudes of 945–975 m in the northwestern Copper River Basin (Williams, 1984; Williams and Galloway, 1986, p. 1) and in the adjacent Denali area (Nichols and Yehle, 1961, p. 1067), but up to altitudes of only 745 m in the eastern part of the basin (Ferrians and Schmoll, 1957; Nichols and Yehle, 1969).

Character of sediments

The deposits of Glacial Lake Atna consist of numerous diamicton units, which are interbedded with stratified lacustrine sediments (Ferrians, 1963a, 1963b). The character, stratigraphic relations, and distribution of these diamicton units indicate that many of them were deposited by turbidity currents and subaqueous mudflows. Others were formed by rapid deposition of massive, fine-grained sediment with numerous ice-rafted stones included. They range in thickness from a few centimeters to 50 m. Most of the diamicton is nonsorted and till-like in character.

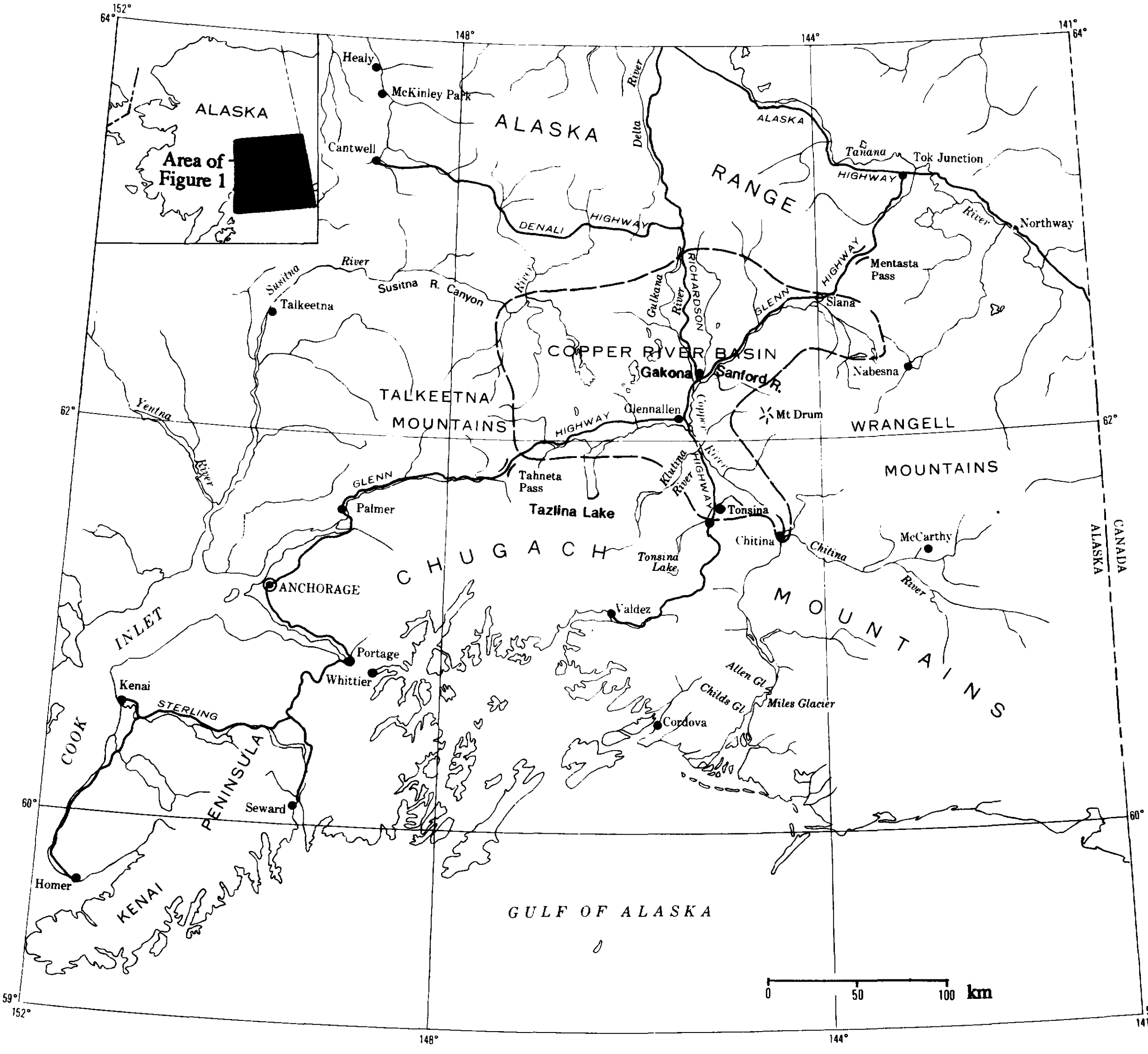


Figure 1. Index map showing setting of Copper River Basin (approximate boundary shown by dashed line).

Other diamicton deposits are poorly sorted and relatively fine grained, with the coarser fraction generally limited to sizes smaller than cobbles. Locally, these poorly sorted deposits show well-developed graded bedding. Numerous units of nonsorted till-like diamicton alternate in vertical sequence with bedded lacustrine sediments and with poorly sorted diamicton. Locally, the alternation repeats itself several times within a stratigraphic thickness of a meter. Generally, the contacts between these units are sharp; the beds immediately underlying the diamicton units generally are not deformed, even though deformation is common within the entire lacustrine sequence. Nichols (1960) described two types of contorted bedding in horizontal zones within the thin-bedded lacustrine sand, silt, and clay, and he concluded that this deformation was caused by slumping generated by earthquakes. Studies of the orientation of phenoclasts in the laminated lacustrine deposits by Schmoll (1961) indicated that there are two preferred orientations of the clasts. The dominant orientation, which is nearly horizontal, resulted from ice-rafted stones being dropped on the relatively firm bottom of the lake. The second preferred orientation, which is weak and which varies between samples of different ages, probably was caused by water currents. According to Eyles (1984), late Quaternary glaciolacustrine deposition in the Copper River Basin and in Lake Ontario, Canada, is analogous to distal glaciomarine sedimentation on the continental shelves.

Age

The age of Glacial Lake Atna is indicated by radiocarbon dates of samples collected from high river-bluff sections at Gakona in the northeastern Copper River Basin. Organic material buried with sediments deposited during the early phase of lake formation when the ancestral Copper River valley was being flooded was dated at greater than 38 ka (sample W–531). Organic material immediately overlying the lacustrine sediments and buried by cliff-head dune deposits was dated at 9.4±0.3 ka (W–714).

Because of the scientific importance of dating the initiation of the lake and the last major glaciation in the Copper River Basin, a very large sample of the older organic material was collected and submitted to the Groningen Laboratory, which uses the carbon-enrichment method to extend the range of radiocarbon dating. Unfortunately, the results were not conclusive. One measurement gave an absolute age of 43.44±0.25 ka (GrN–4086), but because this age was younger than two dates on nonenriched parts of the same sample (>46 ka and >49 ka, GrN–4165 and GrN–4448, respectively) the laboratory concluded that the sample was contaminated during the enrichment process. Consequently, the analysis was repeated and the sample yielded an absolute age of 58.6±1.1 ka (GrN–4798).

Under the circumstances, it is difficult to accept the age determination of 58.6±1.1 ka without corroborating evidence, but the series of dates by the Groningen Laboratory does suggest that the lake started to form before 50 ka, and the 9.4±0.3 ka age determination provides a good minimum date for lake draining. Additional age determinations of organic material collected from bluffs along the Sanford River, 17 km east of Gakona, indicate that the lake level was below 655 m between 31.3±1 ka (W–843) and 28.3±1 ka (W–1343). Thorson and others (1981) and Thorson (1984) presented evidence that subaerial conditions existed in the northwestern corner of the Copper River Basin at an altitude of approximately 700 m between 32 and 21 ka. Also, lake deposits exposed at 670 m altitude in a bluff along the Copper River at Slana, 80 km northeast of Gakona, are dated at about 17 ka (17.6±0.4, W–1134) (Schmoll, 1984). All of these dates, in conjunction with the stratigraphic relations, indicate that glacial Lake Atna and the last major glaciation in the Copper River Basin are, at least in part, comparable in age to the Wisconsin glaciation of central North America.

REFERENCES CITED

Coulter, H.W., Hopkins, D.M., Karlstrom, T.N.V., Péwé, T.L., Wahrhaftig, Clyde, and Williams, J.R., compilers, 1965, Map showing extent of glaciations in Alaska: U.S. Geological Survey Miscellaneous Geologic Investigations Map I–415, scale 1:2,500,000.

Emery, P.A., Jones, S.H., and Glass, R.L., 1985, Water resources of the Copper River Basin, Alaska: U.S. Geological Survey Hydrologic Investigations Atlas HA–686, 3 sheets.

Eyles, Nicholas, 1984, Sediments in large late Quaternary glaciolacustrine basins (Copper River, Alaska and Great Lakes, Canada): Application to pre-Quaternary sequences: Geological Society of America Abstracts with Program, v. 16, no. 5, p. 282.

Ferrians, O.J., Jr., 1963a, Till-like glaciolacustrine deposits in the Copper River Basin, Alaska: Geological Society of America Special Paper 73, p. 151.

———1963b, Glaciolacustrine diamicton deposits in the Copper River Basin, Alaska, *in* Geological Survey Research 1963—Short papers in geology and hydrology, 1963: U.S. Geological Survey Professional Paper 475–C, p. C121–C125.

———1971a, Preliminary engineering geologic maps of the proposed trans-Alaska pipeline route, Gulkana quadrangle: U.S. Geological Survey Open-File Report 71–102 (494), scale 1:125,000, 2 sheets.

———1971b, Preliminary engineering geologic maps of the proposed trans-Alaska pipeline route, Valdez quadrangle: U.S. Geological Survey Open-File Report 71–104 (495), scale 1:125,000, 2 sheets.

———1984, Pleistocene glacial history of the northeastern Copper River Basin, Alaska: Geological Society of America Abstracts with Programs, v. 16, no. 5, p. 282.

Ferrians, O.J., Jr., and Nichols, D.R., 1965, Copper River Basin, *in* Schultz, C.B., and Smith, H.T.U., eds., Guidebook for Field Conference F, central and south-central Alaska: Nebraska Academy of Sciences, International Association for Quaternary Research, VII Congress, Lincoln, Nebr., p. 93–114.

Ferrians, O.J., Jr., Nichols, D.R., and Williams, J.R., 1983, Copper River Basin, *in* Péwé, T.L., and Reger, R.D., eds., Guidebook to permafrost and Quaternary geology along the Richardson and Glenn Highways, between Fairbanks and Anchorage, Alaska: Alaska Division of Geological and Geophysical Surveys Guidebook 1, p. 137–175.

Ferrians, O.J., Jr., and Schmoll, H.R., 1957, Extensive proglacial lake of Wisconsin age in the Copper River Basin, Alaska: Geological Society of America Bulletin, v. 68, no. 12, p. 1726.

Karlstrom, T.N.V., and others, compilers, 1964, Surficial geology of Alaska: U.S. Geological Survey Miscellaneous Geologic Investigations Map I–357, scale 1:584,000.

Mendenhall, W.C., 1905, Geology of the central Copper River region, Alaska: U.S. Geological Survey Professional Paper 41, 133 p.

Moffit, F.H., 1954, Geology of the eastern part of the Alaska Range and adjacent area: U.S. Geological Survey Bulletin 989–D, 218 p.

Nichols, D.R., 1960, Slump structures in Pleistocene lake sediments, Copper River Basin, Alaska, *in* Geological Survey Research 1960—Short papers in the geological sciences 1960: U.S. Geological Survey Professional Paper 400–B, p. B353–B354.

———1965, Glacial history of the Copper River Basin: International Association for Quaternary Research, 7th Congress, Fairbanks, Alaska, 1965, Proceedings, Abstract volume, Lincoln, Nebraska Academy of Sciences, p. 360.

Nichols, D.R., and Yehle, L.A., 1961, Mud volcanoes in the Copper River Basin, Alaska, *in* Raasch, G.O., ed., Geology of the Arctic: Toronto, University of Toronto Press, v. 2, p. 1063–1087.

———1969, Engineering geologic map of the southeastern Copper River Basin, Alaska: U.S. Geological Survey Miscellaneous Geologic Investigations Map I–524, scale 1:125,000.

Richter, D.H., Smith, R.L., Yehle, L.A., and Miller, T.P., 1979, Geologic map of the Gulkana A–2 quadrangle, Alaska: U.S. Geological Survey Geologic Quadrangle Map GQ–1520, scale 1:63,360.

Schmoll, H.R., 1961, Orientation of phenoclasts in laminated glaciolacustrine deposits, Copper River Basin, Alaska, *in* Geological Survey Research 1961—Short papers in the geological and hydrological sciences: U.S. Geological Survey Professional Paper 424–C, p. C192–C195.

———1984, Late Pleistocene morainal and glaciolacustrine geology in the upper Copper River-Mentasta Pass area, Alaska: Geological Society of America Abstracts with Programs, v. 16, no. 5, p. 332.

Schrader, F.C., 1900, A reconnaissance of a part of Prince William Sound and the Copper River district, Alaska: U.S. Geological Survey 20th Annual Report, 1898–99, Pt. VII—Exploration in Alaska in 1898, p. 341–423.

Schrader, F.C., and Spencer, A.C., 1901, The geology and mineral resources of a portion of the Copper River district, Alaska: U.S. Geological Survey Special Publication, 94 p.

Thorson, R.M., 1984, Pattern and chronology of late Quaternary glaciation, northwest Copper River Basin: Geological Society of America Abstracts with Programs, v. 16, no. 5, p. 337.

Thorson, R.M., Dixon, E.J., Jr., Smith, G.S., and Batten, A.R., 1981, Interstadial Proboscidean from south-central Alaska: Implications for biogeography, geology, and archeology: Quaternary Research, v. 16, no. 3, p. 404–417.

Williams, J.R., 1984, Late Wisconsin glacial retreat and lake levels, western Copper River Basin, Alaska: Geological Society of America Abstracts with Programs, v. 16, no. 5, p. 340.

———1985, Engineering-geologic map of the southwestern Copper River Basin and upper Matanuska River valley, Alaska: U.S. Geological Survey Open-File Report 85–143, scale 1:125,000, 2 sheets.

Williams, J.R., and Galloway, J.P., 1986, Map of western Copper River Basin, Alaska, showing lake sediments and shorelines, glacial moraines, and location of stratigraphic sections and radiocarbon-dated samples: U.S. Geological Survey Open-File Report 86–390, scale 1:250,000.

Williams, J.R., and Johnson, K.M., compilers, 1980, Map and description of late Tertiary and Quaternary deposits, Valdez quadrangle, Alaska: U.S. Geological Survey Open-File Report 80–892C, scale 1:250,000, 2 sheets.

Yehle, L.A., 1980, Preliminary surficial geologic map of the Valdez C–1 quadrangle, Alaska: U.S. Geological Survey Miscellaneous Field Studies Map MF–1132, scale 1:63,360.

———1981, Preliminary surficial geologic map of the Valdez B–1 quadrangle, Alaska: U.S. Geological Survey Miscellaneous Field Studies Map MF–1364, scale 1:63,360.

Carter, L. David, Hamilton, Thomas D., and Galloway, John P., eds., 1989, Late Cenozoic History of the Interior Basins of Alaska and the Yukon: U.S. Geological Survey Circular 1026.

History of Late Pleistocene and Holocene Vegetation in the Copper River Basin, South-Central Alaska

By Thomas A. Ager

This paper provides an outline of vegetation history for the Copper River Basin of south-central Alaska (fig. 1) during the late Pleistocene and Holocene. Vegetation reconstructions are based upon analyses of pollen and spore assemblages from lacustrine sediment cores and outcrop samples from scattered sites within the basin and in adjacent areas. The chronology of vegetation changes is based upon radiocarbon-dated organic materials from these samples.

The glacial and glaciolacustrine deposits of the Copper River Basin have been the subject of several investigations during the past four decades. Results of these studies have been summarized recently by Ferrians and others (1983), Hamilton and Thorson (1983), and Williams and Galloway (1986).

Palynological investigations were begun in this region by Hansen (1953), who analyzed arboreal pollen in shallow peat cores from roadside muskegs found along the Richardson and Glenn Highways. The cores lacked radiocarbon control, however, and very few palynological data were included in the publication. What is presently known about the Quaternary vegetational history of the Copper River Basin is based upon recent work by Connor (1984), Ager and Brubaker (1985), and Ager (unpub. data). Sirkin and Tuthill (1987) summarized results of palynological and glacial geologic studies in the lower Copper River valley, but most of that area is outside the region included in the present paper.

The small number of sites investigated within the Copper River Basin is inadequate to permit as detailed a reconstruction of vegetation history as has been carried out in other areas of Alaska (Ager and Brubaker, 1985). The data are sufficient, however, to permit construction of a fragmentary history for the basin during the late Pleistocene, and of a nearly complete Holocene history. The pre-Holocene record is pieced together from radiocarbon-dated outcrop samples from scattered sites in the basin. These samples were collected by J.R. Williams, O.J. Ferrians, Jr., C.L. Connor, and T.A. Ager (all with the U.S. Geological Survey). The largest set of dated outcrop samples used in this palynological investigation was collected by J.R. Williams during fieldwork that spanned more than three decades (Williams and Galloway, 1986).

