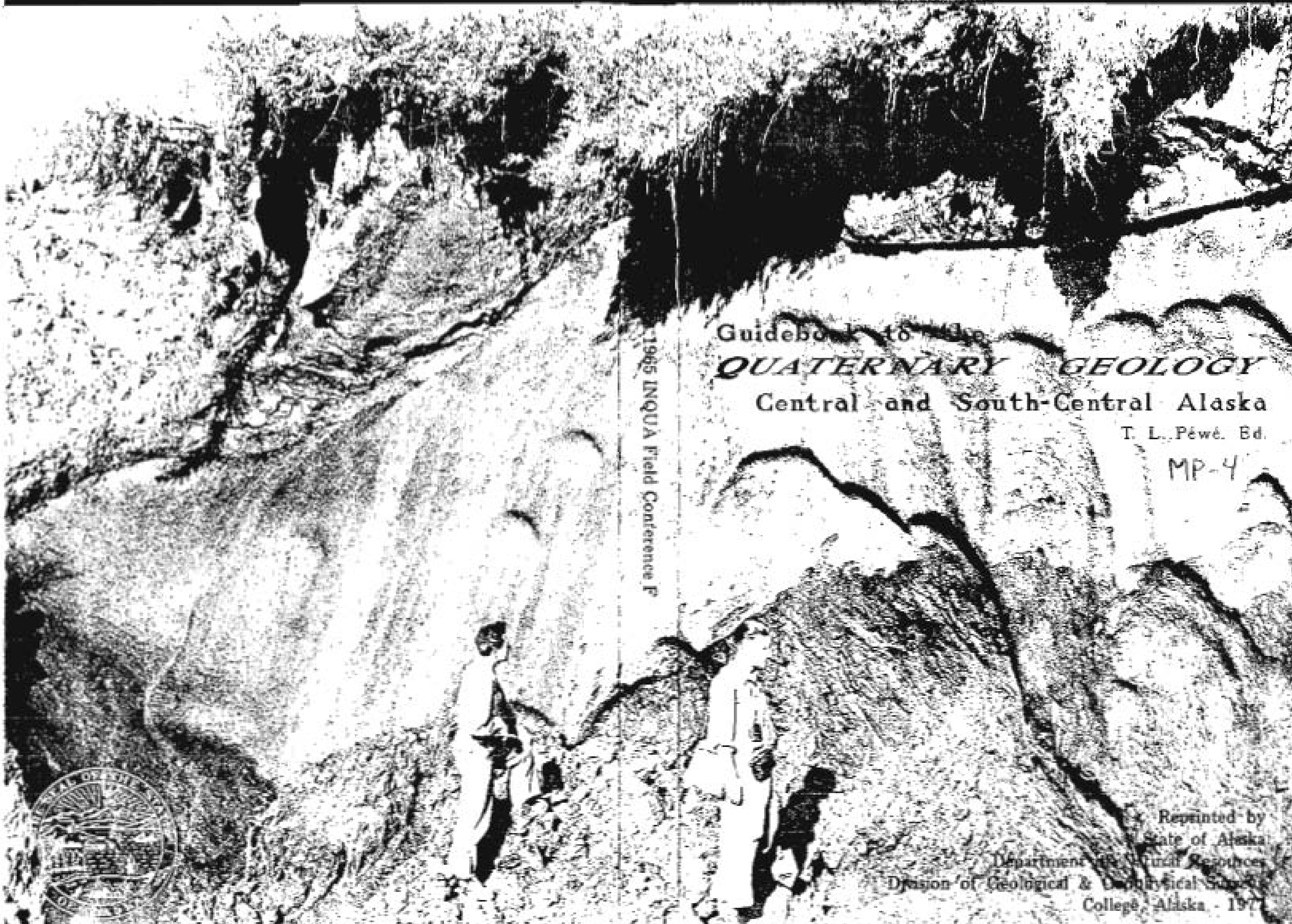

1965 INQUA Field Conference F

Guidebook to the

QUATERNARY GEOLOGY

Central and South-Central Alaska

T. L. Péwé, Ed.

MP-4

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Department of Natural Resources
Division of Geological & Geophysical Surveys
College, Alaska - 197

PREFACE

This book summarizing the Quaternary geology of central and south-central Alaska and containing descriptions of localities between Fairbanks and Anchorage was prepared in 1965 for Field Conference F of the VII Congress of the International Association for Quaternary Research. Its considerable popularity among scientists and nonscientists led to a rapid exhaustion of the limited number of first-edition copies.

In response to a continuing demand for the guidebook, the Alaska Division of Geological and Geophysical Surveys has reprinted the guidebook with the permission of the Nebraska Academy of Sciences. Special appreciation is extended to C. Bertrand Schultz of the Nebraska Academy of Sciences for granting this permission. Troy L. Pewe has updated Figures 1-2, 1-10, and 1-15 for later publications to reflect changes in nomenclature and information collected from the Fairbanks area after 1965. Otherwise the guidebook is unchanged from its first edition.

INQUA

INTERNATIONAL
ASSOCIATION FOR QUATERNARY RESEARCH

VIIth CONGRESS

Guidebook for Field Conference F

CENTRAL AND SOUTH-CENTRAL ALASKA

Start: Fairbanks, Alaska, August 18
End: Anchorage, Alaska, August 29

Conference Organizer

Troy L. Péwé

Conference Leaders

Troy L. Péwé*
Fairbanks area
Central Tanana River Valley
Delta River area, Alaska Range

Oscar J. Ferrians, Jr.,* and
Donald R. Nichols*
Copper River Basin

Thor N. V. Karlstrom*
Upper Cook Inlet area
Matanuska River Valley

Lawrence R. Mayo
Richard D. Reger
Leslie A. Viereck

*Contributor to Guidebook

U.S.A.
1965

VIIIth INQUA CONGRESS

GUIDEBOOKS FOR FIELD CONFERENCES

EDITORS
for Guidebooks:

C. Bertrand Schultz
University of Nebraska

and

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Reprinted by
State of Alaska
Department of Natural Resources
Division of Geological & Geophysical Surveys
College, Alaska - 1977

Cover Photo: *Large foliated ice wedge exposed in retransported loess of Wisconsin age. Dome Creek, 12 km north of Fairbanks, Alaska (photo 1038 by T.L. Pewe, August 1954).*

CENTRAL AND SOUTH CENTRAL ALASKA

SUMMARY OF ITINERARY

OUTLINE OF TECHNICAL FIELD CONFERENCE F

Technical Field Conference F, central and south-central Alaska, from August 18 p.m. to August 29 p.m. is composed of two parts. Part 1 consists of a three day stay in the Fairbanks area of central Alaska. During the three days local field trips will be made to the first 19 Stops of the Field Conference. Also during the three days a symposium will be held titled, "Arctic environment and processes."

Part 2 of the Field Conference consists of a seven day trip to Anchorage from Fairbanks over the Richardson and Glenn Highways. The seven sections of Part 2 are as follows: 1. Middle Tanana River valley; 2. Delta River area, Alaska Range; 3. Gulkana Glacier; 4. Denali Highway and part of Copper River Basin; 5. Copper River Basin; 6. Copper River Basin and Matanuska River valley; 7. Matanuska River valley and Anchorage area.

INTRODUCTION[1]

by

Troy L. Péwé

University of Alaska and
U.S. Geological Survey,
College, Alaska

General Statement

Marine, fluvial, lacustrine, glacial, eolian, and periglacial deposits of Quaternary age are present in the central and southern parts of Alaska and are being formed today. Marine sediments are accumulating along the coast, and glaciers are widespread. Geological processes active in cold regions—periglacial processes such as solifluction, altiplanation, and formation of permafrost—are known throughout much of the area. Dust is being blown from active valley trains and outwash fans and being deposited as loess over adjacent terrain.

Five major areas are considered: the Fairbanks area, the central Tanana River area, the Delta River area of the Alaska Range, the Copper River Basin, and the Upper Cook Inlet area and Matanuska River valley. The Fairbanks area is typical of unglaciated interior Alaska and is characterized by extensive eolian deposits and considerable perennially frozen ground. The central Tanana River valley is a broad river valley with heavily silt-laden glacial streams. The Delta River area of the Alaska Range is characterized by many glaciers and by deposits recording more extensive glaciation in the past. The Copper River Basin has a most interesting record of alternating glacial and lacustrine deposits. The Upper Cook Inlet area and Matanuska River valley record multiple glaciation, as well as provide classic evidence of landslides generated by the Good Friday Earthquake of 1964.

[1] Publication authorized by the Director, United States Geological Survey.

Organization and Acknowledgements

The overall organizer and editor of this guidebook has been Troy L. Péwé. Péwé has been responsible for the sections of the guidebook dealing with the Fairbanks area, central Tanana River valley, and the Alaska Range. Leslie Viereck, Research Botanist, Northern Forest Experiment Station, U.S. Forest Service, University of Alaska, contributed all discussions of vegetation in the sections dealing with the Fairbanks area, the central Tanana River valley, and the Alaska Range. Michael Blackwell, graduate student in geology at the University of Alaska, contributed unpublished information in the Harding Lake and Birch Lake areas and commented on the road log from the Salcha River to Banner Creek. Lawrence Mayo, U.S. Geological Survey, compiled the glaciology section of the Gulkana Glacier Stop.

The archeological statements in sections dealing with the Fairbanks area and the Alaska Range were contributed by Fred Hadleigh-West, Department of Anthropology and Geography, University of Alaska.

The section on the Copper River Basin was prepared by Oscar J. Ferrians, Jr., and Donald R. Nichols. Ferrians was responsible for the road log from Mile 183.3 on the Richardson Highway to Mile 124.8 on the Richardson Highway including the Gakona section Stop. Nichols was responsible for information on permafrost at Mile 130, Richardson Highway, and for the road log along the Richardson Highway from Mile 124.8 to 112.5 and the Glenn Highway from Mile 189 to 170. John R. Williams, U.S. Geological Survey, provided basic information for parts of figures 8-5 and 9-4 and for the road log from Mile 170 to 121 on the Glenn Highway from his unpublished maps and field notes.

The Upper Cook Inlet area and Matanuska River valley was compiled by T. N. V. Karlstrom.

Other leaders for Conference F are: Lawrence R. Mayo, U.S. Geological Survey; Richard D. Reger, Geology Department, University of Alaska; and Leslie A. Viereck, U.S. Forest Service.

References used in compiling the area resumés and the road logs are listed in a general bibliography and in bibliographies at the end of each individual section. References are not inserted with the text except where controversial points are discussed.

All radiocarbon dates are used as years before present. All are cited by laboratory and number. If the date is previously published, the reference is listed in the bibliography. Dr. Jan Lundqvist of the Geological Survey of Sweden kindly made arrangements for radiocarbon dating of remains of carcasses of Pleistocene age from the Fairbanks area so that the information would be available for this guidebook.

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FAIRBANKS AREA

by

Troy L. Péwé

RESUME OF THE QUATERNARY GEOLOGY OF THE FAIRBANKS AREA

The Fairbanks area is in central Alaska approximately 100 miles (160 km) south of the Arctic Circle (fig. 1-1).[1] The area has a continental climate characterized by an extreme range between summer and winter temperatures. The mean annual temperature is 26.1°F (−3.3°C), and the mean annual precipitation is 11.7 inches (29.7 cm).

The Fairbanks area is on the north side of the broad Tanana River valley near the base of the hills that constitute part of the Yukon-Tanana upland. The southern part of the area lies within the Tanana River flood plain at an altitude of 400 feet (120 m) and the rest lies within the

[1] This guidebook has 53 figures, numbered consecutively from 1 through 53, but with a prefix for each number to indicate the day of the trip when the figure is first used. Thus fig. 4-16 would indicate that figure 16 appears in the itinerary for DAY 4. Figures for Days 1, 2, and 3 are all listed as DAY 1.

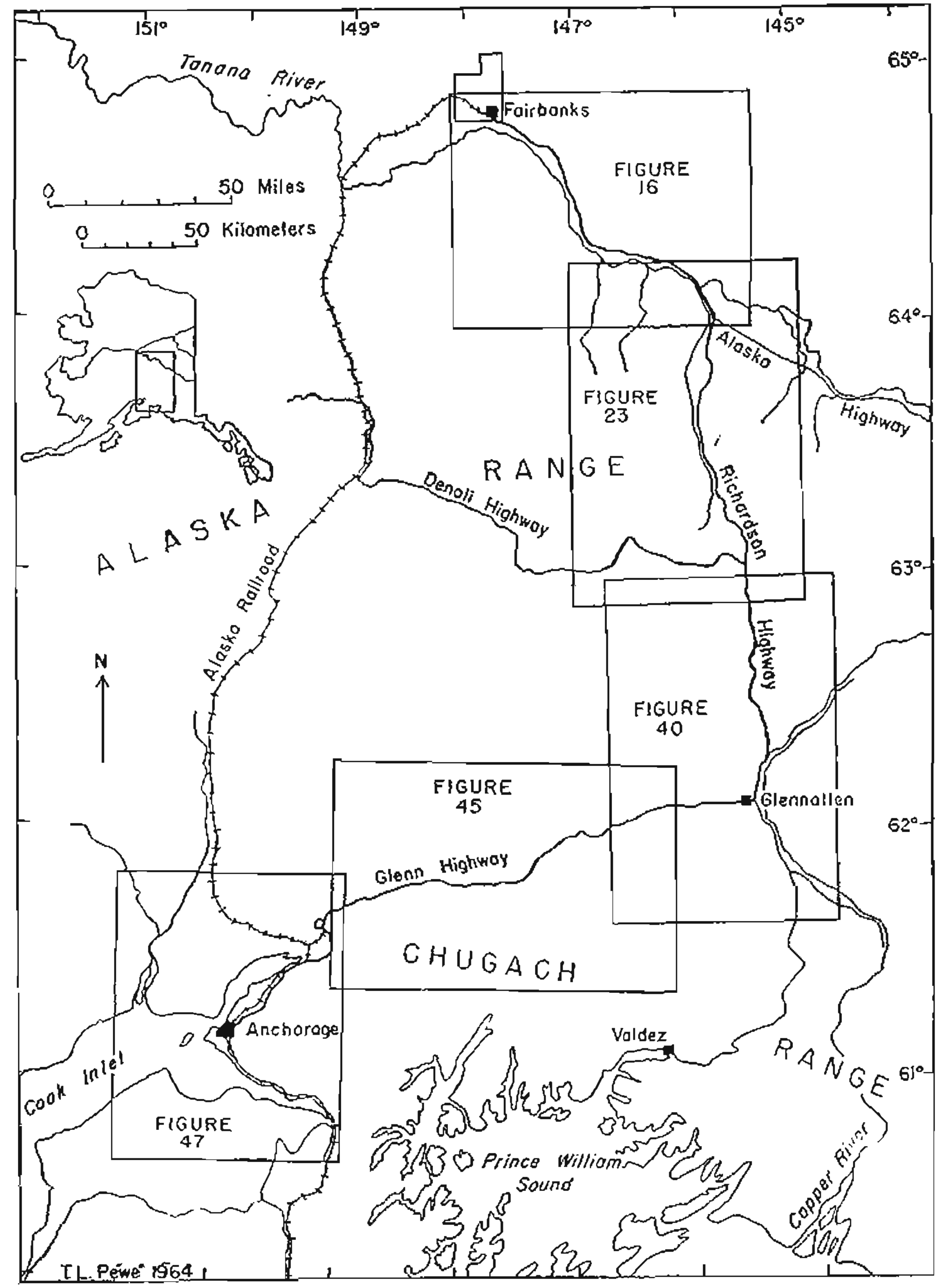

Figure 1-1. Index map showing route of Alaska Field Conference F, INQUA, 1965.

upland. The upland is a maturely dissected area of accordant rounded ridges 2000 to 3000 feet (600 to 900 m) in altitude.

Central Alaska has not been glaciated except in small local mountain masses, but glaciers from the Alaska Range approached within 50 miles (80 km) of Fairbanks during glacial advances, and heavily loaded rivers deposited several hundred feet of silt, sand, and gravel in the Tanana Valley. Aggradation of the trunk valley raised base level and caused tributaries from the Yukon-Tanana upland to aggrade their lower valleys. More than 400 feet (120 m) of sediment was deposited in creek valleys of the upland in the vicinity of Fairbanks. Silt was blown from the Tanana River flood plain and was deposited as loess. Wind-blown silt ranging in thickness from a few inches on summits to more than 150 feet (45 m) on middle slopes blankets the ridges of the uplands.

A complex series of events took place in Quaternary time. The deposits show a record of alternating deposition and erosion of silt and gravel, the formation and destruction of permafrost, and climatic fluctuations ranging from a climate warmer than that which exists now to one colder than the present.

In late Pliocene and/or early Pleistocene time gold placers were formed in creek valleys of the upland; later, great alluviation of coarse angular local gravel occurred in these valleys in response to rising base level of the Tanana River valley. The gravel has been stained brown by percolating ground water. This early period of gravel deposition was followed by erosion and removal of most of the coarse angular local gravel. Streams reconcentrated much of the gold in the earlier placers and deposited additional gold placers. Many of the stream channels of the second gold concentrations are offset from the location of the earlier channels, and therefore, some of the first placer accumulations still exist as fragmentary bench deposits (fig. 1-2). A second cycle of much gravel alluviation followed. The younger gravel is not stained as dark a brown as the older gravel deposit. Both gravel deposits are unconformably overlain by loess, or if in valley bottoms, by retransported loess rich in organic remains.

It is thought that the gravel deposits are Quaternary in age because they contain tusks and large bones of mammoth. Identifiable wood remains are rare, but white spruce is recorded. The poorly sorted, angular gravel grades into and in some instances overlies solifluction deposits, deposits thought to have originated under conditions of a rigorous climate.

If the age of the gravel is Quaternary, it is indeed early Quaternary, because the deposits are deeply buried under two or more loess deposits of considerable antiquity. Because of the antiquity, faunal content, and suggestion of origin under rigorous climatic conditions (solifluction), the two gravel deposits of the creeks in the Fairbanks area are thought to be very early Pleistocene in age, perhaps Nebraskan(?) or Kansan(?) (fig. 1-2).

Throughout the unglaciated part of most of central Alaska there exists a widespread 1 to 10 foot thick (0.3 to 3 m) deposit of poorly sorted and poorly stratified bedrock debris (solifluction deposit) formed by local mass movement of the mantle some time in the past when the climate was more rigorous than now.

In the Fairbanks area a solifluction deposit exists at an elevation as low as 500 feet (150 m) above sea level. The material, which has been called "slide rock" by the gold miners, lies on lower slopes and extends down to and in some instances overlies the auriferous coarse gravels

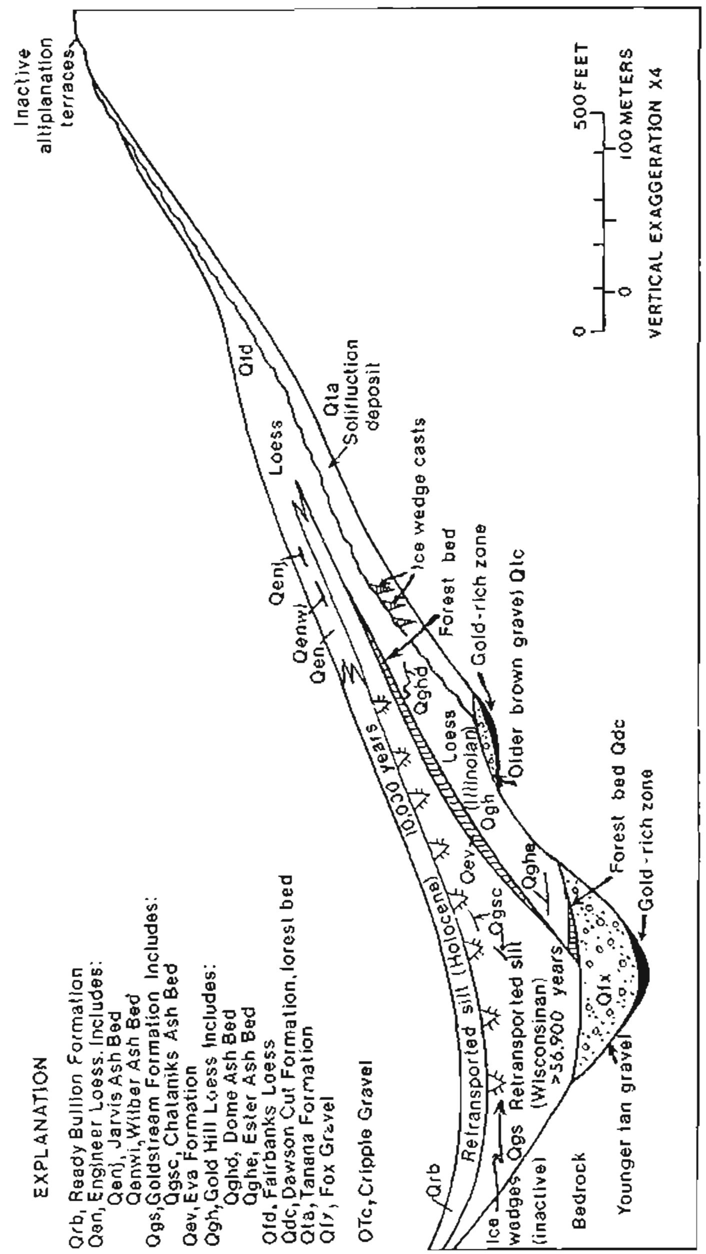

Figure 1-2. Schematic composite cross-section of a creek valley in central Alaska in the vicinity of Fairbanks illustrating stratigraphic relations of the Quaternary deposits. (Modified from figure 4 of U.S. Geological Survey Professional Paper 862 entitled, "Quaternary Stratigraphic Nomenclature in Unglaciated Central Alaska" by Troy L. Péwé, 1975, and figure 20 in U.S. Geological Survey Professional Paper 835 entitled, "Quaternary Geology of Alaska", by Troy L. Péwé, 1975.)

that are early Quaternary in age. Like the gravels, the solifluction deposits underlie all loess and retransported loess deposits. In at least one locality there are solifluction deposits of two ages. In at least two localities fossil ice wedges are present in the solifluction deposits. It is thought that the inactive solifluction deposits are early Quaternary in age.

In later Quaternary time (Illinoian) the hills were blanketed with loess derived from the flood plain of the Tanana River and glacial outwash plains south of the Fairbanks area. Much of this windblown silt was retransported to creek valley bottoms, incorporated much organic debris, including vertebrate remains, and became perennially frozen. Indications of the antiquity of this loess are: (1) its position unconformably beneath a younger silt deposit, the base of which is older than 39,000 years (GXO360) (fig. 1-2); (2) joints that were heavily stained by iron oxide and cemented before the deposit became perennially frozen; and (3) evidence from fossil ice wedges that the loess, having become frozen, was thawed, then again perennially frozen.

Following this loess depositional period there was an erosional period when most of the retransported silt in creek valley bottoms and some of the loess on the hillside slopes and hilltops were removed. The loess was deeply gullied and block slumping occurred. Long parallel gullies more than 30 feet (10 m) deep and 600 feet (180 m) long formed on almost all loess-covered slopes in the Fairbanks area. Permafrost thawed and perhaps disappeared during this warm interval.

In Wisconsin time additional loess was deposited on the uplands. During this period of accumulation much loess was retransported to valley bottoms to form an organic-rich fetid perennially frozen deposit — locally termed "muck" (fig. 1-2). This valley bottom facies of loess of Wisconsin age is 10 to 150 feet (3 to 46 m) thick and contains abundant vertebrate and plant fossils, including partial carcasses of vertebrates that were entombed in the silt and perennially frozen. The most common vertebrate remains in the muck of Wisconsin age, in order of their abundance, are those of bison, mammoth, and horse. The retransported silt (valley bottom facies) of Wisconsin age contains many ice wedges 1 to 10 feet (0.3 to 3 m) wide and up to 30 feet (10 m) high.

The gullies and ridges of pre-Wisconsin age cut in loess of middle and upper slopes in the Fairbanks area were rounded and subdued by the blanket of loess deposited over these undulations in Wisconsin time.

About 5000 to 8000 years ago there occurred a short warming interval that caused the permafrost table to be lowered a few feet and the top of the ice wedges to melt down about 1 to 10 feet (0.3 to 3 m). Loess and the valley bottom facies (retransported silt) of the loess that was deposited since and during the thawing lies unconformably over the thawed-down flat-topped ice wedges and retransported silt of Wisconsin age (fig. 1-2). This Recent silt is 1 to 25 feet (0.3 to 8 m) thick, and all but the upper 4 or 5 feet (1.5 m) is perennially frozen. The silt contains no bones of extinct animals.

On the hilltops the loess deposited in Illinoian, Wisconsin, and post-Wisconsin time constitutes one relatively uniform loess layer, and to date it has not been possible to differentiate this loess into layers of separate ages. All the upland loess is grouped together under the name Fairbanks Loess (fig. 1-2).

Permafrost exists nearly everywhere in the Fairbanks area except beneath hilltops and moderate to steep south-facing slopes (fig. 1-3). Sediments of the flood plain are perennially frozen to depths of as much

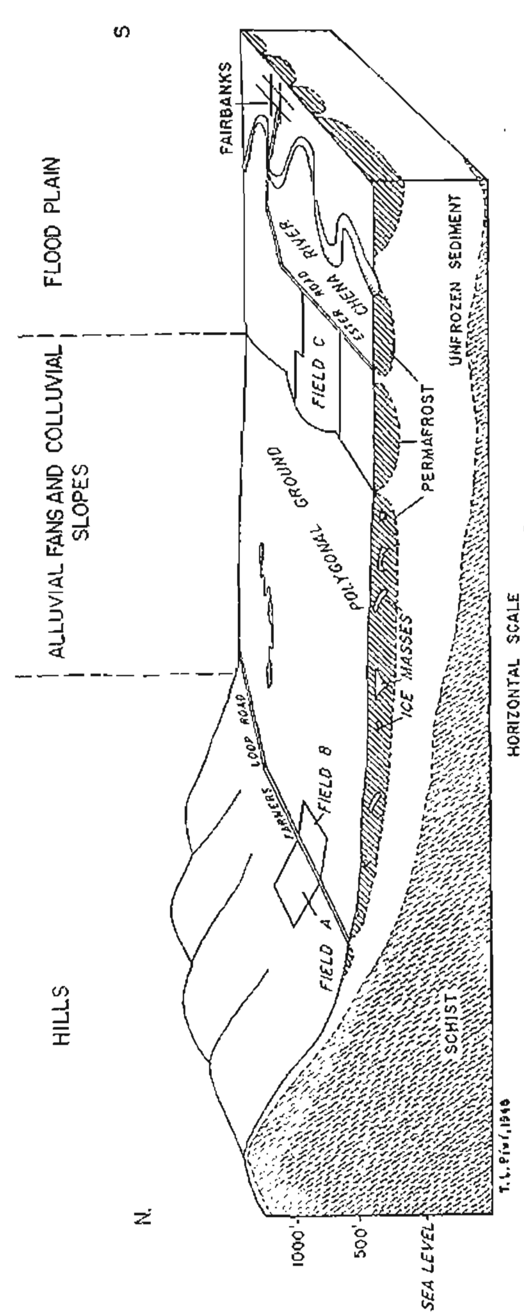

Figure 1-3. Diagrammatic sketch showing character and distribution of permafrost in the Fairbanks area, Alaska. Field A is not underlain by permafrost. Field B is underlain by permafrost containing large ice masses; thermokarst mounds and pits may form in the field when the ice melts. Field C is underlain by permafrost which does not contain large ice masses; thermokarst mounds and pits will not form at the surface. (From Péwé, 1954)

as 265 feet (81 m), but not everywhere is permafrost encountered in a single layer. The thickness of the frozen layer varies widely and in many areas permafrost is lacking. Thawed areas occur beneath existing or recently abandoned river channels, sloughs, or lakes. Elsewhere layers of frozen sand and silt are intercalated with unfrozen layers of gravel. Depth of permafrost in the undisturbed areas may be from 2 to 3 feet (0.6 to 0.9 m) or more than 4 feet (1.2 m) on the slip-off sides of rivers.

Fires, clearings, and construction since 1903 have increased the depth to permafrost 25 to 40 feet (8 to 12 m) in many places. Ice in the perennially frozen sediments of the flood plain consists of granules in cement between the grains. Large ice masses are absent.

Permafrost in the retransported valley bottom silt of the creek valleys and lower slopes reaches a thickness of at least 175 feet (76 m) near the flood plain but decreases toward the hills, pinching out at the base of steep south-facing slopes, but extending nearly to the summit of north-facing slopes. Permafrost in these sediments contains large masses of clear ice occurring as horizontal sheets, vertical sheets, wedges, and saucer-shaped and irregular masses. The ice masses are of foliated ice (ice wedges) and range from less than 1 foot to more than 15 feet (0.3 to 5 m) in thickness and from 1 to 50 feet (0.3 to 15 m) in length. Much of the ice is arranged in a polygonal or honeycomb network enclosing silt polygons 10 to 40 feet (3 to 12 m) in diameter.

Temperature of the permafrost in the Fairbanks area at a depth below the effect of seasonal temperature fluctuations (30 to 50 feet) (8 to 15 m) is about 31°F (−0.5°C). Permafrost is forming in the area today under favorable circumstances. If the vegetation cover is removed the permafrost thaws (degrades). Thawing of the permafrost in the flood plain with low ice content results in little or no subsidence of the ground. In creek valley bottoms, thawing of ice-rich retransported loess (muck) results in great differential subsidence of the ground.

Permafrost in the area is thought to be Wisconsin in age. It is thought that earlier permafrost in the area disappeared in Sangamon time, and present permafrost is the result of the rigorous climate of the latest glacial stage.

The vegetation in the Fairbanks area is a complex mosaic which results from a long history of forest fires, from differences in slope exposure and parent material, and from a complicated pattern of permafrost. On well-drained upland soils where permafrost is lacking or at depths of more than 4 feet (1.2 m), large areas are covered by relatively young stands of paper birch *(Betula papyrifera)* and aspen *(Populus tremuloides)* which have developed directly after forest fires or cutting. An understory of young white spruce *(Picea glauca)* in many of the aspen and birch stands shows that they will eventually be replaced by stands of white spruce. The white spruce and the white spruce/paper birch types are widespread on well-drained upland soils that have not been burned in the past 200 years. Associated with the white spruce is a sparse shrub layer of highbush cranberry *(Viburnum edule)*, rose *(Rosa acicularis)*, alder *(Alnus* spp.), and willows *(Salix* spp.), and a thick moss layer of *Hylocomium splendens*, *Pleurozium schreberi*, and *Rhytidiadelphus triquetrus*.

Upland areas closely underlain by permafrost are usually occupied by black spruce *(Picea mariana)* in either open or dense stands. Black spruce often seed indirectly after

fire in this type of area but may be preceded by stands of alder or paper birch. Associated with the black spruce are the shrubs, Labrador tea *(Ledum groenlandicum)*, bog blueberry *(Vaccinium uliginosum)* and dwarf birch *(Betula glandulosa)*. Ground cover consists of mountain cranberry *(Vaccinium vitis-idaea)*, *Sphagnum* spp., and other mosses, and lichens (especially *Cladonia* and *Peltigera* species). In such areas frozen ground is usually within 1 to 3 feet (0.3 to 1 m) of the surface, even during the late summer months.

On the flood plain, permafrost is lacking under the youngest alluvial deposits, especially on the slip-off slopes. Stands of willow *(Salix* spp.), balsam poplar, or white spruce occur on these sites. These areas have been relatively protected from fire by rivers and sloughs and because of this some of the most extensive stands of commercial white spruce in interior Alaska occur along these rivers. On most of the older alluvial deposits permafrost is close to the surface and the vegetation consists of slow-growing black spruce and larch *(Larix laricina)* or sedge and sphagnum bogs.

Permafrost forms in the alluvial deposits partly as a result of the insulating effect of the vegetation. White spruce stands develop thick layers of mosses which result in a thick organic layer. This layer acts more efficiently as an insulator during the hot summer periods when the moss is dry than it does during the wet or cold periods when the moss is frozen and saturated. As a result the soil becomes colder and eventually perennially frozen. This creates a shallow root zone with wet, cold conditions, a situation that is more suitable for black spruce than for white spruce, and the white spruce is replaced by black spruce as the older trees die. Continued swamping may occur with sphagnum mosses replacing the forest mosses, and eventually the black spruce may be replaced by either sphagnum or sedge bogs. The importance of the insulating effect of the vegetation is clearly shown by the rapid lowering of the permafrost table after clearing or other disturbance of the vegetation.

This relationship of forest type to the presence of permafrost is a close one in the Fairbanks area, and the vegetation can be used as a general indication of the permafrost conditions. Black spruce, larch, and bogs nearly always indicate the presence of frozen ground within a few feet (about 0.5 m) of the surface. White spruce and aspen usually indicate a permafrost-free area or one in which the active layer is several feet thick. Paper birch occurs on sites free of permafrost, but it also occurs over permafrost when the active layer has been temporarily deepened as a result of burning or clearing.

Locality Descriptions

Stop 1-1:[1] University of Alaska (fig. 1-4). Sec. 6, T. 1 S., R. 1 W.

Solifluction deposits

In many excavations on the main campus of the University of Alaska a 5-foot (1.5 m) thick solifluction layer of bedrock (schist) debris is exposed. In some exposures it is evident that two solifluction layers are present (fig. 1-5). The basal layer contains well-formed ice wedge casts of clean eolian (?) sand. The upper part of the casts are drawn out or destroyed by the upper solifluction layer. Both solifluction

[1]This guidebook has 66 stops, numbered consecutively from 1 through 66, but with a prefix for each number to indicate the day of the trip when the stop should occur. Thus **Stop 4-20** would indicate that Stop 20 is included in the itinerary of DAY 4. Stops for Days 1, 2, and 3 are all listed as DAY 1.

layers are composed of clayey, silty schist debris, but the upper layer is finer grained. The solifluction layers are overlain unconformably by loess 1 (0.3 m) to more than 25 feet (8 m) thick.

Present interpretation of the solifluction layers is that the lower layer indicates a cold period with a climate much more rigorous than now, or even than in late Pleistocene time. After the solifluction layer formed, ice wedges developed—either in the same cold period or in a separate later cold period. After a warm period during which the ice wedges thawed and were replaced by sand, another cold period produced the second solifluction layer. The age of the solifluction layers is assigned to early Pleistocene because: (1) they are evidence for a rigorous climate, (2) they are overlain by one or two loess deposits of middle to late Pleistocene age, (3) in valley bottoms, the solifluction deposits grade into coarse creek gravels which contain Pleistocene fossils and are overlain by silt deposits of Illinoian and Wisconsin age.