The Holocene record for the region is based upon lacustrine sediment cores from Seventymile Lake (Ager and Brubaker, 1985), Grizzly Lake (Ager, unpub. data), Tangle Lakes in the Gulkana Upland (Ager and Sims, 1981; Schweger, 1981), and Watana Pond (local usage) in the middle part of the Susitna River valley (Ager, unpub. data) (fig. 1). The pollen records from these cores are supplemented by radiocarbon-dated outcrop samples from several localities (Williams, 1986; Williams and Galloway, 1986; Ager, unpub. data).

The oldest known pollen samples of Pleistocene age from the region are from outcrops on the Dadina River (Connor, 1984) and Tyone Creek (Williams and Galloway, 1986). The samples from the Dadina River locality (fig. 1) yielded pollen assemblages dominated by *Picea* (spruce), *Betula* (birch), and *Alnus* (alder), along with Ericaceae (heaths), Cyperaceae (sedge), and various herbs (Connor, 1984). The sampled units have yielded infinite radiocarbon ages of >33.5 (I–12,321) and >40 ka (I–12,320; I–12,322). Connor (1984) interpreted these samples to represent boreal forest vegetation that occupied the lowlands of the Copper River Basin during a warm climate interval within the mid-Wisconsin interstadial, presumably during the early part of the interstadial. Interpretations of the chronology for a core from Imuruk Lake in western Alaska (Shackleton, 1982) suggested that an unusually warm interval may have occurred in Beringia in early mid-Wisconsin time. Other evidence from Beringia in support of Shackleton's interpretation was recently presented by Matthews and Schweger (1985) and in this volume.

Outcrop samples from Tyone Creek (fig. 1) collected by J.R. Williams (Williams and Galloway, 1986) also are associated with an infinite radiocarbon age of >35 ka (W–357), and their pollen assemblages are dominated by *Picea*, *Betula*, *Alnus*, and Cyperaceae (Ager, unpub. data). These assemblages are quite similar to those described by Connor (1984) from Dadina River. It is likely that the Tyone Creek and Dadina River pollen assemblages are correlative. Pollen data from both sites represent lowland boreal forest vegetation, and that they suggest climate conditions quite similar to those of the present day. It remains in the realm of speculation whether this interval of boreal forest development represents an unusually warm interval during the early mid-Wisconsin interstadial, or whether it represents an older interglacial interval.

An outcrop sample from Nelchina River (fig. 1) collected by J.R. Williams yielded a radiocarbon determination of >38 ka (W–842) (Williams and Galloway, 1986). The sampled horizon is overlain by till and overlies gravel outwash. The dominant pollen types in that sample (Ager, unpub. data) were herbaceous taxa such as Cyperaceae, Caryophyllaceae, and Gramineae. This assemblage suggests that treeless herbaceous tundra, probably a moist sedge meadow, covered at least some lowlands within the Copper River Basin before 38 ka. A nearby outcrop along the Nelchina River yielded a peat sample from a stratigraphically lower position than that which yielded a tundra pollen assemblage. The lower sample was from a sandy deltaic unit underlying the outwash unit. The peat yielded a radiocarbon determination of >38 ka (W–295). The pollen assemblage contained in the lower sample from Nelchina River was dominated by spruce pollen. This suggests boreal forest vegetation similar to that from Tyone Creek. It is not known, however, if the Nelchina sample with spruce pollen is the same age as those with boreal assemblages from Tyone Creek or Dadina River.

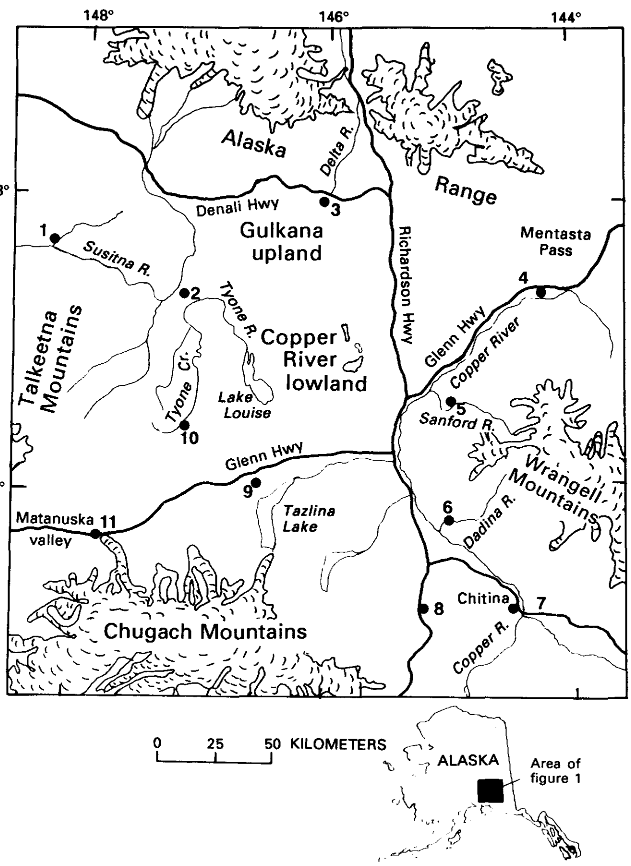


Figure 1. Map of Copper River Basin and adjacent areas, south-central Alaska. Numbers refer to localities discussed in text: 1, Watana Pond (Ager, unpub. data); 2, Tyone Bluff (Thorson and others, 1981); 3, Tangle Lakes (Ager and Sims, 1981; Schweger, 1981); 4, Grizzly Lake (Ager, unpub. data); 5, Sanford River (Ferrians and others, 1983; Ager, unpub. data); 6, Dadina River (Connor, 1984); 7, Chitina area (Ager, unpub. data); 8, Seventymile Lake (Ager and Brubaker, 1985); 9, Nelchina River (Williams and Galloway, 1986; Ager, unpub. data); 10, Tyone Creek (Williams and Galloway, 1986; Ager, unpub. data); 11, Matanuska Glacier (Williams, 1986; Ager, unpub. data).

An interstadial deposit from the Sanford River (fig. 1) was sampled for pollen analysis in 1985 by O.J. Ferrians, Jr. (U.S. Geological Survey). The slightly peaty fine sand and silt unit had been previously sampled for radiocarbon dating, and those samples had yielded bracketing ages of 31.3±1 ka (W–843) near the base of the unit and 28.3±1 ka (W–1343) near the top of the unit (Williams and Galloway, 1986; Ferrians and others, 1983). The sediments were deposited late within the mid-Wisconsin interstadial, at what may have been the only time during the entire Wisconsin when the shoreline of the large proglacial lake that occupied much of the basin dropped sufficiently to expose the Sanford River site subaerially, thus permitting vegetation to colonize.

The pollen data from eight samples collected from this interstadial unit provide important new evidence about the nature of lowland vegetation of this region during the final phase of the Wisconsin interstadial. The assemblages are dominated by pollen of Cyperaceae, Gramineae, Caryophyllaceae, *Artemisia*, *Phlox*, and other herbs. This indicates that lowland areas of the Copper River Basin that are now occupied by boreal forest (mostly spruce) were vegetated by treeless alpine tundra between about 28 and 31 ka. This evidence is consistent with limited plant macrofossil evidence reported previously from a site of approximately the same age at Tyone Bluff on the western edge of the basin (fig. 1; Thorson and others, 1981).

No evidence has yet been found that serves as a basis for reconstructing full-glacial vegetation in the region. Most of the Copper River Basin was covered by glacier ice and lake waters, limiting potential sites for plant colonization to nunataks and relatively small areas of lowlands above lake level. In view of the pollen evidence from Sanford River interstadial deposits, it is highly likely that any vegetation types that developed on nunataks during full glacial times were herbaceous tundra communities.

The vegetation that developed during late Wisconsin deglaciation can be reconstructed from samples taken at several localities in the Copper River Basin (Williams and Galloway, 1986). The Holocene record is based upon a combination of samples from outcrops and two lacustrine sediment cores (Ager and Brubaker, 1985; Ager, unpub. data).

The earliest known vegetation to develop in the basin during the waning of late Wisconsin glaciers was herb-shrub tundra dominated by Cyperaceae and dwarf *Betula*. This vegetation is known to have developed in the region as early as 13.28±0.4 ka (W–583)(Williams and Galloway, 1986; Connor, 1984; Ager, unpub. data). A somewhat older age of 13.9±0.4 ka (I–3796) was reported for the lower Copper River valley by Sirkin and Tuthill (1987).

Betula percentages abruptly increased during deglaciation, indicating the development of shrub tundra dominated by dwarf birch. Shrub tundra and shrub-herb tundra communities dominated the lowland landscape of the basin until early Holocene time, when *Populus* (balsam poplar or aspen trees) and *Alnus* (alder) shrubs rapidly invaded. The lowlands and lower slopes of uplands developed vegetation cover of poplar and willow stands, alder and willow shrub communities, and shrub tundra.

Current evidence indicates that *Picea* first invaded the Copper River Basin from the north by at least 9 ka. *Picea* apparently invaded from the Tanana River valley via Mentasta Pass and the Delta River valley (fig. 1) and spread rapidly westward and southward (Ager and Brubaker, 1985). *Picea glauca* (white spruce) was present at Tangle Lakes (fig. 1) by about 9.1 ka (Schweger, 1981). *Picea* reached the valleys of the northern Chugach Mountains (fig. 1) soon thereafter. Pollen evidence and radiocarbon ages bracket the arrival of *Picea* in that area between 8.48±0.13 ka (I–11,397)(Williams and Galloway, 1986; Ager, unpub. data) and 9.02±0.2 ka at 70 Mile Lake (W–5336) (Ager and Brubaker, 1985). Recently published evidence from the upper Matanuska River valley (fig. 1) indicates that *Picea* reached that area at least as early as 8.95±0.15 ka (I–14,096) (Williams, 1986). A buried spruce stump exposed in place near Chitina (fig. 1) yielded a radiocarbon date of 7.23±0.15 ka (W–5335, Ager, unpub. data). That age provides a minimum age for *Picea* in the Chitina and lower Copper River valleys.

With the establishment of spruce-dominated boreal forest in the lowlands of the Copper River Basin, the available evidence suggests that only minor vegetation changes occurred thereafter in the region. There is some suggestion of lowering of treeline in the vicinity of Seventymile Lake during the Neoglacial (Ager and Brubaker, 1985), but there is also puzzling evidence from the Gulkana Upland that suggests a local spruce reinvasion near altitudinal tree line at about the same time (Ager and Sims, 1981).

The studies done thus far provide only an outline of the region's history (fig. 2), but they reveal that the Quaternary deposits of the Copper River Basin contain a rich record of past environments and climates. Future palynological research that is needed in this region includes detailed studies of the Quaternary records in the vicinity of Mentasta Pass, and the central and northwestern areas of the basin.

Much of the data discussed in this paper were derived from analyses of samples collected by J.R. Williams and O.J. Ferrians, Jr. (U.S. Geological Survey). I gratefully acknowledge their willingness to share their samples and field data.

REFERENCES CITED

Ager, T.A., and Brubaker, L., 1985, Quaternary palynology and vegetational history of Alaska, *in* Bryant, V.M., Jr., and Holloway, R.G., eds., Pollen records of late-Quaternary North American sediments: American Association of Stratigraphic Palynologists Foundation, Dallas, p. 353–384.

Ager, T.A, and Sims, J.D., 1981, Holocene pollen and sediment record from the Tangle Lakes area, central Alaska: Palynology, v. 5, p. 85–98.

Connor, C.L., 1984, Late Quaternary glaciolacustrine vegetational history of the Copper River Basin, south-central Alaska: University of Montana, Ph.D. thesis, 115 p.

Ferrians, O.J., Jr., Nichols, D.R., and Williams, J.R., 1983, Copper River Basin, *in* Péwé, T.L., and Reger, R.D., eds., Richardson and Glenn Highways, Alaska: Guidebook to permafrost and Quaternary geology: Guidebook 1, International Conference on Permafrost, 4th, University of Alaska, Fairbanks, July 18–22, 1983: State of Alaska Department of Natural Resources, Division of Geological and Geophysical Surveys, p. 137–175.

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: v. 1, The late Pleistocene: Minneapolis, University of Minnesota Press, p. 38–52.

Hansen, H.P., 1953, Postglacial forests in the Yukon Territory and Alaska: American Journal of Science, v. 251, p. 505–542.

Matthews, J.V., Jr., and Schweger, C.E., 1985, Old Crow tephra: Its significance for understanding the early and middle Wisconsinan climate history of eastern Beringia, *in* Harington, C.R., ed., Climate change in Canada: Syllogeus, no. 55, p. 453–459.

Schweger, C.E., 1981, Chronology of Late Glacial events from the Tangle Lakes, Alaska Range, Alaska: Arctic Anthropology, v. 18, p. 97–101.

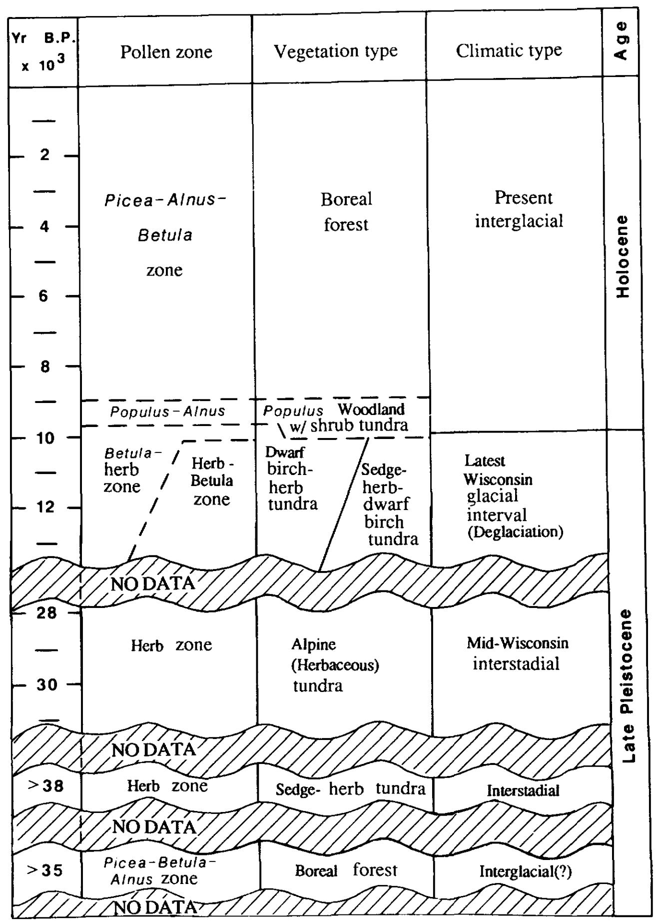


Figure 2. Summary of late Pleistocene and Holocene vegetational history of Copper River Basin and adjacent areas, south-central Alaska.

Shackleton, J., 1982, Environmental histories from Whitefish and Imuruk Lakes, Seward Peninsula, Alaska: Columbus, Ohio State University, Institute of Polar Studies Report 76, 50 p.
Sirkin, L.A., and Tuthill, S.J., 1987, Late Pleistocene and Holocene deglaciation and environments of the southern Chugach Mountains, Alaska: Geological Society of America Bulletin, v. 99, p. 376–384.
Thorson, R.M., Dixon, E.J., Jr., Smith, G.S., and Batten, A.R., 1981, Interstadial proboscidean from south-central Alaska: Implications for biogeography, geology, and archaeology: Quaternary Research, v. 16, p. 404–417.
Williams, J.R., 1986, New radiocarbon dates from the Matanuska Glacier bog section, *in* Bartsch-Winkler, Susan, and Reed, K.M., eds., Geologic studies in Alaska by the U.S. Geological Survey during 1985: U.S. Geological Survey Circular 978, p. 85–88.
Williams, J.R., and Galloway, J.P., 1986, Map of western Copper River Basin, Alaska, showing lake sediments and shorelines, glacial moraines, and location of stratigraphic sections and radiocarbon-dated samples: U.S. Geological Survey Open-File Report 86–390, scale 1:250,000, 30 p.

Carter, L. David, Hamilton, Thomas D., and Galloway, John P., eds., 1989, Late Cenozoic History of the Interior Basins of Alaska and the Yukon: U.S. Geological Survey Circular 1026.

Pleistocene Vertebrate Localities in the Yukon

By C. Richard Harington

Among ice-age vertebrate localities in Canada, those in unglaciated parts of the Yukon are most productive of fossils (Harington, 1985). The National Museum of Natural Sciences (NMNS) has been carrying out a long-term project of collecting and studying Pleistocene vertebrate fossils in the Yukon since 1966. Its collections from the Old Crow area have been greatly augmented by those of the Northern Yukon Research Program and the Yukon Refugium Project, both of which began about 1975. The following sections present previously unrecorded faunal lists (table 1) and radiocarbon ages (table 2) on Yukon Pleistocene vertebrates, as well as comments on evidence for early humans at Yukon localities (fig. 1).