Ice wedge casts

The ice wedge casts or fossil ice wedges are a few inches to 2 feet (10 to 60 cm) wide at the top and 1 to 5 feet (0.3 to 1.5 m) long. In plan view, they form polygons 5 to 15 feet (1.5 to 5 m) in diameter. The sand filling is well sorted and the grains are frosted. Seventy-five percent of the grains are between 0.1 to 0.3 mm in diameter. If the sand were fluvial in origin, it is thought that the solifluction layer would have been eroded away by the stream. Dunes bordering the Tanana River valley were present in the Fairbanks area in early Pleistocene time as they were along the Tanana River valley near Big Delta in late Pleistocene time.

Loess

The Fairbanks Loess is 1 to 80 feet (0.3 to 25 m) thick on the University of Alaska campus on College Hill. The color is tan with dark carbonaceous and iron-stained bands. The silt is well sorted (fig. 1-6), 80 to 98 percent of the particles falling in the silt-size range. Mechanical composition varies little with aerial extent or with depth (fig. 1-7). The silt grains are angular and fresh, although slightly iron stained. A typical sample from College Hill contains abundant quartz, considerable muscovite and feldspar, and less calcite, chlorite, cloritoid, clinozoisite, epidote, garnet, hornblende, opaque minerals, tourmaline, and zircon. A heavy-mineral analysis of loess from College Hill is listed in Table 1.

The loess is massive, has no stratification except where slightly retransported. In such loess, faint stratification consists of iron-oxide stained horizons, organic films, vague color bandings of short lateral extent, and volcanic ash layers. The loess stands in sheer cliffs and is readily subject to gullying.

Vertebrate remains are abundant in the retransported loess of valley bottoms, but few bones of Pleistocene vertebrates have been found in the loess of the hillside locations because of lack of exposures and because most of the bones have been transferred to valley bottoms. *Bison* sp. and Mammoth bones have been found in the Fairbanks Loess on College Hill.

The silt on College Hill comprises loess of both Wisconsin and Illinoian age but no break, such as is present on lower valley slopes (fig. 1-2), can be found here between the two loesses.

Volcanic ash

At least four thin white vitric volcanic ash layers are interbedded with the loess and retransported loess of

the Fairbanks area. They range in thickness from one-tenth of an inch to 6 inches (3 mm to 16 cm) and have sharp contacts with the silt above and below. Chemically and petrologically they are similar, but stratigraphically the ash layers are distinctive. The Ester Ash Bed is a gritty 6-inch-thick (16 cm) gray to white pure glass (index 1.53) layer exposed near the base of Illinoian(?) loess near Ester, Gold Hill, and on the University of Alaska campus. Younger ash layers, not yet formally named, are exposed in many mining and natural exposures of loess and retransported loess.

Soil

Loess on hilltops and upper and middle slopes gives rise to a brown soil termed the Subarctic Brown Forest soil. This soil is limited to the interior part of the state, and is characterized by a surface organic mat overlying an A_1 horizon. The A_2 horizon is pale brown and overlies a yellowish-brown color B horizon which grades into a yellowish-gray C horizon. The A horizon is about 0 to 4 inches (10 cm) thick and the B horizon is up to 20 to 22 inches (50 or 55 cm) thick. The color of the B horizon is described as one that has turned brown from the influence of weathering without evidence of illuviation of either iron or humus compounds. The absence of iron and humus from the B horizons and the presence of an A_1 horizon indicate that there is no podzol development. The upper horizons are not dark and mixed with organic debris because organisms necessary for mixing are absent and because of the slow rate of decay of organic matter.

Textural studies indicate no important changes in clay content from one soil horizon to another, or even throughout the parent loess. Weathering and clay translocation are not primary factors in the origin of the soil. The same general suite of minerals is found throughout the soil profile, and the clay minerals are not due to weathering in place. The presence of minerals subject to easy alteration emphasizes the youthfulness of the soil in a weathering sequence. The brown and red colors in the B horizons are due to coatings of free iron oxide.

Thin textural bands occur in the B horizon in the Subarctic Brown Forest soils of the Fairbanks area. The bands are parallel to the surface and are high in fresh clay-sized particles. The textural bands are not generic and not a part of soil de-

Table 1. Heavy-mineral analysis of upland silt from College Hill, College, Alaska.† (From Péwé, 1955, Table 2, p. 710)

Mineral	Numerical Frequency in Percentages
Opaque minerals*	37.0
Zoisite	10.3
Hornblende	10.1
Garnet**	9.6
Epidote	7.2
Unknown fragments	6.0
Clinozoisite	3.8
Augite	3.1
Sphene	3.1
Tourmaline	2.8
Hypersthene	2.6
Zircon	1.9
Apatite	1.4
Rutile	.7
Enstatite	.2
Tremolite	.2

*Approximately 70 per cent magnetite and ilmenite.

**All isotropic grains listed as garnet.

†Separation made with acetylene tetrabromide; permanent mount made in Canada balsam; 415 grains counted.

velopment. They probably are depositional, the result of slight changes in source and deposition of loess. Frost action also has been suggested as a process involved in the concentrating of the clay. The Subarctic Brown Forest soils are youthful and relatively stable in their environment.

Archeology

On the edge of the hill now occupied by the University is the Campus Site, one of the best-known early sites in North America. The site was discovered in the early 1930's and was excavated periodically over a period of about 5 years. The chief reason for its fame lies in the fact that this was the first site in the Americas in which there was found a typological identity in artifact types with some known from the Old World. Specifically, it was determined that the prepared microcores and blades characteristic of this assemblage were similar to those excavated in the Gobi Desert. Exact dating of the occupancy here is not clear. In the past 10 years a number of other related sites have been found.

The Campus Site deposits, like most of those of Interior Alaska, are shallow. There probably is some mixture of later materials with those of the core and blade tradition. One of the sites at Donnelly Dome (Stop 5-32) which was excavated in the summer of 1964, appears to be a very clear representation of this tradition, and it is hoped that radiocarbon dates will be obtained from it.

Stop 1-2: Gold Hill. Sec. 3, T. 1 S., R. 2 W.

An exposure 1.5 miles (2.5 km) long and 0.2 mile (0.3 km) wide on Gold Hill was created by placer gold mining operations in 1949-53. Here is exposed a rather complete stratigraphic section of Quaternary deposits (fig. 1-8). Thawing and slumping has now destroyed the details, but the 202-foot (62 m) thick section of Fairbanks Loess is still present, as are the brown gravels of early Pleistocene age in tailing piles.

Stop 1-3: Ester placer mining area. Sec. 8, T. 1 S., R. 2 W.

Quaternary stratigraphy

Large-scale placer gold mining with dredges began here in the late 1920's and was terminated July 31, 1964. Millions of dollars worth of gold was recovered from this area. The gravel and gold was eroded from Ester Dome and deposited within 1 to 6 miles (1.5 to 10 km) of its source. The gold is fine grained, while the gravel is coarse, subangular and of local rock types: quartz, micaceous schist, chlorite schist, phyllite slate, gneiss. Almost all gravel exposed consists of tailing piles formed when the gravel was discharged from the dredge. The gravel here has been moved only slightly by mining operations, but has been overturned and the coarse and fine fractions separated during gold extraction.

To the north from this vantage point gravel exposed in mining operations in Ready Bullion Bench and Ready Bullion Creek,[1] Eva Bench and Eva Creek, and Ester Creek can be seen. Gravel near the "Island" and Cripple Creek can be seen to the south (fig. 1-9). Even casual examination reveals that the gravels of Ready Bullion and Eva Benches, as well as the gravel near the "Island" between Ester and Cripple Creeks, is distinctly browner (more

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[1]Bullion Creek in figure 1-4 and on U.S. Geol. Survey Fairbanks D-3 topographic quadrangle map (1949) should be Ready Bullion Creek. The new edition of the topographic map in press has this correction.

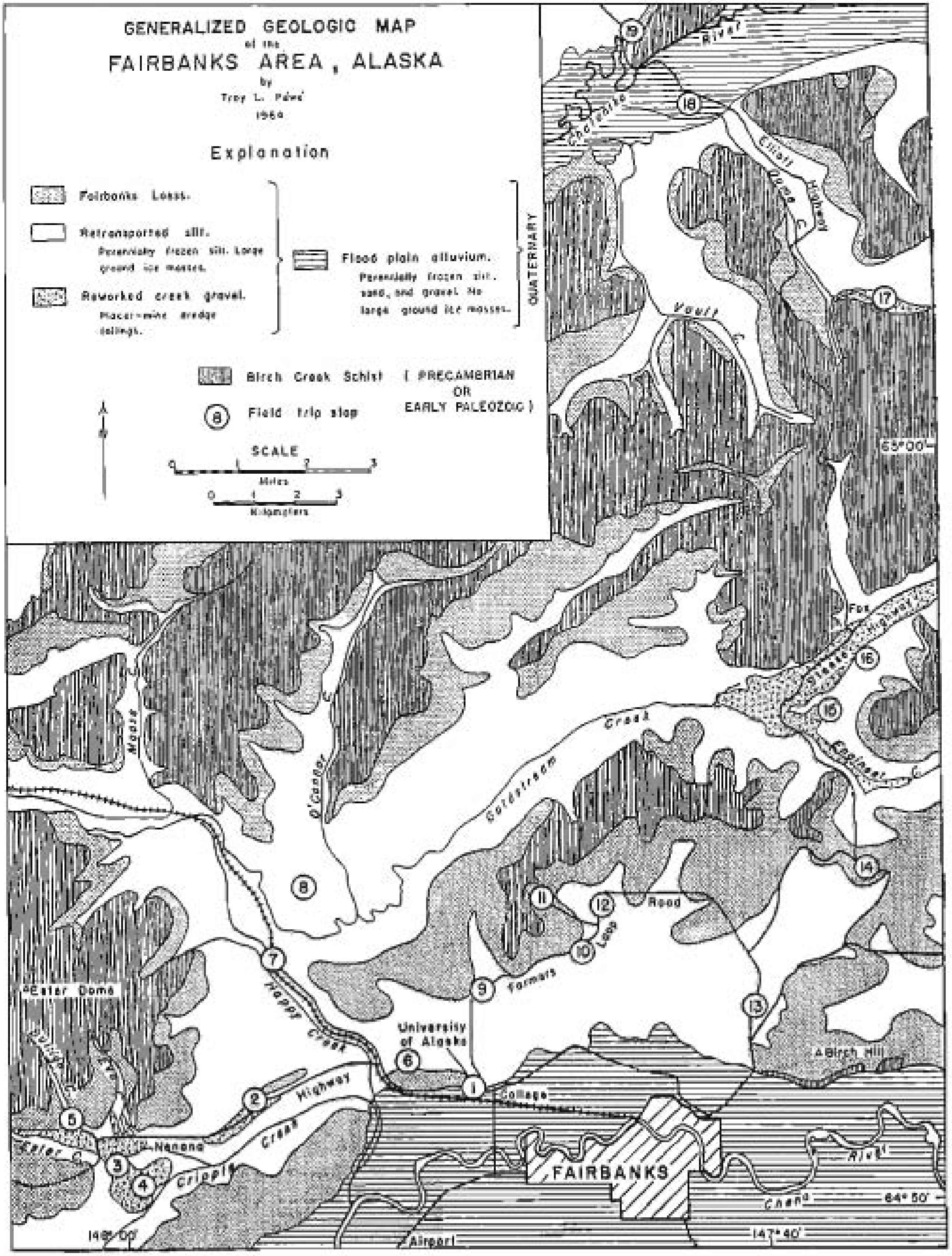


Figure 1-4. Generalized Geologic map of the Fairbanks area, Alaska, showing Field Trip Stops.

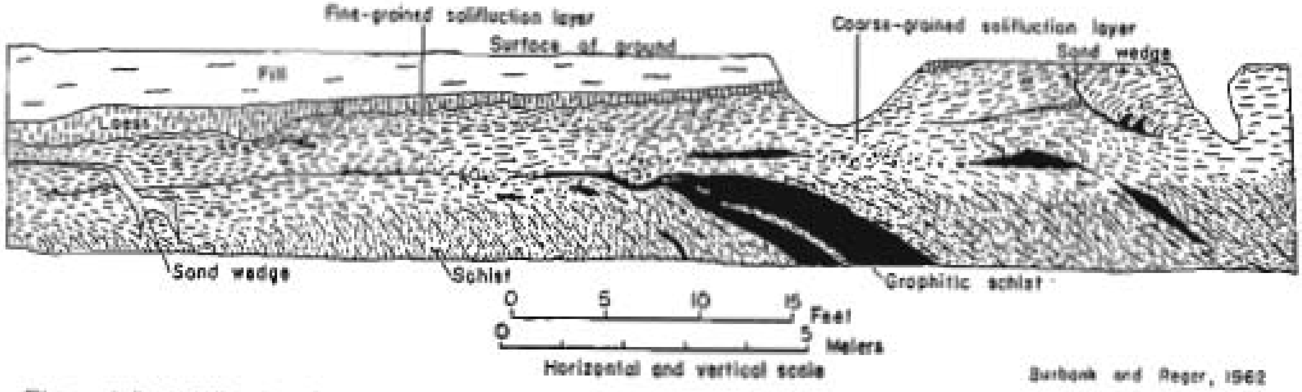


Figure 1-5. Solifluction deposits and ice wedge casts exposed in an excavation for the Duckering Building, University of Alaska campus, College, Alaska.

heavily stained with iron-oxide) than the grayish-tan gravel of Ready Bullion, Eva, Ester, and Cripple Creeks. The darker brown gravel deposits are on benches or otherwise not directly related to modern drainage. They are auriferous gravels of the earliest gravel stage in the Fairbanks area (fig. 1-2) and are thought to be Nebraskan(?) in age.

During the early gravel stage, Eva and Ready Bullion Creeks drained south across present Ester Creek into Cripple Creek and then eastward over what is now known as Gold Hill (fig. 1-9). The gravel exposed in Gold Hill placer cut (Stop 1-2) is of this age.

After a period of gravel alluviation and slight shifting of creek channels the streams cut down, forming new bedrock channels and leaving the old auriferous gravel as remnants on benches (fig. 1-2). Gold was concentrated on bedrock in these new channels; later there occurred alluviation of gravel to a thickness of from 10 to more than 100 feet (3 to

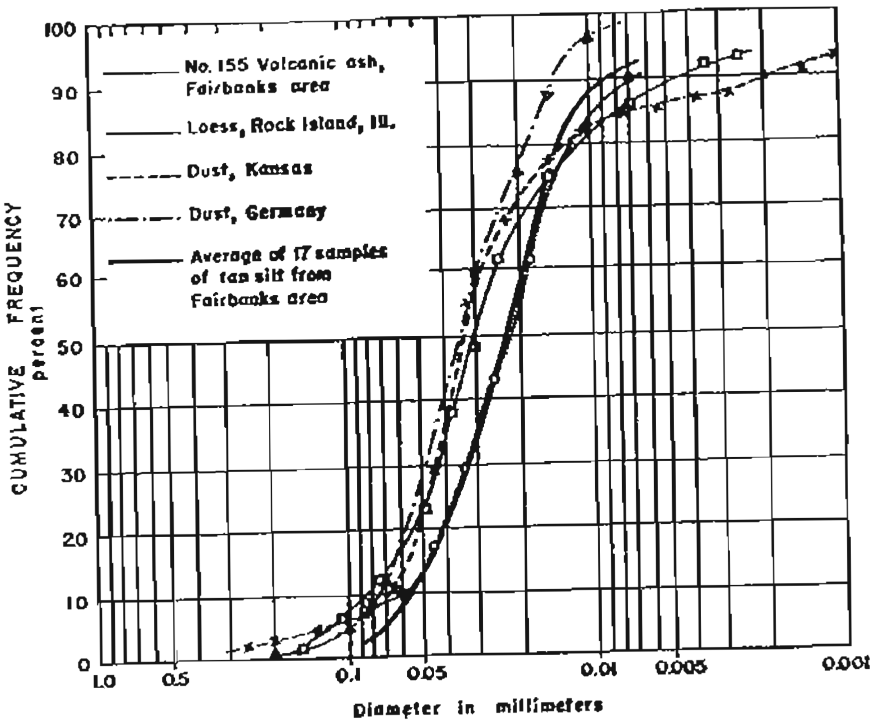


Figure 1-6. Comparison of cumulative-frequency curves of upland silt from Fairbanks area, with cumulative-frequency curves of volcanic ash from Fairbanks area (Ester "Island" section), loess from Rock Island, Illinois and modern wind-deposited dust from Germany (Zeuner, 1949, p. 27) and Kansas (Swineford and Frye, 1945, p. 252). Alaskan and Illinoian samples collected by T. L. Péwé and analyzed by the Corps of Engineers, United States Army, Rock Island, Illinois. (From Péwé, 1955, fig. 11)

30 m). This later gravel is early Pleistocene in age, perhaps Kansan(?). Where old and new gold channels cross, the placer gold concentration is generally higher than elsewhere.

Subsequent to the second gravel period, loess was deposited. This loess, plus retransported loess in valley bottoms, masked the auriferous gravel channels until they were revealed by shaft mining and prospect drilling; and later by dredging.

Vegetation

Various stages in revegetation of the gravel of dredge tailing piles can be seen in the Ester Creek area. The successional sequence varies with the coarseness of the original deposit and with the relation of the area to the water table. On the coarser deposits left by the dredges birch, aspen, balsam poplar and willows become established directly, whereas on the finer deposits there forms first a nearly continuous cover of light seeded forbs and grasses including narrow leafed fireweed (*Epilobium angustifolium*), *Erigeron* spp., Yarrow (*Achillea borealis*), several Cruciferae, fox tail (*Hordeum jubatum*), hairgrass (*Deschampsia* spp.), *Polygonum alaskanum*, and many others. Willows, balsam poplars, and birches soon replace this earlier weedy stage. In a few areas on the tailings white spruce have become established.

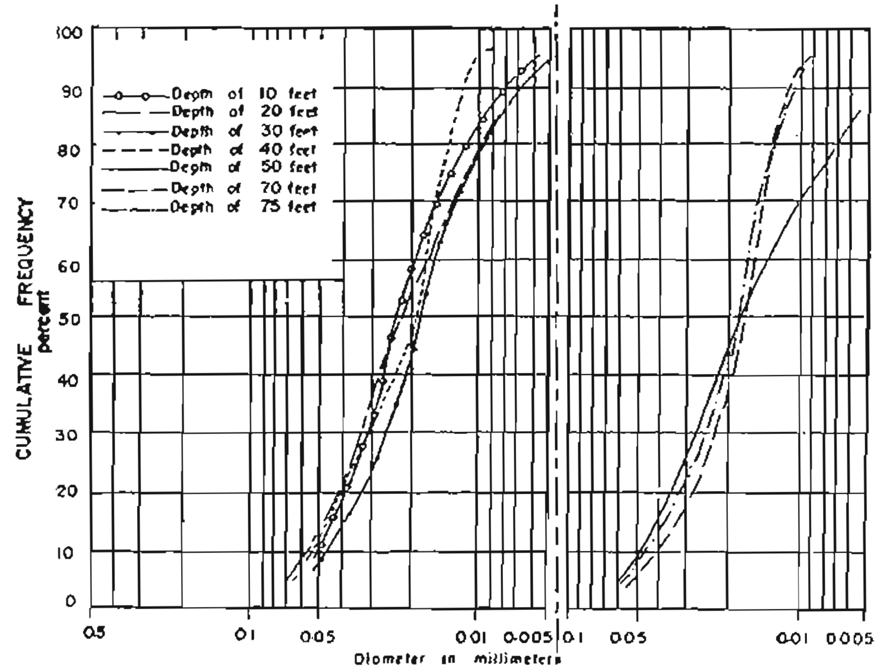


Figure 1-7. Cumulative-frequency curves of upland silt from drill hole near the top of College Hill, Fairbanks, Alaska. (From Péwé, 1955, fig. 8)

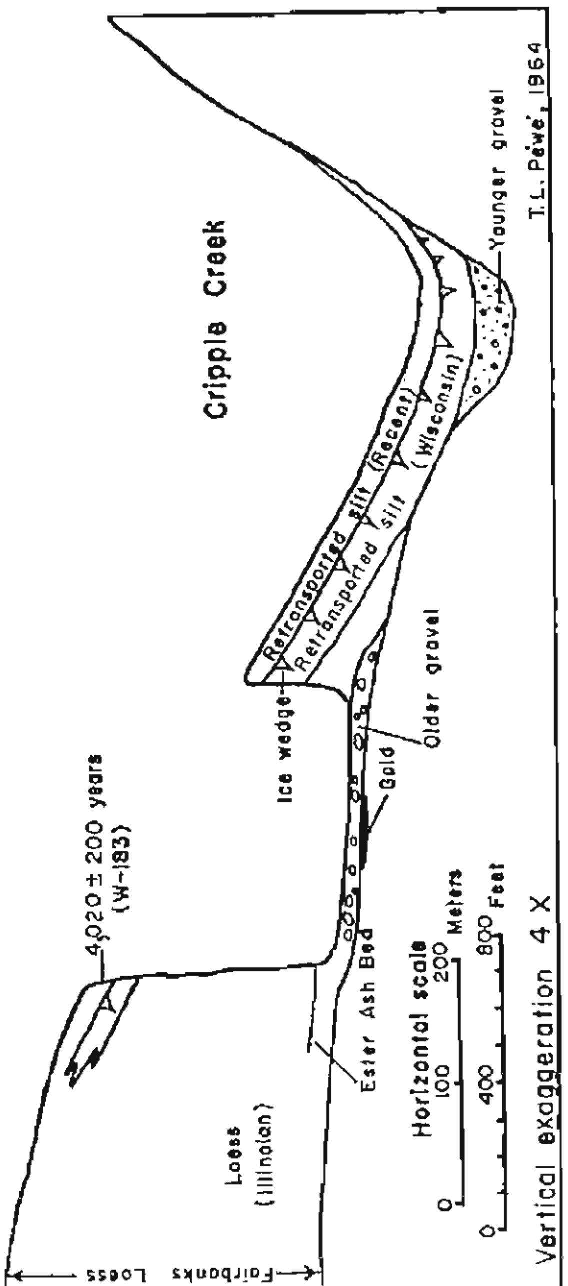


Figure 1-8. Diagrammatic cross section of the Quaternary stratigraphy at Gold Hill, 8 miles (14 km) west of Fairbanks, Alaska. (After Péwé, 1952)

In the depressions, especially where small ponds have been formed, the tailings have been colonized by willows, horsetails (*Equisetum* spp.), and several species of sedge (*Carex* spp.).

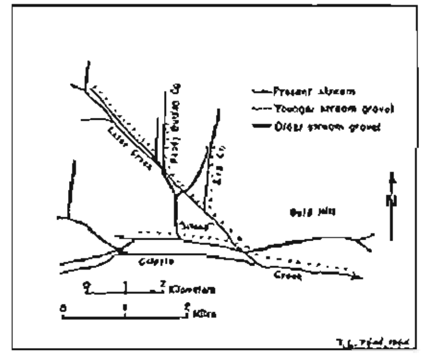


Figure 1-9. Map of present and earlier Quaternary stream channels in the Ester area, 10 miles (16 km) west of Fairbanks, Alaska.

Stop 1-4: Cripple Creek. Sec. 8, T. 1 S., R. 2 W.

Undisturbed gravel

In the walls of the dredge pond 10 to 20 feet (3 to 6 m) of sandy, silty, grayish-brown angular gravel of Kansan(?) age is exposed. The gravel is imbricate and interbedded with sand lenses and layers.

Gold dredge

Dredge No. 10 of the United States Smelting, Refining, and Mining Company was the last of the many gold dredges operating in the Fairbanks area to be shut down. This dredge is the largest in the Fairbanks area and was shut down on July 31, 1964, bringing gold dredging operations in the Fairbanks area to an end.

Placer gold was discovered in 1902 about 16 miles (26 km) north of Fairbanks and, within a few years, the region became one of the greatest gold-producing regions of Alaska. Early placer mining was by underground methods and by the early 1920's most of the richer deposits were exhausted. About $70,000,000 was produced by underground mining. In the middle Twenties a revival of mining in the Fairbanks gold region was stimulated by the initiation of large-scale operations with huge gold dredges and hydraulic stripping of the frozen overburden. Since 1928 more than $150,000,000 worth of gold (at the present price) was mined in the region by the United States Smelting, Refining, and Mining Company.

The mining of gold-bearing gravel in central, northern, and western Alaska is different from the conventional dredging of placer gold in the temperate and tropical latitudes. In the far North not only is most of the gravel perennially frozen, but it is overlain by 10 to 150 feet (3 to 46 m) of perennially frozen ice-rich retransported silt. This barren silt must be removed and the gravel must be thawed before it can be dredged.

After the location and amount of gold is determined by drilling, the frozen retransported silt (muck) overburden is removed by "stripping." Stripping consists of washing away the daily accumulation of thawed silt with water under pressure through giant hydraulic nozzles. The fine-grained retransported loess is carried away easily by the small streams of water under pressures ranging from 50 to 150 pounds per square inch (0.3 to 1 kg/cm^2). Water for the removal of much of the frozen silt was brought 90 miles (145 km) through ditches from the Chatanika River. Water for the operations at Cripple Creek was pumped from the nearby Chena River.

After the frozen retransported silt is removed, the perennially frozen gold-bearing gravel is thawed by introducing water into the ground at average summer temperatures and low pressures through pipes extending to bedrock. The water percolates back to the surface thawing the ground as it moves. Stripping and thawing is limited to the summer months.

After the gravel is thawed the ground is dredged by standard floating bucket-line dredges of the California type. The buckets on Dredge No. 10 are 10 cubic feet (0.28 m^3) in size and approximately 10,000 cubic yards (7500 m^3) of auriferous gravel can be transported daily into the dredge. The dredge is able to dig 72 feet (22 m) beneath the level of the water. Dredge No. 10 was built in 1939, and like all the dredges in the Fairbanks area, was operated electrically. Once the gold-bearing gravel is in the dredge, the gold is separated from the gravel by mechanical shaking and by use of amalgam. The barren gravel is taken up the stacker and dropped to form dredge tailing piles characteristic of the area.

Stop 1-5: Ready Bullion Creek. Sec. 6, T. 1 S., R. 2 W.

Stratigraphy

Silt of Illinoian, Wisconsin, and post-Wisconsin age (fig. 1-10) and gravel of two ages is represented in the exposures at Ready Bullion Creek and Ready Bullion Bench. A thin deposit of coarse brown auriferous gravel lies on a bedrock bench west of the present channel of Ready Bullion Creek. This gravel is early Pleistocene, Nebraskan(?), in age, and the younger gravel of Ready Bullion Creek is thought to be perhaps Kansan(?) in age (fig. 1-10).

In the Ready Bullion Bench exposure a greenish silt (loess) of Illinoian age overlies the gravel and is in turn unconformably overlain by organic-rich retransported silt (muck) of Wisconsin age. The base of the retransported silt is older than 38,000 years. In fresh exposures the perennially frozen Illinoian loess is green, quite in contrast to the overlying black to gray fetid silt of Wisconsin age. The green color is due to the presence of ferrous iron, the result of downward percolating ground water when the sediment was thawed probably in the warm interval prior to the deposition of loess and muck of Wisconsin age.

Overlying the Wisconsin silt is a thin (2 to 5 feet thick) (0.6 to 1.5 m) organic-rich silt layer of Recent age. It is bedded, much less massive than the Wisconsin silt, and lies unconformably over the muck and flat-top ice wedges of Wisconsin age. The base of the silt has been dated here as 8,000 years; the date was determined on retransported sticks.

Retransported silt (muck)

Typical fetid "muck" is exposed here during summer "stripping" operations. The dark silt has the same cumulative size-grain analysis curve, mineral composition, and chemical composition as the loess. It is loess that was retransported to valley bottoms and incorporated minute fragments of burned vegetation as well as larger plant remains. It is poorly stratified and dips downslope parallel to the surface. Much, if not most of the silt, has been transported from uphill by slope wash (probably spring snow melt water); however, presence of folded beds in the silt indicate that some sort of soil creep or solifluction played a role in transporting the silt and vegetation debris.

Vegetation

Many well-preserved macro- and micro-plant specimens are present in the frozen silt. No species have been found in the Illinoian and Wisconsin silts that are not growing in the area today. Pollen analyses show a higher percentage of tundra pollen in the frozen silts than in the modern

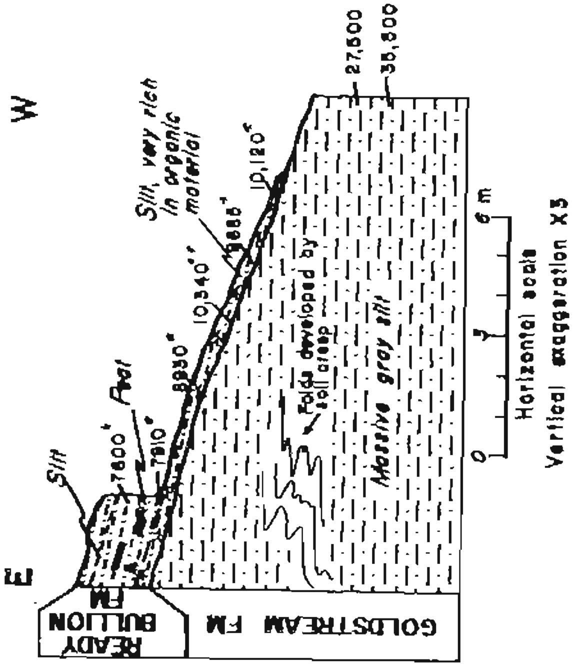


Figure 1-10. Diagramatic section of perennially frozen Quaternary sediments exposed in the upper part of the east wall of Ready Bullion bench placer mining cut on Ready Bullion Creek, 17 km west of Fairbanks, Alaska. Numbers are radiocarbon years; those with an asterisk (*) are average of two or more runs by different laboratories. Radiocarbon years given in the type section for Ready Bullion Formation have been run twice or more, and in all cases the date has averaged younger. Two asterisks (**) indicate date of samples probably a few years too old in as much as it was not run by other laboratories. (Figure 12 from U.S. Geological Survey Professional Paper 862 entitled, "Quaternary Stratigraphic Nomenclature in Unglaciated Central Alaska" by Troy L. Péwé, 1975. Detailed information concerning source of dates, laboratories, and possible errors are listed in Table 3 of the professional paper.)

bogs in the area. This is suggestive of a lower tree line. Excellently preserved peat layers, forest beds, and beaver dams have been observed in this exposure.

Animal remains

Vertebrate and invertebrate remains are common in the retransported silt of Wisconsin age. Freshwater snails occur in local pond deposits and screening of sediments reveals a host of insect remains, small vertebrate bones, and teeth (mice, etc.). Bones of the larger vertebrates occur but are not as numerous as in the other exposures in the Fairbanks area. Recent collecting from the sediments of Wisconsin age here have yielded a skull of *Bison superbison crassicornis*, bones of *Citellus undulatus*, horse, and mammoth. The presence of *Citellus* (ground squirrel) suggests a lower tree line in Wisconsin time because this animal is a near or above tree line creature. No *Citellus* live in the Fairbanks area today.

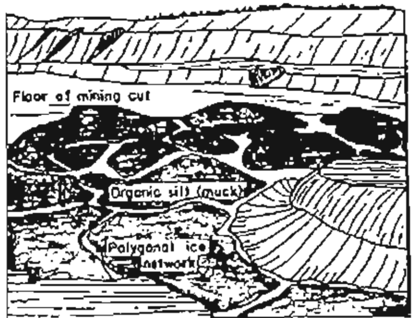
Currently the frozen silt in the Ready Bullion Creek exposure is the site for the collection of uncontaminated frozen Pleistocene silt for analyses of bacteria. Attempt is being made by scientists at the Arctic Aeromedical Laboratory of the U. S. Air Force to reactivate bacteria, frozen for at least 10,000 years.

Ground ice

At least four types of ground ice have been noticed at this exposure. The ice filling the pores in the silt, pore ice, is common in organic silt and inorganic silt. The second type of ice is segregated ice or Taber ice. Ice segregations a few millimeters to a centimeter thick are common in the organic-rich silt but less common in the inorganic silt of Illinoian age. Such segregated ice is commonly associated with small concentrations of vegetation. The amount of ice in the ground ranges from 25 to more than 100 percent by dry weight.

Another type of ice that has been known to occur here is a clear ice mass considered to be buried aufeis.

One of the most striking types of ice present in the ground is the large foliated ice masses termed ice wedges. Such masses may be 1 to



(Sketched from a photograph)

Figure 1-11. Foliated ground ice exposed in a polygonal network in retransported organic-rich silt of Wisconsin age near Fairbanks, Alaska.

10 feet (0.3 to 3 m) wide and 5 or 10 feet (1.5 or 3 m) high. Some have the typical wedge shape but most are very irregular masses that may be inclined or horizontal. In plan view it can be seen that they are part of a polygonal network of massive ground ice (fig. 1-11).

The flat top of the ice wedges is particularly conspicuous and important in the interpretation of the geologic history and climatic changes. An ice wedge which is actively growing has a thermal contraction crack, or vestige of a crack, extending from the ice wedge to the surface. Also, an ice vein generally extends part of the distance from the wedge to the surface. The ice wedges in Fairbanks no longer crack in winter and are considered inactive. Between 8,000 and 4,000 years ago the

climate warmed and the ice wedges in the Fairbanks area melted down; flat tops are now present on the ice wedges. The ground over the ice wedges slumped and was downwarped. Cooling of the climate in the last few thousands of years has caused the slumped and downwarped sediments to become refrozen.

The exposures of frozen silt and ice wedges are ephemeral and constantly changing. Features described above at Ready Bullion Creek (fig. 1-10), and at other mining or natural exposures of frozen silt in the Fairbanks area, may, or may not be observable today.

Stop 1-6: University of Alaska Agricultural Experiment Station. Sec. 1, T. 1 S., R. 2 W.

Thermokarst mounds

The thawing of permafrost with massive ground ice creates thermokarst topography, an uneven surface which contains mounds, sinkholes, tunnels, caverns, and short ravines. The thawing may result from artificial or natural removal of the vegetation or from warming of the climate. Thermokarst mounds in the Fairbanks area are found only in the areas of alluvial fans and colluvial slopes underlain by the retransported organic-rich silt containing large foliated ground ice masses (fig. 1-4).

Thermokarst mounds in the Fairbanks area are polygonal or circular hummocks 10 to 50 feet (3 to 15 m) in diameter, 1 to 8 feet (0.3 to 2.5 m) in height, and are composed of loess. They are commonest in cultivated fields but a few are found in abandoned fields now reforested. In some fields the mounds are separated by trenches 1 to 5 feet (0.3 to 1.5 m) wide. Trenches are formed by the melting of a network of ice masses in the ground, leaving mounds in the intervening areas. The ground surface sinks as ice near the surface melts.

The field at Stop 1-6 is on the north-facing slope of the Agricultural Experiment Station and has the best developed mounds and the most detailed record in the Fairbanks area.