Old Crow Basin

The Old Crow Basin (area 2 on fig. 1) is the richest area for collecting Pleistocene vertebrate remains in Canada, having yielded some 40,000 specimens representing nearly 60 mammal species, at least 7 fish species (Crossman and Harington, 1970; Cumbaa and others, 1981), and at least 33 bird species (G. R. Fitzgerald, NMNS, oral commun., 1985) during the last 20 years. Approximately 150 fossil localities are known within the basin. Remains of taxa found in place between Lower Lake (pre-Sangamon?) and Upper Lake (late Wisconsin) clays, as well as those recovered by intensive collecting at a productive point bar (loc. 11A) are listed in table 1 (Harington, 1977, 1985; Morlan, 1980b, 1984; Jopling and others, 1981). Abundant remains of large-horned bison that apparently died catastrophically about 12 ka have been collected in place above the Upper Lake clay at locality 11(1) (see Harington (1977) for locations of numbered vertebrate localities in the Old Crow Basin). Woolly mammoths survived in the basin until about that time, because a partial skeleton of a young individual collected in 1985 from locality 92 yielded a radiocarbon age of 13.82±0.84 ka (Beta–13867). Among the most interesting mammals represented are Jefferson's ground sloth, giant pika, giant beaver, short-faced bear, short-faced skunk, hyena, scimitar cat, gigantic camels, and the primitive Soergel's muskox.

When did people first reach the Yukon? Jopling and others (1981) claim that specimens of bone interpreted to be artifacts from the reworked Lower Lake beds at locality 12 are of pre-Sangamon age. A caribou antler pestle has been dated by radiocarbon at about 24.5 ka at the Chalk River accelerator mass spectrometer (AMS) of Atomic Energy Canada Ltd. (R.E. Morlan, Archaeological Survey of Canada, oral commun., 1984). However, recent AMS radiocarbon determinations on several artifacts that were thought to provide evidence for late Pleistocene human occupation of the Yukon have all given late Holocene ages (Nelson and others, 1986). For example, a caribou tibia fleshing tool originally dated at about 27 ka (Irving and Harington, 1973) and the jaw of a domestic dog from locality 11A (Beebe, 1980b) are now dated at 1.35±0.15 and 2.11±0.4 ka, respectively. A fragmentary mandible of a child from locality 11A, considered to be of possible Pleistocene age (Irving and others, 1977), needs similar checking, as does a caribou antler punch from Hunker Creek in the Dawson area, because a caribou antler in the same collection and in similar condition yielded a radiocarbon age of 23.9±0.47 ka (I–8580) (Harington, 1975). Despite the fact that no Pleistocene living floors with artifacts have been found so far, there is still strong circumstantial evidence for human Pleistocene occupation of the Yukon (for example, Morlan, 1980a, Harington, 1980).

Bluefish Caves

Bluefish Caves I–III (area 3 on fig. 1) have yielded a Pleistocene vertebrate fauna. Apparently the caves resulted from solution and congelifraction of ancient mixed limestone breccia. The fossils, from a loess bed underlying peaty material, yield radiocarbon ages between about 25 and 12 ka. Relatively recent specimens of small horse (12.9±0.1 ka, GSC–2881), mammoth (15.5±0.13 ka, GSC–3053), and saiga antelope have been collected here. The partial skull of the saiga gave an AMS radiocarbon age of about 13 ka (Jacques Cinq-Mars, Archaeological Survey of Canada, oral commun., 1986), suggesting the existence of steppe-like conditions toward the close of the last glaciation. Evidently most of the fossils accumulated in grassy tundra surroundings (Cinq-Mars, 1979, 1982). Artifacts found in the loess unit indicate that humans occupied the caves, perhaps hunting and butchering many of the large mammalian species, near the end of the last glaciaton.

Sixtymile

Since 1975, a gold placer site (Sixtymile locality 3) has yielded hundreds of excellent specimens, including remains of the relatively rare western camel (23.32±0.64 ka, Beta–8864), helmeted muskox, and tundra muskox (21.16±0.28 ka, Beta–13869). Carcasses of black-footed ferret from this site, and arctic ground squirrel (Walker,

Table 1. Ice Age vertebrate taxa recorded for various fossil localities in the Yukon Territory

[x, present;–,not collected or identified so far; cf., closely comparable specimen identified; *, extinct; **, extinct in Yukon Territory or reintroduced]

	1. Herschel Island (7 localities)	2. Old Crow basin (stratigraphic)			Locality 11A	3. Bluefish Caves I-III	4. Sixtymile locality 3	5. Dawson locality 10	6. Dawson localities 32-33	7. Dublin Gulch locality 1	8. Big Creek localities 1-3
		Reworked lower lake beds	Base of interlake beds	Upper interlake beds	Mixed	Loess zone		Hunker Creek (80 Pup)	Gold Run Creek		
	(Sangamon? to late Wisconsin)	(latest Illinoian?)	(Sangamon? 120 ka)	(early Wisconsin)	(early Pleistocene to Holocene)	(approx. 25 - 10 ka)	(middle to late Wisconsin)	(late Wisconsin)	(middle to late Wisconsin)	(middle Wisconsin)	(middle Wisconsin)
Fishes	-	x	x	x	x	x	-	-	-	-	-
Whitefish (*Coregonus* sp.)	-	x	x	x	-	-	-	-	-	-	-
Broad whitefish (*Coregonus nasus*)	-	x	x	x	-	-	-	-	-	-	-
Inconnu (*Stenodus leucicthys*)	-	-	cf.	x	-	cf.	-	-	-	-	-
Arctic grayling (*Thymallus arcticus*)	-	-	x	x	-	-	-	-	-	-	-
Pike (*Esox* sp.)	-	-	-	-	x	cf.	-	-	-	-	-
Sucker (*Catostomus* sp.)	-	x	x	x	-	-	-	-	-	-	-
Longnose Sucker (*Catostomus catostomus*)	-	x	-	x	-	-	-	-	-	-	-
Burbot (*Lota lota*)	-	-	x	x	-	-	-	-	-	-	-
Sculpin (*Cottus* sp.)	-	-	-	x	-	-	-	-	-	-	-
Birds	-	x	x	x	x	x	x	-	-	-	-
Loon (*Gavia* sp.)	-	-	-	-	x	-	-	-	-	-	-
Horned Grebe (*Podiceps auritus*)	-	-	-	-	x	-	-	-	-	-	-
Ducks and geese (Anatidae)	-	x	x	x	x	-	-	-	-	-	-
Goose (*Chen* sp.)	-	-	cf.	-	-	-	cf.	-	-	-	-
American widgeon (*Mareca americana*)	-	-	-	-	cf.	-	-	-	-	-	-
Oldsquaw (*Clanglua hyemalis*)	-	-	-	-	x	-	-	-	-	-	-
Scoter (*Melanitta* sp.)	-	-	-	-	x	-	-	-	-	-	-
Ptarmigan and grouse (Tetraonidae)	-	cf.	x	x	x	x	-	-	-	-	-
Shorebirds (Charadriiformes)	-	-	cf.	-	-	-	-	-	-	-	-
Perching birds (Passeriformes)	-	-	x	x	-	-	-	-	-	-	-
Mammals	x	x	x	x	x	x	x	x	x	x	x
Shrews (Soricidae)	-	x	x	x	x	x	-	-	-	-	-
*Plains shrew (*Planisorex dixonensis*)	-	-	-	-	cf.	-	-	-	-	-	-
Shrew (*Sorex* sp.)	-	-	-	x	-	-	-	-	-	-	-
Humans (*Homo* sp.)	-	-	-	-	x	-	-	-	-	-	-
*Jefferson's ground sloth (*Megalonyx jeffersonii*)	-	-	-	-	x	-	-	-	-	-	-
*Giant pika (*Ochotona* cf. *whartoni*)	-	x	x	x	x	-	-	-	-	-	-
Pika (*Ochotona princeps*)	-	x	x	-	x	-	-	-	-	-	-
Snowshoe hare (*Lepus americanus*)	-	x	x	x	x	x	-	-	-	-	-
**Arctic hare (*Lepus arcticus*)	-	x	x	x	x	x	-	-	-	-	cf.
Woodchuck (*Marmota monax*)	-	-	-	x	-	-	-	-	-	-	-
Arctic ground squirrel (*Spermophilus parryi*)	-	x	x	x	x	x	x	-	-	-	-
Beaver (*Castor canadensis*)	-	x	x	x	x	-	-	-	-	-	-
*Giant beaver (*Castoroides ohioensis*)	-	x	x	x	x	-	-	-	-	-	-
*Primitive lemming (*Predicrostonyx* sp.)	-	-	-	-	x	-	-	-	-	-	-
Collared lemming (*Dicrostonyx* sp.)	-	x	x	x	x	x	-	-	-	-	-
Brown lemming (*Lemmus sibiricus*)	-	x	x	x	x	x	-	-	-	-	-
Red-backed vole (*Clethrionomys rutilus*)	-	-	x	x	-	x	-	-	-	-	-
Muskrat (*Ondatra zibethicus*)	-	x	x	x	x	x	-	-	-	-	-
Heather vole (*Phenacomys intermedius*)	-	-	-	x	x	-	-	-	-	-	-
Singing vole (*Microtus miurus*)	-	x	x	x	x	x	-	-	-	-	-
Meadow vole (*Microtus pennsylvanicus*)	-	-	-	x	-	-	-	-	-	-	-
Tundra vole (*Microtus oeconomus*)	-	cf.	cf.	-	-	-	-	-	-	-	-
Yellow-cheeked vole (*Microtus xanthoganthus*)	-	x	cf.	x	x	x	-	-	-	-	-

Large whale (Cetacea cf. *Balaena mysticetus*)	x	-	-	-	-	-	-	-	-	-	-
Coyote (*Canis latrans*)	-	-	-	-	x	-	-	-	-	-	-
Wolf (*Canis lupus*)	-	cf.	x	x	x	x	x	x	cf.	-	-
Domestic dog (*Canis familiaris*)	-	-	-	-	x	-	-	cf.	-	-	-
Arctic fox (*Alopex lagopus*)	-	x	x	x	x	x	-	x	-	-	-
Red fox (*Vulpes vulpes*)	-	-	-	-	-	x	x	-	-	-	-
*Yukon short-faced bear (*Arctos simus yukonensis*)	-	-	-	-	x	-	-	x	x	-	-
Bear (*Ursus* sp.)	-	-	-	x	x	cf.	cf.	-	-	-	x
Black bear (*Ursus americanus*)	-	-	-	-	x	-	-	-	-	-	-
Brown bear (*Ursus arctos*)	-	-	-	-	-	-	cf.	-	-	-	x
Ermine (*Mustela erminea*)	-	-	-	x	x	x	-	-	-	-	-
Least weasel (*Mustela nivalis*)	-	-	-	-	-	x	-	-	-	-	-
**Black-footed ferret (*Mustela nigripes*)	-	-	-	-	-	x	x	-	-	-	-
Noble marten (*Martes nobilis*)	-	-	-	-	x	-	-	-	-	-	-
American marten (*Martes americana*)	-	-	-	-	x	x	-	-	-	-	-
Fisher (*Martes pennanti*)	-	-	x	-	x	-	-	-	-	-	-
Wolverine (*Gulo gulo*)	-	-	x	-	x	-	x	-	-	-	-
**Badger (*Taxidea taxus*)	-	-	-	-	-	-	-	x	x	-	-
*Short-faced skunk (*Brachyprotoma obtusata*)	-	-	x	-	-	cf.	-	-	-	-	-
Otter (*Lutra canadensis*)	-	-	-	-	x	-	-	-	-	-	-
*Hyaena (*Adcrocuta* sp.)	-	-	-	-	x	-	-	-	-	-	-
Mountain lion (*Felis concolor*)	-	-	-	-	-	x	-	-	-	-	-
Lynx (*Felis canadensis*)	-	-	-	cf.	x	-	-	-	-	-	-
*American lion (*Panthera leo atrox*)	-	-	-	x	x	x	x	x	x	x	x
*American scimitar cat (*Homotherium serum*)	-	-	-	-	x	-	-	-	-	-	-
Small seal (*Phoca* sp.)	x	-	-	-	-	-	-	-	-	-	-
*American mastodon (*Mammut americanum*)	-	-	-	-	x	-	-	-	x	-	-
*Mammoth (*Mammuthus* sp.)	x	x	x	x	x	x	x	x	x	cf.	x
*Southern mammoth (*Mammuthus meridionalis*)	-	-	-	-	x	-	-	-	-	-	-
*Woolly mammoth (*Mammuthus primigenius*)	x	cf.	x	x	x	cf.	x	x	x	cf.	x
**Horse (*Equus* sp.)	x	x	x	x	x	x	x	x	x	x	x
*Large horse (*Equus verae*)	-	x	x	x	x	-	cf.	cf.	-	-	-
*Scott's horse (*Equus scotti*)	-	-	-	-	cf.	-	-	cf.	-	-	-
*Small horse (*Equus lambei*)	x	-	-	-	x	x	x	x	x	x	x
**Kiang (*Equus (Asinus) kiang*)	-	-	-	-	-	-	-	-	x	-	-
*Large camel (Camelini)	-	x	-	-	x	-	-	-	-	-	-
*Western camel (*Camelops hesternus*)	-	-	-	-	-	-	x	-	-	-	-
**Wapiti (*Cervus elaphus*)	-	-	-	-	x	x	x	-	-	-	-
Moose (*Alces* sp.)	-	-	-	x	x	x	x	x	x	x	x
*Giant moose (*Alces latifrons*)	-	-	-	-	x	-	-	-	-	-	-
American moose (*Alces alces*)	-	-	-	-	x	cf.	x	x	cf.	cf.	-
Caribou (*Rangifer tarandus*)	-	x	x	x	x	x	x	x	x	x	x
**Bison (*Bison* sp.)	x	-	x	x	x	x	x	x	x	x	x
*Large-horned bison (*Bison priscus*)	cf.	-	-	-	x	cf.	x	x	x	cf.	x
*Alaskan bison (*Bison alaskensis*)	-	-	-	-	-	-	-	x	x	-	-
**Saiga antelope (*Saiga tatarica*)	-	-	-	-	-	x	-	-	-	-	-
*Muskoxen (Ovibovini)	x	x	x	-	x	x	x	-	x	-	-
*Soergel's muskox (*Soergelia* sp.)	-	x	-	-	x	-	-	-	-	-	-
*Helmeted muskox (*Symbos cavifrons*)	x	-	-	-	x	-	x	-	cf.	-	-
*Staudinger's muskox (*Praeovibos priscus*)	-	-	-	-	x	-	-	-	-	-	-
**Tundra muskox (*Ovibos moschatus*)	x	-	-	-	x	x	x	-	-	-	-
Dall sheep (*Ovis dalli*)	-	-	-	-	-	x	x	x	-	x	x

Note: In addition to species mentioned here: flat-headed peccary (*Platygonus compressus*) has been recorded from Old Crow locality 74 (Beebe, 1980a); steppe mammoth (*Mammuthus armeniacus* or *M. columbi*) is known from various Old Crow localities, including locality 22, as well as Dawson locality 8 (Quartz Creek) (Harington, 1977, 1980); dhole (*Cuon* sp.) is recorded from Old Crow locality 14N (Harington, 1977); western bison (*Bison bison occidentalis*) is reported from Old Crow River (Gilmore, 1908) and the Dawson area (Harington, 1977); black-footed ferret (*Mustela nigripes*) is tentatively considered to be conspecific with *Mustela eversmanni*(Old Crow localities 65 and 83); and *Bootherium sargenti* is considered to be a female of and conspecific with the Helmeted muskox (*Symbos cavifrons*).

Table 2. Radiocarbon ages for fossil vertebrates and plant macrofossils from the Yukon

Fossil	Radiocarbon age	Laboratory number
Old Crow Basin		
Mammoth	13,820±840	Beta-13867
Domestic dog	2,110±40	TO-276
Bluefish Caves		
Small horse	12,900±100	GSC-2881
Mammoth	15,500±130	GSC-3053
Sixtymile		
Western camel	23,320±640	Beta-8864
Tundra muskox	21,160±280	Beta-13869
Black-footed ferret	39,560±490	TO-214
Arctic ground squirrel	47,500±1,900	Beta-16157
Brown bear	36,500±1,150	Beta-16162
Mastodon	24,980±1,300	Beta-16163
Spruce wood	26,080±300	Beta-13870
Dawson Area		
Caribou	23,900±470	I-8580
Brown bear	41,000±1,050	Beta-16159
Short-faced bear	29,600±1,200	I-11037
Grass (ground squirrel nest)	12,200±100	GSC-2641
Alaskan bison	>39,000	I-5405
Long-horned bison	22,200±1,400	I-3570
Mammoth	32,350±1,750	I-4226
Dublin Gulch		
Horse	31,450±1,300	I-10935
Big Creek		
Willow wood	48,100±1,100	GSC-3032-2
Ketza River		
Bison	26,350±280	TO-393
Herschel Island		
Small horse	16,200±150	SFU:RIDDL-765

1984) from Sixtymile locality 4, a few kilometers away, gave AMS radiocarbon ages of 39.56±0.49 ka (TO–214) and 47.5±1.9 ka (Beta–16157), respectively (Youngman 1987). Brown bear bones from Sixtymile locality 3 and from Dawson locality 16 (Hunker Creek) yielded radiocarbon ages of 36.5±1.15 ka (Beta–16162) and 41±1.05 ka (Beta–16159), respectively, indicating that this species lived in the west-central Yukon before the peak of the last glaciation. Of particular interest is a radiocarbon age of 26.08±0.3 ka (Beta–13870) on a spruce stump in growth position—one of many such stumps in similar stratigraphic position at the site. In this regard, it is worth noting that a mastodon tooth from nearby Sixtymile locality 5 yielded a radiocarbon age on collagen of 24.98±1.3 ka (Beta–16163). Since American mastodons evidently preferred open spruce forest habitat, I am examining the possibility that spruce pockets existed in this region just before the peak of the last glaciation, if not later.

Dawson Area

Ice-age vertebrate remains near Dawson, like those at Sixtymile, Dublin Gulch, and Big Creek, are mainly exposed during placer mining for gold. Approximately 50 localities are recorded in the region. Most of the fossils, when found in place, occur in frozen organic silt just above the surface of gold-bearing gravel, and most date to the latter half of the Wisconsin glaciation (about 30 to 15 ka). Generally, fossils from this area are more complete than those from Old Crow Basin. Faunas from Hunker Creek (80 Pup) and Gold Run Creek are typical of the Dawson area.

Collecting at 80 Pup (Dawson loc. 10) began in 1973 and has continued to the present. During that period, about 260 ice-age mammal bones have been collected. Approximately 90 percent of the finds comprise large-horned bison (46 percent), small horse (19 percent), mammoth (11 percent), Dall sheep (11 percent), and caribou (3 percent). Of 14 bones found in place, their average height above the surface of the gold-bearing gravel is 1.5 m. Rarely, bones are found up to 5 m above the gravel. A humerus of a short-faced bear found in place in frozen organic silt 3 cm above the surface of the gold-bearing gravel yielded a radiocarbon age of 29.6±1.2 ka (I–11037), indicating the presence of this unusually long-limbed species in the region before the peak of the last glaciation (Harington, 1980). Most specimens are incomplete, but many good skulls of large-horned bison, Dall sheep, woolly mammoth, and one of an American lion, have been collected here. An interesting find in 1984 consisted of a nearly complete cranium of a young woolly mammoth with teeth and tusks (the latter about 53 cm long) from the interface of the gold-bearing gravel and the organic silt. Several skulls of arctic ground squirrels have been recovered, as well as their droppings and nesting grasses from nests frozen in the organic silt some 3.7 m above the surface of the gold-bearing gravel. Grasses from a ground squirrel nest on nearby Dominion Creek yielded a radiocarbon age of 12.2±0.1 ka (GSC–2641), which may give an idea of the age of the nests at 80 Pup. Pathological signs are occasionally noted too, for example a bison mandible with lesions on the ascending ramus, and a series of eight bison thoracic vertebrae with offset neural spines.

Gold Run Creek (Dawson locs. 31 through 33) is about 50 km southeast of Dawson. The area remained unglaciated during the Pleistocene, and constituted the southeastern extremity of the Beringian refugium. Two important specimens from this locality were described more than 70 years ago: the type skull of the Yukon short-faced bear (Lambe, 1911) and the type skull of the small horse (Hay, 1917). Other specimens of interest are badger, American mastodon, a nearly complete skull of a mammoth which was exhibited in Dawson in 1907 (Quackenbush, 1909), the first kiang horse described from Beringian Pleistocene deposits, and Alaskan bison (Harington and Clulow, 1973).

The badger, represented by a humerus, is the first record for the Yukon, and is significant because it indicates a former grassland link between the northern prairies and the interior of the Yukon and Alaska. Unfortunately a 1986 attempt at AMS radiocarbon dating of another badger specimen from Dominion Creek (Dawson loc. 28) failed. It is still a priority to establish when badgers entered and lived in the Yukon, because of the important paleoenvironmental and paleoclimatic implications. The deeply buried Alaskan bison horncore yielded an age of >39.9 ka (I–5405), one of the few non-finite dates on Pleistocene mammal bones from the region. If a North American ancestor for the giant bison (*Bison latifrons*) is sought, the Alaskan bison would be a good choice. Large-horned bison and mammoth bones from Gold Run Creek gave radiocarbon ages of 22.2±1.4 ka (I–3570) and 32.35±1.75 ka (I–4226), respectively.

Dublin Gulch

In 1967 I collected an astragalus of a small ungulate at Dublin Gulch, north of Mayo. About 70 bones of ice-age mammals subsequently were collected during the summer of 1977, presumably from an organic unit underlying glacial till and colluvium, and all but seven (kept as a representative sample in the Paleobiology Division of NMNS) were sent to the Kluane Museum of Natural History at Burwash Landing after identification. Species most commonly represented are small horse (38 percent), bison (36 percent), Dall sheep, and caribou (10 percent each). A horse metatarsal from the site has a radiocarbon age of 31.45±1.3 ka (I–10935), indicating that this fauna occupied the Mayo area during the last interstadial. This locality, like Big Creek and Ketza River, is of greater interest than usual because it has been glaciated.

Big Creek

So far, Pleistocene vertebrate remains have been collected from three tributaries of Big Creek: (1) Revenue Creek, where American lion, large-horned bison, and parts of a woolly mammoth skeleton were recovered in 1979 from nonglacial peaty beds that may be of last interstadial age (willow wood from the frozen peat yielded a radiocarbon age of 48.1±1.1 ka; GSC–3032–2); (2) Happy Creek, where mammoth, horse, caribou, large-horned bison, and Dall sheep were recorded in 1980 and 1983; and (3) Boliden Creek, where brown bear, woolly mammoth, small horse, caribou, bison, and Dall sheep remains were recorded in 1983. The brown bear cranium is in excellent condition and should be radiocarbon-dated.

Ketza River

About 40 fragmentary specimens of mammals, including hare, ground squirrel, collared lemming, mammoth, small horse, moose, and bison were collected in 1986 and 1987 from Wisconsin interstadial deposits underlying till of the McConnell glaciation. *Bison* bone found in place yielded an AMS radiocarbon age of 26.35±0.28 ka (TO–393). This is the most southerly Pleistocene vertebrate locality yet recorded in the Yukon.

Herschel Island

In contrast to Ketza River, Herschel Island Pleistocene vertebrate sites are the most northerly in the Yukon, and are the only ones that have produced in-place marine mammal remains. Although none of the marine mammal bones has been dated, a possible bowhead whale specimen was in place in "preglacial" (pre-early Wisconsin?) marine sand, and a seal scapula fragment came from oxidized organic deposits (including marine mollusk shells) approximately 5 m above present sea level, which I suspect are of last interglacial age (Harington, 1977). On the northeastern coast of the island, a partial cranium of a small horse from silty sand that postdates the Buckland glaciation of inferred early Wisconsin age has yielded an AMS radiocarbon age of 16.2±0.15 ka (SFU:RIDDL–765). Because seven scattered fossil sites were discovered during a relatively short period of time, a systematic investigation for ice-age vertebrate remains is planned for Herschel Island.

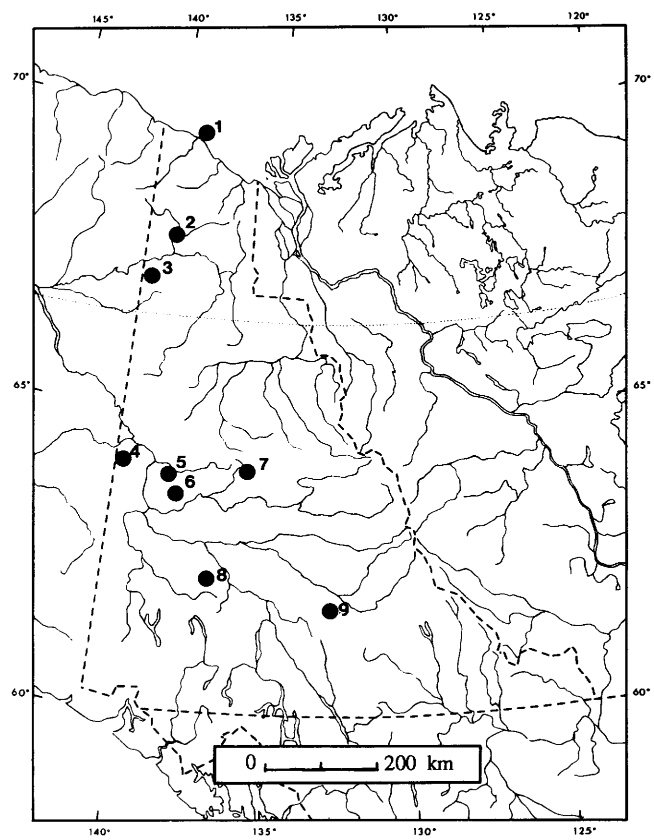


Figure 1. Yukon Territory (dashed outline), showing main Pleistocene vertebrate sites referred to in text: 1, Herschel Island; 2, Old Crow Basin; 3, Bluefish Caves; 4, Sixtymile; 5, Dawson area (80 Pup); 6, Dawson area (Gold Run Creek); 7, Dublin Gulch; 8, Big Creek; 9, Ketza River.

Acknowledgments

I am most grateful to the following people for their assistance: *Old Crow Basin*—P. Lord, C. Thomas, and G.R. Fitzgerald for field assistance; B.F. Beebe, R.E. Morlan, and W.N. Irving for their great paleobiological contributions; W.N. Irving and T.N. Friesen for supplying their list of taxa from locality 11A; *Bluefish Caves*—J. Cinq-Mars for supplying a current list of species from the loess unit (R.E. Morlan identified the microtine rodent remains and P.M. Youngman the small carnivores) and generously allowing me to quote some radiocarbon dates he obtained; *Sixtymile*—W. Yaremcio, A. Brisebois, and particularly C. and L. McDougall for their great generosity in donating important specimens and for their continued hospitality; the late R. Hughes, a most helpful coworker on the geology of the site; *Dawson (80 Pup)*—E. Raguth, V. Hall, and particularly T. Kosuta for donation of important specimens and hospitality; *Dawson (Gold Run Creek)*—L. Ross, J. Lamontagne, S. Mynott, and particularly E. Schink for donating important specimens and hospitality; W.Y. Watson and F.V. Clulow were instrumental in purchasing many specimens from the site and ultimately donating the bulk of them to the NMNS in July 1980; *Dublin Gulch*—F. Taylor, and particulary D. Duensing and R. Holway for collecting the specimens; M. Rogers (Kluane Museum, Burwash Landing) for donating a representative sample to the NMNS; R.E. Morlan for providing a copy of his 1977 field notes on the locality; *Big Creek*—R. McCauley and particularly K. Djukastein and G. MacPherson for donating important specimens and hospitality; Stephen Morison, O.L. Hughes, and R.W. Klassen for stratigraphic and radiocarbon information, and R.L. Debicki for arranging transportation to the site in 1980; *Ketza River*—L.E. Jackson, Jr., for discovering the locality, collecting and transferring important fossils to the NMNS, and supplying critical stratigraphic information; *Herschel Island*—D. Gill, M. Bouchard, V. Rampton, J. Hunston, R. Collier, S. Dallimore, and S. Wolfe; and the Canadian Wildlife Service for allowing me to accompany their biologists on a flight from Old Crow to Herschel Island in 1973.

REFERENCES CITED

Beebe, B.F., 1980a, Pleistocene peccary, *Platygonus compressus* Le Conte, from Yukon Territory, Canada: Canadian Journal of Earth Sciences, v. 17, p. 1204–1209.

———1980b, A domestic dog (*Canis familiaris* L.) of probably Pleistocene age from Old Crow, Yukon Territory, Canada: Canadian Journal of Archaeology, v. 4, p. 161–168.

Cinq-Mars, J., 1979, Bluefish Cave I: A Late Pleistocene Eastern Beringian cave deposit in the northern Yukon: Canadian Journal of Archaeology, v. 3, p. 1–32.

———1982, Les grottes du Poisson-Bleu: Geos, v. 11, no. 1, p. 19–21.

Crossman, E.J., and Harington, C.R., 1970, Pleistocene pike, *Esox lucius*, and *Esox* sp., from the Yukon Territory and Ontario: Canadian Journal of Earth Sciences, v. 7, p. 1130–1138.

Cumbaa, S.L., McAllister, D.E., and Morlan, R.E., 1981, Late Pleistocene fish fossils of *Coregonus*, *Stenodus*, *Thymallus*, *Catostomus*, *Lota* and *Cottus* from Old Crow Basin, northern Yukon, Canada: Canadian Journal of Earth Sciences, v. 18, no. 11, p 1740–1754.

Gilmore, C.W., 1908, Smithsonian exploration in Alaska in 1907 in search of Pleistocene fossil vertebrates: Smithsonian Miscellaneous Collection, v. 51, no. 3, p. 1–38.

Harington, C.R., 1975, A bone tool found with ice age mammal remains near Dawson City, Yukon Territory: Arctic Circular, v. 23, no. 1, p. 1–5.

———1977, Pleistocene mammals of the Yukon Territory: Edmonton, University of Alberta, Ph.D. thesis, 1060 p.

———1980, Radiocarbon dates on some Quaternary mammals and artifacts from northern North America: Arctic, v. 33, no. 4, p. 815–832.

———1985, Comments on Canadian Pleistocene mammals: Acta Zoologica Fennica, v. 170, p. 193–197.

Harington, C.R., and Clulow, F.V., 1973, Pleistocene mammals from Gold Run Creek, Yukon Territory: Canadian Journal of Earth Sciences, v. 10, p. 697–759.

Hay, O.P., 1917, Description of a new species of extinct horse, *Equus lambei*, from the Pleistocene of Yukon Territory: Proceedings of the United States National Museum, v. 53, p. 435–443.

Irving, W.N., and Harington, C.R., 1973, Upper Pleistocene radiocarbon-dated artefacts from the northern Yukon: Science, v. 179, p. 335–340.

Irving, W.N., Mayhall, J.T., Melbye, F.J., and Beebe, B.F., 1977, A human mandible in probable association with a Pleistocene faunal assemblage in eastern Beringia: A preliminary report: Canadian Journal of Archaeology, v. 1, p. 81–93.

Jopling, A.V., Irving, W.N., and Beebe, B.F., 1981, Stratigraphic, sedimentological and faunal evidence for the occurrence of pre-Sangamonian artefacts in northern Yukon: Arctic, v. 34, no. 1, p. 3–33.

Lambe, L.M., 1911, On *Arctotherium* from the Pleistocene of Yukon: Ottawa Naturalist, v. 25, p. 21–26.

Morlan, R.E., 1980a, Taphonomy and archaeology in the Upper Pleistocene of the northern Yukon Territory: A glimpse of the peopling of the New World: National Museum of Man Mercury Series, Archaeological Survey of Canada Paper No. 94, p. 1–380.

———1980b, Yukon Refugium Project, *in* Marois, R.J., ed., Archaeological Survey of Canada Annual Reviews 1977–1979: National Museum of Man Mercury Series, Archaeological Survey of Canada Paper No. 95, p. 65–68.

———1984, Biostratigraphy and biogeography of Quaternary microtine rodents from northern Yukon Territory, Eastern Beringia: Carnegie Museum of Natural History Special Publication No. 8, p. 184–199.

Nelson, D.E., Morlan, R.E., Vogel, J.S., Southon, J.R., and Harington, C.R., 1986, New dates on northern Yukon artifacts: Holocene not upper Pleistocene: Science, v. 232, p. 749–751.

Quackenbush, L.S., 1909, Notes on Alaskan mammoth expeditions of 1907 and 1908: Bulletin of the American Museum of Natural History, v. 26, p. 87–130.

Walker, L., 1984, The "moon-egg" that contained an arctic ground squirrel: BIOME, v. 4, p. 1.

Youngman, P.M., 1987, Freeze-dried mummies are hot items with paleontologists: BIOME, v. 7, no. 2, p. 4.

Carter, L. David, Hamilton, Thomas D., and Galloway, John P., eds., 1989, Late Cenozoic History of the Interior Basins of Alaska and the Yukon: U.S. Geological Survey Circular 1026.

Arctic Microtine Biochronology—Current Status

By Charles A. Repenning

The microtine rodents include the lemmings, muskrats, meadow mice, and similar small-eared, usually short-tailed mice; the name derives from their small ears. They live, and have lived, only in the Northern Hemisphere. The microtines are polyphyletic and include five lineages that derive from different, large-eared ancestors within the Family Cricetidae. Most of these lineages have evolved faster than any other mammal whose history is known; as a result, they are noted for recognizable changes in shorter periods of time than are other mammals. This makes their biochronology more precise for interpreting geologic time.

The biochronology of the microtine rodents of the Northern Hemisphere spans the past 6.5 m.y. and was first reconstructed about 10 years ago (Repenning and Feijfar, 1977). Subsequently, its factual foundations and its network of correlations with other methods of inferring geologic time have been greatly enlarged and the confidence of dates based upon the microtine biochronology has been accordingly strengthened.

Because of varying precision of temporal control, the 6-m.y.-long microtine biochronology is stronger during some intervals than in others. However, on the average, this biochronology can distinguish time differences as brief as 250 k.y., and this is sufficient to identify magnetic polarity events. Microtine dates combined with polarity determinations are significantly refined in dating precision; in many cases the uncertainty is reduced by half. Under ideal conditions, with both associated magnetic and climatic sequences, this biochronology enables a correlation of climatic changes on the continents with the oxygen-isotope stages in the oceans. In such cases, the uncertainty in dating is nearly reduced to the errors necessitated by the assumptions of constant depositional rates in oceanic deposits.

Most microtine rodents evolved in the Arctic region, and the microtine biochronology is very applicable there. The Beringian segment of the Arctic Ocean borderland is unique in that there the "microtine east" meets the "microtine west." Thus Beringia holds many of the biochronologic keys of the entire Arctic Ocean borderland, and these keys are significant also to the biochronologies of other parts of the Northern Hemisphere at lower latitudes. Because of this intermediate position, an understanding of only North American (Repenning, 1987) or of only Eurasian microtine biochronology is not enough to fully comprehend the temporal significance of Beringian fossil microtines.