The surface of the field was smooth before clearing in 1908. By 1922 pronounced individual and connected depressions had formed and by 1926 some trenches between mounds were as much as 5 feet (1.5 m) deep. Cultivation stopped a year or two later because the irregular topography formed by the pits and mounds was dangerous to the operation of farm machinery. The field then was seeded to pasture. By 1938 the mounds were 3 to 8 feet (1 to 2.5 m) high and about 20 to 50 feet (6 to 15 m) in diameter. Rockie studied the field in November 1938, and in order to determine whether the ice was actively melting, a tractor bulldozer was used to remove the upper part of every hummock and fill each pit until the land surface assumed approximately a uniform slope. The surface remained smooth for nearly a year, but in July 1939 irregularities began to form. In succeeding years polygonal mounds formed as the ground surface subsided over melting ice.

Mounds in the test area smoothed in 1938 were as large and as high as those in the part of the field that had not been smoothed when the writer first studied the field in 1947. Maximum mound height was 8 feet (2.5 m). Comparisons of aerial photographs with those taken 10 years earlier reveal that in 1948 mounds were about the same size and shape and in the same position as in 1938. Probing with a soil auger on July 14, 1948, revealed no ice or frozen ground at a depth of 9 feet (3 m) below the surface of a trench.

(Pictures of the stages in the development of this thermokarst mound field can be seen in figures 75, 76, 78, and 79 of Péwé, 1954).

Two stages of plant succession relating to the original abandonment of the field in 1926 and to the leveling in 1938 can be seen. In the younger section, paper birch, balsam poplar, willow *(Salix bebbiana)*, and alder *(Alnus crispa)* occur in an open stand with an understory of weedy species such as fireweed *(Epilobium angustifolium)*, horsetail *(Equisetum arvense)*, *Erigeron* spp., and several grasses.

In the older section the birches are more dense, there are a few large balsam poplars, and there is almost no understory vegetation except for scattered horsetail *(Equisetum arvense)*, a few mosses, and occasional clumps of the lichen *(Peltigera aphthosa)*. In both stages of the succession there are occasional white spruce, some of which were established at the same time as the birch and poplars, and some that have become established more recently.

Stop 1-7: Happy. Sec. 27, T. 1 N., R. 2 W.

Permafrost

Happy siding on the Alaska Railroad is in the middle of a typical sediment-filled creek valley in the Fairbanks area. Borings show 187 feet of retransported organic silt over 40 feet (12 m) of coarse creek gravel over bedrock. A cross section (fig. 1-12) illustrates a typical saucer-shaped permafrost lens up to 90 feet (27 m) thick in the valley bottom. Permafrost pinches out upslope to the east and west. Large masses of foliated ice occur in the silt (muck) and cause severe difficulties with railroad and highway construction and farming. The Alaska Railroad line is severely distorted in this area. Thawing of permafrost in the summer causes subsidence and seasonal frost action in the winter causes heaving. It is possible to see undulations in the railroad track despite efforts to block up the track to maintain a level road.

Stop 1-8: Pingos. Sec. 23, T. 1 N., R. 2 W.

Five small open-system pingos occur in Goldstream Valley near the junction with O'Connor Creek (fig. 1-4). They occur on the lower end of a low-angle alluvial fan from O'Connor Creek. Alpha pingo, the largest and southernmost of the group, is an elliptical mound about 350 feet (100 m) long, 200 feet (60 m) wide, and 34 feet (10.4 m) high. The pingo has a breached "crater" 20 feet (6 m) deep in the center. Concentric cracks 1 to 3 inches (3 to 8 cm) wide and 10 feet (3 m) long occur on the flanks, and radial cracks 2 to 150 feet (0.6 to 45 m) long radiate outward from the crest of the pingo (fig. 1-13). The pingo is composed of silt. Permafrost occurs at a depth of 3 to 10 feet (1 to 3 m).

The vegetation on the pingos is a mixture of balsam poplar, aspen, white spruce, and paper birch. The area surrounding the pingos was originally black spruce, muskeg, and bog that had been cleared for farming. Some of the original vegetation can be seen to the west between the road and the line of tall spruce that marks the recent alluvial deposits of Goldstream. Many of the lower parts of the abandoned field are reverting to bog with the establishment of dwarf birch *(Betula glandulosa)*, leather leaf *(Chamadaphne calyculata)*, Labrador tea *(Ledum groenlandicum)*, and sphagnum mosses *(Sphagnum* spp.).

Beta pingo lies about 400 feet (120 m) to the northeast of Alpha pingo and is about one-half the size. It has no depression in the center. On the south side, which is steep and

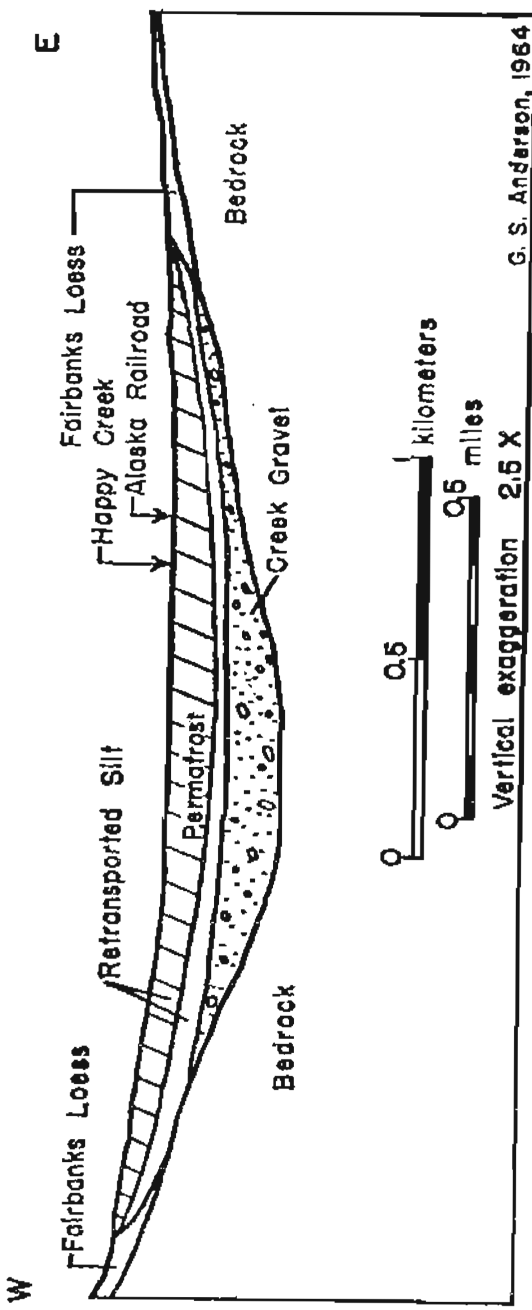


Figure 1-12. Cross section of permafrost and unconsolidated sediments of Quaternary age in Happy Creek valley, 7 miles (11 km) northwest of Fairbanks, Alaska. (Cross section extends westward from the 750 ft contour line in sec. 26, R. 2 W., T. 1 N. to the "nose" of the 1000 ft contour line on the south side of Happy Creek. Source map: Péwé, 1958b.)

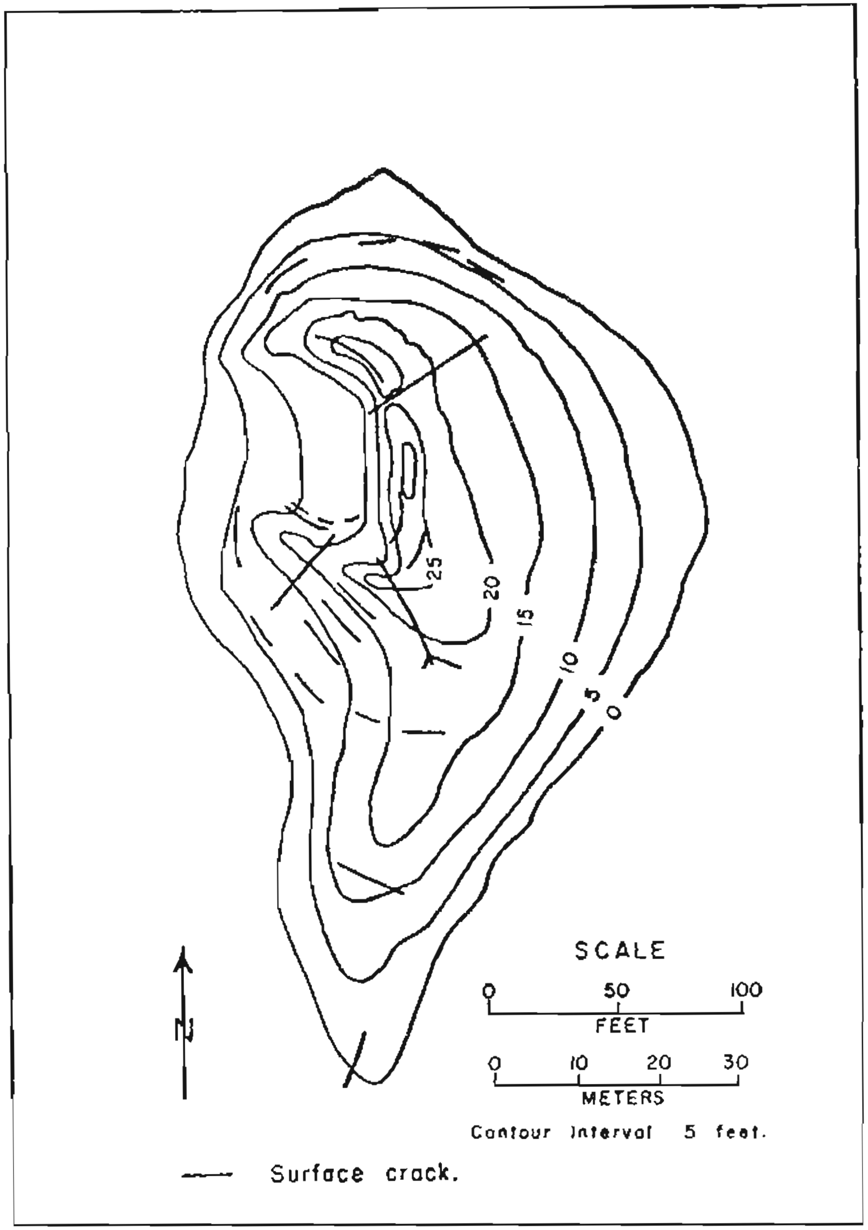


Figure 1-13. Sketch map of Alpha Pingo, Goldstream Valley, Fairbanks, Alaska, 1964. Compiled by geology students, University of Alaska, by pace and compass method. Datum at base of pingo approximately 600 feet (180 m) above sea level.

actively slumping, growing trees are deformed and several have split trunks. It is not known if the pingo surface is rising or if the south side is slumping because of thawing at the base. The rate of movement is not known, but in September 1964, studies were started to learn more about type and rate of surface deformation.

Three smaller pingos lie north and west of Beta pingo. All are on an alluvial fan of retransported organic-rich silt of Wisconsin and Recent age. In a nearby drill hole the silt is 90 feet (27 m) thick and overlies 50 feet (15 m) of creek gravel. The permafrost is 140 feet (43 m) thick. Although studies have only begun, it is apparent that ground water under the fan feeds the pingos and that they are relatively young, probably no older than 6,000 to 8,000 years. A somewhat typical Subarctic Brown Forest soil 7 to 13 inches (18 to 33 cm) thick is present on the pingos. The soil was not formed on undisturbed loess but on retransported loess that probably had a bog soil developed prior to the arching of the pingos.

Stop 1-9: Golf course. Sec. 30, T. 1 N., R. 1 W.

Thermokarst pits and mounds

Thermokarst pits are steep walled pits 5 to 20 feet (1.5 to 6 m) deep and 3 to 30 feet (1 to 9 m) across. Commonly, they are larger at depth than at the surface. They occur mainly in cultivated fields on the alluvial fans and colluvial slopes near the contact where the permafrost free slopes. The pits are started by melting of ground ice and commonly are enlarged and modified by surface water that is diverted into small cracks and subterranean passageways and then flows 6 to 20 feet (2 to 6 m) beneath the surface. The passageways do not collapse readily because of the structural properties of the loess, and, consequently, such caverns lie undetected for many years before the roof collapses or surface water wears a small opening through the roof. Many thermokarst pits occur in cultivated fields in the Fairbanks area, some in frozen ground and some in thawed ground.

This Stop involves two cleared fields: one lies north, and the other south, of the Farmers Loop Road and are now occupied by a golf course. The area to the north lies on or near the border of the perennially frozen retransported organic-rich silt and the unfrozen silt on the hillside (fig. 1-4). The field north of the road was cleared in 1910 and has more thermokarst pits than any other field in the Fairbanks area. Two pits are reported to have been 3 feet (1 m) in diameter when they originated in 1936. Records of earlier thermokarst pits are not known. The diameters of these pits when examined in 1948 were more than 20 feet (6 m). Two other pits 5 feet (1.5 m) in diameter were exposed in 1946. Three new pits 10 to 15 feet (3 to 4.5 m) in diameter were opened in 1949. Since this time several pits have opened and others have been filled. The field south of the road was cleared in 1939 and no pits are known to have formed.

In addition to thermokarst pits, thermokarst mounds have formed as a polygonal network of ground ice masses melted. Broad mounds 50 to 100 feet (15 to 30 m) in diameter and 2 to 4 feet (0.6 to 1.2 m) high have formed and are slowly growing higher as the ice continues to melt. These are best displayed in the field south of the road. The golf course was established in 1946 and is perhaps not only the farthest north golf course, but the only one in which thermokarst mounds and pits as natural hazards cause serious difficulties.

The radio transmitter building and tower were built south of the road in 1939. The building is of reinforced concrete and, although the underlying ground has settled considerably due to melting of ground ice, the building has not been seriously deformed. The excellent construction of the building has prevented it from being wracked; however, it does settle as a unit. Although it can now be seen to have settled, it was difficult to demonstrate this in the early years. One early evidence of movement was that the well in the building continued to "rise." The casing of the well is frozen into 200 feet (60 m) of permafrost, and yet, the pump, which is set on top of the casing, continues to "rise" from the floor. After it has risen about 1 or 2 feet (0.3 m or 0.6 m) off the floor it is necessary to cut off the casing and reset the pump on the floor. Contrary to popular opinion it is not the rising of the casing out of the ground that causes the pump to leave the floor, but it is the sinking of the ground and building around the stable casing that causes the displacement.

The transmitter tower has remained stable. The massive concrete piers are evidently frozen firmly into the permafrost and resist seasonal frost action.

Stop 1-10: Thermokarst topography. Sec. 28, T. 1 N., R. 1 W.

On the south side of the Farmers Loop Road, in an area of perennially frozen retransported organic-rich silt (fig. 1-4), is a very well-developed thermokarst terrain. This area has not been studied in detail, but from reconnaissance observations it appears that the thawing of ground ice is creating cave-in lakes, thermokarst mounds, and thermokarst pits. The cave-in lakes are lakes that have been initiated by thawing of ground ice and have enlarged by the retreat of the banks. Whether or not some of the mounds may be pingos is not known. The thermokarst topography is actively forming inasmuch as many of the trees indicate recent deformation of the ground. The cover of trees includes both black spruce and white spruce as well as aspen and balsam poplar. The forest cover is less than 50 years old in this area.

Stop 1-11: Thermokarst pits. Sec. 20, T. 1 N., R. 1 W.

On this east-facing wooded slope at the contact of the perennially frozen silt and unfrozen loess, several large thermokarst pits formed in 1961-64. These pits lie in the bottom of a shallow drainage line and are 10 to 20 feet (3 to 6 m) wide and 5 to 15 feet (1.5 to 4.5 m) deep. Small depressions were present prior to 1961, and 1 pit formed in 1961, 2 in 1962, 1 in 1963, and 1 in 1964. Surface water is channeled into these steep-walled depressions and disappears in openings at the base of the pits. Water emerges from the ground downslope some 100 to 200 feet (30 to 60 m) from the pits.

No clearing of the forest has taken place in this area; therefore, perhaps the natural thawing near the permafrost boundary has initiated melting of massive ground-ice bodies. This melting, plus erosion by diverted surface water, forms cavities, and later the roof collapses.

This type of "gully formation" may have been instrumental in locating, initiating, and forming the numerous parallel gullies in the loess so common in the Fairbanks area—gullies that were formed prior to Wisconsin time, perhaps during Sangamon time (see Stop 1-13).

Stop 1-12: Pingo. Sec. 21, T. 1 N., R. 1 W.

An equidimensional open-system pingo approximately 100 feet (30 m) across and 20 feet (6 m) high on the down-slope side lies in a drainage

line next to Farmers Loop Road. The pingo is in silt and has a slight depression on the crest. Some 4 or 6 years ago a large trench 15 feet (4.5 m) deep was dug in the center of the pingo for the foundation of a home. It is not known why construction was not begun, or if the exploration reached the permafrost table. The cut is entirely in retransported silt. An auger hole 8 feet (2.5 m) deep in the bottom of the cut reveals organic-rich silt, but no permafrost was encountered.

Stop 1-13: Alaska Field Station. Sec. 36, T. 1 N., R. 1 W.

The Alaska Field Station of U. S. Army Cold Regions Research and Engineering Laboratory (USA CRREL) on Farmers Loop Road was established by the Corps of Engineers in 1946 to obtain information related to problems in design and construction in permafrost regions. The Alaska Field Station is built on retransported perennially frozen ice-rich silt overlying creek gravel. Permafrost is 176 feet (54 m) thick and does not extend to bedrock. Ground water flowing from the permafrost-free slope on Birch Hill lies between permafrost and bedrock and is under artesian pressure. A well drilled at the station in 1946 produced a flowing artesian well which is now capped.

Some of the studies under way include testing of different types of piling and pile installation methods for structural support, testing of anti-frost heaving devices, and insulation under pavements.

Stop 1-14: Gullies in loess. Sec. 19, T. 1 N., R. 1 E.

In this vicinity the Steese Highway cuts through a series of parallel ridges of loess. The ridges are broad rounded features 30 to 50 feet (9 to 15 m) high and 300 to 600 feet (90 to 180 m) long. They are best seen from the air and are well shown in figure 1, Plate 3, of Péwé, 1955.

Inasmuch as this spectacular gullying is not a local feature, but is present in areas of thick loess in much of the Tanana Valley and elsewhere, the gully cutting is believed to be of significance in the Quaternary history. Sections reveal that since gullying and production of steep-walled trenches, the ridges and valley bottoms in loess have become rounded, and a 4- to 6-foot (1.2 to 1.8 m) thick layer of additional loess was draped over the undulating topography. Current interpretation is that this gullying occurred with thawing of permafrost in Sangamon time.

Stop 1-15: Engineer Creek. Sec. 6, T. 1 N., R. 1 E.

Stratigraphy

Tailing piles (fig. 1-4) of the younger gravel lie in the valley of Engineer Creek. Near the junction with the gravel of the same age in Goldstream Valley, Engineer Creek cuts across a bench deposit of older gravel on the south side of Goldstream Valley. This gravel lies 10 to 30 feet (3 to 9 m) higher than the younger gravel and is composed almost entirely of white quartzite cobbles and boulders. The source of the quartzite is not known. Overlying the bench gravel is 100 feet (30 m) of Illinoian loess which in turn is overlain by retransported organic-rich silt (muck) of Wisconsin age.

Slump over ice wedges

When exposures were created by placer mining, all the numerous ice wedges in the retransported organic-rich silt of Wisconsin age were overlain by downwarped and downfaulted sediments (fig. 1-14). Downwarping of the sediments was due to the down melting of the ice wedge tops in post-Wisconsin time.

The ice wedge tops in the Fairbanks area are now flat. Currently it is thought, based on radiocarbon dating, that this downwarping occurred between 8,000 and 4,000 years ago. Occasionally erosion by small intermittent streams exposes ice wedges and slump structures today.

tary purposes, a quick, mechanical technique of tunnel excavation in perennially frozen ground. The project also seeks to test refrigeration and other systems for slope protection and retardation of deformation, and to investigate various spoil-removal methods.

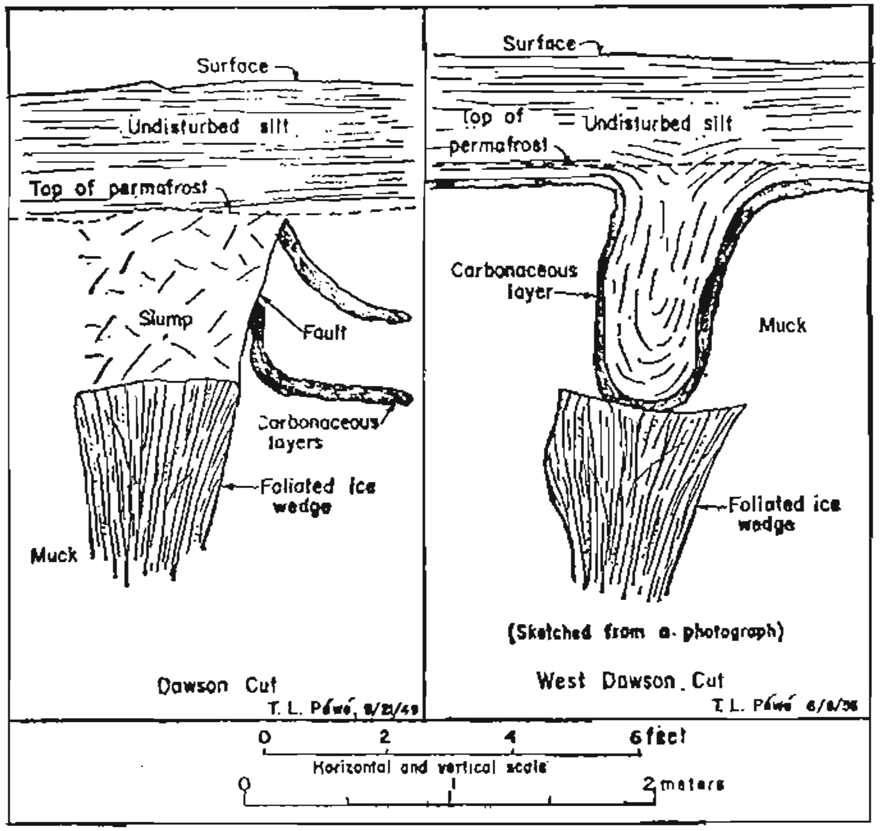


Figure 1-14. Slump structures over foliated ice wedges near Fairbanks, Alaska.

Stop 1-16: Permafrost tunnel. Sec. 31, T. 2 N., R. 1 E.

Among studies undertaken in Alaska by the U. S. Army Cold Regions Research and Engineering Laboratory is one to find, for military purposes, a quick, mechanical technique of tunnel excavation in perennially frozen ground.

In a relatively short period of time (16 days), a tunnel 155 feet (47 m) long was dug horizontally into the frozen silt, using an Alkirk Cycle Miner. A drag bit was developed to withstand the high abrasion of the silt and the impact of the gravel.

Excavated material is hauled out of the adit by a Joy 10 SC AC shuttle car, loaded by a standard Joy 8 BU loader.

The gravel tailing piles near the entrance of the tunnel illustrate many stages of revegetation. Willows, alders, balsam poplars, and birch seed in directly on the coarse tailings without any previous pioneer stage. In some localities crustose and fruticose lichens are growing in between the rocks. In the wet depressions aquatic and semi-aquatic plants have become colonized.

Stop 1-17: Dome Creek. Sec. 5, T. 2 N., R. 1 E.

Placer gold mining in the early 1950's produced many excellent exposures of retransported organic-rich silt of Wisconsin age. From these deposits came the pair of enormous Mammoth tusks 13 feet 7 inches (3.2 m) (outside curve) long that weigh approximately 350 pounds (160 kg) apiece and are exhibited in the University of Alaska museum. The tusks were found with a fairly complete skull, other bones, and much well preserved Mammoth hair. Some of the hair is in the Department of Geology exhibit cases at the University of Alaska, and has been dated at 32,700 ± 980 years (St 1632).

It was from the silt in this exposure that a partial carcass of *Bison superbison crassicornis* was recovered in 1951. The carcass consisted of a head, complete with hide, horns, and one ear, four legs with hooves, and much torso hide about 3 mm thick. The carcass showed evidence of some transportation, yet movement must have been only a short distance, otherwise the carcass would have been completely destroyed. A date of more than 28,000 (L-127) years was obtained on a piece of the carcass in 1951. In 1965 a date of 31,400 (+2040 or −1815) years (st. 1721) was obtained by the radiocarbon laboratory of the Geological Survey of Sweden. A photograph of the carcass is on display in the Department of Geology at the University of Alaska. Pieces of the fur and hide of a female superbison recovered from Fairbanks creek has been dated as 11,950 ± 135 years (St 1633) and is on display at the Department of Geology, University of Alaska.

Small streams crossing the slumped silt cliffs at Dome Creek have exposed large massive ice masses that may be observed today.

Stop 1-18: Chatanika River Valley. Sec. 23, T. 3 N., R. 1 W.

Although the Chatanika River is a small stream, it was large enough to carry away silt deposited into it by tributaries in Wisconsin time. Rounded flood plain gravel lies close to the surface and extends to a depth of approximately 180 feet (55 m). Placer gold was mined by subsurface methods here and many "spoil" piles of gravel from the underground workings are present on the surface.

Stop 1-19: Chatanika River bluff. Sec. 15, T. 3 N., R. 1 W.

Stratigraphy

A 50-foot (15 m) high cutbank of the Chatanika River 4 miles (6.5 km) downstream from the Elliott Highway bridge exposes retransported organic silt and ice wedges of Wisconsin age overlain unconformably by a thick deposit of retransported silt and buried forest beds of post-Wisconsin age (fig. 1-15). The perennially frozen silt of Wisconsin age is massive to poorly bedded. The faint bedding is emphasized by a concentration of ice 1/16 to ¾ inch (1 to 10 mm) thick at the bedding planes. A white volcanic ash layer 1/16 to ⅓ inch (1 to 10 mm) thick lies about 10 feet (3 m) below the top of the Wisconsin sediments. The ash layer is a good marker to

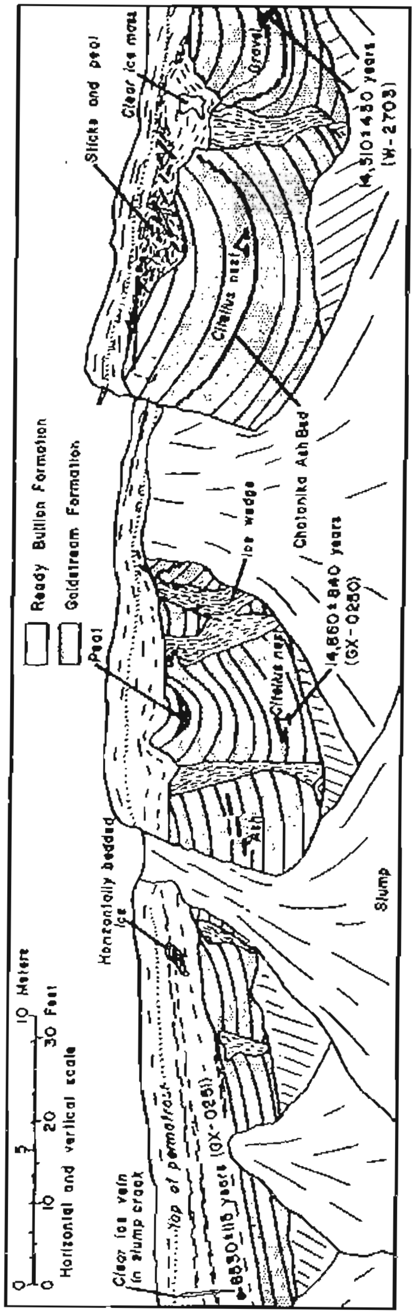


Figure 1-15. Bank of the Chatanika River 2 km downstream from Elliot Highway bridge, 40 km north of Fairbanks, Alaska. Goldstream Formation (Wisconsin) with ice wedges and Chatanika Ash Bed, is overlain by the Ready Bullion Formation (Holocene). (From figure 10 in U.S. Geological Survey Professional Paper 862 entitled, "Quaternary Stratigraphic Nomenclature in Unglaciated Central Alaska", by Troy L. Péwé, 1975. Details of radiocarbon dates and laboratories can be found in Table 4 of the professional paper.)

demonstrate how the beds have been upturned to an almost vertical position next to the massive ice wedges. The ash also outlines the minute faults that are present in the frozen silt. The slope of the upper surface of the Wisconsin silt is slightly steeper than the modern slope. A radiocarbon date of 14,860 ± 840 years (GX-0250) was obtained on a ground squirrel nest recovered about 15 feet (4.5 m) below the top of the Wisconsin sediments (fig. 1-15).

Unconformably overlying the sediments of Wisconsin age are post-Wisconsin, bedded, perennially frozen sediments rich in peat layers. These sediments elsewhere in the Fairbanks area are normally about 3 to 6 feet (1 to 2 m) thick but here they thicken to 25 feet (8 m) downslope. This is the thickest recorded deposit of post-Wisconsin silt in the Fairbanks area. A radiocarbon date of 8,530 ± 115 years (GX-0251) was obtained on peat 15 feet (4.5 m) below the surface (fig. 1-15).

Ground ice

Large foliated ice wedges up to 10 feet (3 m) wide and 18 feet (5.5 m) long strongly deform the sediments of Wisconsin age, including the volcanic ash layer. The wedges have flat tops and slump structures over the ice wedges are poorly to fairly well developed.

In the sediments of post-Wisconsin age are ice masses up to 1 foot (0.3 m) in diameter and 4 to 5 feet (1.2 to 1.5 m) long. The ice is clear and some is horizontally bedded. Such ice has formed in fault or slump cracks as the bank retreats by landslide slumping.

Flora

The sediments of post-Wisconsin age contain many peat and forest beds. Superficial examination reveals that the flora is similar to that forming on the surface today — a vegetation characterized by willow, alder, black spruce, dwarf birch, grass, and sphagnum moss, all indicative of a high permafrost table. Tundra tussocks present on the surface are also found in the frozen sediments.

Fauna

Nests and coprolites of the ground squirrel *(Citellus undulatus)* occur in the Wisconsin sediments but not in the post-Wisconsin sediments. In other exposures, dried carcasses of the squirrels have been found. Approximately 250 nests (with bones) have been recovered from the Fairbanks area and are now under preliminary study.

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MIDDLE TANANA RIVER VALLEY

by

Troy L. Péwé

RESUME OF THE QUATERNARY GEOLOGY OF THE MIDDLE TANANA RIVER VALLEY

The middle Tanana River valley or lowland (fig. 4-16) is a large tectonic trough bounded on the south by the towering Alaska Range and on the north by the rounded hills of the Yukon-Tanana upland. It is a structural basin, the floor of which is below sea level in much of the trough. Quaternary deposits 300 to 700 feet (91 to 230 m) thick are in

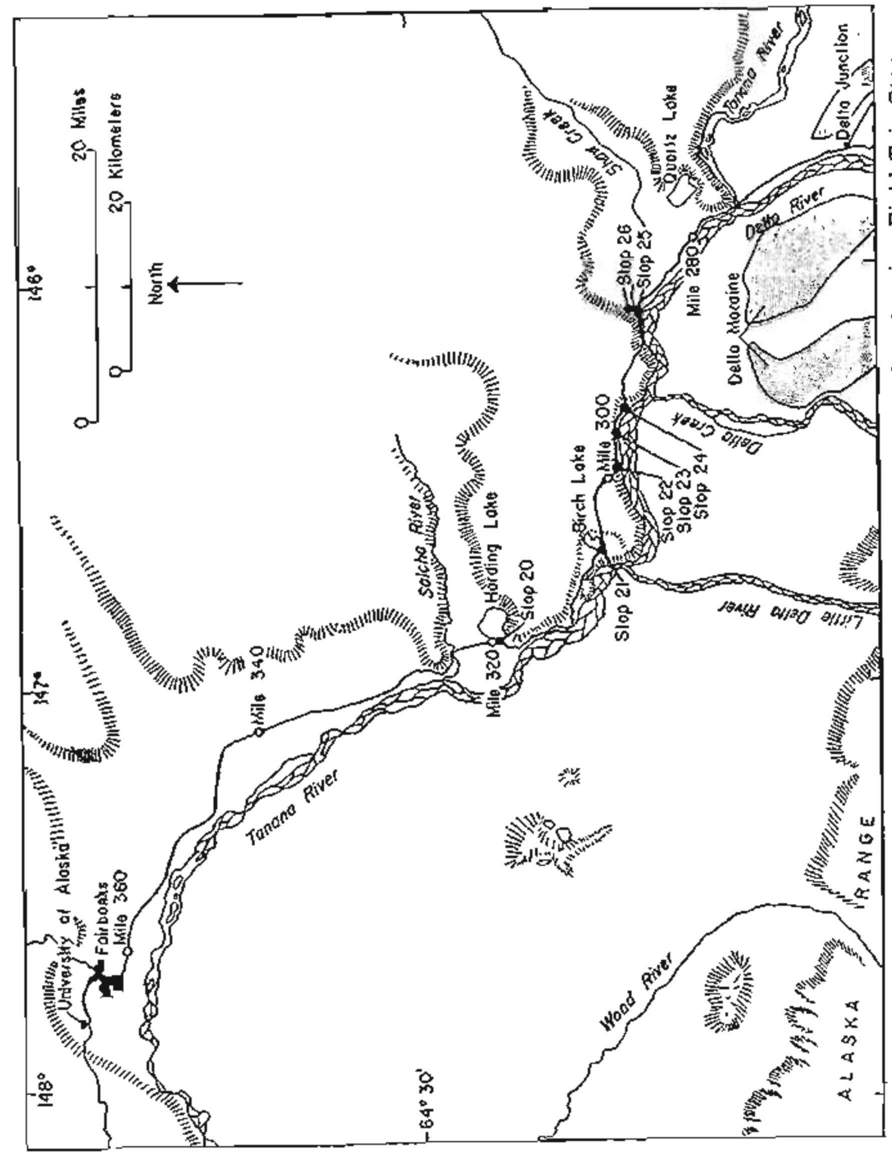

Figure 4-16. Index map of the middle Tanana River Valley, Alaska, showing Field Trip Stops.

large part an accumulation of fluvial and glaciofluvial sediments derived from the rising Alaska Range. This deposition pushed the Tanana River to the north against the Yukon-Tanana upland. The sediments bury a fairly rugged topography, the hilltops of which now protrude through the Quaternary sediments and form small knobs above the plain. Well-drained fans 30 miles (50 km) long flank the Alaska Range.

This report is concerned with Quaternary geology along the north side of the Tanana River valley from Fairbanks to Big Delta, a distance of approximately 90 miles (150 km) (fig. 4-16). The Tanana River is 230 miles (370 km) long and has an average discharge of 35,000 cubic feet (100 m³) per second on August at Big Delta. Up stream from Birch Lake it receives glacial sediment from several glacial streams and is a large braided stream. It has a gradient of 6 to 7 feet per mile (1 to 1.2 m per km) between Birch Lake and the Delta River. The river becomes less braided and more meandering as it flows to the Fairbanks area. Downstream from Fairbanks it is distinctly a broad meandering stream with few islands, compared to the area upstream from Fairbanks.

The north wall of the Tanana Valley is formed by low hills of the Yukon-Tanana upland. The hills are heavily blanketed with eolian deposits of sand and silt that have been blown, and are still being blown, from the flood plain of the Tanana River. The Quaternary geology of this segment of the valley is known only in a reconnaissance fashion. Detailed studies are underway which should tie together the two better known areas—the unglaciated Fairbanks area and the glaciated Delta River area.

The middle to late Quaternary history of the valley is a record of alternating deposition and erosion of eolian and fluvial deposits and of climatic fluctuations. Both geologic processes and climatic changes are related to glacial advances and retreats from the Alaska Range on the south side of the valley.

The oldest Quaternary event known at present is a cold period recorded by a solifluction deposit. The deposit, which contains fossil ice wedges, is thought to be pre-Illinoian in age and lies on bedrock under eolian sediments at numerous exposures from Fairbanks (fig. 1-5) to the Delta River.

The next event was the advance of glaciers in Illinoian time (Delta Glaciation) accompanied by aggradation of the Tanana Valley. As the Tanana River aggraded, the south-flowing streams of the Yukon-Tanana upland were dammed in many places, forming lakes that are held in on one side by fluvial deposits and flanked by the bedrock hills on the other sides. Lakes held against the bedrock hills on the north side of the Tanana River valley that are thought to have formed at this time are: Harding, Birch, Quartz, and others.

The unvegetated outwash plains and valley trains were exposed to wind action, and sand and silt picked up by the wind were dropped on the north valley wall. Sand is most common in the Delta River section of the valley, thinning and disappearing to the northwest, where loess predominates. Ventifacts were formed at this time on the bluffs facing the Tanana River near its junction with the Delta River.

As the glaciers withdrew, downcutting probably predominated, and some of the fill of the Tanana Valley was removed, leaving terraces on the valley sides. With lowering of local base level much of the eolian

sediment on the valley walls was gullied and carried away.

In Wisconsin time (Donnelly Glaciation) the river again aggraded and windblown silt was deposited on the valley walls. Eolian sand deposition at this time was limited to lowlands adjacent to the Delta River. Tree line was 1500 to 1800 feet (450 to 550 m) lower, and trees were limited mainly to creek and valley bottoms. The lowering of tree line has been substantiated by the great reduction of tree pollen and a rise in pollen of tundra species, and by the presence of above-and near-tree-line animals that lived in this area at that time.

In post-Wisconsin time the amount of loess deposition declined, except perhaps in the lower Delta River area.

Road Log and Locality Descriptions

363.2[1] Start in Fairbanks D-2 Quadrangle[2].

Cushman and Airport Way, Fairbanks. Mileposts are on the left side going toward Anchorage. The Highway is on the flood plain of the Tanana River. From Fairbanks to Mile 330 the flood plain is a flat plain with meandering streams and a complex network of shallow swales. The surface layer of silt is 1 to 20 feet (0.3 to 6 m) thick and the total thickness of alluvium is 300 to 700 feet (91 to 230 m). The shallow swales are filled with about 30 feet (9 m) of clayey silt. The depth of permafrost is 2 to 4 feet (0.6 to 1.2 m) on the older parts of the flood plain and more than 4 feet (1.2 m) on the inside of meander curves near the river. Depth to permafrost may be 25 to 40 feet (8 to 12 m) in some cleared areas. Permafrost is discontinuous and there are many unfrozen lenses, layers, and vertical zones. The ground ice content of permafrost is low and no large ice masses are known.

The drainage is excellent and permeability is high, except locally in silt or where ground is perennially frozen. Depth to the water table is about 10 to 15 feet (3 to 4.5 m) where permafrost is absent. The surface of this area is good for agriculture if the soil is fertilized. The City of Fairbanks and the military reservations are on the flood plain.

Because of active aggradation of the Tanana River and the wide braided nature of the stream, large areas along the river and on the islands in the rivers are in various stages of recolonization by vegetation. Extensive areas of willows (*Salix alaxensis*, *S. bebbiana*, and *Salix* spp.) and alders (*Alnus crispa*) exist between Fairbanks and Big Delta. Large stands of balsam poplar range in age from 30 to 100 years. Later stages in plant succession, especially white spruce stands of all stages, occur along the river flood plain adjacent to the highway. Older terraces of black spruce, larch, and bogs are conspicuous at many points along the river but are absent in localities where the river is adjacent to the hills of the Yukon-Tanana upland.

363.2 to 347.2. Most of the types of vegetation along the first 25 miles (40 km) of the highway are a result of recent disturbance and fire. Nearly all of the stands are successional, and because of the poor drainage, stands of willows, black spruce, larch, and paper birch are the most common.

355. Enter Fairbanks D-1 Quadrangle.

356.1. Right; telephone poles heaved by seasonal freezing of the

[1]Numbers are miles from Valdez on the Richardson Highway.

[2]U. S. Geological Survey Topographic Quadrangle map, scale 1:63,360.

ground. Poles are braced at the base, but not successfully.

352. Right; water well here penetrated 265 feet (81 m) of permafrost, greatest thickness recorded on Tanana River flood plain.

350.0. Village of North Pole. Santa Claus Lodge.

349.7. Enter Fairbanks C-1 Quadrangle.

347.2. Highway on dike for next 2.8 miles (4.5 km). The dike, 3 miles (5 km) long and 12 feet (3 m) high, is composed of Tanana River gravel and was built in 1939-40 to divert flood waters of the Tanana River from entering the Chena Slough which flowed through the City of Fairbanks. This diversion has caused aggradation. By 1960, the river had aggraded to within 3 to 4 feet (about a meter) of the top of the dike on the east (river) side. The road along the top of the dike is one of the best in the Tanana River valley and no large transverse frost cracks occur.

On the river side of the dike are extensive stands of willow and alder that have resulted from the aggradation. These stands are 15 to 20 years old and are flooded annually.

344.9. End of dike. Moose Creek Bluff is a bedrock hill of schist and gneiss of Precambrian(?) age poking through the alluvial fill of the Tanana Valley. Cliff-head dunes of Recent age occur on top. Rock paintings in red pigment were discovered on the south side in 1940. The largest figure was 25 inches (53 cm) high. All are now destroyed except one slab which is at the University of Alaska museum. The age of the paintings is unknown.

337.5. South entrance of Eielson Air Force Base. South of this point a statistical study was made of the prominent transverse frost cracks in the highway over a distance of 5,280 feet (1610 m); the study showed that the cracks are spaced an average of 105 feet (32 m) apart. These are thermal contraction cracks in seasonally frozen ground (seasonal frost cracks). Cracking of the seasonally frozen ground in the interior of Alaska occurs only in certain restricted environments at present. It occurs in central Alaska, as far as known, only in areas that are vegetation free and/or that are kept snow free during the winter, such as roads and pathways near Fairbanks. In seasonally frozen ground the cracks which traverse the roads and paths narrow rapidly and disappear as they pass into adjacent areas covered with vegetation and unpacked snow. Inasmuch as the highway is unprotected by vegetation and snow cover, it is subject to stresses that cause cracking, stresses formed by low temperature and rapid cooling of the ground.

Under natural conditions, no thermal contraction cracking of the ground is known today in the seasonally frozen ground or the permafrost of interior Alaska. Ice wedges that exist today in perennially frozen ground in central Alaska formed in the past under a more rigorous climate. The average diameter of the polygons formed by thermal contraction cracking of permafrost in central Alaska is approximately 100 feet (30 m).

336.1. The skyline to the east in the Big Delta C-6 Quadrangle shows an ancient altiplanation terrace(?).

336. The road passes through an extensive stand of black spruce and larch that developed after a fire in this area. The trees are about 35 years old.

334.05. Here and at many other places along the Richardson Highway to Big Delta the old 4-inch-diameter (10 cm) pipeline which was used during World War II to supply fuel from Skagway to interior Alaska can be seen.

330. Bedrock hill of the Birch Creek Schist. On the steep, dry, unstable south-facing slopes at this location and elsewhere along the highway two species of sagebrush, *Artemisia frigida* and *A. krushiana* are growing. Aspen and white spruce are common on the stabilized slopes. Several stands of river-bottom white spruce are visible along this section of the highway.

329.2. Enter Big Delta C-6 Quadrangle.

328.5. Right; confluence of the clear water of the Little Salcha River and the silt-rich water of the Tanana River.

328.3. Rise onto lower terrace of the Tanana River. This terrace is underlain by rounded, moderately well sorted sandy gravel capped with as much as 8 feet (2.5 m) of frozen silt at the scarp. The silt thickens to the east, and may have buried the upper terrace of the Tanana River. The top of the terrace (including the silt overburden) averages 15 feet (4.5 m) above the modern flood plain of the Tanana River.

328. Road parallels the scarp of low terrace.

327.6. Little Salcha River. This stream is incised into the lower terrace of the Tanana River.

327.5 to 326.2. The road is built on ice-rich perennially-frozen silt as much as 12 feet thick.

326.2. Enter Big Delta B-6 Quadrangle.

325.8. Leave lower terrace, descend to present flood plain of Tanana River.

325.7. Rise onto lower terrace. Gravel of lower terrace exposed in borrow pit. Sandy cobble gravel overlain by 3 to 6 feet (1 to 2 m) of silt.

325.2. Salcha School. Lower terrace is about 10 feet (3 m) above modern flood plain.

325. Left; bluff of Birch Creek Schist. Tanana River on right. Terraces absent. Several successional stands of willow, alder, and balsam poplar are visible along the river.

324.6 to 323.2. Road is on modern flood plain of Tanana-Salcha River confluence. Terraces are absent or concealed.

323.3. Salcha River bridge. Large 10-inch (30 cm) diameter pipeline on the left side of bridge is new (1954-55) fuel pipeline from Haines to Fairbanks. It can carry 7 types of fuel at one time.

323.3 to 322.5. Modern flood plain of Salcha River. Road is occasionally inundated during high water.

322.5. Scarp of lower terrace of Salcha River.

322.4. Scarp of upper terrace of Salcha River. Relief is subdued locally by thick silt.

322.4 to 320.2. Road is on the higher of the two terraces recognized on the northeast side of the Tanana Valley in the Big Delta (B-6) Quadrangle. The Salcha River has apparently been graded to the Tanana River, and similar sequences of terraces are recognized along both rivers. Studies of lithology and sedimentary structure indicate that the material underlying the highway was deposited by the Salcha River. One to 1.5 miles (1.6 to 2 km) to the west, the gravel is of Tanana River origin.

The upper terrace is mantled with from a few inches to 6 feet (several cm to 2 m) of silt, probably loess. Several fossil ice wedges are found in the gravel of the upper terrace; none are recognized in the lower terrace.

The ages of the two terraces are thought to be Illinoian (upper) and Wisconsin (lower) in age.

320.1. Borrow pit on left exposes gravel of upper terrace, mantled by several feet of sand and silt (partially stripped). Finer sediment is firmly cemented by iron oxide and displays intense convolutions. Imbricate structures in surface zone are similar to those found in modern ice-pushed ridges of Harding Lake, suggesting that an arm of Harding Lake may have once occupied this area. Borrow pit on right exposes a fossil ice wedge extending several feet (more than a meter) into the gravel of the upper terrace.

319. Stop 4-20[1]: Harding Lake landing and picnic ground.

Aggradation by the Tanana River and its major tributaries have apparently dammed several of the valleys of the Yukon-Tanana upland during late Pleistocene time. This damming has resulted in the formation of several lakes: Harding, Birch, Quartz, and perhaps a dozen others. Harding Lake is unique among these lakes in that its maximum depth exceeds, by a factor of 3 or 4, that of any of the other lakes. Also, the bottom of Harding Lake shows considerable relief, whereas the bottom profiles of the other lakes are generally smooth. A typical east-west profile of Harding Lake is shown in figure 4-17.[2] For comparison, a typical profile of Birch Lake (Stop 4-21) is included.

Ice-pushed ridges (ice ramparts) are well developed along the shores of Harding Lake. Along the north shore, as many as seven distinct ridges, averaging about 3 feet (1 m) in height, have been counted. These ridges indicate higher stands of the lake — the sharpness of the ridges suggests that they are only a few hundred years old.

A detailed bathymetric survey of the lake reveals that submerged beach features also are present. Analysis of the depth of these features (ice-pushed ridges and wave-built terraces) indicates that the west margin of Harding Lake has been depressed as much as 6 feet (2 m) relative to the east shore. Age and significance of this warped strandline are not known. The Salcha River valley has been the epicenter of at least one major earthquake during the last 30 years.

Extensive stands of paper birch, often with an understory of white or black spruce surround the lake. These stands are about 100 years old and have originated during a period when fires seemed to have been especially extensive in the Tanana Valley. Many of the upland stands between here and Big Delta are about this age.

318. Excellent view of Mt. Hayes (13,832 feet [4,219 m] altitude) and peaks of central Alaska Range.

317.9. Road descends to lower terrace of Tanana River. Terrace is underlain by moderately well sorted sandy gravel overlain by 2 to 4 feet (about 1 m) of silt. Terrace is 10 feet (3 m) above river level.

317. Right; road parallels edge of 10 to 15 foot-high (3 to 5 m) scarp of lower terrace. Scarp of upper terrace can be seen 1000 feet (300 m) to the left.

316.7. Road descends to modern flood plain of Tanana River. Gravel exposed here is typical Tanana River sediment—from gneiss, quartz diorite, schist, and volcanic rocks. Schist is weathered and friable.

316.5. Scarp of the upper terrace 800 feet (250 m) on the left. Upper

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[1]This guidebook has 66 stops, numbered consecutively from 1 through 66, but with a prefix for each number to indicate the day of the trip when the stop should occur. Thus **Stop 4-20** would indicate that Stop 20 is included in the itinerary of DAY 4.

[2]This guidebook has 53 figures, numbered consecutively from 1 through 53, but with a prefix for each number to indicate the day of the trip when the figure is first used. Thus fig. 4-16 would indicate that figure 16 appears in the itinerary for DAY 4.

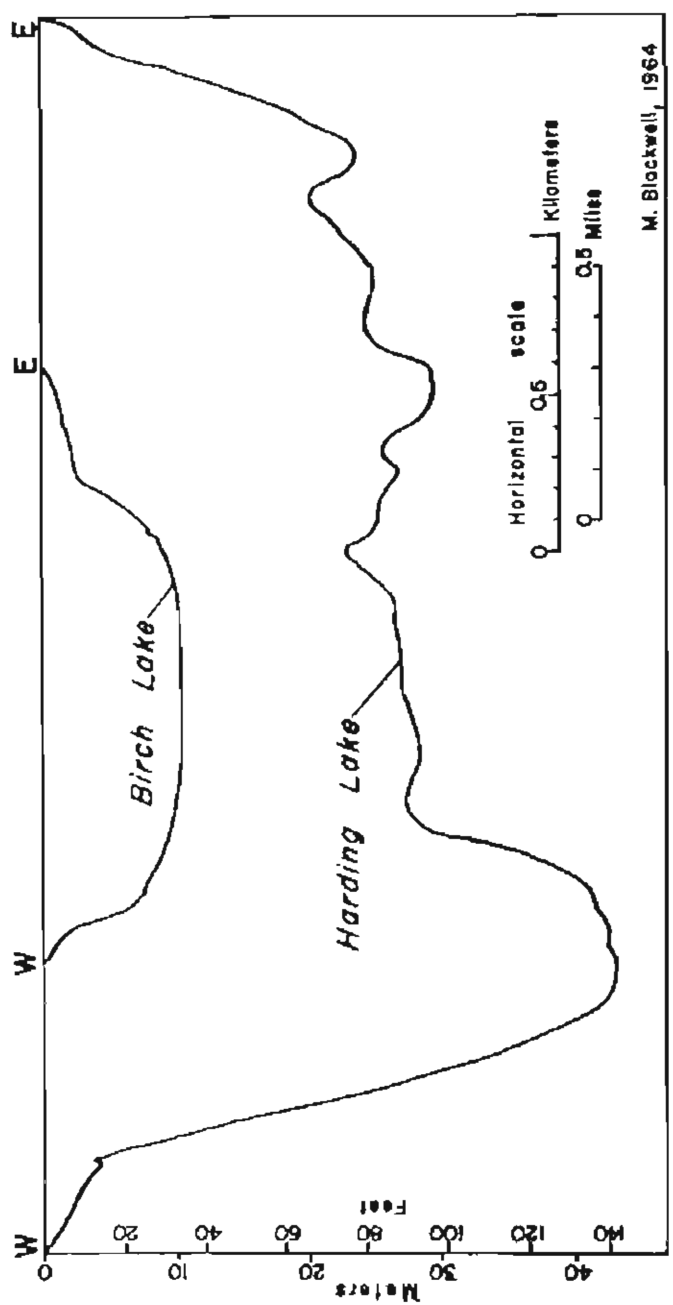

Figure 4-17. Bottom profiles of Birch and Harding Lakes, Alaska.

terrace is 15 to 18 feet (4.5 to 5.5 m) above the lower, and is covered with about 3 feet (1 m) of silt at the scarp. This silt mantle thickens rapidly toward the east.

314.4. Ventifacts of quartz occur on west slope of hill, 1 mile (1.6 km) east of road and 400 feet (120 m) above the present level of the Tanana. Farthest north occurrence of ventifacts of Delta(?) age. Ventifacts are buried by 2.5 feet (7.6 m) of loess.

313.7. Descend from lower terrace to modern Tanana flood plain. Terrace scarp is rock-defended at this point.

313. South-facing bedrock slopes are covered with sagebrush *(Artemisia frigida)*.

312.8. Left; basalt dikes and faults in Birch Creek Schist.

312.7. Left; landslide, probably inactive, in loess and colluvium on slope.

310.8 to 308.8. The road overlies ice-rich permafrost of colluvial loess derived from the loess-covered hill on the right. The roadway is consequently poor. The vegetation along this section of road is also a good indication of the presence of permafrost, there being mostly stands of black spruce, larch, and paper birch.

308.4. Right; bedrock is coarse-grained granite and quartz monzonite which is weathered to a depth of more than 20 feet (6 m). A borrow pit in this weathered zone reveals a well developed solifluction layer 5 feet (1.5 m) thick of grus. This layer is overlain with 1 to 5 feet (0.3 to 1.5 m) of loess.

308.3. Descend to lower terrace of Tanana River.

308.4 to 308. Black spruce *Picea mariana* forest on left is growing on frozen, organic-rich silt overlying the gravel of the lower Tanana terrace. Depth to permafrost is 1 foot (0.3 m) or less.

307.4. Ascend to upper terrace of Tanana River. Borrow pit on right is in Tanana River gravel. A ventifact has been found near the surface on the upper terrace under the 4-foot (1.2 m) thick silt cover near the southwest end of the lake.

307.2. This very rough stretch of road passes through a dense stand of black spruce that is about 90 years old and underlain by permafrost at a depth of 2 to 6 inches (4.5 to 25 cm) under the vegetation mat.

306.8. Stop 4-21: Birch Lake (fig. 4-17).

This lake has apparently been dammed by aggradation of the Tanana River in late Pleistocene time (see Stop 4-20).

The bottom sediment of the lake is a well-sorted organic silt nearly identical in lithology and grain size to the loess on the surrounding hills. Cobbles of Tanana River gravel have been recovered 475 feet (145 m) off the west shore.

On the right side of the road is exposed the sharp unconformity between the bedrock and the overlying loess. Mineralogical studies indicate that the silt is compositionally distinct from the quartz monzonite bedrock, thus excluding the possibility that the silt may be of residual origin, as postulated by Taber (1943).

305.9. A large number of finished stone tools and flakes of slate and chert were found 6 to 18 inches (25 to 46 cm) below the surface in loess when borrow pits were opened in the grus in 1947.

304.8. Left; borrow pit in weathered quartz monzonite, overlain by 4 feet of loess.

300.7. Enter Big Delta B-5 Quadrangle.

300. Stop 4-22: Mouth of Canyon Creek.

The two Tanana River gravel terraces apparently converge near here: the lower terrace (mantled by 17

feet (4.3 m) of silt) is only 3 feet (1 m) above the present level of the river, whereas the upper terrace is represented by scattered patches of lime-coated cobble gravel 17 feet (4.3 m) above the river. A log at the base of the silt overlying the lower gravel has been dated as 3,005 ± 75 years old (GX-0277).

The hill on the immediate left consists of bedrock, overlain by at least 3 feet (1 m) of fine-grained, uniform, eolian sand which is in turn covered by as much as 10 feet (3 m) of loess. The contacts of these beds slope gently toward the Tanana River.

299.3. Gravel of upper terrace overlying Birch Creek Schist. Gravel is 3.5 feet (1.1 m) thick and is covered by 5 inches (13 cm) of fine sand and 4.5 feet (1.4 m) of loess. This gravel is sporadically exposed on the left for a distance of a few hundred yards beyond Canyon Creek. Fine-grained sand usually overlies the gravel, but farther to the east (0.3 mile [0.5 km] beyond Canyon Creek) the sand lies directly on bedrock. This gravel is believed to be part of the upper Tanana terrace.

296.5. Stop 4-23: Tanana River Overlook.

River is 1¼ miles (2 km) wide. The gradient of the braided stream here is 6 feet per mile (1 m per km). Most of the successional stands visible along the river from this point are of alder and willow. An extensive white spruce stand can be seen in the distance on the south side of the river.

295.9. Left; borrow pit shows solifluction layer of bedrock debris, capped by retransported silt and sand.

295.6. Right; pingos occur in forest 100 feet (30 m) from road.

295.4. Banner Creek. One to 10 feet (0.3 to 3 m) of silt over creek gravel. Banner Creek gravel may very well correspond to the lower terrace gravel of the Tanana River. A log from 4 feet (1.2 m) below the top of the Banner Creek gravel, 220 feet (67 m) downstream from the highway bridge, has been dated as 3920 ± 75 years old (GX-0257).

295. Richardson Roadhouse. First established in 1907 after the gold strike on Tenderfoot Creek. The Roadhouse has been moved toward the hills three times as the Tanana River has encroached.

294.99. Stop 4-24: Richardson Roadhouse Gravel Pit.

Exposed in a borrow pit near the Richardson Roadhouse is an iron-stained, well stratified, poorly sorted angular to sub-rounded creek gravel with pebbles and cobbles 1 to 6 inches (2.5 to 15 cm) in diameter. No detailed study has yet been made of this exposure; however, the face exhibits the following:

Unit I. (basal unit). From the floor of the pit to a height of about 12 feet (3.5 m)—well bedded, iron oxide stained, pebble to cobble gravel with "fossil" ice wedges 1 to 4 feet (0.3 to 1.2 m) deep and .5 to 2 feet (15 cm to 0.6 m) wide.

Unit II. One to 10 feet (0.3 to 3 m) of gravel similar to Unit I with lenses and layers of sand and alluvial silt. Péwé found a weathered mammoth tusk 2 feet (0.6 m) from the top of the cut gravel surface.

Unit III. Loess 1 to 20 feet (0.3 to 6 m) thick draped over slope which was cut across the stratified gravel.

The gravel differs in size, sorting, roundness, and lithology from the Tanana River gravel at the base of the bluff nearby. The creek gravel in the pit contains many angular fragments of a porphyritic rhyolite. This rock is known to crop out a short distance away in Banner Creek Valley. A study of pebble orientations reveals that the direction of transport of the gravel was to the southwest, the direction of flow of the nearby Banner Creek today.

Preliminary interpretation of this exposure is that the gravel represents a high level creek gravel deposit of early to middle Pleistocene age and is capped by loess of late Pleistocene to Recent age. The relationship of the section to the terraces of the Tanana River is not yet known.

292.9. Top of hill, elevation of 1350 feet (422 m) or 450 feet (137 m) above river level. Three feet (1 m) of loess overlies a 2-inch (5 cm) thick gray sand layer which in turn overlies the Birch Creek Schist. Many wind-faceted quartz fragments occur in sand layer. The quartz fragments are residual from weathered bedrock. On top of Tenderfoot Hill ¼ mile (0.5 km) to the north at an elevation of 1535 feet (468 m), 650 feet (198 m) above the river, an excavation exposes 3.5 feet (1.1 m) of loess over an inch-thick gray sand layer containing quartz ventifacts. This in turn rests upon bedrock of the Birch Creek Schist.

292. Valley of Tenderfoot Creek. Gold strike occurred here in 1905. Placer gold was recovered by drift mining. Tailing piles of creek gravel can be seen in the trees on the left in the bottom of the valley. The road overlies ice-rich colluvial silt; thawing of permafrost produces a very irregular highway surface.

290.9. Right; coarse gray sand exposed in truncated ridge. This is eolian sand partly retransported from the hillside by wash. This is the first good exposure of this sand along the road between Fairbanks and Big Delta and represents the sand facies of eolian deposits of Delta age.

290.1. Tenderfoot Creek. Right; remains of old gold mine including elevated sluice boxes and steam power plant.

289.5. Left; 4 feet (1.2 m) of loess overlying eolian sand.

287.5. Left; 4 feet (1.2 m) of loess overlying clean gray crossbedded eolian sand with sharp contact. (Excellent stop for sand sampling.) Physical characteristics of the sand listed under discussion at Milepost 282.5.

287.2. Stop 4-25: Shaw Creek Bluff.

Geologic history — To the south from Shaw Creek Bluff can be seen the panorama of the Alaska Range and the broad Tanana River valley (fig. 4-18). On the right is Mt. Hayes (13,832 feet; 4219 m) and associated peaks. To the left of Mt. Hayes is a pass through the Alaska Range, the Delta River Pass occupied by the north-flowing Delta River and the Richardson Highway. The Granite Mountains form the foothill mountain group on the left of the pass. A small, dark, pyramidal peak, Donnelly Dome, stands out in front of the Range near the Delta River. In the middle ground on the right are low forested hills formed on the Birch Creek Schist. In the center middle ground is hummocky terrain of the 200-foot (60 m) high terminal moraine of the Delta Glaciation. To the far left (north of the Tanana River) is Shaw Creek Flats (fig. 4-18). The braided silt-laden Tanana River lies at the base of the bluff 100 feet below. From this spot the middle to late Quaternary history of the area can be reviewed, with emphasis on the eolian deposits (figs. 4-19, 4-20).

In Delta (Illinoian) time a piedmont glacier from the Alaska Range pushed north along the Delta River and terminated 8 miles (13 km) south of Shaw Creek Bluff, at the position of the terminal moraine today. Broad gravel plains extended northward from the glacier to Shaw Creek Bluff and into the Shaw Creek Flats. The braided Tanana River and outwash streams wandered over these plains. Winds from the south blew sand from the outwash plain—sand which cut ventifacts on the

the plain (Jack Warren Road and Quartz Lake) and cut and polished ventifacts on the south-facing bedrock slopes and hilltops from Shaw Creek Bluff to at least 20 miles (32 km) to the west (fig. 4-19). Ventifacts were formed at elevations ranging from river level to the tops of hills at least 650 feet (200 m) above river.

Probably throughout the time of the Delta Glaciation sand and silt continued to be blown from the plains northward and westward onto the hills. Sand dunes were formed on the south-facing slopes of the hills north and west of Shaw Creek Flats as the sand was stopped by the hills. Some dunes migrated over the low hills and exist on the north side of the hills. Approximately 40 square miles (100 km^2) of sand dunes were formed at this time.

The sand facies of the eolian deposits covers about 800 square miles (2000 km^2) north of Tanana River and thins away from the Shaw Creek Bluff area. At about 15 to 20 miles (25 to 30 km) away from the former ice front, windblown silt dominated. In the Fairbanks area 60 miles (96.5 km) away, for example, no sand facies of the eolian deposits of Delta time are known, but thick deposits of silt of Delta age are present (fig. 1-2).

After the Delta Glaciation, the glaciers retreated, tree line rose, eolian deposition decreased, and it is thought that base level was lowered as the Tanana River cut into its valley fill. Much of the sand cover on the lower hills was gullied, as was the loess in the Fairbanks area (Stop 1-14).

In late Quaternary (Wisconsin) time the Donnelly Glaciation occurred in the area and valley glaciers again pushed north from the range to the Tanana River Valley. In the Delta River area the glacier terminated near Donnelly Dome. Strong winds again blew sand and silt from the valley trains and outwash plains to cut ventifacts. Ventifacts formed during this glaciation occur much farther south than those associated with the more extensive Delta Glaciation (fig. 4-20). The ventifacts occur on moraines of Delta age and are within 1 to 5 miles (1.6 to 8 km) of the front of the Donnelly ice. Sand dune formation was restricted, being limited to areas near the Delta River (fig. 4-20). Windblown silt, however, was abundant, and areas north of the Donnelly glacier were blanketed with loess. Near Shaw Creek Bluff and Shaw Creek Flats the loess was draped over the sand dunes and the gullied sand blanket of the hill slopes. In post-Wisconsin time additional loess was deposited and covered dunes and moraines of Wisconsin age.

Shaw Creek Bluff section — Exposed in the road cut at Shaw Creek Bluff are unconsolidated sediments overlying the Birch Creek Schist which illustrate, in part, the geologic history of the area (fig. 4-21).

Silt and sand are spread on the hillside, and thicken downslope. Overlying the Birch Creek Schist is a 1 to 2 foot (0.3 to 6.1 m) thick calcareous solifluction layer with no ventifacts and rare pods of gray sand. Overlying the solifluction debris is a gray, coarse- to medium-grained sand layer a few inches (several centimeters) thick on the upslope end of the exposure and thickening more than 5 feet (1.5 m) downslope. During retransportation of the sand downslope many ventifacts were incorporated. The ventifacts are cut and polished and are from ¼ of an inch (0.6 cm) to more than 12 inches (30.5 cm) in diameter. The ventifacts have a 1 mm thick lime coat on their lower sides.

Overlying the sand with a fairly sharp break is loess less than a foot

(0.3 m) thick upslope but thickening to 5 feet (1.5 m) downslope. A Subarctic Brown Forest soil profile about 20 inches (51 cm) thick is present on the loess. No ventifacts occur in the loess.

The present interpretation is that the solifluction layer indicates a cold period and glacial advance of pre-Delta time. The ventifacts and the sand are of Delta age. In the warm period that followed the sand deposition and formation of ventifacts, the ground was not perennially frozen and ground water deposited lime on the ventifacts. Subsequently, in Donnelly time, the sand and ventifacts were buried by a loess blanket deposited on the hills.

296.7. Stop 4-26: Shaw Creek Road.

Description of the section—On the bluffs on the west side of Shaw Creek Flats at an altitude of 950 feet (289.2 m), only 50 feet (15.2 m) above the present level of the Tanana River, a borrow pit cut across two low ridges at mile 0.6 on Shaw Creek Road has exposed an informative section 170 feet (51.8 m) long in which are sediments of middle to late Quaternary age (fig. 4-22). Lying unconformably over bedrock is a calcareous layer of solifluction debris .5 foot (0.15 m) to more than 4 feet (1.2 m) thick composed of weathered and fractured bedrock that has been transported a short distance down the gentle bedrock surface toward Shaw Creek Flats.

Wedges and pockets 1 to 4 feet (0.3 to 1.2 m) long and .5 to 1 foot (0.2 to 0.3 m) wide of nonbedded medium-grained sand mixed with smaller rock fragments occur as ice-wedge fillings in the solifluction material. This sand fill contains no ventifacts, is more poorly sorted, and contains more silt than the overlying bedded sand. In the shallow valley between the two ridges the solifluction deposit is complex. There appears to be a gully filled with solifluction debris and large pockets and masses of pebbly sand with weathered rock fragments. There also occur a few stream pebbles of dark mafic rock, perhaps a dike rock, unlike the bedrock of the exposure. The pebbles are wind polished but are not wind cut.

Unconformably overlying the bedrock and the solifluction layer, as well as the sand wedges and pockets, is a well developed and extensive layer 1 to 4 inches (2.5 to 10 cm) thick composed of ventifacts. The ventifacts are vein quartz from the bedrock, are well cut and polished, and form almost a continuous horizontal sheet. The ventifacts are ½ to 6 inches (0.2 to 15.2 cm) in diameter. The largest and most striking ventifact is also the only one of exotic lithology—it is a 6-inch (15.2 cm) diameter ventifact of black chert found in the gully area overlying pebbles of transported mafic rock. Ventifacts and rock fragments in the solifluction and fractured bedrock have a lime coating on their lower sides.

Overlying the ventifact layer is a well-sorted medium-grain cross-bedded gray eolian sand layer which is a few millimeters thick in the center of the exposure in the shallow valley and as much as 3.5 feet (9.3 m) thick under the ridges (fig. 4-22). Several filled burrows of *Citellus undulatus* are present in the sand. The burrows are filled with sand or sand and loess. Near one burrow Richard Hill Péwé found a skull of a ground squirrel. It was identified as *Citellus undulatus* by Dr. G. D. Guthrie, vertebrate paleontologist of the University of Alaska, who stated that the mature skull was smaller than that of modern ground squirrels.

Blanketing the landscape is a 4-foot (1.2 m) thick layer of loess which uncomformably covers sand,

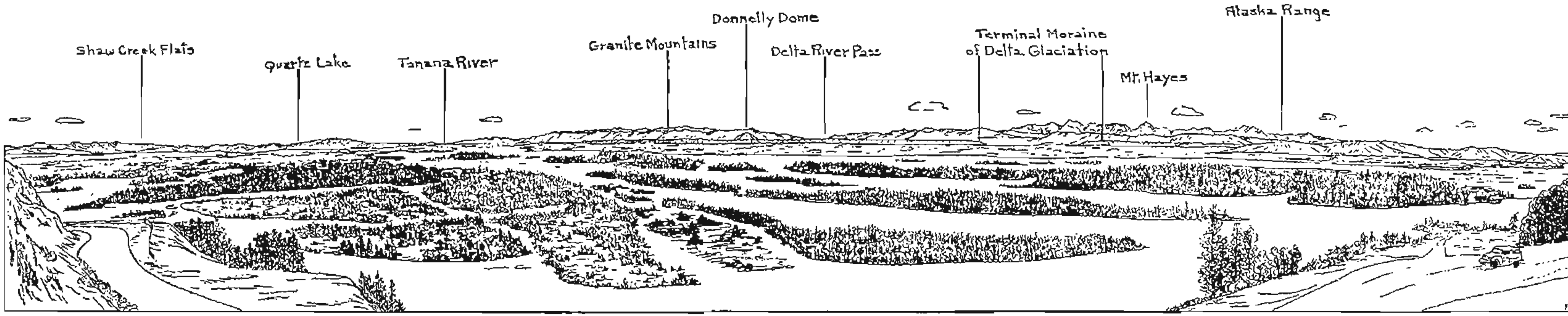

Figure 4-18. Panoramic view from Shaw Creek Bluff of the Alaska Range, glacial moraines, and the broad, braided Tanana River. The terminal moraine of the Delta Glaciation is represented by a 200-foot high ridge on the south side of the Tanana River, 8 miles south of the Shaw Creek Bluff. The glacier that deposited the moraine emerged from the Delta River Pass in the Alaska Range. (Sketch by Mary Jean Péwé.)

ventifacts, and bedrock. A well-developed Subarctic Brown Forest soil 24 inches (61 cm) thick is present on the loess. The loess, which maintains a vertical face, is perforated with holes of cliff swallow. Two feet (0.61 m) below the surface at the south end of the section is a thin ¼ inch (0.6 cm) thick charcoal layer. Radiocarbon dating places the age of this charcoal at 2565 ± 290 years (GX-0254).