Twenty years ago the Arctic Ocean borderland was an area completely devoid of information about pre-Pleistocene, or even pre-Illinoian, land mammal faunas. No information was known about Arctic microtine rodents older than latest Pleistocene until 1971, when Guthrie and Matthews described the Cape Deceit fauna from the Seward Peninsula. At that time, there was only a suspicion in a few minds that microtine rodents might provide a worldwide biochronology, and thus the temporal significance of the Cape Deceit fauna could not be realized. In addition, the taxonomic composition of the fauna had little chance of being fully recognized.

The breakthrough in Beringian microtine biochronology came in 1979 through the efforts of A.V. Sher, the principal student of the Kolyma lowlands in eastern Yakutia, U.S.S.R., and his associates, particularly V.S. Zazhigin, the microtine specialist. Sher directed an interdisciplinary study of the Krestovka section in the Kolyma lowlands that identified many significantly primitive microtine faunas. These were tied to paleomagnetic determinations and to climate indicators. Today the Krestovka section is the "key" to the microtine biochronology of Beringia and the Arctic Ocean borderland (see fig. 1).

The microtine faunas of the Krestovka section that have sufficient temporal control to be of use in biochronology are from that part of the section that is between 3 and 2 Ma. Faunal assemblages younger than 2 Ma are present in the Krestovka section, but paleomagnetic correlations are not clear, and the faunas are not sufficiently definitive to indicate age. The useful faunas are the biochronologic basis for understanding the age of the Cape Deceit fauna and of a fauna from Fish Creek on the Alaskan North Slope (Repenning and others, 1987; see fig. 1).

Through the Krestovka section an intercontinental biochronologic correlation network has been established that is the foundation for future research. The network also provides insight into the evolutionary pattern of several lineages that explains anomalous faunal relations that have hampered interpretation of arctic faunas, as well as of faunas in lower latitudes. These anomalies include the earlier record in Asia of many lineages, the abrupt appearance of many genera, as *Microtus*, *Phenacomys*, *Mictomys*, *Pitymys*, *Allophaiomys*, and *Dicrostonyx* in both Europe and North America with no record of where they evolved, and the asynchronous appearance of evolutionary lineages in different parts of the Arctic Ocean borderland.

The earliest records of at least four microtine genera are found in the Krestovka section of western Beringia. These records are between 200 and 400 k.y. older than the oldest records in Europe and North America (Repenning,

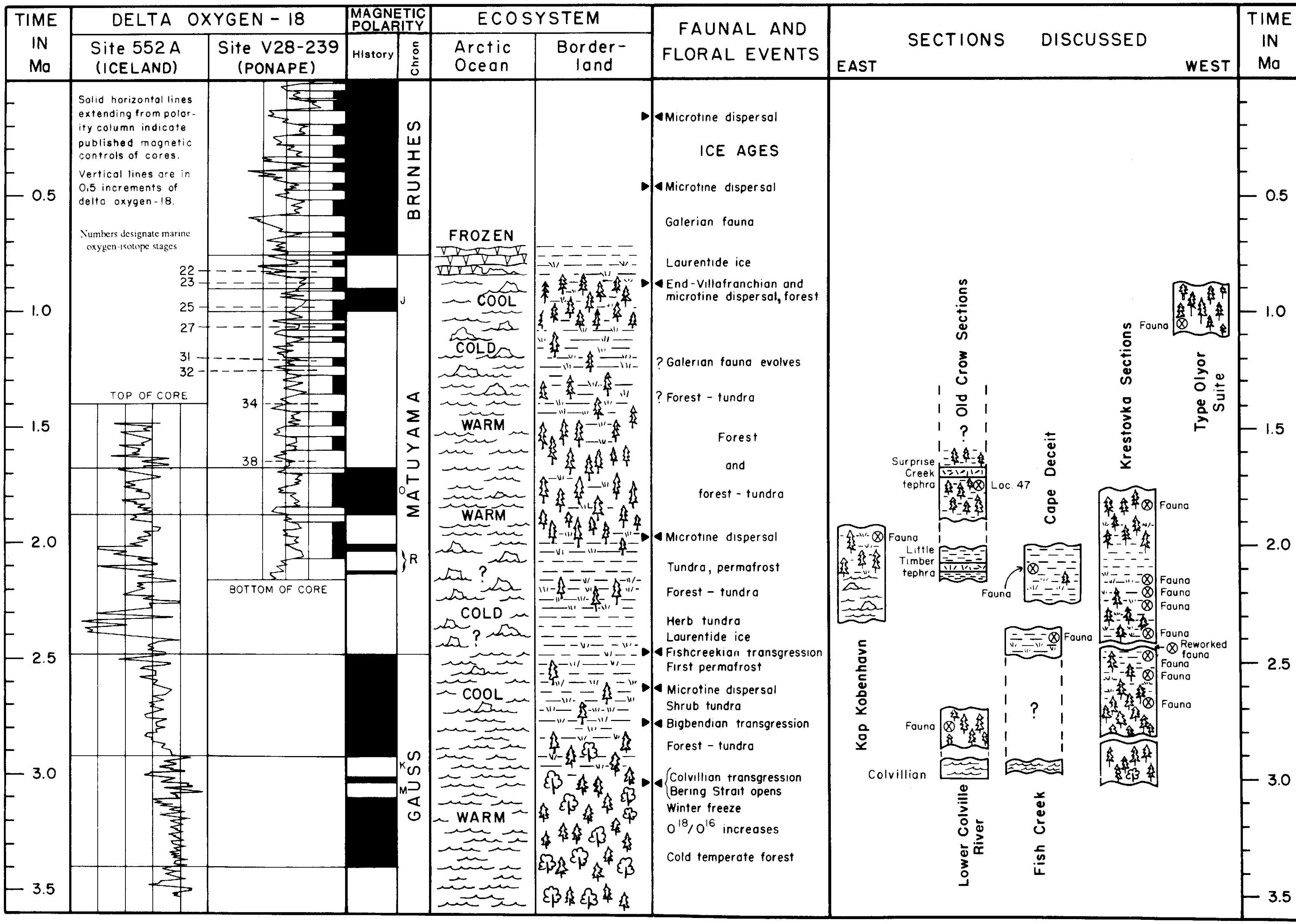


Figure 1. Current status of arctic microtine biochronologic correlations. Oxygen isotope records replotted from original data (Iceland: Shackleton and others, 1984; Ponape: Shackleton and Opdyke, 1976) to a common temporal scale; no paleomagnetic control intended below the Olduvai event for site V28–239 nor above it for site 552 A. J, Jaramillo Normal-Polarity Subchron; O, Olduvai Normal-Polarity Subchron; R, the two brief events of the Reunion Normal-Polarity Subchron; K, Kaena Reversed-Polarity Subchron; and M, Mammoth Reversed-Polarity Subchron.

1984), which clearly confirms previous interpretations that these four genera must have been immigrants to lower latitudes, presumably from Siberia (Kowalski, 1966).

Delays in the dispersals of some microtines from Asiatic Beringia to North American Beringia can now be demonstrated, and similar delays for dispersal of some large mammals from Beringia to other areas of the Northern Hemisphere are also beginning to be recognized as a result of Sher's study in the Kolyma lowlands. Such understanding of the origins and dispersals of faunas provides a clearer concept of the historic framework of biochronology, particularly the biochronology of the Arctic Ocean borderland. Many mammalian types evolved in this region, and they have an older temporal significance here than in lower latitudes.

In direct contrast to what was possible as little as three years ago, it is highly likely that any adequate microtine fauna found in the Canadian-Alaskan arctic can be convincingly dated to within about 200 ka if it is between 3 and 2 Ma. Dating with even greater precision is possible if the paleomagnetic pattern is known; dating with still greater precision is feasible if the climatic pattern is also known.

However, the same cannot be said of microtine faunas between 2 and 1 Ma, because there is no as yet "key" section for this period of time. Even in the well-known areas of Europe and the conterminous United States, mammalian biochronology is weak for the period between the Olduvai and Jaramillo Normal-Polarity Subchrons of the Matuyama Reversed-Polarity Chron. The beginning and the end of this time are well dated and are easily identifiable in the lower latitudes by the immigration of microtines that completely alter the faunal composition, but there is as yet no well-documented biochronology for the intervening 500 k.y. These immigrants came from the north, and their ancestors are found in older sediments of Beringia. Microtines of the older immigration have been found in Beringian deposits older than 2 Ma. Those of the immigration 850 ka are still poorly known in the Arctic, but should be found in Beringian deposits that formed between 2 and 1 Ma. At present only one Beringian locality in Canada (Old Crow locality 47; R.E. Morlan, oral commun., 1985), and one in Yakutia (type locality of the Olyor Suite; Sher, 1971) are known that date from this time span, but studies of these two localities are already improving biochronologic understanding of this vague period (see fig. 1). In fact, the correlation of events during this period, at present, is better understood in the Arctic Ocean borderland than it is in lower latitudes (Repenning, 1987).

This long period of poorly known microtine history ended 850 ka with the southward dispersal of arctic microtines and other mammals. In North America, this dispersal included both large and small mammals and appears to have resulted from the southward advance of the first of the several "Nebraskan" ice sheets. In Eurasia, the southward dispersal of microtines was accompanied by a conspicuous dispersal of large mammals from Beringian and, presumably, other Siberian areas. This was called the "End-Villafranchian" dispersal event by Azzaroli (1983). The large-mammal fauna was present in western Beringia before the "End-Villafranchian" dispersal event.

There were two subsequent dispersals into the conterminous United States, and a greater number into Europe. These introduced new species to the faunas of these low-latitude areas, but the species were of lineages whose evolution is difficult to understand. The evolution of some seems to be evident in the younger faunas of the Old Crow Basin, Yukon Territory, Canada, but much work remains before this is understood. The dates of the arrival in the conterminous United States of these new species, coupled with an understanding of their stage of evolution derived from study of the Old Crow or other material, may assist in approximating the age of these arctic faunas and help establish the later microtine biochronology of the Arctic Ocean borderland.

Figure 1 illustrates the current status of microtine correlations in the Arctic Ocean borderland, based upon the microtine biochronology as correlated with climatic and paleomagnetic evidence. Kap Kobenhavn is in northernmost Greenland, and the Old Crow sections are in eastern Beringia in the Yukon Territory, Canada. Study of these two sites is in progress, and there appears to be considerable time that is not represented. Fish Creek and the lower Colville River are in the central part of the North Slope of Alaska, and Cape Deceit is on the Seward Peninsula of Alaska, central Beringia. The Krestovka sections and the type section of the Olyor Suite are in eastern Yakutsk A.S.S.R., western Beringia.

REFERENCES CITED

Azzaroli, Augusto, 1983, Quaternary mammals and the "End-Villafranchian" dispersal event—A turning point in the history of Eurasia: Paleogeography, Paleoclimatology, Paleoecology, v. 44, p. 117–1139.

Guthrie, R.D., and Matthews, J.V., Jr., 1971, The Cape Deceit fauna—Early Pleistocene mammalian assemblage from the Alaskan arctic: Quaternary Research, v. 1, p. 474–510.

Kowalski, Kazimierz, 1966, The stratigraphic importance of rodents in the studies on the European Quaternary: Folia Quaternaria, Krakow, no. 22, 16 p.

Repenning, C.A., 1984, Quaternary rodent biochronology and its correlation with climatic and magnetic stratigraphies, *in* Mahaney, W.C., ed., Correlation of Quaternary chronologies: Norwich, England, Geo Books, p. 105–118.

———1987, Biochronology of the microtine rodents of the United States, *in* Woodburne, M.D., ed., Cenozoic mammals of North America: Geochronology and biochronology: Berkeley, University of California Press, p. 236–268.

Repenning, C.A., Brouwers, E.M., Carter, L.D., Marincovich, Louie, Jr. and Ager, T.A., 1987, The Beringian ancestry of *Phenacomys* (Rodentia: Cricetidae) and the beginning of the modern Arctic Ocean borderland biota: U.S. Geological Survey Bulletin 1687, 31 p.

Repenning, C.A., and Feijfar, Oldrich, 1977, Holarctic correlations of microtid rodents, *in* Sibrava, Vladimir, ed., Quaternary glaciations in the Northern Hemisphere: Report 4, On the session in Stuttgart, Federal Republic of Germany, September 1976: Prague, Czechoslovakian Geological Survey, p. 234–259.

Shackleton, N.J., and Opdyke, N.D., 1976, Oxygen-isotope and paleomagnetic stratigraphy of Pacific core V28–239: Late Pliocene to latest Pleistocene: Geological Society of America Memoir 145, p. 449–464.

Shackleton, N.J., and others, 1984, Oxygen isotope calibration of the onset of ice-rafting and history of glaciation in the North Atlantic region: Nature, v. 307, p. 620–623.

Sher, A.V., 1971, Pleistocene mammals and stratigraphy of the far northeast U.S.S.R. and North America: Moscow, U.S.S.R. Academy of Science (in Russian; translation in International Geology Review, 1974, v. 16, nos. 7–10, separate pagination, 284 p.)

Sher, A.V., Kaplina, T.N., Kouznetsov, Yu.V., Virina, E.I., and Zazhigin, V.S., 1979, Late Cenozoic of the Kolyma Lowland: XIV Pacific Science Congress, Tour Guide XI, Academy of Sciences U.S.S.R., Moscow, 115 p.

Carter, L. David, Hamilton, Thomas D., and Galloway, John P., eds., 1989, Late Cenozoic History of the Interior Basins of Alaska and the Yukon: U.S. Geological Survey Circular 1026.

Comment on the Present Status of Dating Loess by Thermoluminescence

By Glenn W. Berger

T.L. Péwé (this volume) has pointed out the ambiguities in the chronology of sediments (mostly loess and ash) in the Fairbanks region, and has mentioned the present confusion about the reliability of the thermoluminescence (TL) dating methods as applied to loess. This comment addresses problems of TL methodology and suggests probable causes for these problems.

Considering all the published indirect arguments (for example, stratigraphic and paleoclimatic) for assigning ages to the Sheep Creek tephra (Hamilton and Bischoff, 1984) and Old Crow tephra (Westgate and others, 1985), one is left with the realization that the true ages could be anywhere within the respective time ranges of 60 to 110 ka and 80 to 160 ka (Hughes, this volume; Matthews and others, this volume). Adding to this confusion are reported discrepant TL age estimates for the Old Crow tephra of about 85 ka (Wintle and Westgate, 1986) and about 150 ka (Berger and Huntley, 1985). Clearly, it is important to resolve the questions about the accuracy of present TL methods, because TL appears to offer the only direct means for dating the relevant deposits.

Methodology and discordancies

The view here is that improper application of previously introduced techniques (Wintle and Huntley, 1980), coupled with lack of rigor in tests for anomalous fading and for sensitivity changes, has created the present confusion about the reliability of the current TL dating methods.

Consider first the data in figure 1, where all published results using two TL methods are plotted. Almost all of the partial bleach data are from Norton and Bradford (1985). A few of these are incorrect, probably because of failure to recognize and correct for anomalous fading, overbleaching of laboratory samples, and use of linear extrapolations to dose-response curves for older samples. These are my inferences based on experiments with fine-grained sediments (see Berger, 1986), because no reasons for the incorrect results were discussed by Norton and Bradford. Whatever the causes for the scatter of these few partial bleach results, there is far more discordancy in the regeneration results of Debenham (1985).

Now this discordancy is interesting because these data are from the same set used by Debenham to hypothesize a physical phenomenon (decay of luminescence centers) to explain the discordancy of another part of this data set (those results for deposits thought to be older than 80 to 100 ka). This hypothesis predicts that all TL results (by any method) for deposits younger than about 50 ka should undershoot the correct age by <10 percent. Debenham did not comment on the evident inconsistency of a large number of his own results (fig. 1) with this prediction of his hypothesis.

My view in this note is that uncorrected sensitivity changes (for example, Rendell and others, 1983) and (or) anomalous fading have led to the incorrect results in figure 1, and very probably to all of the remaining discordancies reported by Debenham (notwithstanding that some of the "known ages" of Debenham's samples are questionable (Mejdahl, 1986)). Thus, Occam's Razor has not been applied to the interpretation of these (and other) published discordancies.

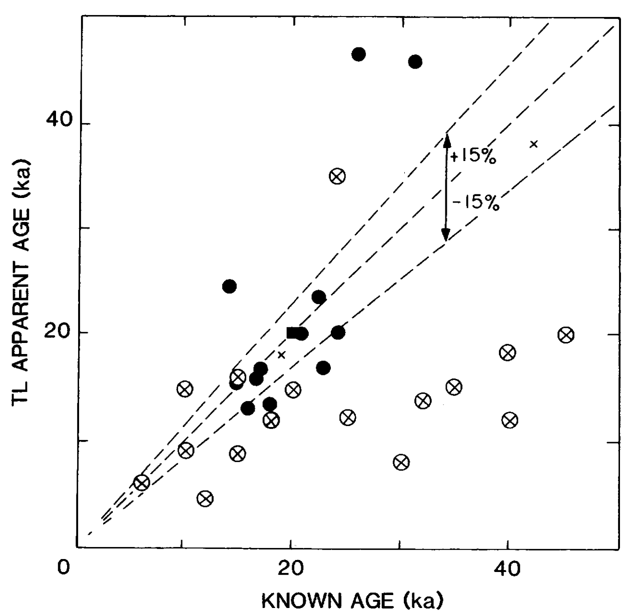


Figure 1. Plot of published thermoluminescence (TL) age estimates versus known ages (mostly associated C^{14} dates, when accurate to less than 10 percent) for loess. Results from partial bleach method (dot, from Norton and Bradford, 1985; square, from Wintle and Brunnacker, 1982) and regeneration method (x, from Prosynska, 1983; circle with an x, from Debenham, 1985). Results between outermost dashed lines are accurate to within 15 percent of the known age of the sample.