Geologic history—A record of alternating periods of cold and warmer climates from middle Quaternary to present times is represented by the sediments in this exposure. The solifluction deposits were formed under a rigorous climate of middle Quaternary time. Solifluction deposits formed at an altitude of 950 feet (289.2 m) at this time; today solifluction deposits are forming at an altitude of 3000 to 3500 feet (910 to 1070 m) in the Yukon-Tanana upland. The ice wedges that formed in the solifluction deposits may either be interpreted as forming in the later part of the cold period, or in a separate cold period after the solifluction deposits became stabilized. In either instance, the mean annual air temperature was at least 3°C to 4°C colder than now.

Evidently some sand was blown on to the hills in the later part of the cold period. Streams existed in the small gully in the center of the section, and rocks foreign to the area were brought in from upslope.

The cold period was followed by a warmer period, as indicated by the melting of the ice wedges and by the filling of voids with overlying sand and with solifluction material from the collapsing sides of the voids. During this time a thin blanket of residual fragments of vein quartz collected on the surface as the result of long weathering of the bedrock and the solifluction deposit.

With the return of a more rigorous climate, a cold period thought to be associated with the Delta Glaciation, sand was blown across the bluff, cutting and polishing the rock fragments. The cutting and polishing is especially well preserved on the quartz fragments and other resistant rocks. Eventually, accumulating sand covered the bedrock, building dunes and a sand blanket, thereby terminating most of the ventifact cutting by burial. Ground squirrels were abundant at this time, indicating near-tree-line conditions.

With the amelioration of the climate, glaciers retreated and sand deposition ceased or remarkably decreased. The squirrels moved to other areas as the tree line rose. The sand blanket was gullied and in part removed by erosion.

The loess layer is thought to be associated with the advance of glaciers in Donnelly time, a time when the outwash plains and valley trains were less vegetated. The loess was draped over the landscape during this time of more rigorous climate. In the last few thousand years, with the withdrawal of glaciers and warming of climate, loess deposition was reduced and a well-developed Subarctic Brown Forest soil formed on the loess. The upper layers of the loess are relatively young.

285.5. On the poorly drained Shaw Creek Flats, which are underlain by permafrost at shallow depths, are extensive stands of larch (*Larix laricina*). In the center of the flats is a nearly treeless bog which appears to be a stabilized strangmoor or string bog. The vegetation of the bog consists mainly of sedge tussocks (*Eriophorum vaginatum*), sedges, (*Carex* spp.), and the shrubs, dwarf birch (*Betula glandulosa*), and

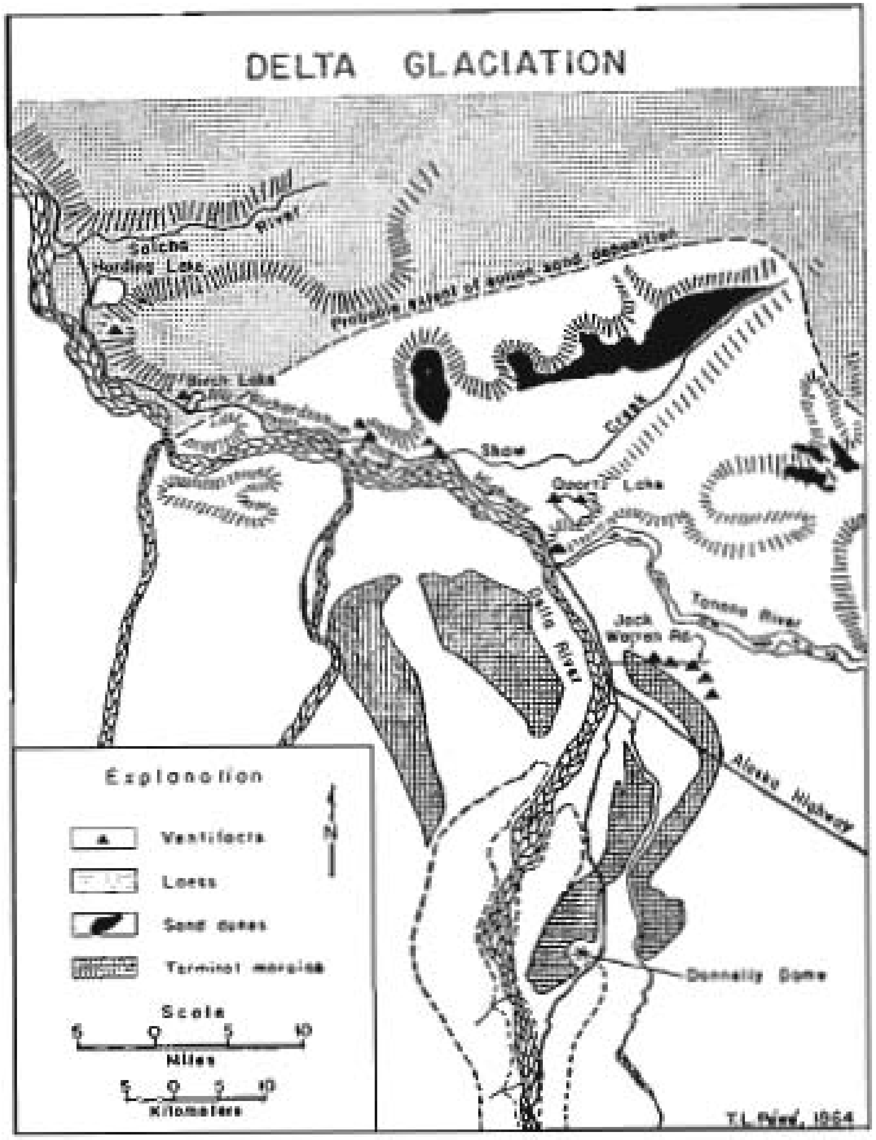

Figure 4-19. Distribution and types of eolian deposits in the Big Delta area, Alaska in relation to glacial advance: Delta Glaciation.

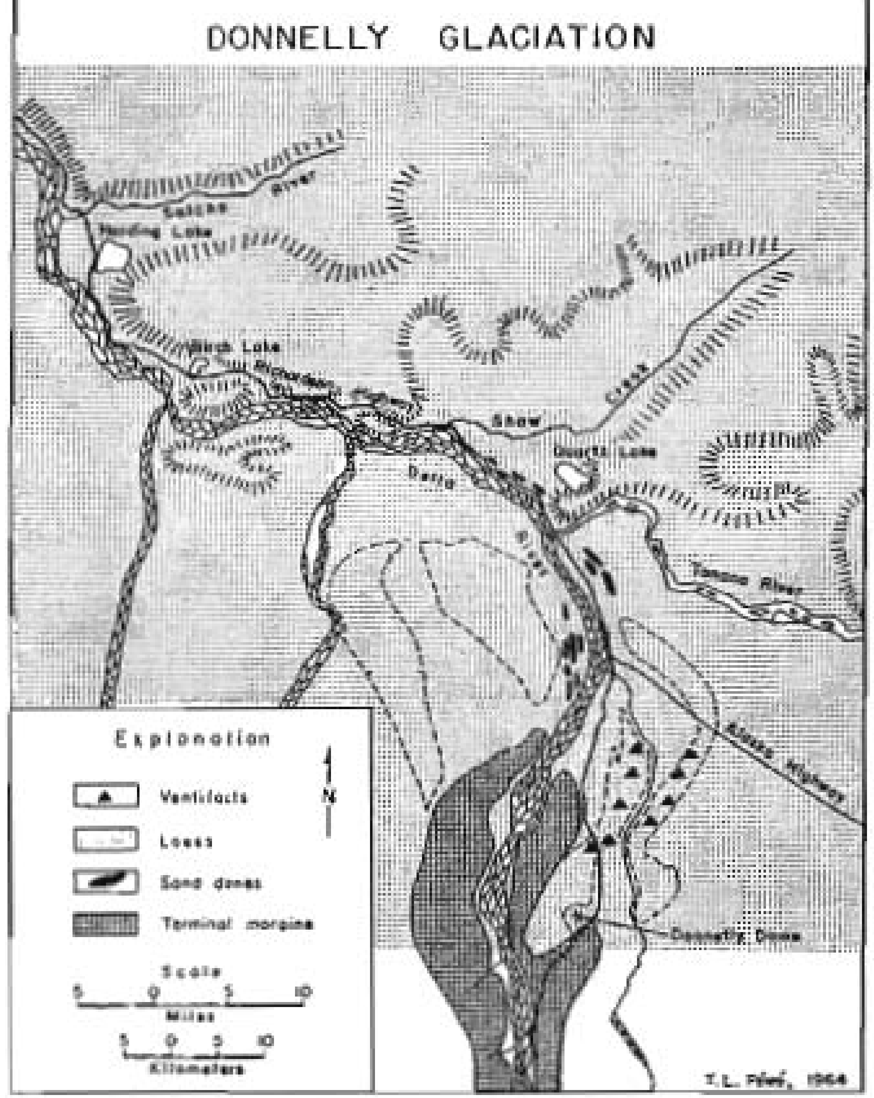

Figure 4-20. Distribution and types of eolian deposits in the Big Delta area, Alaska in relation to glacial advance: Donnelly Glaciation.

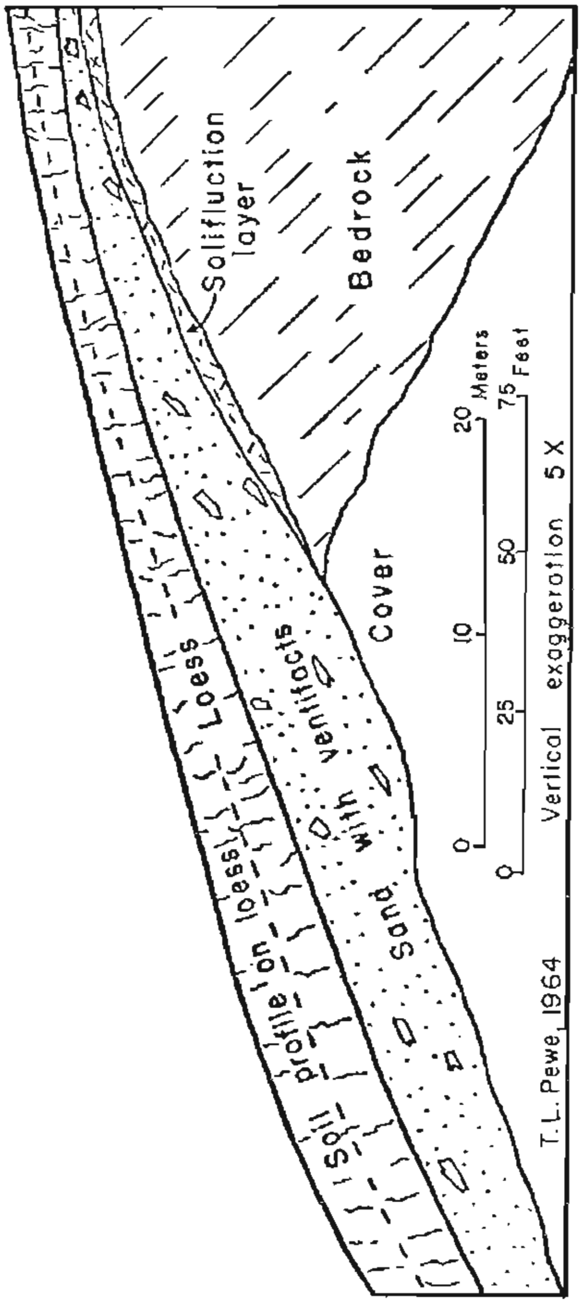

Figure 4-21. Road cut exposing loess, eolian sand, ventifacts, and solifluction deposits overlying bedrock. Shaw Creek Bluff, 20 miles (32 km) west of Delta Junction, Alaska.

leatherleaf *(Chamadaphne calyculata)*. From the air distinct patterns of ridges and depressions can be seen. At least the edges of the bog have been burned, as is evidenced by a few charred stumps. Most of the black spruce and larch visible from the highway are from 50 to 100 years old.

285. Enter Big Delta A-3 Quadrangle.

The sand dunes along the edge of the Tanana River were deposited more than 8,000 years ago, probably in late Wisconsin time. They represent the northernmost sand dune area associated with the glacial advance of Donnelly time in the lower Delta River area (fig. 4-20).

277.8. Left; pumping station for 10-inch (25.4 cm) fuel line. Right; section exposed in Tanana River terrace scarp is as follows:

Top	
0 to 4 feet (0 to 1.2 m)	loess with forest-fire layers. Radiocarbon date on charcoal near base of loess is 8040 ± 190 years (GX-0255).
4 to 9 feet (1.2 to 2.7 m)	tan, fine-grained, crossbedded eolian sand (dune).
9 to 13 feet (2.7 to 4.0 m)	buff fluvial sandy silt.
13 to 19 feet (4.0 to 5.8 m)	Tanana River rounded pebble gravel.

283. Enter Big Delta A-4 Quadrangle.

282.5. Sand dunes of late Wisconsin and Recent age. For the next 5 miles (8.1 km) (282.5 to 277.7) the road follows a linear, "cliff head" type sand dune area on the edge of a low terrace next to the Tanana River (fig. 4-20). The dunes are 5 to 10 feet (1.5 to 3 m) high and occupy a belt about 100 to 200 yards (90 to 180 m) wide, forming a dry ridge between the river and the extensive swampy Shaw Creek Flats to the north. It is an excellent location for the highway. The dunes are covered with 1 to 5 feet (0.3 to 1.5 m) of loess, but the sand is well exposed in road cuts.

The physical properties of the tan sand exposed here and the gray sand of Illinoian age exposed at milepost 287.5 near Shaw Creek Bluff (Stop 4-25) have been examined by members of the Pleistocene Geology class at the University of Alaska and are listed in Table II.

277.7. End of sand dunes.

277.5. Two miles (3.2 km) north (left) lies Quartz Lake (fig. 4-18). Quartz Lake is similar to Harding, Birch, and other lakes on the north side of the central Tanana Valley inasmuch as it is thought to have formed as the result of aggradation of the Tanana River damming small valleys of the Yukon-Tanana upland during late Pleistocene time. Quartz Lake is a shallow (40 ft [12 m]) lake held in on the west side by Tanana-Delta River alluvium (fig. 4-22). Examination of the gravel alluvium in the ice-shoved ramparts of the lake reveals numerous cobbles up to 6 inches (15 cm) in diameter, of quartz, diorite, granite, and other rocks of the Alaska Range. The cobbles are severely wind faceted, grooved, and polished. The presence of wind-cut cobbles of Alaska Range lithologies at this location suggests that in Delta time the outwash plain from the glacier extended to the lake area, dammed local drainage from the

Table II. Physical properties of eolian sand exposed at Shaw Creek Flats and near Shaw Creek Bluff, central Alaska.

	Shaw Creek Flats (Mile 282.5 on Richardson Highway)	Near Shaw Creek Bluff (Mile 287.5 on Richardson Highway)
Mineralogy	73% quartz	63% quartz
Roundness	approx. 75% angular to subangular grains	approx. 75% angular to subangular grains
Sphericity	32% grains have high sphericity	75% grains have high sphericity
Luster	89% of grains have dull luster	75% of grains have dull luster
Sorting	well sorted	well sorted
Mean diameter	1.47 mm	2.27 mm
Color	tan	gray

hills, and formed the lake. The cobbles were wind cut in Delta time when the broad outwash plain was scantily vegetated and sand could be transported easily. These wind-cut cobbles are thought to be the same age as ventifacts examined at earlier stops in the vicinity of Shaw Creek Bluff. No detailed work has been done on the origin and age of Quartz Lake, and much work, especially pollen analyses, is necessary to provide final support for this hypothesis of origin and age.

275.7. Left; former location of Bert and Mary's lodge. For several years a large log cabin, deformed by thawing of permafrost, existed at this spot. Photographs of the cabin have appeared in various publications. The roadhouse was torn down in 1964. The deformed foundation remains.

275.5. Right; hill of Birch Creek. Schist covered with Wisconsin loess that has slumped.

275.3. Tanana River Bridge. Mouth of the braided Delta River emptying into the Tanana.

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DELTA RIVER AREA, ALASKA RANGE

by

Troy L. Péwé

RESUME OF QUATERNARY GEOLOGY OF THE DELTA RIVER AREA, ALASKA RANGE

The Alaska Range is a great glacially sculptured arcuate mountain wall extending west and southwest 600 miles (170 km) from the Canadian border to the Aleutian Range. It is composed of a core of Precambrian or lower Paleozoic schist and gneiss, the higher mountains being held up by granitic intrusions of Mesozoic age. The range is flanked by, and in part made up of, sedimentary and volcanic rocks of Paleozoic and Mesozoic age. On the lower flanks of the range, and underlying adjoining lowlands, are poorly consolidated coal-bearing gravels and clay sediments of Tertiary age. The range is extremely rugged and contains Mt. McKinley, 20,300 feet (6,191.5 m) in altitude, the highest mountain on the North American continent. In the central Alaska Range, Mt. Hayes, Mt. Deborah, Mt. Hess, and Mt. Kimball are prominent peaks. The Delta River rises on the south side of the range in the Tangle Lakes and flows north. The Richardson Highway (fig. 5-23) crosses the range through the Delta River Pass.

The range is characterized by spectacular valley glaciers 1 to 40 miles (1.6 to 65 km) long. The glaciers are largest and most numerous on the south side of the range, as they are nourished chiefly by air masses moving northward to northeastward from the northern Pacific Ocean. On the south side of the central part of the range, snowline today is 5,500 feet (1,670 m) above sea level; on the north side it is 6,500 feet (1,980 m).

At least four Quaternary glaciations, each successively less extensive than the former, are recorded in the Delta River area in the central Alaska Range. The glaciers pushed south and north from the crest of the range, and some of the ice on the south side found egress to the north through the Delta River Pass. On the north side, the glaciers largely remained as valley glaciers, spreading out terminal bulbs on the lowland of the Tanana River valley. On the south side, the glaciers coalesced, forming large piedmont glaciers or ice sheets covering the lowlands and pushed south into the Copper River basin.

The earliest glacial advance recognized is the Darling Creek Glaciation of early Quaternary age. It is identified from patches of drift 2,000 to 3,000 feet (610 to 920 m) above the floor of the Delta River and from isolated erratics up to 15 feet (4.6 m) in diameter in the Amphitheater Mountains on the south side of the Alaska Range. The succeeding glacial advance was the Delta Glaciation, recognized by fairly well preserved, breached moraines on the north side of the range. On the south side the glaciers of this advance did not cover all the small peaks of the foothills. This advance is thought to be middle to late Quaternary in age and extended well into the Copper River basin.

The next two glacial advances took place in late Quaternary (Wisconsin) time and are closely related in extent and age. On the north side they have been grouped into one broad glaciation termed the Donnelly Glaciation; on the south side, they have been named the Denali I and Denali II Glaciations. These deposits on both sides of the range are characterized by fresh knob and kettle topography. On the south side, especially in the vicinity of the Tangle Lakes, are hundreds of square miles (several hundred square km.) of fresh ice-contact features formed when broad sheets of the ice stagnated. In very late Quaternary time (post-Wisconsin) there was a rather small advance of the glaciers. It is especially well represented on the south side of the range.

Glacial advances of the last three centuries have left moraine loops within a few hundred yards (a few hundred m) to a few miles (several km) of the present glaciers.

Road Log and Locality Descriptions

274.3.[1] South edge of flood plain of Tanana River. Large yellow markers on right side of the road are mileage markers for buried 10-inch (25.4 cm) diameter pipeline extending from Haines to Fairbanks.

274.4. Outwash fan of Donnelly Glaciation of Wisconsin age. Elongate sand dunes 1 to 1.5 miles (1.6 to 2.4 km) long are on left side of the road .5 to 1 mile (0.8 to 1.6 km) to the east (fig. 4-20). Can be seen on the Big Delta A-4 Quadrangle.

268.2. Stop 5-27: Left; Jack Warren Road.

Terminal moraine of Delta age 1¼ miles (2 km) from Richardson Highway. In front of moraine is outwash plain covered with well-developed ventifacts (fig. 4-19).

267.0. Old channel of Jarvis Creek. Channel active in late Wisconsin time.

266. Delta Junction. End of the Alaska Highway, Milepost 1422.

264.8. Jarvis Creek. On windy days dust is blown from this small flood plain. A ridge several feet high (more than a meter) of loess has been built on the north side of the creek. The vegetation succession stages are somewhat different on the Delta River and its tributaries than on the finer alluvial deposits of the Tanana. Early succession stages are more long persisting and include both the dwarf and narrow-leafed fireweed (*Epilobium latifolium*) and *E. angustifolium*), several legumes, the most conspicuous of which are *Hedysarum* spp., *Astragalus* spp., and *Oxytropis* spp., and the shrubs *Dryas drummondii* and *D. octopetala*, buffalo berry (*Shepherdia canadensis*), and several willow species. This early pioneer stage is often replaced by a nearly continuous cover of forbs and grasses, of which *Festuca* spp., *Elymus inovatus*, and *Agropyron* spp. are the most common. These early successional stages are an important source of food for the buffalo herd that was introduced in the Big Delta area in the 1920's.

Balsam poplars invade and replace the meadow stage which in turn is succeeded by white spruce. On most of the coarse gravel deposits white spruce may persist indefinitely, or it may eventually be replaced by black spruce in a similar manner to that described for the Fairbanks area.

Directly to the south Donnelly Dome can be seen, 16 miles (26 km) away.

264.6. Edge of outwash plain from the main lobe of the ice advance of Donnelly age.

263.2. Enter Mt. Hayes D-4 Quadrangle.

(Refer to U.S. Geological Survey Map I-394 by T. L. Péwé and G. W. Holmes, "Geology of the Mt. Hayes D-4 Quadrangle, Alaska," 1964.)

262.7. Stop 5-28: Right; Federal Aviation Agency Station Overlook.

South from here can be seen a panorama of the Alaska Range with

[1]Miles from Valdez on the Richardson Highway.

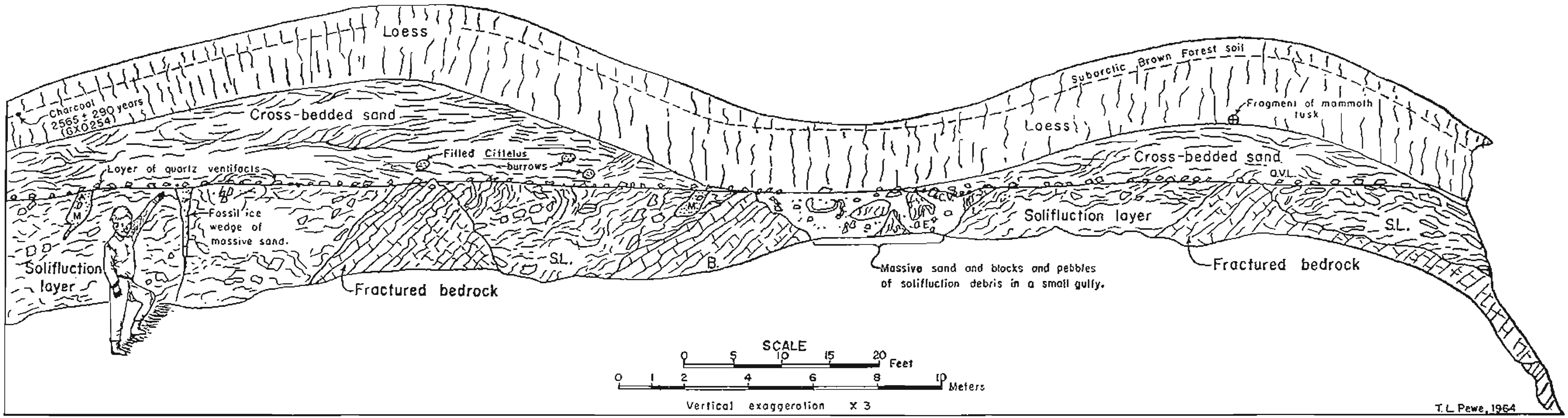

Figure 4-22. Exposure in a barrow pit at mile 0.6 (1.0 km) on Shaw Creek Road, Central Alaska, of a solifluction layer with fossil ice wedges of sand overlain by an extensive ventifact layer. The ventifacts are overlain by eolian sand which is in turn blanketed by 4 feet (1.2 m) of loess. SYMBOLS: QVL—quartz ventifact layer; SL— solifluction layer; B—fractured bedrock; CV—well developed chert ventifact; E—stream pebbles of basic dike rock wind polished but not cut; M—massive unbedded sand. Boy is pointing to layer of quartz ventifacts.

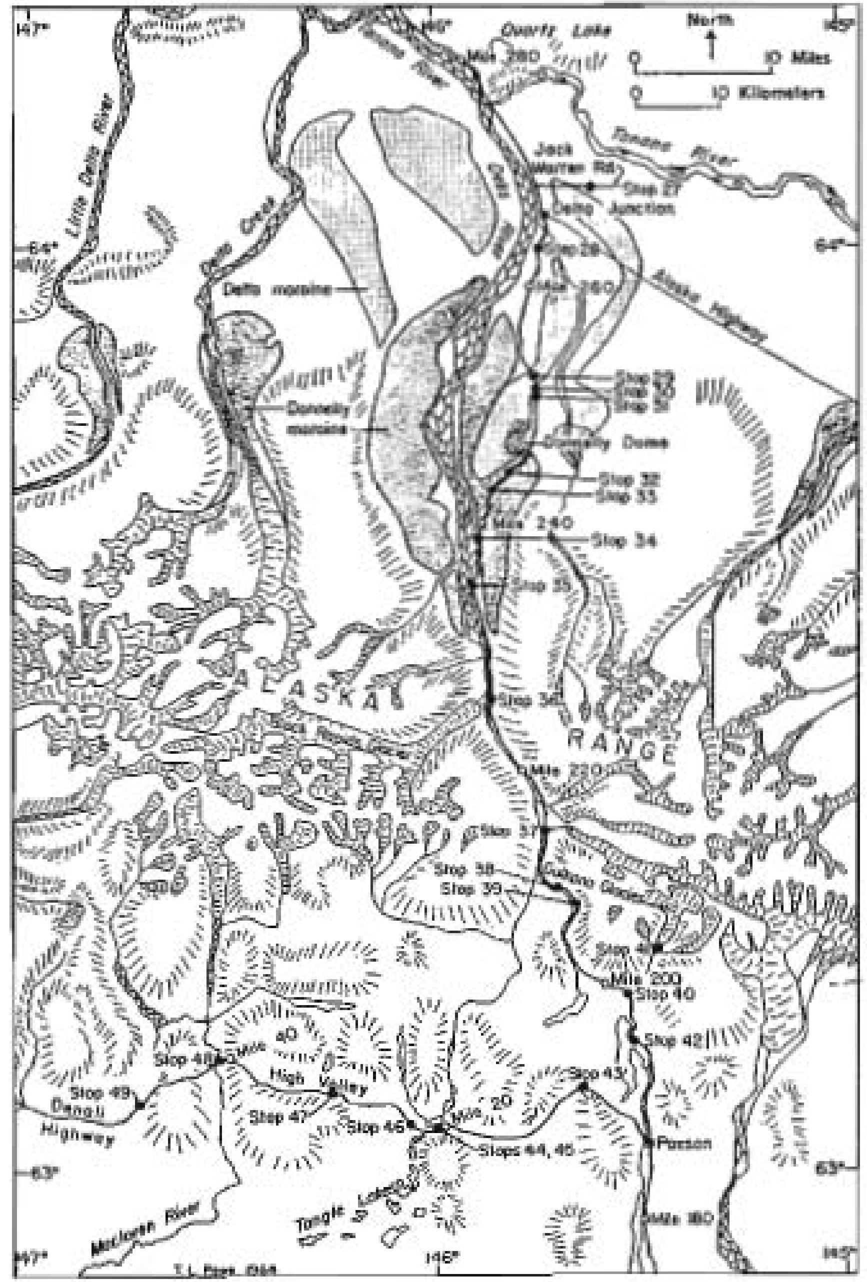

Figure 5-23. Index map of the Delta River area, Alaska Range, showing Field Trip Stops.

Mt. Hayes looming as the most prominent peak. Where the braided Delta River comes into view it has cut through the terminal moraine of the Donnelly Glaciation (Wisconsin). Directly west from the overlook, the subdued moraine of Illinoian age forms the crenulate skyline.

261.8. Left; entrance to Ft. Greely.

258.8. The gradient of the outwash plain from Ft. Greely to the Donnelly terminal moraine is 55 feet per mile (10.5 m per km). The outwash sediments range in size from 4- to 6-inch (10 to 15 cm) diameter cobbles at the end of the outwash plain at Milepost 264.4 to boulder size, 1 or 2 feet (0.3 to 0.6 m) in diameter, at front of the moraine at Milepost 256.5.

On this section of outwash are nearly continuous stands of balsam poplar and aspen, in many places with an understory of white spruce, that have resulted from a fire about 50 years ago. The area will probably develop a white spruce stand similar to that through which the highway passed just north of Big Delta.

At the edge of the Donnelly moraine the successional stands change to those of paper birch and young spruce, perhaps due to the increase in silts in the moraine.

257.5. Right; entrance to Meadows Road. This road leads to outcrops of loess more than 40 to 50 feet (12 to 15 m) thick along the edge of Delta River flood plain. Radiometric dates of approximately 7,000 years have been obtained on wood at the base of the loess.

256.5. Front of the Donnelly moraine. Good view of Granite Mountain to left.

254. Start climbing the subdued moraine of Delta age.

252.7. Stop 5-29: Moraine of the Delta Glaciation.

The roadcut exposed light yellowish-brown silty gravel composed of cobbles and boulders of gneiss, granite, diorite, dark volcanic rocks, with some limestone and schist. The percentage of schist fragments in the till of Delta age is only 1 to 10 percent in contrast to the large percentage of such fragments, 20 to 25 percent, in the less weathered till of Donnelly age. The Delta Glaciation is thought to be pre-Wisconsin in age because the morainal topography is considerably subdued and the moraines fragmentary. Also, the boulder count is low, and many large boulders on the moraine are weathered into pinnacled masses. The kettles are partly filled and modified. The number of kettles per square mile on the Delta moraine is 3 (1.1 per sq. km), in contrast to 16 kettles per square mile (6.3 per sq. km) on the Donnelly moraine. However, there is only one bog per square mile (0.4 per sq. km) in the kettles on the Donnelly moraine as compared to 2 (0.8 per sq. km) on the Delta moraine.

251.5. Stop 5-30: Edge of the Delta Moraine and Outwash of Donnelly age.

South from this location can be seen a broad, flat lowland of outwash gravel of Donnelly age that emanated from a lobe of the Donnelly glacier as it stood at the base of Donnelly Dome 4 miles (6.4 km) to the south. The highway can be seen on the front of the terminal moraine of Donnelly age in the distance. To the right, extending from Donnelly Dome, is a moraine of Delta age that is the same moraine that swings around to the locality of this Stop. The elevation of this moraine can be matched with the fairly well developed shoulders of the Dome. In Delta time the glacier completely surrounded Donnelly Dome which protruded as an nunatuk 700 feet (214 m) above the glacial ice. The Dome itself is composed

of the Birch Creek Schist of Precambrian or early Paleozoic age and is a stream and glacially modified fault block. It lies on the upthrown side of a fault which extends from the Dome eastward into the Granite Mountain. Granite Mountain is a fault block of quartz monzonite. The fault line along the front of Granite Mountain is quite prominent and cuts the moraines of Wisconsin age which were deposited at the base of the mountain by small glaciers originating in the block. The flat remnants present on the top of Granite Mountain have been interpreted as an exhumed erosional surface upon which Tertiary gravel and coal-bearing deposits were laid down.

During Wisconsin time strong winds blew across the unvegetated outwash plain to the south. The winds carrying sand buffeted the edge of the moraine of Delta age, cutting, grooving, and polishing boulders and cobbles, forming ventifacts. Many of these ventifacts occur on the moraine from this spot north to Ft. Greely (fig. 4-20).

251.2. Stop 5-31: Polygonal ground and fossil ice wedges.

Large-scale polygons on the outwash plain of Wisconsin age (Donnelly Glaciation) are outlined by a network of intersecting trenchlike depressions 1 to 3 feet (0.3 to 0.9 m) deep and 3 to 6 feet (0.9 to 1.8 m) wide. The polygonal outline is accentuated by differences in vegetation between the centers of the polygons and the trenches. The polygons are 80 to 130 feet (24.4 to 39.6 m) in diameter and are 3 to 6 sided; the greatest number are 4 sided (fig. 5-24).

Wedge-shaped masses of relatively fine-grained sediments underlie the slight surface depressions which mark the polygon boundaries (fig. 5-25). These wedge-shaped masses, or wedges, crosscut poorly stratified glacial outwash gravel. The wedges have a wide upper part, a narrow middle section, and an extremely irregularly shaped lower part. They extend from 3 to 9 feet (0.9 to 2.7 m) below the ground surface and all bend and curve to varying degrees. Some of the wedges widen and narrow again; many of them terminate in a sharp hook, and some terminate in large footlike masses or bulges.

The sediments of the wedge consist of brownish silt in the upper part and greenish-gray silt in the middle and lower part. Mixed with this silt are pebbles and cobbles the same size as found in the outwash gravel. The outwash sediments adjacent to the wedges are slightly different from the undisturbed outwash material, inasmuch as fine material appears to have been added. Adjacent to the sediment wedges, generally on one side, there is a zone of iron staining which is widest at the top and narrows downward. It is the light-gray to tan silt in the wedge lying next to the brown or reddish gravel that so strikingly outlines the wedges in the field.

The hypotheses considered for the origin of the large-scale patterned ground in the Donnelly Dome area fall into two major groups: (1) desiccation-crack hypothesis and (2) thermal-contraction crack hypothesis. The first hypothesis is illogical because of the absence of clay-sized minerals. The thermal-contraction crack hypothesis proposes that the polygonal pattern is produced by cooling and tension cracking of the ground in the winter as the result of volume reduction owing to contraction of ice-cemented sediments. Two variants to be considered under this hypothesis include: (1) seasonal crack polygons and (2) ice-wedge polygons.

Polygons do not crack in the seasonally frozen ground in the Donnelly Dome area today. It would ap-

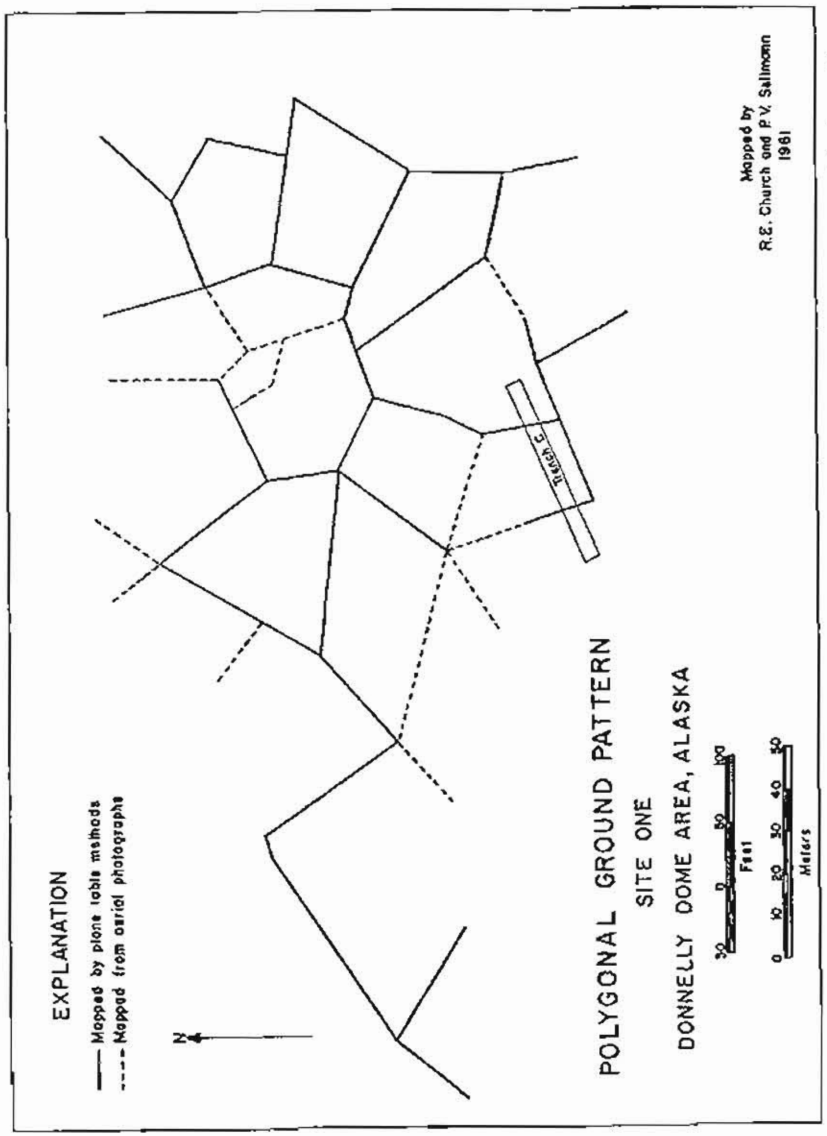

Figure 5-24. Plane table map of large-scale polygons, Site 1, Donnelly Dome area, Alaska. (From Church, Péwé, and Andresen, 1965, fig. 6)

pear that for seasonal contraction cracks to form under natural conditions, the snow and/or vegetation cover in the past must have been different, or the ground must have been colder. With greater cooling of the ground in the past, permafrost would have been more widespread, and therefore it is thought that conventional ice wedges would form in the contraction cracks in the permafrost rather than sediment wedges in the seasonally frozen ground.

The origin of the large patterned ground in the Donnelly Dome area can be explained by the ice-wedge hypothesis theory. Ice wedges formed in the area during Wisconsin time, subsequently melted, and the voids filled with sediment.

The climate in the Donnelly Dome area at the time the large-scale contraction crack polygons formed was colder and more rigorous than today. The tree line was 1,500 to 1,800 feet (460 to 550 m) lower and snowline

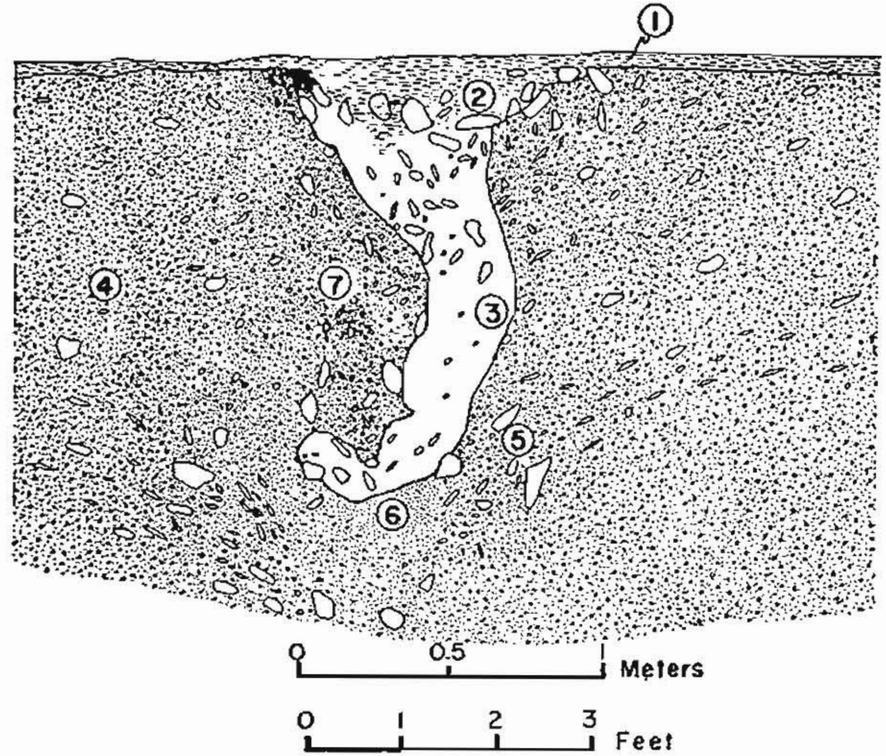

Figure 5-25. Diagrammatic sketch of a fossil ice wedge, Donnelly Dome area, Alaska. Circled numbers indicate the types of sediment: (1) silt mantle, (2) fill of upper part of wedge, (3) fill of middle and lower part of wedge, (4) undisturbed outwash gravel, (5) disturbed outwash gravel, (6) sand, and (7) iron-stained sediments. No vertical exaggeration. (From Church, Péwé, and Andresen, 1965, fig. 17)

was 1,500 feet (460 m) lower. In Wisconsin time the air temperature was at least 5.4°F (3°C) lower, or at least 21.6°F (−5.8°C), in contrast to the mean annual air temperature of 27°F (−2.8°C) today.

Study of the large-scale patterned ground in the Donnelly Dome area has demonstrated that extensive areas of fossil ice-wedge polygons can occur in coarse-grained sediments in regions where permafrost is actively growing, such as in central Alaska, and where large ice wedges are still present in the fine-grained sediments. Such an association supports the suggestion that the permafrost and ice wedges thaw more rapidly in coarse-grained sediments than in ice-rich fine-grained sediment.

247.8. Rise upon edge of moraine of Donnelly Glaciation.

247.4. Normal(?) fault, the movement having occurred in post-Wisconsin time. Road crosses a fault-line scarp approximately 10 feet (3 m) high. Fault extends east-west across the terminal moraine of Donnelly age with the upthrown side on the south. Donnelly Dome is bounded on the north by this fault.

246. Knob and kettle terrain of the terminal moraine of the Donnelly Glaciation. The aspect of this terrain is typical of the topography of terminal moraines of Wisconsin age throughout Alaska, especially of moraines of middle to late Wisconsin age.

The vegetation on the Donnelly moraine is typical for that in the Alaska Range near the altitudinal limit of trees. Both black and white spruce are found on the moraine but in a very scattered distribution pattern. Between the trees are shrubs of dwarf birch *(Betula glandulosa)*, willows *(Salix* spp. especially *S. pulchra)*, blueberry *(Vaccinium uliginosum)*, and Labrador tea *(Ledum groenlandicum* and *L. decumbens)*. Between the shrubs and on exposed knobs are mats of crowberry *(Empetrum nigrum)*, alpine bearberry *(Arctostaphylos alpina)*, and lichens, especially *Cladonia* spp. Many of the spruce on the Donnelly moraine are less than 100 years of age which indicates that under present climatic condition an invasion by the trees is in progress. Small clumps of balsam poplar are typical on recently exposed glacial moraines in the Alaska Range, and these seen on the Donnelly moraine may have persisted since the original colonization by plants more than 7,000 years ago.

245.2. Stop 5-32: Donnelly Till.

The light yellowish-brown silty to sandy till is characteristic of the Donnelly Glaciation in the Delta River area of the Alaska Range. The schist fragments are much more common in till of this age then in the more weathered till of the Delta Glaciation. Directly to the north from this stop near Donnelly Dome is a massive push moraine. On this moraine, and others nearby, a number of important archaeological sites have been found in the past two years.

In the summer of 1964, the University of Alaska carried out excavations and site surveys in the area which have resulted in the accumulation of important new data on early man in Alaska. Some twenty sites were discovered, at least two of which were of an early core and blade tradition related to the Campus site and probably predating the primarily coastal Denbigh Flint Complex. One site contained notched points and is clearly separate from the core and blade sites and from the several sites which yielded lanceolate points of the general Plano category.

In the past there has been a great deal of game here, probably including, in late Wisconsin or early Recent times, herds of horses and bison. In the Donnelly region game is not now abundant. This may be due in part to its ready accessibility to hunters. Unfortunately no faunal remains have been found in these sites. Radiocarbon dates from several sites are pending.

244. Enter Mt. Hayes C-4 Quadrangle.

243.5. Stop 5-33: Picture Stop of Mt. Hayes and the Alaska Range.

241.3. Right; view of the Alaska Range, Delta River, the alluvial fans. Black Rapids glacier in distance. On left, the immediate skyline is the lateral moraine of the Donnelly Glaciation. Moraine is 650 feet (198 m) above the river. Lateral moraine on other side of the valley is of Donnelly age.

240.2. View south shows several rock knobs (roches moutonnées).

239.9. Right; two loess-covered roches moutonnées of Birch Creek Schist. Quartz veins still retain glacial polish.

238.15. Stop 5-34: Loess section.

Left; roches moutonnées of Birch Creek Schist on left of lake. Right; 300 feet (91 m) off the road, is an exposure in the cliff along the Delta River. The following section is exposed:

0 to 3 feet (0 to 0.9 m)	Loess with numerous organic laminations and fossils of pulmonate snails. The following snails have been identified from the loess, *Succinea avara* Say, *Discus chronkhitei* (Newcomb), and *Euconulus fulvus alaskensis* Pilsbry.
3 to 7 feet (0.9 to 2.1 m)	Unfossiliferous loess with some organic layers.
7 feet (2.1 m)	Ash Bed Jarvis Creek, ¼ inch (6 mm) thick.
7 to 12 feet (2.1 to 3.6 m)	Loess with little organic material.
12 to 28 feet (3.6 to 8.5 m)	Light yellowish-brown sandy to silty boulder till of Donnelly age.
28 to 41 feet (8.5 to 12.5 m)	Sandy outwash gravel. (Probably pro-Donnelly outwash.)
41 to 70 feet (12.5 to 21.4 m)	Cover

Good view of the section can be seen looking north from Mile 237.

237.5. Flood plain of the Delta River. Good place to observe wind blowing dust from the flood plain. Strong persistent winds and wide vegetation-free river flood plains in the past, as well as present, make the Delta River an ideal area for watching wind as a geologic process. Winds from the south and east have blown great clouds of silt from the Delta River and Jarvis Creek flood plains since at least the time of Delta Glaciation and have, therefore, blanketed the adjacent terrain with loess (See isopach loess map, Péwé and Holmes, 1964). Clouds of silt are blown 1,000 to as much as 4,000 feet (300 to 1,200 m) above the land surface today and cover hundreds of square miles in the area. As might be expected, areas to the leeward of the source areas (flood plains) are most heavily blanketed with silt.

The area west of the Delta River is heavily covered with loess. Smaller areas on the east side of the Delta River that break the regular flow of the wind are also heavily covered.

Deposits of windblown sediments are thicker near the flood plains. The leaves and limbs of the trees near the river are covered with fine silt during much of the summer. In many areas the trunks of the trees have small cones of fresh silt at their base, the result of the silt washing down the tree from the limbs and leaves. The floor of the forest is dusty, and a gradual accumulation of silt requires new root shoots to be extended laterally at higher and higher intervals on the white spruce trees. A constant regeneration of the forest floor of vegetation is required. Deposits of loess are 1 to 40 feet (0.3 to 12 m) thick and consistently have much more forest vegetation in their upper part than in their lower. This suggests that the vegetation in the upper parts has not yet had a chance to decay and disappear as it has in the lower.

Almost all the loess in the immediate area is pre-Wisconsin in age. Radiocarbon dates available indicate that the Jarvis Ash Bed near the middle of most of the loess sections is 2,000 to 4,600 years old. Two dates at the base of the loess section are approximately 6,000 and 7,000 years.

235.9. From here to Ruby Creek (fig. 5-26) the scarp 60 to 100 feet (18 to 30 m) high on the left was cut by the Delta River in an old alluvial fan of Ruby Creek. Since the glacier of Wisconsin age withdrew from the Delta River valley, streams have built large alluvial fans of gravel into the valley. These fans are truncated as the river wanders from one side of the valley to the other. The fans are capped with a veneer of loess 1 to 40 feet (0.3 to 12 m) thick.

235. Start up younger fan of Ruby Creek.

234.8. Stop 5-35: Yardang Site (fig. 5-27).

The alluvial fan is composed of schist, gneiss, and quartz from the Birch Creek Schist and a large percentage of pebbles and cobbles of coal and orange-brown nonsiliceous siltstone of Tertiary age from the head of Ruby Creek in the Jarvis Creek coal field. The fan perhaps took thousands of years to reach its present size.

Ruby Creek wandered over its gravel fan removing any loess that accumulated. About 6,000 years ago, however, Ruby Creek began to entrench its fan, perhaps because of downcutting of the Delta River or more likely because of the shift of the Delta River to the east side of the valley, nipping the fan and thereby shortening the course of Ruby Creek. For the last 6,000 years, loess has been accumulating on most of the gently sloping fan burying successive generations of white spruce forests. The loess is unconsolidated and possesses crude vertical jointing. It is tannish gray and mottled with iron oxide and organic material. The silt has rather distinct laminations parallel to the surface of the alluvial fan which are caused by the presence of forest layers or iron oxide staining. White spruce stumps up to a maximum diameter of 1 foot (0.3 m) are common in the loess. The organic-rich layers indicate that the loess has been deposited on the forest floors, thereby burying successive forest layers. As much as 16 feet (4.9 m) of loess has been deposited on the fan in the last 6,000 years. A radiocarbon date of 5,999 ± 250 years (I-646) was obtained on a spruce stump 2 inches (5.1 cm) above the base of the loess (fig. 5-27).

Between 2,000 and 4,000 years ago the Jarvis Ash Bed was deposited on the fan and became buried by subsequent loess accumulation. The

ash bed has a relatively uniform thickness of 1 to 5 mm and consists mostly of glass. At the time of ash deposition the Ruby Creek flood plain extended about 500 feet (150 m) north of its present location, and the ash was, therefore, not preserved there. Approximately 2,300 years ago, Ruby Creek moved to the south side of its flood plain, and loess began to accumulate on the old gravel of the inactive creek flood plain. For the last 2,300 years Ruby Creek has not swung to the north, and 10 feet (3 m) of loess has accumulated over the gravel surface and artifacts, burying forests as it accumulated.

Strong wind action in the Delta River valley has attacked the unprotected loess deposit once the vegetation was broken during road construction. From the unconsolidated loess, yardangs 3 to 5 feet (0.9 to 1.5 m) high and 4 to 10 feet (1.2 to 3 m) long have been carved. These features are similar to the yardangs cut out of slightly more consolidated material elsewhere in the world. Because of their poor resistance to erosion most of them are short lived.

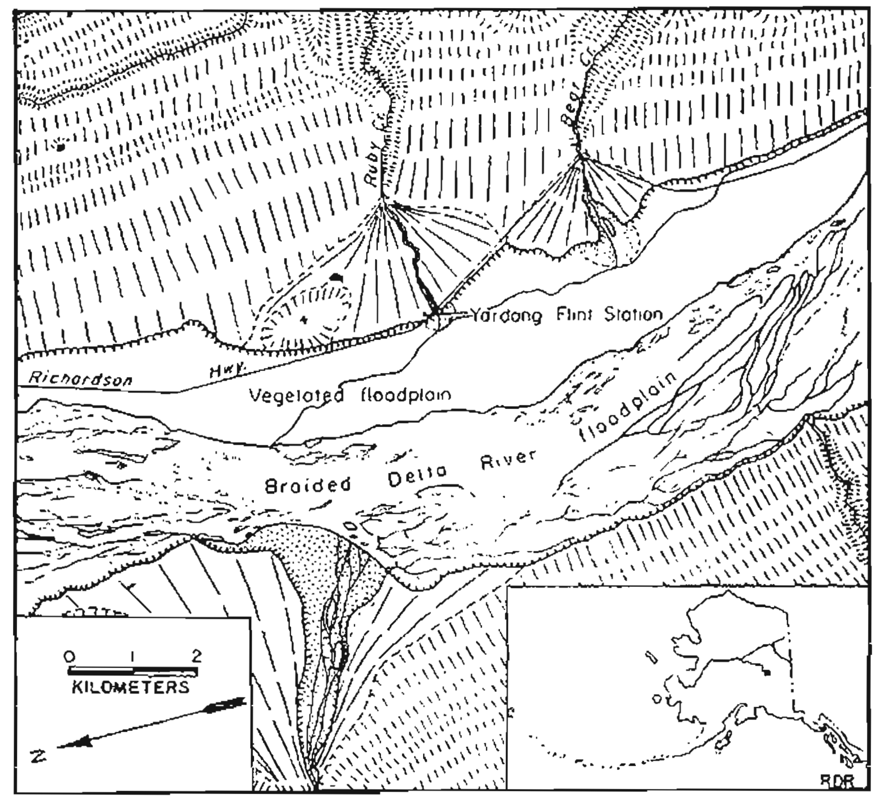

Figure 5-26. Landform map of part of the Delta River Valley in the vicinity of the Yardang Flint Station, central Alaska Range. (From Reger, Péwé, West, and Skarland, 1965, fig. 1)

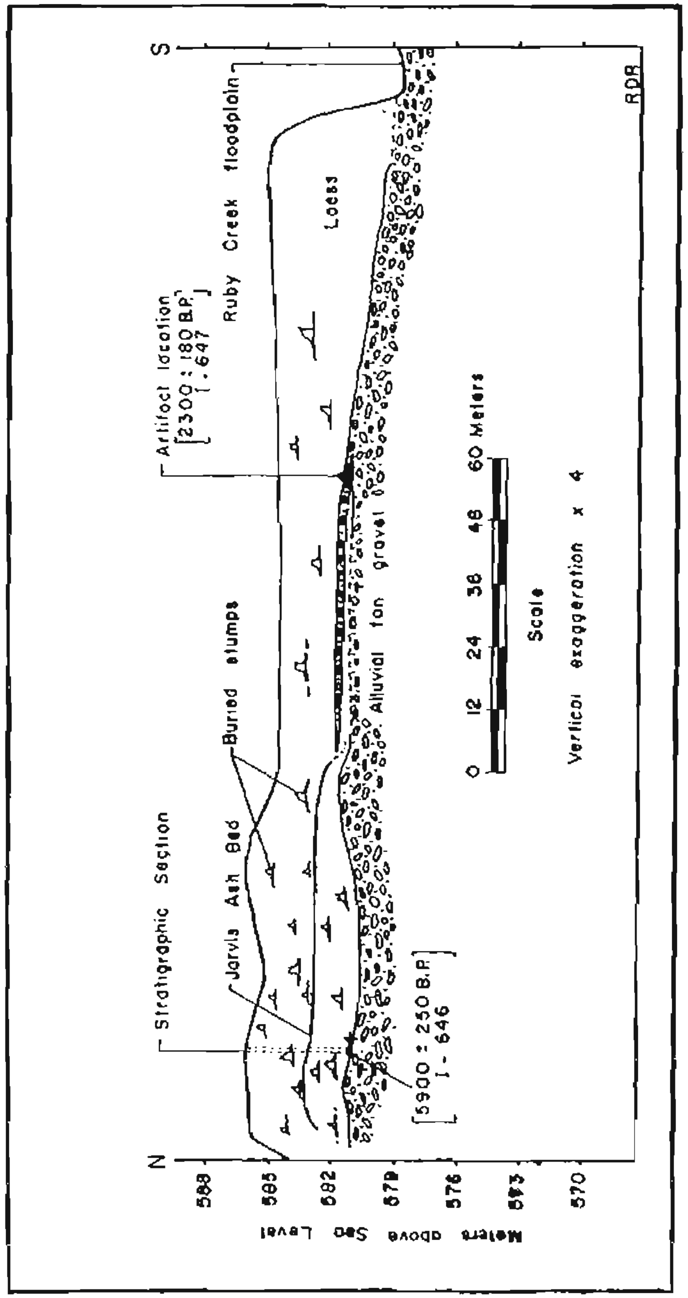

Figure 5-27. Stratigraphy of the Yardang Flint Station along the Richardson Highway in the Delta River Valley, central Alaska Range. (From Reger, Péwé, West, and Skarland, 1965, fig. 3)

Twenty-four artifacts, concentrated in an area 6 inches (15 cm) in diameter were found in an organic-rich layer ¼ inch (6 mm) thick a few inches (several cm) above the gravel-loess contact in the southern part of the section exposed at Stop 5-35. The organic material in the artifact layer was determined by radiocarbon analysis to be 2,300 ± 180 years (I-647) old. The cultural layer at the Yardang Flint Station represents the first location in interior Alaska where artifacts this old are absolutely dated and put into their geologic environment. However, the artifacts were all scrapers, and no blades or points were present. No definite conclusions can be derived from the study of these pieces. It is significant, however, that the people who occupied this temporary site were not of the Athapaskan culture, a culture which is present in the interior today. The Athapaskans made most of their instruments from bones and antlers and did not do the type of flaking exhibited by the artifacts at this locality.

Bridge at Ruby Creek. From the bridge it is possible to see on the left the ridge between Ruby Creek and Bear Creek. At an elevation of 4,000 feet (1,219 m), cobbles of drift of Delta age were found. The lateral moraine of Donnelly age in this area is near 3,000 feet (915 m) elevation, approximately at timberline.

234.7. The next mile of the road is on ice-rich silt overlying the old Ruby Creek fan. Poor road.

234 to **233.5.** Alluvial fan of Bear Creek.

233.8. Bear Creek bridge. Upstream from the bridge one may see a yellowish gravel deposit that was banked against the Donnelly lateral moraine.

231.8. Start of Darling Creek fan.

231. Left, on the ridge between Darling Creek and an unnamed creek to the south at bench mark Darling (4,927 feet [1,502 m] elevation) is the type evidence of the Darling Creek Glaciation. Here deeply stained washed gravel of pre-Illinoian age lies on the Birch Creek Schist. Other deposits of this age can be found at the head of Ruby Creek.

230. Left; outcrop of loess and the Jarvis Ash Bed.

229.9. Left; 1950 forest fire scar.

227. Gunnysack Creek and alluvial-fan exposure. Brownish fan gravel is also exposed across the Delta River under the Recent moraines of Black Rapids Glacier.

226.7. Left; 1650(?) moraine of Black Rapids Glacier.

226.6. Stop 5-46: Falls Creek Black Rapids Glacier.

Black Rapids Glacier gained worldwide publicity in 1937 when it advanced spectacularly up to 200 feet (61 m) per day. The glacier did not reach the Richardson Highway as it was feared. Inspection by the geologists in 1949 revealed that earlier Recent advances did, however, reach across the Delta River valley (fig. 5-28).

Near the mouth of Falls Creek occurs a large pod of greenish fine-grained amphibolite cut by a multitude of quartz veins. This rock is more resistant then the surrounding schist and protrudes as a knob; it displays excellent glacial polish and grooving. Grooving records glacial movement of ice down the Delta River valley (from south to north), probably during the Donnelly Glaciation. From the top of the amphibolite mass a good view is obtained of the glacier and the moraines of historic advances. During the earliest advance the glacier pushed across the valley, riding up a few hundred yards on the opposite bedrock valley wall. Clear Creek, which formerly drained directly into the Delta River, was diverted north by the ice and moraines so it now reaches the river

¾ of a mile (more than a km) farther north. Later shifting of the creek caused the till to be washed away over the amphibolite. The Delta River drained around the front of the ice, cutting a bedrock gorge now occupied by Hidden Lake (fig. 5-28). To the north (across the river) one can see the old alluvial-fan gravel of Gunnysack Creek underlying the Recent moraines. The glacier in its advance pushed the Delta River to the east, causing it to truncate the old fan of Gunnysack Creek, forming the exposure along which the Richardson Highway now lies.

Attempt has been made to date this early advance by various methods—dendrochronology, lichenometry, soil development, and radiocarbon dating, but no exact date is yet available. Advances of comparable age are present on Canwell and Castner Glaciers farther up the Delta River valley.

Trees up to 144 years old, and older trees with rotten centers, are growing on the oldest moraine of Black Rapids Glacier (1951). The timberline conditions, severe winds, and shifting of ice-cored moraines prohibit tree growth for at least 15 or 20 years following formation of the moraines. Trees up to 228 years old occur on the outwash plain in front of the moraine but may have been present prior to the advance. The development of the soil indicates that the advance may be more than 200 years old. A log in a moraine of comparable age of Canwell Glacier nearby was dated as, "less than 200 years" (W-268) in 1953; however, Meyer Rubin (written communication, 1964) states that the log could indeed be older than 200 years. The earliest advance is now thought definitely to be more than 200 years.

A dense forest with 350- to 500-year old trees lies adjacent to the early moraine on the west side of the Delta River. This forest is not of first generation trees and is greatly different in maturity from the young forests on the adjacent moraines. The moraine is certainly younger than 400 to 500 years. Péwé believes that the early advance of Black Rapids Glacier, and nearby Canwell and Castner Glaciers, is about 300 years prior to the 1951, 1954, 1957 tree dating and is an advance of about 1650(?). This early advance is compound on Black Rapids Glacier.

An arcuate terminal moraine lies about a mile in front of the 1937 moraine. This moraine is fresh appearing, has no turf cover, and an ice core is present in places. Trees 98 years old were measured in 1951 on this moraine, and trees 102 years old were measured on moraines of comparable age on Castner Glacier. Some time is needed for tree generation in this rigorous climate near timberline, and it is thought by Péwé that this advance therefore occurred about 1830.

The 1937 advance was rapid, and a 300-foot (91 m) ice cliff formed the terminus in 1937. The ice thinned rapidly, and today little ice can be seen from the highway. The terminal moraine is still ice cored. The first spruce trees (4 years old) were found in 1957, 20 years after the advance, growing on the moraine at a place where the ice core was no longer present.

In an attempt to date recent glacial advances in the central Alaska Range, lichenometry was used in 1962 at Black Rapids, Canwell, Castner, Gulkana, and College Glaciers. Lichenometry is used by measuring the diameter of lichen thalli growing on a surface and is based on the slow but constant increase of the plant diameter. The lichen found most applicable in the central Alaska Range is the crustaceous *Rhizocarpon geographicum*. This lichen is abundant, easily recognized, and is

proved reliable for dating elsewhere in the world. Discussion of the lichenometry will be reserved for Stop 6-41: Gulkana and College Glaciers; however, lichens were measured on the Black Rapids and Castner moraines, moraines that have been dated by dendrochronology.

The 1950(?) moraine of Black Rapids Glacier was unsatisfactory for the lichenometry study inasmuch as the older moraine is heavily for-

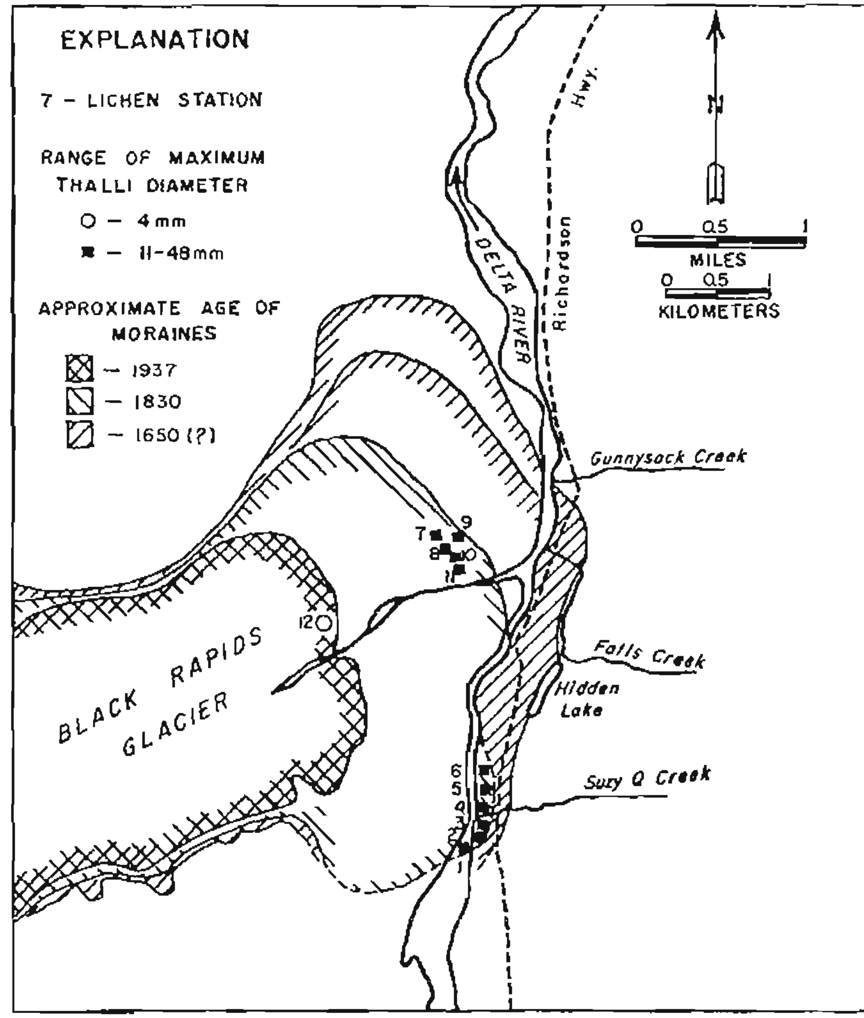

Figure 5-28. Relationship of maximum-diameter *Rhizocarpon geographicum* size ranges to Recent moraines of Black Rapids Glacier, central Alaska Range. (Base after U. S. Army Map Service Mt. Hayes B-4 and C-4, 1951. Scale 1:50,000.) (After Reger, 1964, fig. 5)

ested and elsewhere shows the effects of sand blasting.

The largest *Rhizocarpon geographicum* on the 1830 terminal moraine of Black Rapids Glacier range from 0.43 to 1.69 inches (11 to 43 mm) in diameter and average 0.95 inches (24 mm) (fig. 5-28). Lichens were measured wherever possible on large boulders on that part of the moraine which was not ice cored.

Scattered *Rhizocarpon geographicum* up to 4 mm in diameter were measured on the 1937 terminal moraine, on boulders that were no longer part of the ice-cored moraine.

225.4. Right; glacier overlook. Left; Hidden Lake and diversion channel; *Rhizocarpon geographicum* on boulders in forest.

Enter Mt. Hayes B-4 Quadrangle.

225.1. From 226.5 to 224.4 the road is on the 1650(?) moraine.

224.7. Suzy Q. Creek. The 1830 moraine is a prominent gravel ridge on the right, adjacent to the road. It is an unvegetated gravel pile. Remnants of the 1830 moraine also can be seen across the Delta River valley.

222.9. Right; across (west) the Delta River valley is a hanging, 3-mile (4.8 km) long, U-shaped valley. It is hanging at both ends and was occupied by a glacier in Wisconsin time. View up Delta River valley. First view of Recent moraines of Castner and Canwell Glaciers. Birch Creek Schist of Precambrian and early Paleozoic age forms the valley walls in this area, but Rainbow Mountain in the distance 8 miles (13 km) to the south, just beyond Castner and Canwell Glaciers, is composed of Paleozoic and Mesozoic rocks.

222. Dust can sometimes be seen here being blown from the flood plain of the Delta River.

217.2. Left; Castner Glacier (fig. 5-29). Stagnant ice of the 1830 advance is exposed 3,000 feet (900 m) to east (left) where water discharges from a tunnel under the glacier.

217. The two Recent advances of Castner Glacier, comparable to the 1650(?) and the 1830 advances of Black Rapids Glacier, are recorded by moraines near the highway. A bit of the 1650(?) terminal moraine is plastered on a bedrock hill at Mile 216.7 adjacent to the highway just across lower Miller Creek. The terminal moraine of the 1830 advance lies to the left of the highway at Mile 217 and is ice cored. Trees up to 102 years old were measured in 1951 growing on the ice-cored moraine.

Early stages in revegetation of both the moraine and outwash can be seen at this point. The shrubs are buffalo berry (*Shepherdia canadensis*) and willow (*Salix* spp.). The legumes, *Hedysarum* spp., *Astragalus* spp., and *Oxytropis* spp. occur in abundance as does dwarf fireweed (*Epilobium latifolium*). The scattered grasses are primarily *Festuca rubra* and *Hierochloe alpina*. A few mats of *Dryas* spp. can be observed. Young white spruce have invaded both the outwash and oldest part of the moraine.