Probable assignable causes

The above view is supported by the following arguments. First, when the regeneration dose-response curve becomes sublinear, the practice of several workers to estimate sensitivity changes (Debenham, 1985; Canfield, 1985; Singhvi and others, 1987) has been to compare the slopes of tangents to this curve with those to an additive-dose curve. Esoteric arguments can be made why this practice may be invalid, but it is questionable for a more obvious reason—the choice of tangential slope is subject to very large relative errors, partly due to the limited reproducibility of TL data. Consequently, the uncertainties in these slopes have been as large as 40 percent(!) and typically are 20 to 30 percent (Debenham, 1985; Canfield, 1985; Singhvi and others, 1987). Yet these authors have dismissed this variation as insignificant, even though for sublinear growth curves small changes in the interpolated TL can lead to large changes in the estimated equivalent dose values (for example, Wintle, 1985). The obvious way to test this explanation is to apply the partial bleach method to the same samples that have yielded underestimated regeneration ages, because the partial bleach method is not subject to this type of sensitivity change. This test has not been done, probably for the reasons stated below.

Second, anomalous fading (up to 25 percent) has been reported for the same samples of loess from the Fairbanks region (Berger and Huntley, 1985; Berger, 1987) for which Wintle and Westgate (1986) reported none. Also, several other workers have reported no significant anomalous fading in loess. The reason for this discrepancy is likely methodological rather than geological. In particular, few if any reports have stated how such tests were made, and some workers have not even mentioned such tests, despite the early admonishment of Wintle and Huntley (1982). It is commonly believed that any significant fading in loess would be detectable within a few days of laboratory irradiatons. Yet Berger and Huntley (1985) and Berger (1987) have shown that storage tests up to at least several weeks were required. The routinely used short (room temperature) storage intervals may have caused anomalous fading in loess often to go undetected. An uncorrected 20 to 30 percent fading component may account for much of the underestimation of loess ages by TL dating.

Complicating these aspects of methodology is the misconception that the partial bleach method is invalid when dose-response curves are not linear (for example, Aitken, 1985, p. 266; Wintle and others, 1984). This misconception persists despite the evidence of many successful applications of this technique to waterlaid sediments, when the dose-response curves are unambiguously sublinear, and even saturating (for example, Berger, 1986; Berger and others, 1987). Feldspars dominate the TL signal in both loess and waterlaid sediments, and there is no reason to suppose that the TL properties of the feldspars in these sediments differ in any unknown way. Indeed, these successes motivated the application of this method (results described below) to the loess lying just above the Old Crow ash layer.

Thus, there are many reasonable methodological arguments for not accepting Debenham's hypothesis, and for believing that the regeneration method is in any case presently unreliable for loess older than about 50 ka. Even so, this method continues to be applied without independent age checks (for example, Singhvi and others, 1987). There are thus compelling arguments for applying only the partial bleach technique when coupled with rigorous regression analyses.

Concluding remarks

To close this comment, some new results (Berger, 1987) should be outlined. By careful laboratory techniques, a significant anomalous fading component has been eliminated from both the Old Crow tephra and a contiguous (ash-free) loess. Combined with appropriate extrapolations of saturating-exponential growth curves (contrast with the linear extrapolations of Wintle and Westgate, 1986), this empirical correction for anomalous fading has led to concordant TL age estimates of 110±12 ka for the Old Crow tephra and 108±16 ka for loess just above this ash layer. The average of about 150 ka of Berger and Huntley (1985) is invalid because in those preliminary experiments the results for bulk ash were averaged with those of a glass-rich fraction, and because the plateau test was not satisfied for the experiments on loess.

In conclusion, successful applications of the present TL methods require analytical care. For loess, the partial bleach method of Huntley (in Wintle and Huntley, 1980) is the preferred technique, even though it is more laborious than other methods.

REFERENCES CITED

Aitken, M.J., 1985, thermoluminescence Dating, New York, Academic Press, 351 p.

Berger, G.W., 1986, Dating Quaternary deposits by luminescence—recent advances: Geoscience Canada, v. 13, p. 15–21.

———1987, Thermoluminescence dating of the Pleistocene Old Crow tephra and adjacent loess, near Fairbanks, Alaska: Canadian Journal of Earth Sciences, v. 24, p. 1975–1984.

Berger, G.W., Clague, J.J., and Huntley, D.J., 1987, Thermoluminescence dating applied to glaciolacustrine sediments from central British Columbia: Canadian Journal of Earth Sciences, v. 24, p. 425–434.

Berger, G.W., and Huntley, D.J., 1985, A thermoluminescence date for the Old Crow tephra: Geological Society of America Abstracts with Programs, v. 17, p. 341.

Canfield, H., 1985, Thermoluminescence dating and the chronology of loess deposition in the central United States: Madison, University of Wisconsin, M.S. thesis, 159 p.

Debenham, N.C., 1985, Use of UV emissions in TL dating of

sediments: Nuclear Tracks, v. 10, p. 717–724.
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, Susan, eds., The United States Geological Survey in Alaska: Accomplishments during 1982: U.S. Geological Survey Circular 939, p. 26–29.
Mejdahl, V., 1986, Thermoluminescence dating of sediments: Radiation Protection Dosimetry, v. 17, p. 219–227.
Norton, L.D., and Bradford, J.M., 1985, Thermoluminescence dating of loess from western Iowa: Soil Science Society of America Journal, v. 49, p. 708–712.
Prosynska, H., 1983, TL dating of some subaerial sediments from Poland: PACT Journal, v. 9, p. 539–546.
Rendell, H.M., Gamble, I.J.A., and Townsend, P.D., 1983, Thermoluminescence dating of loess from Potwar Plateau, northern Pakistan: PACT Journal, v. 9, p. 555–562.
Singhvi, A.K., Bronger, A., Pant, R.K., and Sauer, W., 1987, Thermoluminescence dating and its implications for the chronostratigraphy of loess-paleosol sequences in the Kashmir Valley (India); Isotope Geoscience, v. 65, p. 45–56.
Westgate, J.A., Walter, R.C., Pearce, G.W., and Gorton, M.P., 1985, Distribution, stratigraphy, petrochemistry, and paleomagnetism of the late Pleistocene Old Crow tephra in Alaska and the Yukon: Canadian Journal of Earth Sciences, v. 22, p. 893–902.
Wintle, A.G., 1985, Sensitization of TL signal by exposure to light: Ancient TL, v. 3, p. 16–21.
Wintle, A.G., and Brunnacker, K., 1982, Ages of volcanic tuff in Rheinhessen obtained by thermoluminescence dating of loess: Naturwissenschaften, v. 69, p. 181–183.
Wintle, A.G., and Huntley, D.J., 1980, Thermoluminescence dating of ocean sediments: Canadian Journal of Earth Sciences, v. 17, p. 348–360.
———1982, Thermoluminescence dating of sediments: Quaternary Science Reviews, v. 1, p. 31–53.
Wintle, A.G., and Westgate, J.A., 1986, Thermoluminescence age of Old Crow tephra in Alaska: Geology, v. 14, p. 594–597.
Wintle, A.G., Shackleton, N.J., and Lautridou, J.P., 1984, Thermoluminescence dating of periods of loess deposition and soil formation in Normandy: Nature, v. 310, p. 491–493.

Carter, L. David, Hamilton, Thomas D., and Galloway, John P., eds., 1989, Late Cenozoic History of the Interior Basins of Alaska and the Yukon: U.S. Geological Survey Circular 1026.

SUMMARIES OF DISCUSSION SESSIONS

Following the presentation of individual papers, a series of two-hour discussions was held on topics that were considered to be of particular interest to the participants. The following summaries were prepared by individual discussion leaders for incorporation in this volume.

Geochronology

Julie Brigham-Grette

Accurate dating is essential for piecing together the Quaternary stratigraphy of any area and for correlating sequences in separate drainage basins. During our discussions, it was clear that some stratigraphic problems will remain unresolved until we can distinguish between real gaps in the preserved record and disagreements caused by inadequate age control. For example, differences in the glacial record from the Brooks Range, the Alaska Range, and the northern Yukon may reflect differences in the style and intensity of glaciation or differences in the preservation potential of ice marginal features from different areas (Gibbon and others, 1984). Direct matching of series of glacial or periglacial events may mislead us into false correlations unless they can be supported by adequate age control.

The discussion focused largely on the problems and reliability of standard radiocarbon dating, accelerator radiocarbon dating, and thermoluminescence dating. Time did not permit adequate discussion of other dating methods such as amino acid, fission track, K-Ar, uranium series, and paleomagnetism.

Several researchers presented examples of significant discrepancies in radiocarbon dates on organic materials from the same deposit. David Carter reported seven ages ranging from about 13 to 30 ka on different size fractions of a detrital organic deposit from northern Alaska. Successively finer fractions yielded successively older ages (Nelson and others, 1988). Lake sequences in Alberta yield basal ages on moss that are consistently older than wood at the same horizon, according to Lionel Jackson. The basal ages on moss and peat generally are too old because of bicarbonate enrichment soon after deglaciation.

Owen Hughes emphasized that there is a fundamental need to carefully define what reseachers mean by "peat," because this term commonly is used for disseminated organics in fine-grained sediments as well as for densely compressed plant macrofossils. Charles Schweger commented that most palynologists use the classification scheme outlined by Troels-Smith (1955; see also Aaby and Berglund, 1986) and suggested that perhaps Quaternary geologists should adopt the same system. Moreover, we must concern ourselves more with studies of the taphonomy of organic material. Articles by Arundale (1981), Clayton and Moran (1982), Stuckenrath and others (1979), and Fowler and others (1986) were cited as useful guides to sample selection.

Robert Thorson discussed problems in recognizing organic material out of stratigraphic context in terrace sequences along the upper Yukon River drainage. For example, wood and other plant debris may become lodged in a deep thermo-erosional niche at the base of a river bluff. If the bluff face then slumps or settles to close such a niche, the inset deposit may appear stratigraphically older than undisturbed organic layers higher in the bluff.

Discussion then turned to the maximum age limits of radiocarbon dating. David Carter presented a series of unpublished radiocarbon ages from the Flaxman Member of the Gubik Formation on the North Slope of Alaska. Initial analyses of seal bone, mollusks, and whale bone all yielded finite ages in the range of 19.6 to 22.6 ka, but subsequent ages on whale bone collagen yielded ages of >38 and >48 ka and a uranium-series age on the same material was about 75 ka. Thomas Hamilton reported that radiocarbon ages from the permafrost tunnel near Fairbanks and from Epiguruk bluff in northwest Alaska were stratigraphically consistent back to about 37 ka but that the ages became erratic beyond that point. Richard Harington reviewed the dating of a bone tool of presumed Pleistocene age from the Old Crow Basin, Yukon Territory. Initial radiocarbon dating of the apatite fraction of this specimen yielded an age of about 27 ka, but recent accelerator dating of the collagen fraction of the same specimen produced an age of about 1.5 ka (see Harington, this volume). Recent studies (for example, Hassan and Ortner, 1977; Hassan and others, 1977) have discredited dating of bone apatite.

Most participants agreed that the reliable range of the radiocarbon method may be less than that stated by the laboratories, and that accelerator dating has not extended the useful age range but provides only a means of dating small samples. The accelerator technique also enables us to date several fractions of a sample to reveal whether they represent a detrital mixture. Whereas wood may be recommended most often as the best material to date, John Matthews pointed out that it also generally preserves well and, therefore, is commonly redeposited in Arctic stratigraphic sequences. Radiocarbon dating of modern marine shells from the Canadian Beaufort Sea and adjacent Arctic Islands also is needed to regionally assess the reservoir age of such materials.

Thermoluminescence (TL) dating is gradually becoming an important dating tool in Alaska and the Yukon, particularly for subdividing stratigraphy in unglaciated regions. No commercial laboratories do TL dating, and it was agreed that the method is still at the research stage. TL dating of the Old Crow tephra is critically important because the tephra forms a nearly instantaneous time datum for subdividing the stratigraphy of a very broad region of Alaska and the Yukon. Charles Schweger summarized evidence from a variety of sites which place age estimates on the tephra using different dating methods (Schweger and Matthews, 1985). Radiocarbon analyses of samples from the Koyukuk River valley and the Fairbanks region provide minimum age estimates of >56 and >60 ka, whereas a fission-track analysis provides a maximum age limit of about 120 ka. A maximum age limit of 100–125 ka is also provided by the occurence of the Blake Reversed-Polarity Subchron of the Brunhes Normal-Polarity Chron below the tephra at three localities (Shackleton, 1982; Westgate and others, 1985). Extrapolation of sedimentation rates places the age of the Old Crow tephra between at 87 and 105 ka (Schweger and Matthews, 1985) whereas recent TL dating of loess above and below the tephra at Halfway House limits the age to 86±8 ka (Wintle and Westgate, 1986). In addition, a TL age of 110±12 ka has been obtained by Glen Berger (this volume). Given the range of ages for the Old Crow tephra, Schweger proposed that perhaps the best age estimate on the tephra is between 87 and 94 ka. Most participants agreed that this was a reasonable age bracket to use until additional geochronology is available. Accurate TL and fission-track dating of other tephras found in Pleistocene sequences of Beringia is equally crucial to provide synchronous datums for regional correlation.

It was concluded that problems of geochronology, including the reliability of datable material and the need for more accurate methods, provide the greatest obstacle to determining and correlating the late Cenozoic glacial and periglacial record of Alaska and the Yukon.

Neotectonics in Alaska and Northwestern Canada

Robert M. Thorson

Neotectonics was defined as the effects, both direct and indirect, of regional or local crustal movements that are either part of the Neogene stratigraphic record or have influenced it in some way. The discussion revolved around four principal themes: glacio-isostasy, fault zones and relations to placer deposits, non-isostatic regional warping, and Holocene faulting and fault-related activity.

The isostatic record

Thomas Hamilton presented evidence from the Brooks Range for warping of late-glacial shorelines. He concluded that glacial loading caused at least 100 m of differential uplift between upper mountain valleys and the range front. Lionel Jackson confirmed the isostatic deflection of mountain valleys by presenting evidence for upwarping of extensive proglacial shorelines that formed during ice recession in southern Yukon Territory.

Robert Thorson and Thomas Hamilton suggested that shorelines in the Copper River Basin should be substantially warped because glacier complexes around the basin were larger than those of the Brooks Range. Evidence from aerial photographs indicates to Hamilton that strand lines near the Tangle Lakes in the northern Copper River Basin may have stood as high as 900 m altitude owing to crustal depression during the last glaciation. Other participants agreed that crustal warping should have occurred, but they argued that the present shoreline record is too fragmentary and the topographic data base is too imprecise to provide firm evidence for warping. A specific study of shoreline facies and an accurate survey of their altitudes should be a future research priority.

Thorson raised the possibility that young radiocarbon ages on deglaciation at low altitudes around the Gulf of Alaska could reflect isostatic submergence. However, Thomas Ager and several others pointed out that other factors such as Holocene faulting could have caused the apparent hiatus between deglaciation and basal ages.

Fault zones and placer deposits

Stephen Morison provided evidence for hydrothermal alteration and thrusting associated with Neogene fault displacements in the Dawson area. Owen Hughes confirmed such observations, noting that similar features also exist in the Tintina Trench, and that a true fault scarp in this area cuts till of the Reid glaciation of Bostock (1966). Warren Yeend also described surface scarps and fault-related stratigraphy in the Alaskan sector of the Tintina Trench, and cited the work of Weber and Foster (1982) who demonstrated Quaternary faulting near Circle. All of the discussants agreed that there is a strong link between Neogene tectonism and placer development, and that displacements have continued into the Quaternary.

Non-isostatic regional warping

Richard Reger opened the discussion by pointing out that the Alaska Range did not exist as an obstacle to southward drainage from central Alaska during early Tertiary time and that its uplift history dates from the late Cenozoic. David Hopkins stated that the Cenozoic history of all master drainages in Alaska is almost completely unknown. Owen Hughes discussed the high strath terrace along the Yukon River, noting that it rises eastward (downriver) in altitude to about 220 m near the Fortymile River and then descends toward the Yukon Flats. John Williams and Robert Thorson correlate this high strath terrace with the high terrace flanking the Yukon Flats.

Much of the discussion focused on the "White Gravels" described by Stephen Morison in the Yukon. Charles Schweger suggested that the arkosic (grus-like) character of the gravel could reflect the rapid deposition of mechanically disintegrated regolith into newly formed Tertiary basins. Robert Thorson pointed out that similar gravels along the Porcupine River are interbedded with lava flows, zeolitized argillic paleosols, and lignitized organic horizons. John Williams indicated that lava is overlain by white arkosic gravel at the western border of the Yukon Flats, and that the modern Yukon River is incised below these deposits. Lionel Jackson suggested that the petrology of one occurrence of these basalts is consistent with rifting. Julie Brigham-Grette pointed out that significant tectonic activity also occurred in the Canadian Arctic Islands during this interval. The discussants concluded that one or more protracted episodes of substantial uplift, volcanism, and formation of basins occurred broadly across the region during the Tertiary.

Holocene faulting and fault-related activity

Several participants described evidence for large landslides and rockfall avalanches in recently deglaciated terrain. They speculated that many of these events may have been seismically induced, but the instability and steepness of recently deglaciated slopes must have been an important factor as well. James Beget also mentioned that high seismicity might be expected as a natural consequence of unloading during rapid deglacation.

John Williams and Richard Reger described peats of early Holocene age that are presently submerged more than 10 m below sea level near Valdez, suggesting either isostatic or tectonic deformation. This raised the question of the expected lag time between deglaciation and peat formation, especially in coastal areas. Charles Schweger emphasized that peat is a botanical deposit, and is controlled by moisture, acidity, and climate as well as by time. He suggested that throughout northwestern North America 3,000 to 4,000 years may elapse following deglaciation or coastal uplift before peat is able to form or be preserved.

Specific research questions that developed during our discussion included: What was the impact of basin subsidence and movement along flanking border faults on the lacustrine history of the northern Yukon basins? How do regional tectonic warping and glacial isostasy affect the marine transgressive record? How and when did Neogene faulting affect fluvial placer deposits? Are tectonic and volcanic records obtainable from the distribution of seismically initiated rockfalls and volcanic debris flows, respectively? Can the sedimentology, distribution, and fauna of the Bootlegger Cove Formation be used to infer a coastal isostatic record for upper Cook Inlet? Can the regional slope of full-glacial lacustrine shoreline features be used to infer ice thickness and distribution in the Brooks Range and the Yukon? What evidence exists for glacioisostatic deformation in the Copper River Basin? How did regional tectonism influence the Neogene fluvial incision of interior uplands and the formation of high-level placer benches?

Three general research problems were considered to be particularly important:

1. The long-term history of regional drainage systems and interior basins remains obscure. This history should be studied through correlation and dating of high level strath terraces, interstratified lava flows, tephra, and fluvial-lacustrine associations.
2. Isostatic effects of Pleistocene glaciation were significant and measurable throughout Alaska and the Yukon, as indicated by raised marine deposits in coastal areas and warped lacustrine strandlines. Systematic data need to be collected to determine mechanisms of glacio-isostatic loading and the seismic and tectonic effects of subsequent unloading.
3. Interpreting and dating fluvial and tectonic records at placer localities is important for mineral exploration and development and also for broader assessment of Cenozoic stratigraphy in interior basins.

Late Tertiary Events

John V. Matthews, Jr.

The interior basins of Alaska and the Yukon contain a rich record of late Neogene and early Pleistocene time, which has important implications for paleoclimatology and biogeography. Our discussion focused on fauna and flora of the late Neogene, involving such subsidiary questions as (1) the first development of lowland tundra, (2) the character of the early boreal forest, (3) correlaton and dating based on floral and faunal data, and (4) past vegetation zones. Much of the discussion centered on the origin of lowland tundra. David Carter, David Hopkins, Charles Repenning, and others believe that the earliest manifestation of tundra vegetation in the lowlands of the Northern Hemisphere is recorded at the Krestovka site in northern Siberia, and that tundra environments were present shortly afterward at some Alaskan localities as represented in exposures of the Gubik Formation. Matthews maintains that floras of the Beaufort Formation on Meighen Island represent a lowland tundra-like environment that is older than that represented by the Gubik Formation. The flora of the Kap Kobehnavn site in northern Greenland figures prominently in this argument. There is a problem, however, because if Kap Kobenhavn is of Olduvai age (Olduvai Normal-Polarity Subchron of Matuyama Reversed-Polarity Chron) as Repenning suggests, then sites farther south that are presumed to be of similar age should display a much richer boreal flora and fauna. One such site in the Old Crow Basin has flora and fauna that apparently are less diverse than at Kap Kobenhavn. The basic assumption implicit in such arguments is that the latitudinal diversity gradient of the Neogene was similar at least in trend to that of the present. Thomas Ager noted that it is very difficult to know what northern Alaskan equivalents of the Cook Inlet floras should be like because of the unknown effect of latitudinal diversity. In other words, should we expect Seldovian-type (Seldovian Stage of Wolfe and others, 1966) pollen floras from northern Alaska to be the same age as those from the Seldovian type locality in southern Alaska? Charles Repenning stated that the geologic record no longer supports a clear division between forested conditions in east Beringia during the late Tertiary and widespread tundra at the beginning of the Quaternary.

Another topic considered was Neogene climates and their proxy indicators: glaciation, vegetation changes, and dispersal events. Most discussion concerned earliest glacial advances. David Hopkins stated that the earliest record on the Seward Peninsula probably is at 2.4 Ma at California Creek on the Seward Peninsula. Thomas Hamilton mentioned the Gunsight Mountain glacial interval of Hamilton (1979) as a probable equivalent in the Brooks Range, and Troy Péwé stated that the Browne glaciation of Wahrhaftig (1958) was probably also older than 2 Ma. Robert Thorson presented possible evidence in the Alaska Range for a pre-Browne glaciation, which led to a discussion of the record of Tertiary ice rafting in the Yakataga Formation of the Prince William Sound region and the question of possible Miocene glaciations in the Wrangell Mountains. It is clear there were major and possibly successive glacial events in east Beringia prior to the Quaternary.

Marine and terrestrial relationships were discussed briefly, mostly in conjunction with the degree to which Pacific mollusks swamped the resident Arctic (North Atlantic) fauna when the Bering land bridge was first breached in late Pliocene time. An important question is how long an interval was required for total replacement (in other words, are there faunas that contain a mixture of Atlantic and Pacific mollusks?).

A final question raised during the discussion concerned the reason for apparently distinct modern arboreal floras on either side of Bering Strait. The Bering land bridge existed for nearly all of Neogene time, and boreal forest occurred in Alaska and East Siberia for nearly all of that interval. Hopkins stated that exchange of arboreal taxa during that interval is indicated by comparison of North American fossils to those from Siberia. Matthews noted that recent work has shown that some of the shrubs and herbs that were present in Neogene Siberian forests are now known from the Beaufort Formation in North America. So perhaps the east and west Beringian forests were similar during the Neogene. Why are they different today?

One joint Canadian-American research project that would go far to clarifying some of our Neogene problems would involve drilling boreholes in both the Old Crow Basin and the Yukon Flats. Alan Judge stated that his group had been looking into the feasibility of putting in holes on Old Crow Mountain and in the adjacent basin. John Williams described an earlier study at Fort Yukon, and suggested that it might be the best locality for a new borehole. Ideally, boreholes in the Old Crow Basin and the Yukon Flats should be drilled in the same year, using the same scientific sampling team and possibly the same drilling crew.

Early and Middle Pleistocene Events

T.D. Hamilton

We followed international convention in assigning to the early Pleistocene those events that took place between 1.65 Ma and the boundary between the Matuyama Reversed-Polarity Chron and the Brunhes Normal-Polarity Chron at about 790 ka (Richmond and Fullerton, 1986). Middle Pleistocene events are those that occured after this polarity reversal and before the peak of the last interglaciation (marine oxygen-isotope substage 5e) at about 125–130 ka. Our discussion covered glacial and sea-level records, biogeographic and evolutionary changes in flora and fauna, drainage history, tectonic records, development of permafrost and sea ice, and paleosol formation.

Glacial sequences of inferred early through middle Pleistocene age have been reported from the Brooks Range (Hamilton, 1986), Seward Peninsula (Kaufman and Hopkins, 1986), and the Yukon-Tanana Upland (Weber, 1986), but glacial records farther to the south and east generally are more fragmentary owing to active tectonism and to overlap of older drift sheets by younger glacial deposits. Correlations between relatively continuous and fragmentary sequences are tenuous at present. Although several sea-level highstands are known from the Bering and Chukchi Sea regions, the Arctic Coastal Plain close to the Canadian border contains no elevated strandlines younger than late Tertiary and older than Pelukian (substage 5e in age). Expansion of the tundra biome continued during early Pleistocene time and, according to Charles Repenning, evolution of microtine rodents provides a time scale valid to within about 200 ka for some fossiliferous Pleistocene deposits. A general history of drainage evolution in the central and western Brooks Range has been inferred from Thomas Hamilton's detailed mapping of drift sheets deposited by sequential glacial advances of early and middle Pleistocene age. Drainage evolution elsewhere in Alaska and the Yukon during this interval is poorly known, according to Stephen Morison and Warren Yeend, but the widespread occurrence of economically important placer deposits makes this research field of particular interest. Tectonism altered relief and drainage patterns of many areas and impacted others through changes in wind patterns, precipitation values, and regional drainage. Few quantitative data are available on the timing or rates of faulting and other tectonic changes during this interval, but the topic is of great importance to studies of crustal dynamics and to assessment of geologic hazards as well as to Quaternary geology. Development and expansion of permafrost in northern Alaska and Arctic Ocean ice cover probably occurred during late Pliocene time, but details are little known. Studies applicable to permafrost and sea-ice history include paleoecology of fossil taxa, deep thermal profiles in permafrost boreholes, and history of kettle formation in ice-cored glacial deposits. Detailed studies of paleosols in Yukon Territory by Charles Tarnocai show that these fossil soils reflect the ages and environments of their formation. Mapping of similar paleosols should be carried out in central Alaska as an aid to dating surficial deposits and extending correlations across the international boundary.

Absolute ages of early and middle Pleistocene events are provided by K-Ar, thermoluminescence, and fission-track dating and by paleomagnetism; amino-acid analyses provide a means of relative-age dating and correlation. Absolute-dating approaches show that the very extensive Anaktuvuk River-Nome River drift complex of northern and northwestern Alaska formed shortly before 790 ka according to David Hopkins, and TL dating by J.A. Westgate (University of Toronto) and Troy Péwé shows that the Gold Hill Loess around Fairbanks began to form about 450 ka.

A major problem in both Alaska and the Yukon is the age of the youngest glacial event that preceded late Wisconsin glaciation. Data from northern Alaska suggest an age younger than the last interglacial maximum (substage 5e) according to Thomas Hamilton, and a comparable history has been reported from the eastern Canadian Arctic (Andrews and others, 1985). On the other hand, examination of soils in the Yukon by Charles Tarnocai suggest that seemingly correlative drift may be older than substage 5e. This controversial topic was covered more extensively in the following discussion session.

Late Pleistocene Glacial Advances

Jean-Serge Vincent

The workshop participants were asked to present the crucial data which enabled them to recognize the existence and chronology of glacial advances during late Pleistocene time (132 to 10 ka) in their individual study areas. The informal time divisions presented in table 1 were followed to ensure a uniform chronologic framework.

In the Brooks Range, according to Thomas Hamilton, three major ice advances have occurred since marine oxygen-isotope substage 5e. In the southern Brooks Range the oldest and most extensive advance, termed the *Siruk Creek*, occurred prior to deposition of the Old Crow tephra and overlying forest beds. This advance is younger than oxidized interglacial gravels, and probably took place during substage 5d and (or) 5b. The early(?) Wisconsin Chebanika advance (>35 ka) and the late Wisconsin Walker Lake advance, dated at between 25 and 12 ka, followed. In the northern Brooks Range, after a period of weathering and soil formation assigned to marine oxygen-isotope substage 5e, ice of the Itkillik IA glaciation of Hamilton (1986) advanced in phase with the Siruk Creek advance in southern valleys. The subsequent Itkillik IB and Itkillik II glacial advances probably took place during oxygen isotope stages 4 and 2, respectively. Glaciations may well have occurred during Sangamon time (substages 5d and (or) 5b) because at that time the oceans were warmer and the summers cooler on land. Abundant moisture sources, therefore, were available for snowfall, but summer ablation would have been inhibited.

Table 1. Informal time subdivisions of late Pleistocene time used in discussion

Epoch	Glaciation/interglaciation		Age (in ka)	Marine oxygen-isotope stages and substages
	Stage	Substage		
HOLOCENE			0–10	1
PLEISTOCENE				
		Late Wisconsin	10–23	2
	WISCONSIN	Middle Wisconsin	23–65	3
		Early Wisconsin	65–80	4
		Late Sangamon	80–122	5a
				5b
	SANGAMON			5c
				5d
		Early Sangamon	122–132	5e

According to Thomas Hamilton, the glacial record in the Yukon-Tanana Upland (Weber, 1986) is very similar to that of the Brooks Range. He also stated that at Canyon Creek, outwash of the Delta glaciation of Péwé (1953), derived from the Alaska Range, is overlain by sediments that contain bones with uranium-series ages of about 80 ka. This suggests that glaciation of late Sangamon age took place in the Alaska Range as well.

Late Wisconsin glacial advances are well documented in the Alaska Range and the Canadian Cordillera, but the existence of early Wisconsin or late Sangamon ice advances is still subject to controversy. Glacial deposits underlie nonglacial sediments that are older than the time range of radiocarbon dating and that contain the Old Crow and Sheep Creek tephras. These glacial deposits may date from the early part of the late Pleistocene but they could also be older. On the basis of data at the Ash Bend locality in the Yukon, for example, Owen Hughes now believes that the Reid glaciation of Bostock (1966) is much older than the Sheep Creek tephra and may predate the Sangamon. According to Charles Tarnocai, this older age assignment is also supported by the stratigraphic position of the Diversion Creek paleosols, of probable Sangamon age, which overlie Reid till. In the Alaska Range, the age of the pre-late Wisconsin Delta glaciation, generally correlated with the Reid glaciation, is uncertain. Thomas Hamilton believes that two advances of Delta age may have occurred respectively during late Sangamon and early Wisconsin time. However, Troy Péwé argues that the Delta glaciation may be middle Pleistocene in age because (1) interglacial soils are developed on its till, (2) tephras correlated with the Old Crow and Sheep Creek tephras found in loess linked with the ice advances are providing ages of 120 ka or older, and (3) in at least three localities the Blake geomagnetic excursion (Blake Reversed-Polarity Subchron of Brunhes Normal-Polarity Chron), which occurred about 115 ka, is recorded in sediments overlying the till. In the Watson Lake area of the Yukon, Lionel Jackson observed no sediments recording Sangamon through middle Wisconsin events. The existence or nonexistence of an extensive glaciation in the early part of the late Pleistocene remains an open question. Julie Brigham-Grette suggested that the apparent absence of a record of glaciation in the Cordillera during oxygen-isotope stages 5 to 3 is perhaps due to the fact that late Wisconsin ice advances of the McConnell or Donnelly glaciations (of Bostock, 1966, and Péwé, 1953, respectively) were so extensive that they obliterated the earlier glacial record. In the Brooks Range the record is preserved because the earlier ice advances of the late Pleistocene were stronger than the later ones.

The chronology of lacustrine filling of the interior basins of the Yukon seems to indicate that Laurentide ice was strong enough to advance to the Cordilleran mountain front (thus blocking eastward drainage and impounding waters) only in the later part of the late Pleistocene (after about 35 ka). No earlier record of such an event seem to be recorded in the basins or at the ramparts which served as spillways for the glacier-dammed lake basins.

Two late Pleistocene ice advances are documented in the Northern Interior Plains, both on land on the coastal plain (by Jean-Serge Vincent) and offshore on the shelf (by Julie Brigham-Grette). There is a general consensus that the youngest ice advance, the Tutsieta Lake phase of Owen Hughes, took place during late Wisconsin time and was confined on land. The age of the earlier and more extensive late Pleistocene ice advance, which extended to the continental shelf, is controversial. This event, which has been assigned to the Hungry Creek glaciation in the Bonnet Plume area by Owen Hughes and to the Buckland glaciation on the Yukon Coastal Plain by Rampton (1982), could be as old as late Sangamon (marine oxygen-isotope substage 5d or 5b) or as young as middle Wisconsin in age.

Because of the lack of adequate dating techniques for deposits beyond the range of the radiocarbon method, most of our age assignments for deposits older than about 35 ka commonly rest on the assignment of associated organic-bearing nonglacial deposits to the Sangamon interglaciation or the middle Wisconsin nonglacial period. The presence of tephras or geomagnetic events permits better age assignments in some instances. Glacial advances of stages 5 and 4 age may be present in the Brooks Range, but are more dubious in regions farther to the east and south such as the Alaska Range, the northwest Yukon, and the periphery of the Laurentide ice sheet. If a major ice advance did not occur at this time, what then is the age of the well-documented glaciogenic deposits found stratigraphically below glacial deposits of late Wisconsin age? Do these predate the Sangamon or were they laid down during late Sangamon or early Wisconsin time? These important research problems deserve particular attention.

Late Pleistocene Interstadial Intervals

Charles E. Schweger

Interstades must be considered as unique time periods of considerable biological and climatic significance rather than merely as warm periods that bracket glacial stades. To initiate discussion, Charles Schweger described evidence for the Koy-Yukon thermal event, a period when treelines advanced and active layers deepened through much of interior Alaska and the Yukon. Dated at between 50 and 60 ka, it occurred during the insolation maximum that initiated marine oxygen-isotope stage 3. Troy Péwé pointed out that evidence from the Fairbanks area indicates this may have been the warmest climatic interval in the past 500 k.y. If it is of stage 3 age, he asked, why is this event more prominent than the Sangamon (the last major interglacial interval) and how does one use a climatic signal for distinguishing interglacials from interstadials?