216.8. Road crosses the Denali fault. The fault is one of the longest fault zones in Alaska and is expressed topographically as a well-defined arcuate lineament that can be traced without interruption from the southwest part of the Alaska Range, through the crest of the range into Canada, and perhaps into Chatham Strait in southeastern Alaska. The magnitude and direction of displacement are controversial. Offset post-Pleistocene alluvial fans and glacial deposits of Donnelly age occur in the valley across the Delta River, indicating relatively recent movement along this fault. Some of the largest glaciers in the Alaska Range occupy segments of the fault

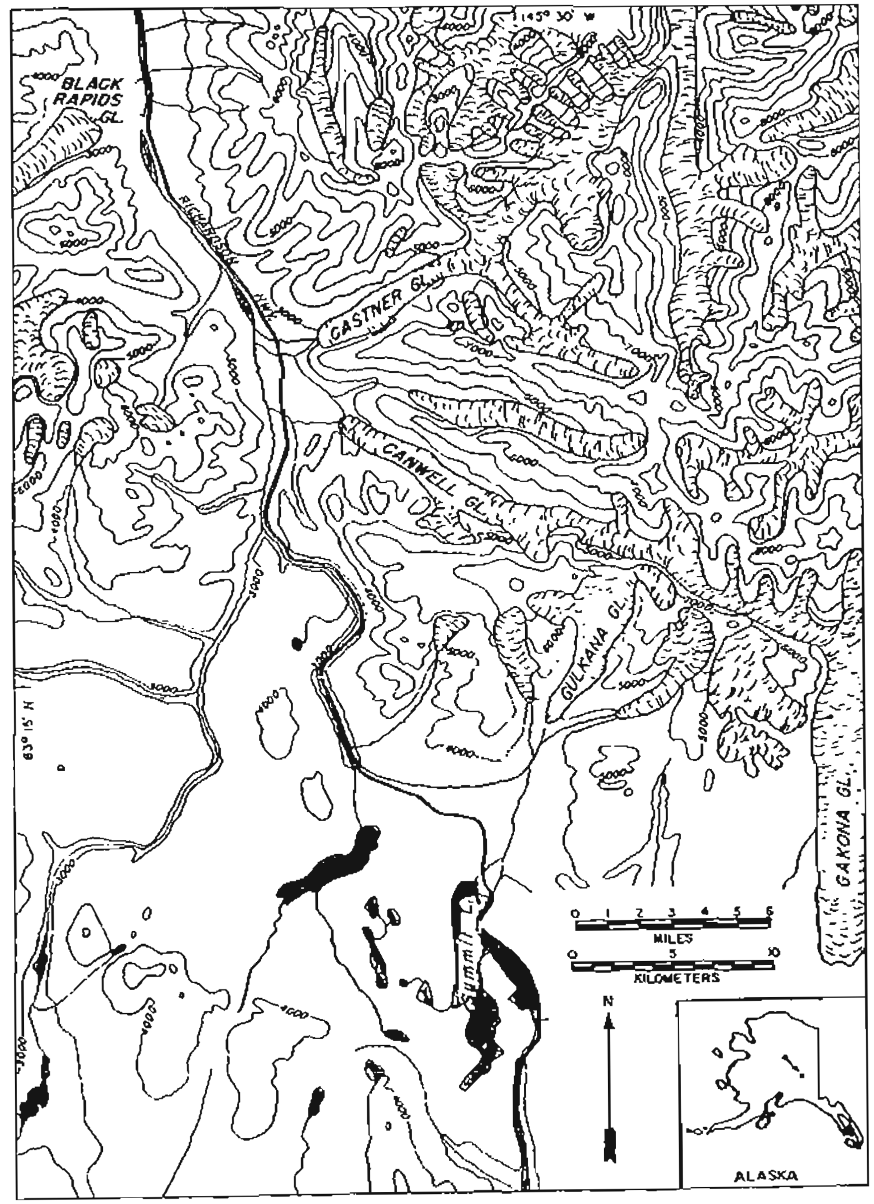

Figure 5-29. Index map of the central Alaska Range showing locations of major glaciers. (Base after U. S. Geol. Survey Mt. Hayes, 1950. Scale 1:250,000). (From Reger, 1964, fig. 2)

trough: Chistochina, Gakona, Canwell, Black Rapids, Susitna, and Muldrow Glaciers.

216.7. Bedrock knob of quartz diorite. Plastered against the rock knob is part of the moraine of the 1650(?) advance of Castner Glacier, a moraine composed of lithologies characteristic of the Birch Creek Schist.

215. Stop 5-37: Left; Canwell Glacier (fig. 5-30).

Canwell Glacier displays well the advances of 1650(?) and 1830. Lateral moraines of the older advance are well preserved up-glacier. A fragment of the older moraine exists at Milepost 215 on the south side of upper Miller Creek (fig. 5-30). Tree stumps up to 159 years old were present in 1951; the trees were probably cut down 10 to 15 years previously. A well-developed forest and turf occurs on this moraine fragment. The largest *Rhizocarpon geographicum* on the south lateral moraine of the older advance range from 5.16 to 6.34 inches (132 to 161 mm) in diameter and average 5.67 inches (144 mm) (fig. 5-30). The thick moss and spruce cover on the fragment of the 1650(?) terminal moraine remaining is unfavorable for lichen growth and none were observed.

The terminal moraine of the 1830 advance is well preserved and displayed near the highway (fig. 5-30). A low prominent gravel ridge 10 to 15 feet (3 to 4.6 m) high is forested with white spruce, willow, and alder. In 1951 the oldest tree recorded was 102 years old. *Rhizocarpon geographicum* is common on the terminal and lateral moraines of this advance. On the south lateral moraine the largest *Rhizocarpon geographicum* are 1.10 to 1.97 inches (28 to 50 mm), but on the terminal moraine they are only .71 to 1.3 inches (18 to 33 mm). The average size for both parts of the moraines is 1.18 inches (30 mm). On this moraine there are few trees, no turf, and lichen-bearing boulders are scant or absent.

214. Left; large area of iron oxide stain on Rainbow Mountain in distance is one of the many caused by weathering of pyrite. Right; at the fork in the large creek directly to the west (right) across the Delta River, is the trace of one of four imbricate thrust faults. This particular thrust, the largest of the four, strikes east-west and dips 34° N. It extends at least 15 miles (2.4 km) to the west and is subparallel to the Denali fault (oral communication, J. H. Stout, 1965).

212.1. Left; rock quarry. Exposed are flows and agglomerates of Pennsylvanian age.

211.95. Left; fossils of late Paleozoic age exposed in steeply dipping limestone at spot painted "USBM." According to Dr. C. L. Rowett, invertebrate paleontologist at the University of Alaska (oral communication), the beds here are early Pennsylvanian (pre-Desmoinesian) in age. Some of the fossils that occur here are:

Aulophyllid coral, genus undet.
Cladochonus cf. *C. texasensis* Moore and Jeffords
Composita sp.
Linoproductus (sensu stricto) sp.
Echinoconchus sp.
Spirifer cf. *S. rockymontanus* Marcou
Spirifer spp.
Aviculopecten sp.
Pseudoparalegoceras n. sp.
Phillipsid trilobite, genus undet.

210.7. Left; the "Green Thumb": active solifluction slope. No trees grow on this slope. The road cannot be satisfactorily maintained where it crosses the solifluction slope.

210. Stop 5-38: Rainbow Mountain.

Left; the till (bedded?) fill indicates that the rock gorge was here prior to the Donnelly Glaciation. Rainbow Mountain is essentially a

strike ridge tending northwest. It rises abruptly more than 4,200 feet (1,280 m) above the valley floor of the Delta River. Much of the western slope of the ridge approximates a dip slope. The mountain is composed of folded and faulted detrital rocks: "graywackes," limestones, andesitic and dacitic pyroclastic rocks and minor andesite flows of Mississip-

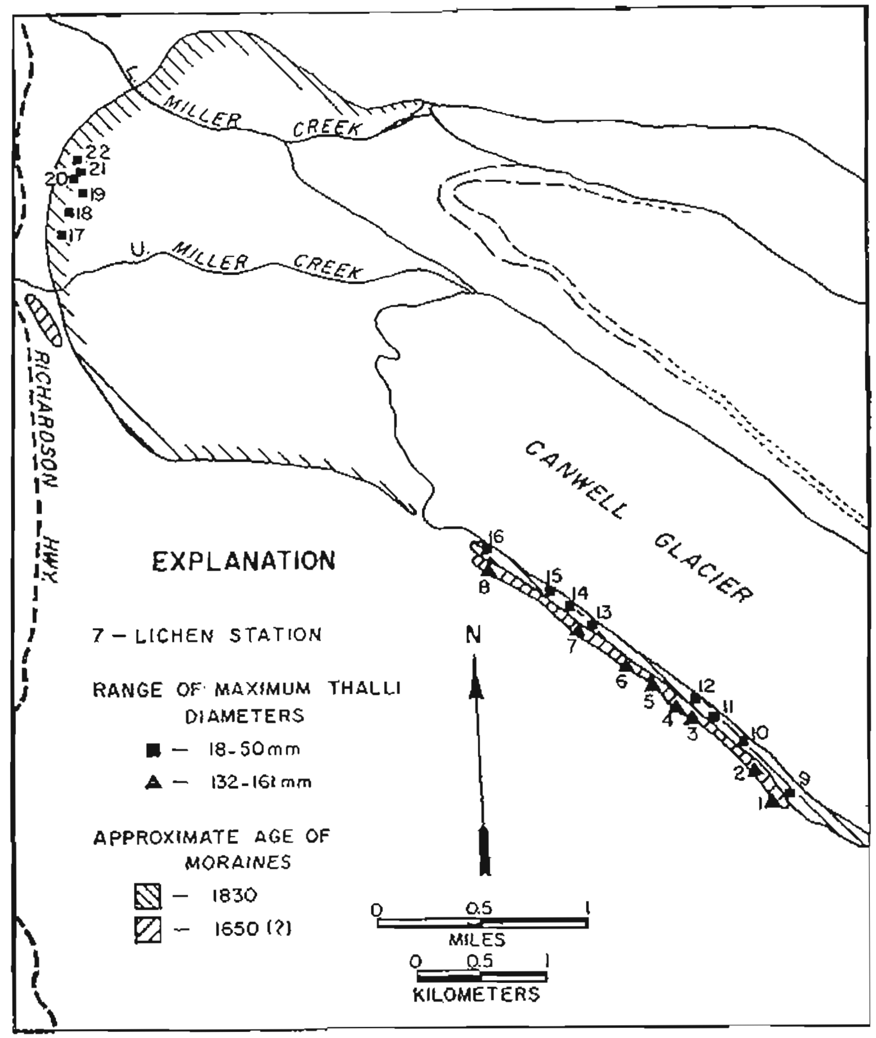

Figure 5-30. Relationship of maximum-diameter *Rhizocarpon geographicum* size ranges to Recent moraines of Canwell Glacier, central Alaska Range. (Base after U. S. Geol. Survey Mt. Hayes B-4. Scale 1:40,000.) (From Reger, 1964, fig. 4)

pian(?) and Pennsylvanian age which have been intruded by andesite, granodiorite, and quartz diorite masses.

The green and maroon colors are predominantly those associated with the volcanic rocks, but the yellow-greens are associated with the siltstones and sandstones. The porphyritic andesite sill-like intrusions are characteristically purplish green and the rhyolites are dark green.

209. Left; stabilized scree slope deposit.

207.7. Stop 5-39: Left; Active Rock Glacier.

It is about a mile (1.6 km) long and originates in empty cirque at top of Rainbow Mountain. Right; Tertiary rock exposed in lower part of valley wall across Phelan Creek.

From Rainbow Mountain to Isabelle Pass the road travels upward through the altitudinal tree line. As the upper limit of trees is approached the trees become scattered in distribution, but in most cases are upright and relatively fast growing. Open stands of white spruce can be seen along the terraces of Phelan Creek to an elevation of 3,000 feet (900 m), but scattered individuals occur at elevations of 3,500 feet (1,100 m) in this area. No old stumps are found above the present tree line, and many of the trees at or near tree line are less than 100 years of age. This indicates that tree line is now as high as it has been in recent times and that it may actually be advancing in some localities.

At tree line and above is a zone of shrubs. This may be narrow on steep slopes, as in the area around Rainbow Mountain, or very extensive, as in the Isabelle Pass-Summit Lake area. This zone consists primarily of the same plants that are found in the open spruce stands at lower elevations. These include the shrubs, dwarf birch, willows, Labrador tea, and blueberry. Scattered through the shrubs is a moss mat of *Hylocomium splendens*. Between the tall shrubs and on exposed ridges are mats of *Cladonia* lichens, crowberry, alpine bearberry, and mountain cranberry (*Vaccinium vitis-idaea*). Because of the abundance of lichens, this vegetation zone is utilized heavily by caribou in the winter—small herds can usually be seen in the winter months in the Isabelle Pass-Summit Lake area feeding on the lichens.

206.8. Right; boulder bed across Phelan Creek. This boulder bed appears to be a 4-foot (1.2 m) thick layer of outwash gravel deposited on a surface cut on Tertiary rock. A large slump block, talus deposit, or rock glacier up to 100 feet (30 m) thick from Rainbow Mountain overlies the outwash. The outwash gravel appears to be a continuation of the outwash gravel on the terrace surface directly ahead, on this side of Phelan Creek. If this interpretation is correct, the "slump block" is post-Donnelly in age.

Left; the contact between the Paleozoic and Tertiary rocks. Directly ahead, across Phelan Creek, at an altitude of 4,000 feet (1,219 m), till of Donnelly age occurs on top of the low hill. (Road from 206.8 to 204.7 is now on flood plain of Phelan Creek, not on terrace as shown on 1958 U.S. Geological Survey Mt. Hayes B-4 topographic map).

206.2. Left; alluvial fan is burying the terrace surface in the distance. The terrace surface is thought to be of late Donnelly (Wisconsin) or early post-Donnelly age. The surface has large-scale polygonal ground similar to that exposed near Donnelly Dome, Milepost 251.2.

205. Left; alluvial fan burying the surface of two terraces. The terrace remnants may be matched with terrace remnants on other side of Phelan Valley. The gradient of these terraces is less than that of the present Phelan Creek.

204.7. Enter Mt. Hayes A-4 Quadrangle.

204. Left; Tertiary rocks, coal-bearing, underlie glacial drift and are subject to landsliding.

203. Lower terrace surface.

203.3. Higher terrace surface.

200.1. Edge of terrace composed of outwash gravel. Road ascends terrace and the terrace is graded to a terminal moraine. The terminal moraine lies in Isabelle Pass and perhaps is a moraine of the Summit Lake advance.

199. Isabelle Pass. Elevation 3,285 feet (991.9 m) above sea level.

198. Moraine of Summit Lake(?) advance.

197.7. Stop 6-40: Gulkana Glacier View.

Richardson Monument. Pingo-like knob in the floor of the Gulkana River valley is a moraine remnant. Surface drainage from Gulkana Glacier wanders back and forth on the outwash plain and sometimes it drains into the Delta-Tanana-Yukon Rivers to the Bering Sea, sometimes into the Copper River and to the Pacific Ocean to the south, and sometimes it goes in both directions. For the last few years the drainage has been artificially diverted into the Delta River.

The Richardson Highway, connecting Valdez and Fairbanks, is the oldest major transportation route in Alaska. It was initiated as the Fairbanks-Valdez Trail in 1901 when Congress appropriated funds for its construction. In 1904 funds were appropriated to survey a wagon road on this route. It was mainly a winter sled and wagon road in the early years, and the first complete truck and auto trip is reported to have been made in 1913. In the 1920's the road was named the Richardson Highway after W. P. Richardson, President, Alaska Road Commission, from 1905 to 1917.

Stop 6-41: Gulkana Glacier.

Gulkana Glacier has been intensively studied since 1960 under a program of integrated investigations directed by the Department of Geology, University of Alaska. Gulkana Glacier is one of the several temperate valley glaciers on the south side of the Alaska Range (fig. 5-29). It lies 4 miles (6.4 km) east of the Richardson Highway at Isabelle Pass, and a gravel road branching northeast from the Richardson Monument at Isabelle Pass extends to within 1 mile (1.6 km) of the present terminus of Gulkana Glacier. College Glacier lies on the east side and West Gulkana Glacier lies on the west side of Gulkana Glacier (fig. 6-31). Gulkana Glacier originates in three adjacent compound cirques at an average elevation of 6,500 feet (2,000 m) above sea level. Ice from these three cirques converges to form a south-flowing valley glacier which extends 2.5 miles (4 km) to an elevation of approximately 3,700 feet (1,140 m) at its terminus. A sequence of arcuate end and parallel lateral Recent moraines encircle the terminal areas of Gulkana, West Gulkana, and College Glaciers (fig. 6-31).

Glacial Geology

The recent moraines which flank the terminus and lower parts of Gulkana and College Glaciers lie above tree line, and therefore cannot be dated by dendrochronology. However, dating has been possible by applying lichenometric information obtained from Recent moraines of Black Rapids and Canwell Glaciers discussed at Stops 5-36 and 5-37.

Lichenometry is a method by which one can date recently exposed rock surfaces or recently active geologic processes in treeless areas by measuring the rate of lichen growth. The dating is based on the slow, but constant increase of the individual plant diameters. The length of life

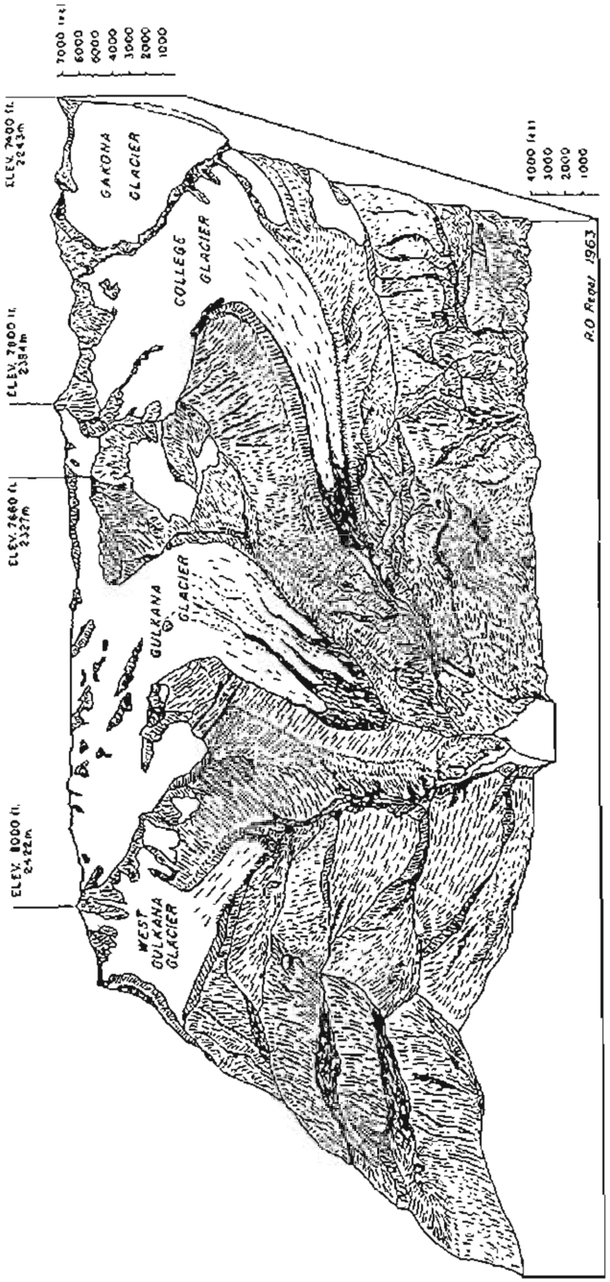

Figure 6-31. Block digram of the Gulkana-College Glacier area, central Alaska Range, Alaska. (From Reger, 1964)

of the lichen species used imposes a limit on the ages of the exposed rock surfaces which can be determined by lichenometry. Crustaceous lichens grow extremely slowly, and the lichen species which has been used for dating in other parts of the world is *Rhizocarpon geographicum*.

To date the moraine of Gulkana and College Glaciers it is necessary to determine thalli growth rates for the central Alaska Range before lichenometry can be used for dating in this region. To establish growth rates, lichen diameters were measured on the Recent moraines of Canwell and Black Rapids, moraines which have been dated by dendrochronology as having formed about 1830 and 1650(?) (figs. 5-28 and 5-30) (see Stops 5-36 and 5-37). At least ten of the largest lichens were measured at each established station in order to determine a lichen growth standard for the central Alaska Range. Lichen was measured on the larger boulders because these were considered more stable than small boulders. The effect of varying lithology was minimized by measuring lichens on diorite or quartz diorite whenever possible.

The average diameter of *Rhizocarpon geographicum* on the Canwell Glacier 1650(?) moraine is 5.67 inches (144 mm). On the 1830 moraine the average diameter is 1.18 inches (30 mm). The average diameter was .94 inches (24 mm) on the 1830 moraine of Black Rapids Glacier. Good measuring sites on the 1650(?) moraine of Black Rapids Glacier have not yet been found. Lichen diameters on the 1830 moraines of both Canwell and Black Rapids Glaciers are well established and compare favorably in size.

Gulkana Glacier has four Recent moraines of which two are considered to correlate with the 1650(?) and 1830 moraines, respectively, of Canwell Glacier because of their position and topographic expression. Using these assumed dates and measuring maximum thalli diameters of *Rhizocarpon geographicum* on these two prominent moraines, a growth rate curve was constructed and compared with the growth rate curve of Canwell Glacier. The curve for *Rhizocarpon geographicum* in the Gulkana-College Glacier area is almost identical with the growth rate curve of this lichen at Canwell Glacier, which supports the basic assumption that the prominent moraines of Gulkana and College Glaciers are equivalent to the 1650(?) and 1830 moraines of Canwell Glacier. Once the growth rate curve for the Gulkana-College Glacier area was established it was possible to date the two Recent moraines in the Gulkana-College Glacier area that apparently have no representation on Canwell Glacier. There moraines were determined to date from approximately 1580(?) and 1875 (fig. 6-32). The average diameter of the lichen on moraines of the 1580(?) advance is 6.7 inches (170 mm) and that of the 1875 advance is 0.39 inches (10 mm).

The oldest Recent advance of Gulkana Glacier for which moraines exist is thought to have occurred about 1580(?). Although buried moraines(?) occur adjacent to Gulkana and College Glaciers and may indicate Recent advances prior to 1580, no good evidence is present. The 1580(?) advance is best documented on the outer part of the east Recent moraine complex of Gulkana Glacier (fig. 6-32). This moraine is flat topped and has incipient stone rings. The boulders are heavily covered with lichen having an average diameter of 6.7 inches (170 mm). The 1580(?) advance of Gulkana Glacier has no known equivalent on College Glacier.

Gulkana Glacier readvanced in 1650(?) and overrode the 1580(?)

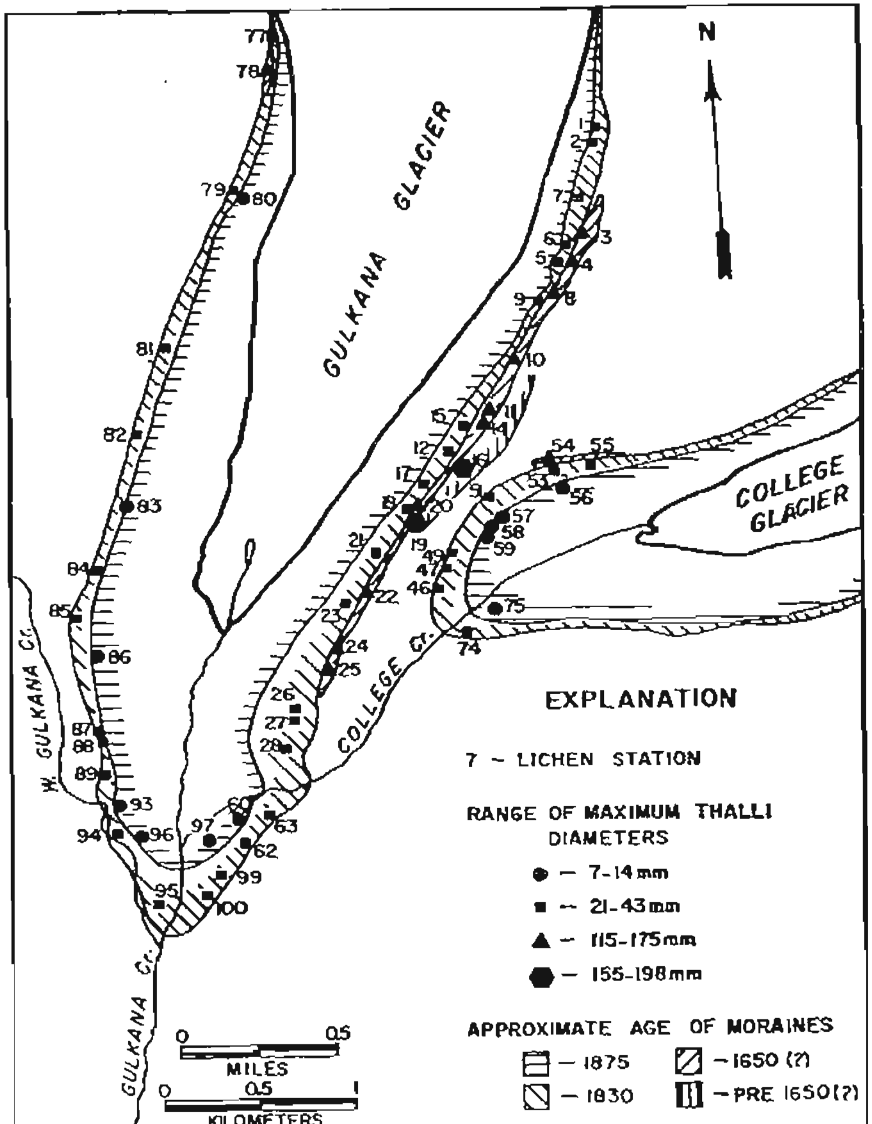

Figure 6-32. Relationship of maximum-diameter *Rhizocarpon geographicum* size ranges to Recent moraines of Gulkana and College Glaciers, central Alaska Range. (Base after air photography by U. S. Navy, 6 August 57, identification no. 004 VVAP-61 DET "T" USN 10/c.) (From Reger, 1964, fig. 7)

terminal moraine. The most prominent 1650(?) moraine in the area is preserved in the east lateral moraine complex of Gulkana Glacier as a narrow continuous ridge. Part of the 1650(?) lateral moraine is also preserved as a small linear patch of till outside the northwest 1830 lateral moraine of Gulkana Glacier (fig. 6-32). Boulders are fresher and less heavily lichen covered than on the 1580(?) moraine. This moraine, as the older, is not ice-cored, and the average diameter of *Rhizocarpon geographicum* is 5.39 inches (137 mm). The exact extent of the 1650(?) advance of Gulkana Glacier is unknown, but it probably was only slightly less than the advance of 1830. Advances of many glaciers in the Alaska Range occurred at this time, leaving comparable moraines.

In 1830 Gulkana and College Glaciers advanced. Gulkana Glacier advanced 1.5 miles (2.5 km) beyond the 1962 ice terminus and overran the 1650(?) terminal moraine, spilling out onto the relatively flat Wisconsin glacial floor. The advance dammed College Creek and partly blocked west Gulkana Creek (fig. 6-32). Lateral moraines of the 1830 advance of Gulkana and College Glaciers are well preserved, but the terminal moraines are fragmentary.

Boulders on the surface of the moraines of this age appear light greenish black because of the cover of lichen. Lateral moraines support no vegetation other than lichen, but the 1830 terminal moraine has willows up to 10 feet (3 m) high growing on it. There is a small ice core in the 1830 moraines as indicated by fresh slumpage features on lateral moraines above an altitude of 4,400 feet (1,320 m).

The major drainage of College Glacier in 1830 was down College Creek Canyon as indicated by terraces and terrace remnants on the canyon walls 5.9 to 7.9 feet (18 to 24 mm) above the present flood plain. This drainage was blocked by Gulkana Glacier, and when the water of the resulting ice-dammed lake reached an elevation of 3,740 feet (1,140 m) the lake and subsequent runoff drained toward the south along the edge of the 1830 terminus of Gulkana Glacier cutting an overflow channel. The 1830 advances of Gulkana and College Glaciers have equivalent advances on many glaciers in the central Alaska Range, as well as glaciers in other parts of Alaska and the world.

About 1875, Gulkana Glacier advanced to a position 1.4 miles (2.2 km) beyond the present terminus, again blocking College Creek and depositing a distinct terminal moraine (fig. 6-32). In the west terminal area of Gulkana Glacier the 1875 terminal moraine occurs as a distinct steep-sided rather arcuate ridge of till about 8.2 feet (2.5 m) high which is plastered on 1830 till. The east 1875 terminal moraine is more massive than in the west terminal area, and its position is brought out by the location of the overflow channel which is cut along the 1875 ice front of Gulkana Glacier. Lateral moraines of the 1875 advance are much less conspicuous than those of the earlier advances. The boulders on these moraines are light gray to black and the lichen cover is thin.

As in 1830, the 1875 advance blocked College Creek, forming a temporary lake. Waters of the ice-dammed lake rose rather rapidly and spilled along the 1875 terminus of Gulkana Glacier cutting a 66-foot (20 m) deep overflow channel in the 1830 till.

Since the culmination of the 1875 advance of Gulkana Glacier the ice surface has lowered and the terminus has retreated until the present. The earliest known photographs of Gulkana Glacier were taken by Moffit of the U.S. Geological Survey in

1910; at the time the ice was still near the 1875 moraine.

Glaciology

Detailed studies made on Gulkana Glacier include meteorology, mass budget, surface flow, sub-glacial topography, and ice structures. The studies were begun in 1960 and continue to the present. The glacier is shaped roughly like a "T", with the horizontal bar representing a complex accumulation zone, and the vertical bar representing the south flowing ablation zone. Accumulation in three main cirques produces three ice streams, 1, 2, and 3 (fig. 6-33), delineated by medial moraines. Ice stream 3 is cleaved by a bedrock bastion into two unequal halves, 3a and 3b. The area of the glacier is 7.7 $miles^2$ (19.9 km^2); the ablation zone being 40% of this area. Two large icefalls are present, Moore Icefall is in the eastern cirque, and Gabriel Icefall is just below the firnline of the western ice stream (fig. 6-33).

The weather on Gulkana Glacier, and presumably along much of the southern flank of the Alaska Range, has the same pattern summer and winter, with only the temperature changing. During all seasons, cold airmasses form in the region when the general airmass movement is stagnant or moving from the north. The humidity is low, and if the sun is high, daily temperatures are high. Occasional violent windstorms of short duration occur during clear weather. When airmass movement is from the south, temperatures are generally from 15° to 40°F (−10° to +5°C) summer or winter, the humidity is high, and precipitation is likely. Superimposed on the regional airmass movement are katabatic drainage winds down the glacier. The average annual precipitation at Paxson, just south of the glacier, is about 16 inches (41 cm) water equivalent. The amount of precipitation increases rapidly with elevation up the Alaska Range to a three year average of more than 120 inches (300 cm) water equivalent per year near the head of the glacier, an eight-fold increase.

The large amounts of precipitation at the head of the glacier, accompanied by little ablation, caused about 110 inches (2.7 m) water equivalent of snow to accumulate during 1959 and 1960 at an elevation of 6900 feet (2100 m); 136 inches (3.5 m) accumulated at the same place in 1961. The 1961 net accumulation consisted of both winter and summer snow (fig. 6-34). The firnline was at about 5700 feet (1700 m) elevation during 1960 and 1961. At the terminus of the glacier, about 160 inches (4.1 m) water equivalent of glacial ice ablated during both 1960 and 1961. The budget gradient thus defined averages 0.09 inches change of ablation or accumulation per foot elevation (7.3 mm/m).

The mass budget, or regimen, of Gulkana Glacier is, in general, negative. Comparison of photographs taken in 1910 and in 1960 shows that the glacier thinned appreciably and the terminus retreated 1.1 miles (1.8 km), with a loss of $204x10^8$ cubic feet ($577x10^6m^3$) water equivalent of ice. The glacial budget was studied most carefully during 1961. That year $3.8x10^8ft^3$ ($10.7x10^6m^3$) water equivalent of ice ablated and $8.6x10^8ft^3$ ($24.5x10^6m^3$) accumulated, which is a strongly positive mass budget. During 1960, $4.2x10^8ft^3$ ($12x10^6m^3$) ablated, and using data from only one snowpit, it is estimated that $5.3x10^8ft^3$ ($15x10^6m^3$) accumulated, which is near equilibrium or a slightly positive mass budget. The 1959 accumulation was the same as 1960 at that snowpit. This one piece of data suggests that the glacier was near equilibrium in 1959 also. Since 1961, the glacier is thought to have

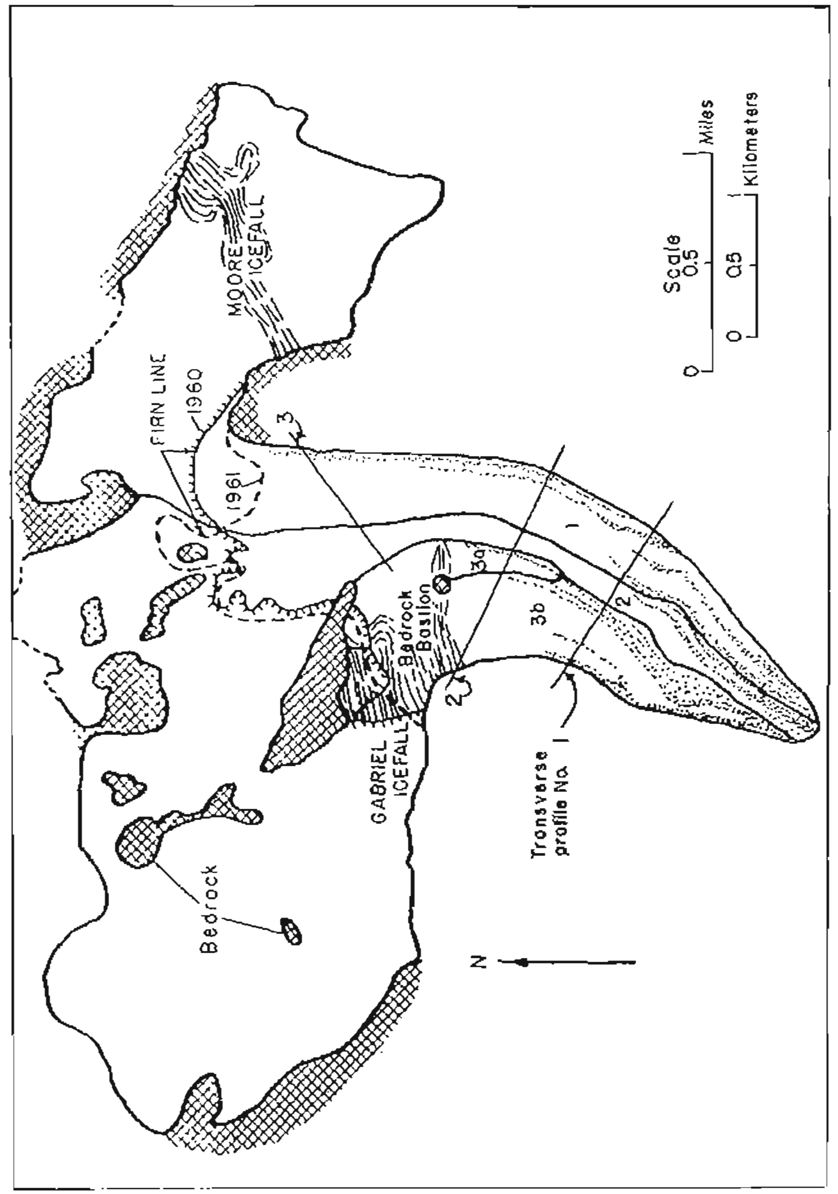

Figure 6-33. Map of Gulkana Glacier, central Alaska Range, showing accumulation area, ablation area, principal ice streams, and transverse profiles along which motion and ice thickness surveys were made. (After Moores, 1962, fig. 2)

had a negative mass budget because the snowline has been much higher than in either 1960 or 1961. The anomalous positive regimen of 1961 was primarily due to an abnormal number of summer snowstorms. During the ablation season, 11 snowstorms occurred, halting ablation of glacial ice for considerable periods and also causing net accumulation of snow near the head of the glacier.