Mary Edwards demonstrated the usefulness of the marine oxygen isotope and Milankovitch insolation records for testing paleoclimatic reconstructions. Marine oxygen-isotope stage 1 (Holocene), stage 3 (in part Koy-Yukon thermal event), and substage 5e (Sangamon) represent three late Pleistocene warm periods with different insolation budgets and duration. The length of the warm period may be of great importance because the migration lag of many boreal forest species may be considerable, as demonstrated by the Holocene record. Because of its long duration, stage 3 might appear to have a warmer climatic signal than substages 5a, 5c, or 5e. In addition, the boreal forest can create its own climate through changes in albedo, surface wind shear effects, and snow depth. Once established, the forest could conceivably enhance or reinforce the results of a warmer and possibly more moist climate. Several participants suggested that the interval represented by marine oxygen-isotope substage 5e may never have seen the full development of boreal forest vegetation, and therefore it is poorly represented in the stratigraphic record. David Hopkins, however, noted that substage 5e is represented on the Seward Peninsula by higher sea levels and an apparent westward advance of boreal treeline (see Hopkins, 1967).

The botanical record of interstadials is rapidly developing and points to the need for more autecological studies. *Typha* (cattail) is known to have greatly expanded its range during the Koy-Yukon thermal event and the early Holocene, indicating higher summer water temperatures and (or) higher nutrient status. *Alnus* seems to have been present in the interior during the Koy-Yukon thermal event but not during other interstadials. Thomas Ager expressed a cautionary note by pointing out that ecotypic evolution may have occurred during the course of the late Pleistocene, making modern autecological data less reliable.

Knowledge of interstadial, interglacial, and stadial faunas is much less precise, and we need to follow a more strict stratigraphic approach to vertebrate studies. Richard Harington added that many of the so-called "arctic" species appear to have existed throughout the late Pleistocene seemingly without regard for the differences between stadial and interstadial conditions. This led David Hopkins to suggest that past interstadial vegetation may have had a more open character than that of true interglaciations.

Thomas Hamilton criticized reliance on very old radiocarbon ages based on isotope enrichment, and several discussants commented on the wide range of ages assigned to the Old Crow tephra in regional chronologies. An age range of 87 to 94 ka for the Old Crow tephra (see "Geochronology" discussion) allows the tephra to be placed in substage 5c, which satisfies the pollen evidence.

There is also special interest in events occurring around 35 ka (late stage 3). Pollen studies over eastern Beringia point out the existence at this time of an east-west, as well as the more familiar north-south, climatic and ecological gradient. This period also seems to have been the time of pronounced westward expansion of the Laurentide ice sheet, with resulting proglacial flooding through northern Yukon basins and the Porcupine River. This episode of ice growth and expansion during middle Wisconsin time challenges our assumptions about interrelations between climate and glacial events. Data presented by John Matthews and Thomas Hamilton suggest that the late Wisconsin (stage 2) may have included warm oscillations. Mary Edwards pointed out that the type of study site will greatly influence the record of paleovegetation, and that many different sites will be needed to adequately represent regional vegetation and therefore synoptic paleoclimate.

Canadian-American Collaborative Research

L. David Carter

In the final discussion session we reviewed ongoing research projects involving Canadian and American scientists, identified topics and specific problems that are best addressed through joint Canadian-American research, and discussed field excursions in Alaska and the Yukon that would provide information necessary for a preliminary evaluation of some of these problems.

Ongoing Canadian-American collaborative research includes studies of the Quaternary stratigraphy and tephrochronology of central Alaska being carried out by Troy Péwé, J.A. Westgate (University of Toronto), and A.G. Wintle (Cambridge University). Their work focuses on the resolution of many long-standing stratigraphic problems through thermoluminescence dating of loess, fission-track dating of tephra, petrographic and geochemical correlation of tephras, radiocarbon dating of shells, wood, peat, and bone, and paleomagnetic stratigraphy of loess. Charles Schweger and John Matthews, in collaboration with Péwé and Westgate, are correlating tephras across Alaska and the Yukon and using them as datums in order to reconstruct spatial variation in paleoecological conditions during instants of geologic time.

Topics and problems that seem particularly appropriate for collaborative research initiatives include:

1. Application of the stratigraphy and sedimentology of placer gravels established for the Yukon (see Morison, this volume) to studies of the origin and distribution of placer gravels in eastern and central Alaska.
2. A taxonomic study of the Cripple Sump fauna of central Alaska. This would augment the ongoing placer research projects in both Alaska and the Yukon.
3. Coring through the deep sediments that fill the Yukon Flats and the Old Crow basins, and instrumenting the coreholes for precise temperature measurements. This will allow comparison of the depositional histories of these tectonic basins, and provide important information on heat flow; permafrost temperature, history, and thickness; and climatic change for northeast Alaska and the Yukon.
4. Tracing high-level terraces of the Yukon River into Alaska and determining their relation to climatic change and tectonism.
5. Comparing the Pleistocene glacial sequence in the Yukon with those established for the Brooks Range, Alaska Range, and Yukon-Tanana Upland.
6. Determining the size and number of late Pleistocene catastrophic floods of the Porcupine and Yukon rivers, and comparing this record with the history of sedimentation in the Bering Sea to identify flood-generated pulses of sedimentation.
7. Developing a soil chronosequence for glacial and nonglacial deposits in Alaska that can be compared with the soil chronosequence established for the Yukon (see Tarnocai, this volume). This would allow correlation of geomorphic surfaces between Alaska and the Yukon.
8. Determining radiometric ages of Cenozoic lavas on the east side of the Yukon Flats that are interbedded

with organic-rich deposits. Samples of the organic-rich deposits are in hand and are being studied by Thomas Ager and Charles Schweger. Dating the lavas will provide a record of vegetation change that can be extended into the Yukon and correlated with the Tertiary floral stages established for the Cook Inlet region.

It was the unanimous opinion of the workshop participants that joint Canadian-American field excursions be organized for the summer of 1988 to evaluate several of the research opportunities enumerated above and examine some of the localities discussed at the workshop that are most critical to the interpretation of late Cenozoic deposits and events in Alaska and the Yukon. The excursions should include examination of key stratigraphic sections on the Stewart, Porcupine, and Yukon Rivers via riverboat, and in the Whitehorse and Dawson areas and the Tanana River valley by automobile or van.

COMBINED REFERENCES CITED

Aaby, B., and Berglund, B.E., 1986, Characterization of peat and lake sediments, *in* B.E. Berglund, ed., Handbook of palaeoecology and palaeohydrology: New York, Wiley, p. 231–246.

Andrews, J.T., Aksu, A., Kelly, M., Klassen, R., Miller, G.H., Mode, W.N., and Mudie, P., 1985, Land/ocean correlations during the late interglacial/glacial transition, Baffin Bay, northwestern North Atlantic—A review: Quaternary Science Reviews, v. 4, p. 333–355.

Arundale, W.H., 1981, Radiocarbon dating in eastern arctic archaeology: A flexible approach: American Antiquity, v. 46, p. 244–271.

Bostock, H.S., 1966, Notes on glaciation in central Yukon Territory: Geological Survey of Canada Paper 65–36, 18 p.

Clayton, L., and Moran, S.R., 1982, Chronology of late Wisconsinan glaciation in middle North America: Quaternary Science Reviews, v. 1, p. 55–82.

Fowler, A.J., Gillespie, R., and Hedges, R.E.M., 1986, Radiocarbon dating of sediments by accelerator mass spectrometry: Physics of the Earth and Planetary Interiors, v. 44, p. 15–20.

Gibbons, A.B., Megeath, J.D., and Pierce, K.L., 1984, Probability of moraine survival in a succession of glacial advances: Geology, v. 12, p. 327–330.

Hamilton, T.D., 1979, Late Cenozoic glaciations and erosion intervals, north-central Brooks Range, *in* Johnson, K.M., and Williams, J.R., eds., The United State Geological Survey in Alaska—Accomplishments during 1978: U.S. Geological Survey Circular 804–B, p. B27–B29.

———1986, Late Cenozoic glaciation of the central Brooks Range, *in* Hamilton, T.D., Reed, K.M., and Thorson, R.M., eds., Glaciation in Alaska—The geologic record: Anchorage, Alaska Geological Society, p. 9–49.

Hassan, A.A., and Ortner, D.J., 1977, Inclusions in bone material as a source of error in radiocarbon dating: Archaeometry, v. 19, p. 131–135.

Hassan, A.A., Termine, J.D., and Haynes, C.V., Jr., 1977, Mineralogical studies on bone apatite and their implications for radiocarbon dating: Radiocarbon, v. 19, p. 364–374.

Hopkins, D.M., 1967, Quaternary marine transgressions in Alaska, *in* Hopkins, D.M., ed., The Bering Land Bridge: Stanford, Calif., Stanford University Press, p. 47–90.

Kaufman, D.S., and Hopkins, D.M., 1986, Glacial history of the Seward Peninsula, *in* Hamilton, T.D., Reed, K.M., and Thorson, R.M., eds., Glaciation in Alaska—The geologic record: Anchorage, Alaska Geological Society, p. 51–77.

Nelson, R.E., Carter, L.D., and Robinson, S.W., 1988, Anomalous radiocarbon ages from a Holocene detrital organic lens in Alaska and their implications for radiocarbon dating and paleoenvironmental reconstructions in the Arctic: Quaternary Research, v. 29, p. 66–71.

Péwé, T.L., 1953, Big Delta area, Alaska, *in* Péwé, T.L., and others, Multiple glaciations in Alaska—A progress report: U.S. Geological Survey Circular 289, p. 8–10.

Rampton, V.N., 1982, Quaternary geology of the Yukon coastal plain: Geological Survey of Canada Bulletin 317, 49 p.

Richmond, G.M., and Fullerton, D.S., 1986, Introduction to Quaternary glaciation in the United States of America, *in* Sibrava, Vladimir, Bowen, D.Q., and Richmond, G.M., eds., Quaternary glaciations in the Northern Hemisphere: Quaternary Science Reviews, v. 5, p. 3–10.

Schweger, C.E., and Matthews, J.V., Jr., 1985, Early and middle Wisconsinan environments of eastern Beringia: Stratigraphic and paleoecological implications of the Old Crow Tephra: Géographie Physique et Quaternaire, v. 39, p. 275–290.

Shackleton, Jonathan, 1982, Environmental histories from Whitefish and Imuruk Lakes, Seward Peninsula, Alaska: Columbus, Ohio State University, Institute of Polar Studies Report No. 76, 49 p.

Stuckenrath, Robert, Miller, G.H., and Andrews, J.T., 1979, Problems of radiocarbon dating Holocene organic-bearing sediments, Cumberland Peninsula, Baffin Island, NWT, Canada: Arctic and Alpine Research, v. 11, p. 109–120.

Troels-Smith, J., 1955, characterization of unconsolidated sediments: Danmarks Geologiske Undersogelse, IV Roekke, v. 3, no. 10, p. 1–73.

Wahrhaftig, Clyde, 1958, Quaternary geology of the Nenana River valley and adjacent parts of the Alaska Range: U.S. Geological Survey Professional Paper 293–A, p. 1–68.

Weber, F.R., 1986, Glacial geology of the Yukon-Tanana Upland, *in* Hamilton, T.D., Reed, K.M., and Thorson, R.M., eds., Glaciation in Alaska—The geologic record: Anchorage, Alaska Geological Society, p. 79–98.

Weber, R.R., and Foster, H.L., 1982, Tertiary(?) conglomerate and Quaternary faulting in the Circle quadrangle, Alaska, *in* Coonrad, W.L., ed., The United States Geological Survey in Alaska—Accomplishments during 1980: U.S. Geological Survey Circular 844, p. 58–61.

Westgate, J.A., Walter, R.C., Pearce, G.W., and Gorton, M.P., 1985, Distribution, stratigraphy, petrochemistry, and paleomagnetism of the late Pleistocene Old Crow tephra in Alaska and the Yukon: Canadian Journal of Earth Sciences, v. 22, p. 893–906.

Wintle, A.G., and Westgate, J.A., 1986, Thermoluminescence age of Old Crow Tephra in Alaska: Geology, v. 14, p. 594–597.

Wolfe, J.A., Hopkins, D.M., and Leopold, E.B., 1966, Tertiary stratigraphy and paleobotany of the Cook Inlet region, Alaska: U.S. Geological Survey Professional Paper 398–A, 29 p.

Carter, L. David, Hamilton, Thomas D., and Galloway, John P., eds., 1989, Late Cenozoic History of the Interior Basins of Alaska and the Yukon: U.S. Geological Survey Circular 1026.

GPO 685-041/9793

SELECTED SERIES OF U.S. GEOLOGICAL SURVEY PUBLICATIONS

Periodicals

Earthquakes & Volcanoes (issued bimonthly).
Preliminary Determination of Epicenters (issued monthly).

Technical Books and Reports

Professional Papers are mainly comprehensive scientific reports of wide and lasting interest and importance to professional scientists and engineers. Included are reports on the results of resource studies and of topographic, hydrologic, and geologic investigations. They also include collections of related papers addressing different aspects of a single scientific topic.

Bulletins contain significant data and interpretations that are of lasting scientific interest but are generally more limited in scope or geographic coverage than Professional Papers. They include the results of resource studies and of geologic and topographic investigations; as well as collections of short papers related to a specific topic.

Water-Supply Papers are comprehensive reports that present significant interpretive results of hydrologic investigations of wide interest to professional geologists, hydrologists, and engineers. The series covers investigations in all phases of hydrology, including hydrogeology, availability of water, quality of water, and use of water.

Circulars present administrative information or important scientific information of wide popular interest in a format designed for distribution at no cost to the public. Information is usually of short-term interest.

Water-Resources Investigations Reports are papers of an interpretive nature made available to the public outside the formal USGS publications series. Copies are reproduced on request unlike formal USGS publications, and they are also available for public inspection at depositories indicated in USGS catalogs.

Open-File Reports include unpublished manuscript reports, maps, and other material that are made available for public consultation at depositories. They are a nonpermanent form of publication that may be cited in other publications as sources of information.

Maps

Geologic Quadrangle Maps are multicolor geologic maps on topographic bases in 7 1/2- or 15-minute quadrangle formats (scales mainly 1:24,000 or 1:62,500) showing bedrock, surficial, or engineering geology. Maps generally include brief texts; some maps include structure and columnar sections only.

Geophysical Investigations Maps are on topographic or planimetric bases at various scales; they show results of surveys using geophysical techniques, such as gravity, magnetic, seismic, or radioactivity, which reflect subsurface structures that are of economic or geologic significance. Many maps include correlations with the geology.

Miscellaneous Investigations Series Maps are on planimetric or topographic bases of regular and irregular areas at various scales; they present a wide variety of format and subject matter. The series also includes 7 1/2-minute quadrangle photogeologic maps on planimetric bases which show geology as interpreted from aerial photographs. Series also includes maps of Mars and the Moon.

Coal Investigations Maps are geologic maps on topographic or planimetric bases at various scales showing bedrock or surficial geology, stratigraphy, and structural relations in certain coal-resource areas.

Oil and Gas Investigations Charts show stratigraphic information for certain oil and gas fields and other areas having petroleum potential.

Miscellaneous Field Studies Maps are multicolor or black-and-white maps on topographic or planimetric bases on quadrangle or irregular areas at various scales. Pre-1971 maps show bedrock geology in relation to specific mining or mineral-deposit problems; post-1971 maps are primarily black-and-white maps on various subjects such as environmental studies or wilderness mineral investigations.

Hydrologic Investigations Atlases are multicolored or black-and-white maps on topographic or planimetric bases presenting a wide range of geohydrologic data of both regular and irregular areas; principal scale is 1:24,000 and regional studies are at 1:250,000 scale or smaller.

Catalogs

Permanent catalogs, as well as some others, giving comprehensive listings of U.S. Geological Survey publications are available under the conditions indicated below from the U.S. Geological Survey, Books and Open-File Reports Section, Federal Center, Box 25425, Denver, CO 80225. (See latest Price and Availability List.)

"Publications of the Geological Survey, 1879- 1961" may be purchased by mail and over the counter in paperback book form and as a set of microfiche.

"Publications of the Geological Survey, 1962- 1970" may be purchased by mail and over the counter in paperback book form and as a set of microfiche.

"Publications of the U.S. Geological Survey, 1971- 1981" may be purchased by mail and over the counter in paperback book form (two volumes, publications listing and index) and as a set of microfiche.

Supplements for 1982, 1983, 1984, 1985, 1986, and for subsequent years since the last permanent catalog may be purchased by mail and over the counter in paperback book form.

State catalogs, "List of U.S. Geological Survey Geologic and Water-Supply Reports and Maps For (State)," may be purchased by mail and over the counter in paperback booklet form only.

"Price and Availability List of U.S. Geological Survey Publications," issued annually, is available free of charge in paperback booklet form only.

Selected copies of a monthly catalog "New Publications of the U.S. Geological Survey" available free of charge by mail or may be obtained over the counter in paperback booklet form only. Those wishing a free subscription to the monthly catalog "New Publications of the U.S. Geological Survey" should write to the U.S. Geological Survey, 582 National Center, Reston, VA 22092.

Note.--Prices of Government publications listed in older catalogs, announcements, and publications may be incorrect. Therefore, the prices charged may differ from the prices in catalogs, announcements, and publications.