In 1961 a study was undertaken to determine the relative importance of the various means by which energy is transmitted to the glacial ice to cause melting. The following energy exchange processes are involved: radiation, condensation, evaporation, and conduction from precipitation and the atmosphere. Solar radiation is commonly accepted to be the most important single energy source for ablation in many areas and a study of solar radiation on Gulkana Glacier at an elevation of 4,800 feet (1,465 m) was undertaken. A net total radiometer continuously recorded the radiation flux throughout the season. During the 53 24-hour periods with reliable radiation data, a total of 8,300 g cal cm^{-2} of heat were received by the surface of the glacier. During the same interval 51.4 inches (131 cm) water equivalent of ice and snow melted, a process requiring 10,400 g cal cm^{-2} of heat. Radiation received was equal to 75% of the energy required to melt the glacial ice and apparently 89% of the energy required to ablate the snow (fig. 6-35).

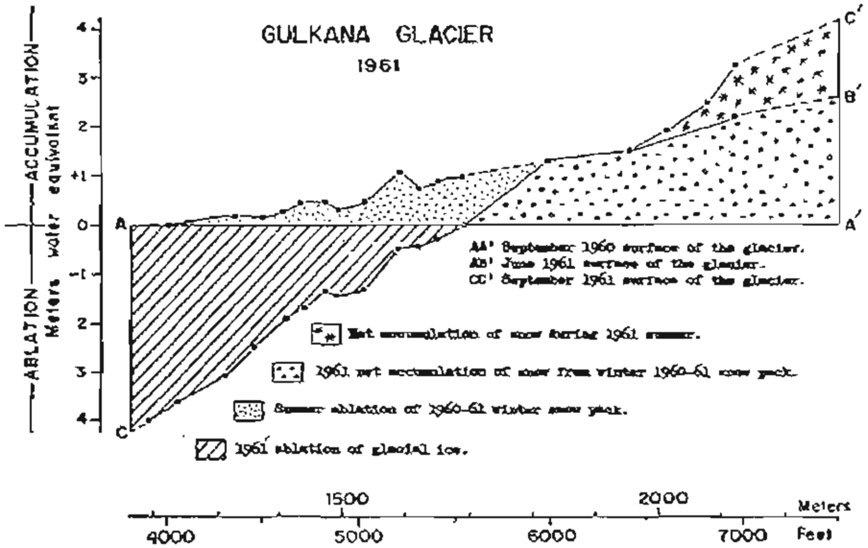

Figure 6-34. The 1961 net budget gradient of Gulkana Glacier. The details of the accumulation and ablation of the snow, and the ablation of the glacial ice are shown. The distribution of winter snow in the ablation zone of the glacier is reflected well by the amounts of ablation of glacial ice. (From Mayo, 1963, fig. 12)

The surface motion in the ablation zone of Gulkana Glacier was studied in 1960 and 1961. On either side of the exposed bedrock bastion, the glacier flows as two ice streams. Below the bastion, these ice streams retain their identity as "velocity ice streams," each with its own axis of maximum velocity (fig. 6-36). These "velocity ice streams" merge near the terminus to form one stream. The velocity measured in the eastern ice stream is greater than 125 feet per year (41 m yr^{-1}) near the firn-line. The velocity decreases downglacier to zero, stagnant ice, near the terminus.

A gravity survey done in 1961 defined the bottom topography under Gulkana Glacier along profiles 1, 2, and 3 (fig. 6-36). A simple "U" shaped valley exists under the eastern ice stream at profile 3. Below the bedrock bastion, however, a medial ridge was found to extend part way down the valley. No ridge exists at the present terminus. Thus, each ice stream flows in its own well defined sub-glacial channel, and the reason for two "velocity ice streams" in the one glacial tongue is apparent.

The surface velocity of Gulkana Glacier varies not only with location, as seen above, but also with time; and the two "velocity ice streams" behave somewhat independently. The velocity varies annually, the summer velocity being locally as much as twice as fast as the winter velocity. This difference is more pronounced at profile 2 than at 1. The "velocity ice streams" are also identifiable farther down-glacier in the summer than in the winter. The increased summer velocity is thought to be caused by meltwater lubrica-

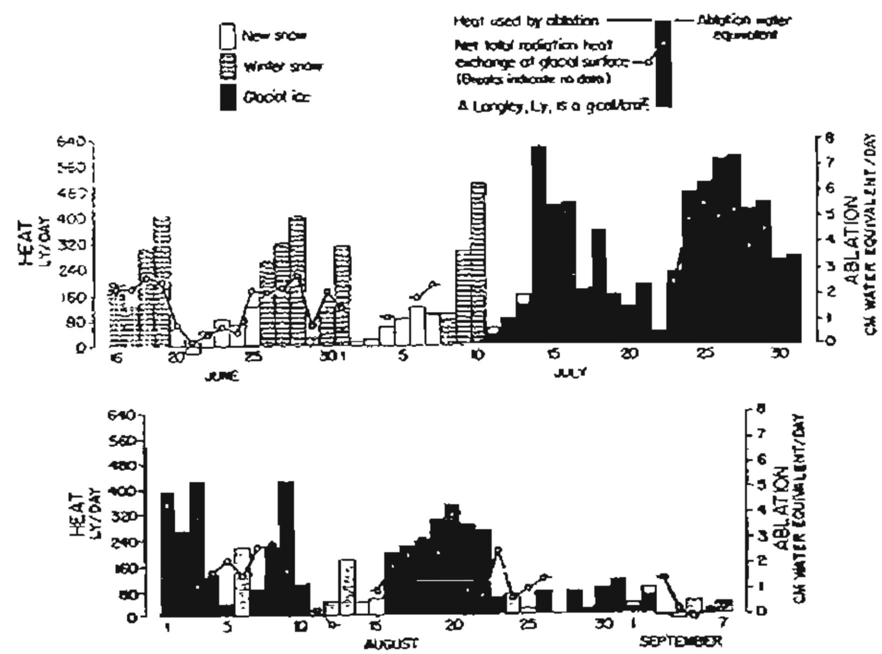

Figure 6-35. Net total radiation and ablation of snow and ice, Gulkana Glacier, Alaska, 1961. (From Mayo and Péwé, 1963, fig. 2)

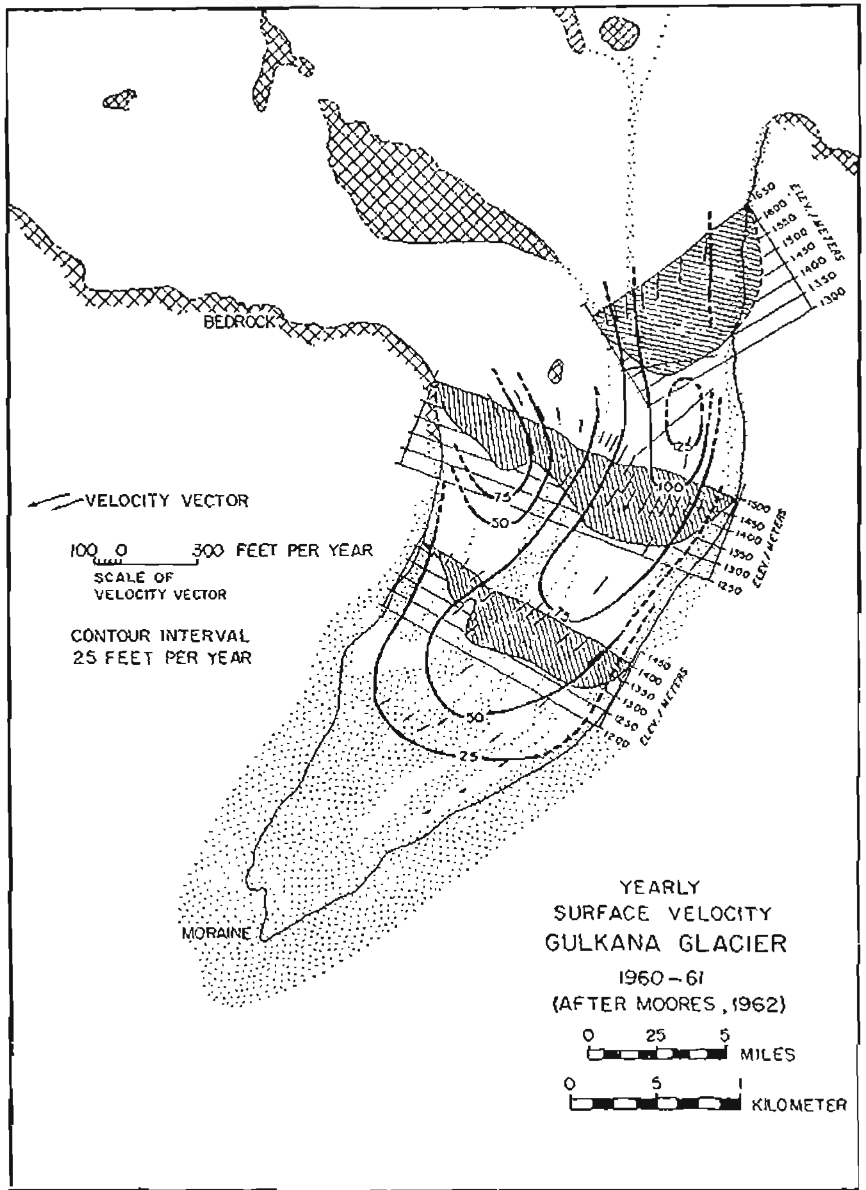

Figure 6-36. Yearly surface velocity (1960-61) and subsurface topography, Gulkana Glacier, central Alaska Range. (From Ostenso, Sellmann, and Péwé, fig. 3)

tion at the base of the glacier. For very short time intervals, one month, in 1961, the western ice stream had its maximum velocity in July above the icefall and in August below the icefall. At profile 2 (fig. 6-33) in 1961, the eastern ice stream reached its maximum velocity in July, the month with greatest surface melting.

The vertical component of surface motion was combined with the ablation data to define the change of thickness of Gulkana Glacier for the 1961 budget year. The eastern "velocity ice stream" thickened and thinned somewhat irregularly with a maximum change of 7 feet (2.1 m) thicker. The western "velocity ice stream" became as much as 16 feet (5.0 m) thicker immediately below Gabriel Icefall. The terminal area of the glacier, which is stagnant, thinned the same amount as the ablation, 15 feet (4.6 m). The thickening in the upper part of the ablation zone far exceeded local thinning, and this is thought to be due to the change of the mass budget from negative before 1960 to positive during 1961.

Gulkana Glacier has a well developed foliation pattern (fig. 6-37) which is displayed by alternating layers of bubble rich and bubble free ice. In the western "velocity ice stream," the foliation develops at the base of Gabriel Icefall as a series of small, arcuate structures. By all gradations, these merge into the larger arcs which occur in the entire ice stream (oral communication, Dr. Donal R. Ragan, University of Alaska). In the eastern ice stream, the pattern is formed above the firnline. The foliation pattern is strikingly parallel to the contours of surface velocity (fig. 6-36), and outlines well the two sub-glacial channels.

195. Stop 6-42: Summit Lake.

To the west and south of Summit Lake lie the Amphitheater Mountains, small knobs mostly of metabasalt protruding 2,000 to 3,000 feet (600 to 900 m) above the lowland which is covered with glacial drift. The lowland in large part is underlain with Tertiary sand and gravel and some coal beds. The Denali Highway traverses the Amphitheater Mountain area from Paxson to the Maclaren River. Glaciers originating on the south side of the Alaska Range pushed over and through these mountains.

Paxson Mountain is an elongate 2,000-foot (600 m) high knob of metabasalt that can be seen clearly to the south from Summit Lake. The overflow gorges, as well as the drift cover on the mountain, record a complex glacial history, a history typical of the Amphitheater Mountains. Isolated erratics on top of Paxson Mountain are thought to be early to middle Quaternary in age. A large prominent weathered overflow gorge or escape notch was cut across the crest of the mountain probably at this time. This gorge is easily visible in the skyline of the mountain. The next glacial advance (middle to late Quaternary age), left escape gorges at lower levels on Paxson Mountain, gorges that are well filled with frost-rived rubble. The Wisconsin advances surrounded Paxson Mountain. Fresh escape notches exist at an altitude of 3,800 to 4,000 feet (1,160 to 1,220 m) and young drift flanks the mountain below this altitude (fig. 6-38). Cirques on Paxson Mountain appear to be of two ages, Wisconsin and pre-Wisconsin.

191. Fish Creek. Here in July or August may be seen a run of bright red salmon which have migrated up from the sea to spawn nearby.

190.5. Right; artificial cut-off of meander. Tertiary sediments are exposed in the bottom of the cut; claystone with unidentifiable plant fragments.

190.5 to 185.6. Paxson Canyon is cut in stratified and unstratified

Figure 6-37. Ice thickness and foliation pattern, Gulkana Glacier, central Alaska Range. (From Ostenso, Sellmann, and Péwé, fig. 4)

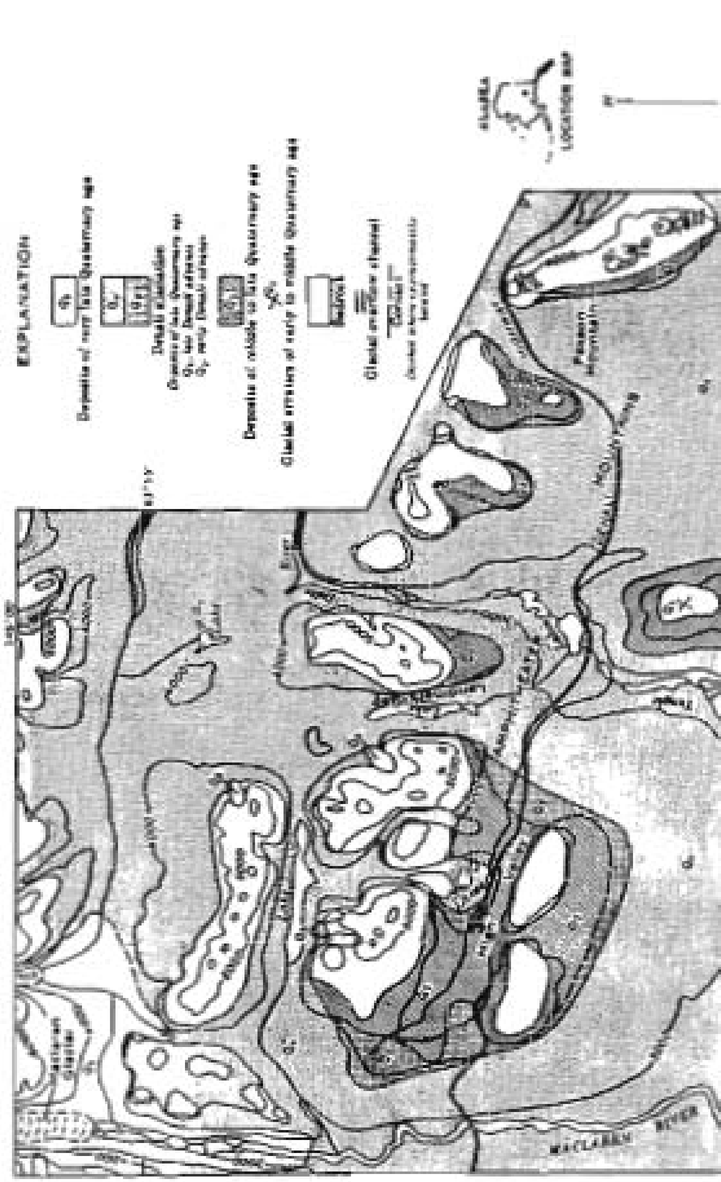

Figure 6-38. Glacial deposits in the headwaters area of the Delta River, Amphitheater Mountains, Alaska. See also figure 7-1. (From Péwé, 1961, fig. 357)

drift. An esker lies on top of right canyon wall.

193. From Milepost 193 to Paxson the highway is partly in Mt. Hayes A-3 Quadrangle and partly in Mt. Hayes A-4 Quadrangle.

185.4. Junction of Denali Highway.

Denali Highway

Mt. Hayes A-4 Quadrangle

0.3[1] Right; Tertiary rocks crop out at base of the cliff. Left; Paxson Mountain.

2.5. Ice-contact deposits of Summit Lake advance. These deposits hold in Summit Lake.

4.1. Right; ice-contact deposits. Paxson Mountain directly ahead with excellent view of escape gorge for overflow water.

The distribution of snow in winter is important in determining the pattern of vegetation in the rough ice-contact deposits. Pockets of grasses (mostly *Festuca altaica*) and sedges (*Carex* spp.) are found where the snow collects on the leeward side of ridges and in depressions. Associated with these snowbed areas are loose mats of lichens, primarily *Peltigara* spp., *Cladonia* spp., and *Stereocaulon* spp.

Alpine tundra of *Dryas octopetala*, low sedges, and other low matted plants can be seen above the shrub zone on the slopes of Paxson Mountain.

6. Left; escape gorge of Wisconsin age on Paxson Mountain (fig. 6-38).

7.2. Stop 7-43: Alaska Range Panorama to North.

From here can be seen the Delta River Pass through which some glacier ice from the south side of the Alaska Range pushed north to the Tanana Valley. Bedrock at this location has been described as greenstone and is metabasalt. It is probably Triassic in age. Location is above timberline in alpine tundra vegetation.

10. Valley filled with ice-contact deposits of Donnelly (Denali II) age. Here the tundra becomes less shrubby in nature, and extensive stands of sedges, cottongrass (*Eriophorum* spp.), and grasses are interspersed with low willows.

On the exposed ridges are low matted shrubs of crowberry (*Empetrum nigrum*), alpine bearberry (*Arctostaphylos alpina*), mountain cranberry (*Vaccinium vitis-idaea*), Labrador tea (*Ledum decumbens*), and lichens.

From this point onward the Denali Highway passes alternately through shrub tundra, low sedge and grass tundra, lichen-covered ridges of low matted tundra, and occasionally, very open stands of black and white spruce.

11. Beginning of Eleven Mile Hill. This hill has become well known because of the great difficulties experienced with permafrost in constructing a satisfactory road on the silty till of Wisconsin age. Thawing permafrost and intense seasonal frost action provides an unstable road bed.

13. Top of Eleven Mile Hill. Road is on silty till of Wisconsin age. The lower slopes in the immediate foreground are blanketed with silty till but the valley bottoms, 5 miles (8 km) ahead, contain stagnant ice deposits. Tops of hills on the right (north) stood above ice of Donnelly age (both early and late Denali age). Directly ahead, 15 miles (24 km) in the distance, can be seen High Valley (fig. 6-38).

16.4. Road on silty till. Good view of large area of ice-contact deposits directly ahead. Gravel pit ahead is in these deposits.

17. Left; ice-contact deposits.

20. Round Tangle Lake in the heart of the esker and kame topography. Such topography is well illustrated

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[1]Miles from Paxson on the Denali Highway.

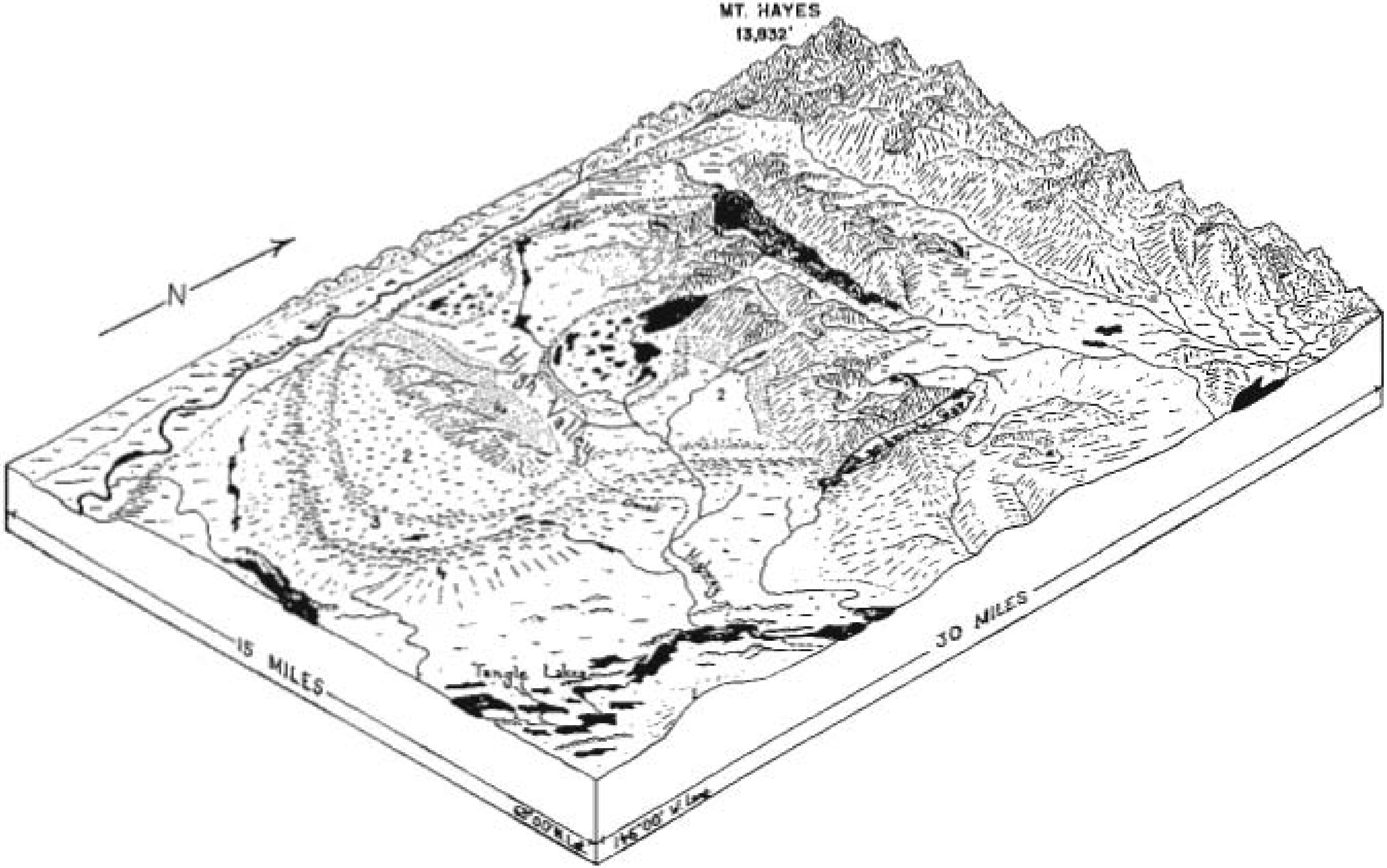

Figure 7-39. Physiographic diagram of Mt. Hayes A-5 topographic quadrangle, Amphitheater Mountains, with Alaskan Range in the distance. Symbols: 1—Glacial erratics of early to middle Quaternary age; 2—Deposits of Illinoian age; 3—Moraine of early Wisconsin age (Denali I Glaciation); 4—Moraine of late Wisconsin age (Denali II Glaciation); 5—Deposits of post-Wisconsin glaciation; RS—Rubble sheet of Wisconsin age. Compare diagram with geologic map of glacial deposits, Figure 6-8. (Diagram redrawn with slight additions by George B. Wharton, 1964, from an unpublished U. S. Geological Survey report by Kachadoorian and Péwé, 1954.)

on the Mt. Hayes A-4, A-5 and Gulkana D-4, D-5 quadrangle topographic maps.

20.4. Stop 7-44: Ice contact deposits and the Tangle Lakes.

The drainage here goes to the north through the Alaska Range into the Tanana and eventually the Yukon River. In the far distance, about 8 miles (13 km) directly north on the west side of the river, can be seen an excellent rock glacier which has formed in post-Pleistocene time.

20.7. Stop 7-45: Tangle Lake campground.

20.8. Enter Mt. Hayes A-5 Quadrangle.

21.2. Tangle River, source of the Delta River.

22.2. Stop 7-46: Road cut in small esker.

From the top of the esker to the left (south) and east can be seen extensive fresh ice-contact deposits. To the right (north) can be seen Landmark Gap (figs. 6-38, 7-39) through which a glacier poured southward from the Alaska Range. Elevation of the ice surface in Wisconsin time as it emerged from the gap was 4,000 feet (1,200 m). In the past, glaciers from the Alaska Range pushed south, and some of the ice was confined to the wide, deep, major river valleys, such as the Maclaren. Much of the ice, however, filtered through gaps and passes in the Amphitheater Mountains and was joined by local glaciers (fig. 7-39).

Several glaciations are recorded, each less extensive than the former. The earliest glacial advance is thought to be early to middle Quaternary in age and covered the 6,000-foot (1,800 m) peaks of the Amphitheater Mountains leaving isolated erratics on the tops of some of the peaks. The second glacial advance pushed south to the Copper River basin but did not cover the peaks of the Amphitheater Mountains. Many overflow channels were cut where drainage escaped over low places in the ridges. This advance, thought to be middle to late Quaternary in age, left an olive-colored silty till covering the lowlands and flanking the lower slopes of the Amphitheater Mountains. This till is now covered by later Quaternary drift sheets except above an elevation of 4,000 feet (1,200 m) and on the floor of High Valley. No morainal forms of this glaciation are preserved in the area. This glacial advance is tentatively correlated with the Delta Glaciation of the lower Delta River.

The next two glacial advances took place in late Quaternary time and are closely related in extent and age. They are grouped together and named the Denali Glaciation after the Denali Highway. The deposits are correlated with the Donnelly Glaciation.

These two advances did not cover the Amphitheater Mountains but moved through gaps and were joined by local cirque ice. Ice was thick in the major valley, such as the Maclaren River valley on the south side of the Range, but thin over interfluves. The ice, therefore, was relatively thin in the lowland of Tangle Lakes and in lowlands elsewhere at altitudes of 3,000 to 4,000 feet (900 to 1,200 m). Upon retreat, much of the ice covering interfluves thinned and stagnated. As the ice stagnated, many ice-contact features formed: eskers, kames, crevasse fillings, and pitted surfaces. The Tangle Lakes at the headwaters of the Delta River are a classic area for such features which are strikingly fresh, and cover an area of several hundred square miles.

In post-Wisconsin time the glaciers in the Alaska Range and the small local glaciers in the Amphitheater Mountains advanced a short distance and then retreated, leaving arcuate moraines at the mouths of short valleys.

On the tops of many of the drift features of the Tangle Lakes area are small archeological sites. Many of the artifacts found here are similar in style and general technique to some of those known in the northern Plains and Southwest. Although none of the diagnostic types may be called common, projectile points have been found here which resemble the Lerma, Agate Basin, Angostura, and Plainview types and perhaps others. Specimens have been found which could be placed within the generalized Clovis category, and notched points of northern affinities also occur.

The major problem with these sites at present is that of dating. Altogether some 25 sites have been discovered, at least 20 of which were found in the summer of 1964 in conjunction with the work in the vicinity of Donnelly Dome. Most of these were surface finds, however, and although some of the chert used is heavily patinated, as yet we have no means of placing any of these finds accurately in time.

The Tangle Lakes evidently provided especially good ecological conditions for herding animals (as did the Donnelly Dome area), but it would be unwise at this juncture to assume that in immediate postglacial times these were caribou alone. The mere fact of the relative abundance of archeological remains gives good evidence of the former presence of large herbivores here, and at the present time caribou are seasonally abundant.

26.5. Start up right (west) lateral moraine of glacier that poured south through Landmark Gap in Wisconsin time.

28.1. Top of prominent lateral moraine of early Wisconsin age (early Denali age, figs. 6-38, 7-39).

29. High Valley (fig. 7-39). (High Valley extends from Milepost 28 to 36.5.) Ahead on the right, across High Valley at elevation 4,855 feet (1,480.7 m), may be seen step terraces cut in bedrock which are believed to be altiplanation terraces. They stand above the upper limit of the Wisconsin ice.

30. Directly ahead is the terminal moraine of the glacial bulb that pushed south through Glacier Gap and terminated in High Valley. Left; inactive rock glacier(?) of Wisconsin age.

30.4. Stop 7-47: Terminal moraine.

Glacial lobe that pushed south through Glacier Gap into High Valley. Road follows crest of terminal moraine (figs. 6-38, 7-39). High Valley is an area that is about 4,000 feet (1,200 m) above sea level and is unique inasmuch as it was ice free in Wisconsin time, although completely surrounded by glaciers. High Valley was glaciated in Delta time, and the floor is covered with a silty till. In Denali (Wisconsin) time ice pushed south from the Alaska Range filling the major valleys and lowlands but pushing only slightly over the edge and spilling a short distance into High Valley. Ice also pushed south through Glacier Gap and terminated as a bulb in High Valley. The ice of this bulb was mostly from local sources, inasmuch as 98 percent of the rock types are from the Amphitheater Mountains.

Lower slopes, especially north-facing slopes, in the High Valley area are blanketed with a 3- to 7-foot (1 to 2.1 m) thick sheet of now inactive rubble that was derived from higher slopes by frost riving during the rigorous climate of the time of Denali Glaciation. Some of these rubble sheets extended out over the silty till of middle Quaternary age.

During this rigorous climate of Wisconsin time, small rock glaciers(?) originated on the north side of the bedrock ridge south of the highway and pushed a short distance into High Valley.

31.2. Dissected rock glacier(?) of Wisconsin age on left. Originates from ridge of metabasalt.

32.2. Road leaves moraine.

32.5. Roadcut through rubble or rock glacier(?).

34.5. Right; extending from ridge of metabasalt is an inactive rubble sheet, 3 to 7 feet (1 to 2.1 m) thick, and overlying silty till of Illinoian age derived from the bedrock ridge by frost riving during the rigorous climate of Wisconsin (Denali) time. The bedrock ridge was not high enough to support glaciers such as those on the north side of the Amphitheater Mountains across High Valley.

35.4. Summit of Denali Highway. Elevation 4,081 feet (1,240.6 m).

35.6. Right; lobe of early Wisconsin age moraine (Denali I) (fig. 7-39). Deposited when Maclaren Glacier pushed over the edge of Maclaren Valley into High Valley a short distance.

36.6. View of Maclaren Valley and Maclaren Glacier. Road can be seen across the valley. Ice-contact deposits in the bottom of the valley at the bridge site.

40.8. Enter Mt. Hayes A-6 Quadrangle.

40.8. Stop 7-48: Pingos.

Pingos in bog on both sides of the road. Pingo on left has been cut in half. Radiocarbon age of lowest peat in contact with ice is 10,565 ± 225 years (GX-0249). Pingo formed in post-Pleistocene time and this gives a minimum date on the age of the late Denali advance.

41.1. Ascend ice-contact deposits. This large mass of washed gravel was deposited in what was probably a very large hole in the stagnant mass of ice. Large kettle holes; the one on the right is approximately 50 feet (15 m) deep.

42. Maclaren River and view of the Maclaren Glacier. The river cuts through ice-contact deposits at this point.

45. Entrance to Crazy Notch.

46. Exit of Crazy Notch. Crazy Notch was cut by a subaerial stream which may have been sub-ice both to the east and west. An esker begins (or ends) at the exit of the notch. Two systems of eskers enter Crazy Notch from higher elevations to the west, and drainage was evidently to the east into the Maclaren Valley. The road is built on eskers for the next 6 miles.

47. A view back to the right permits one to see an excellent rock glacier on the bedrock ridge through which Crazy Notch is cut.

49. Peat pingos 6 feet (1.8 m) high and 15 feet (4.6 m) in diameter occur in the bogs on the right.

50. Stop 7-49: Roadcut in esker.

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COPPER RIVER BASIN

by

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RESUME OF QUATERNARY GEOLOGY OF THE COPPER RIVER BASIN

The Copper River basin is an intermontane basin ranging from 500 to 4,000 feet (150 to 1,200 m) above sea level and is rimmed by 4,500- to 16,500-foot (1,370 to 5,000 m) high peaks of the Alaska Range and the Talkeetna, Chugach, and Wrangell Mountains. Rocks bordering the basin range in age from middle(?) Paleozoic to Tertiary. They consist largely of schist, greenstone, greywacke, slate, shale, and sandstone, locally associated with minor amounts of altered limestone, tuffaceous beds, and basalt flows, and are intruded by a wide variety of igneous rocks. Also, large parts of the Wrangell and Talkeetna Mountains and local areas in the Chugach Mountains are underlain by considerable thicknesses of basaltic and andesitic lava flows. Thin andesite flows are interbedded with Pleistocene deposits in the southeastern part of the Copper River basin.

During one or more early Pleistocene glaciations, glaciers advancing from the surrounding mountains covered the entire basin floor. The ice may have thickened enough to have caused the principal accumulation areas to shift toward the center of the basin, resulting in the formation of a domed ice cap within the basin. Such an ice cap would have drained by radial flow from the basin through the Copper River Canyon and low passes, including Tahneta Pass, Susitna River valley, Delta River valley, and Mentasta Pass. In middle Pleistocene time, ice may have covered all but small areas of the basin floor, but during late Pleistocene time, large areas were ice-free.

In the early stages of each major Pleistocene glaciation, ice advancing in the surrounding mountains dammed the drainage of the Copper River basin, causing the formation of an extensive proglacial lake in the basin. Because the glaciers fronted in deep lake water, there is a general lack of end moraines and associated features in the lower part of the basin (Copper River trough) (fig. 8-40). However, glacial landforms, although modified, are present below former lake levels in the higher parts of the basin (Copper-Susitna Lowland). As the glaciers retreated, their deposits were reworked by lake currents or were buried by lacustrine sediments. The complicated interfingering of lacustrine and glacial deposits and the numerous well- to poorly developed shoreline features at altitudes below 2,650 feet (810 m) indicate that the lake level fluctuated widely as it lowered during retreat of the glaciers of the last major glaciation (Wisconsin).

After retreat of the glaciers and drainage of the lake about 9,000 years ago, permafrost began to form in the lacustrine and glacial deposits, and the rivers began downcutting

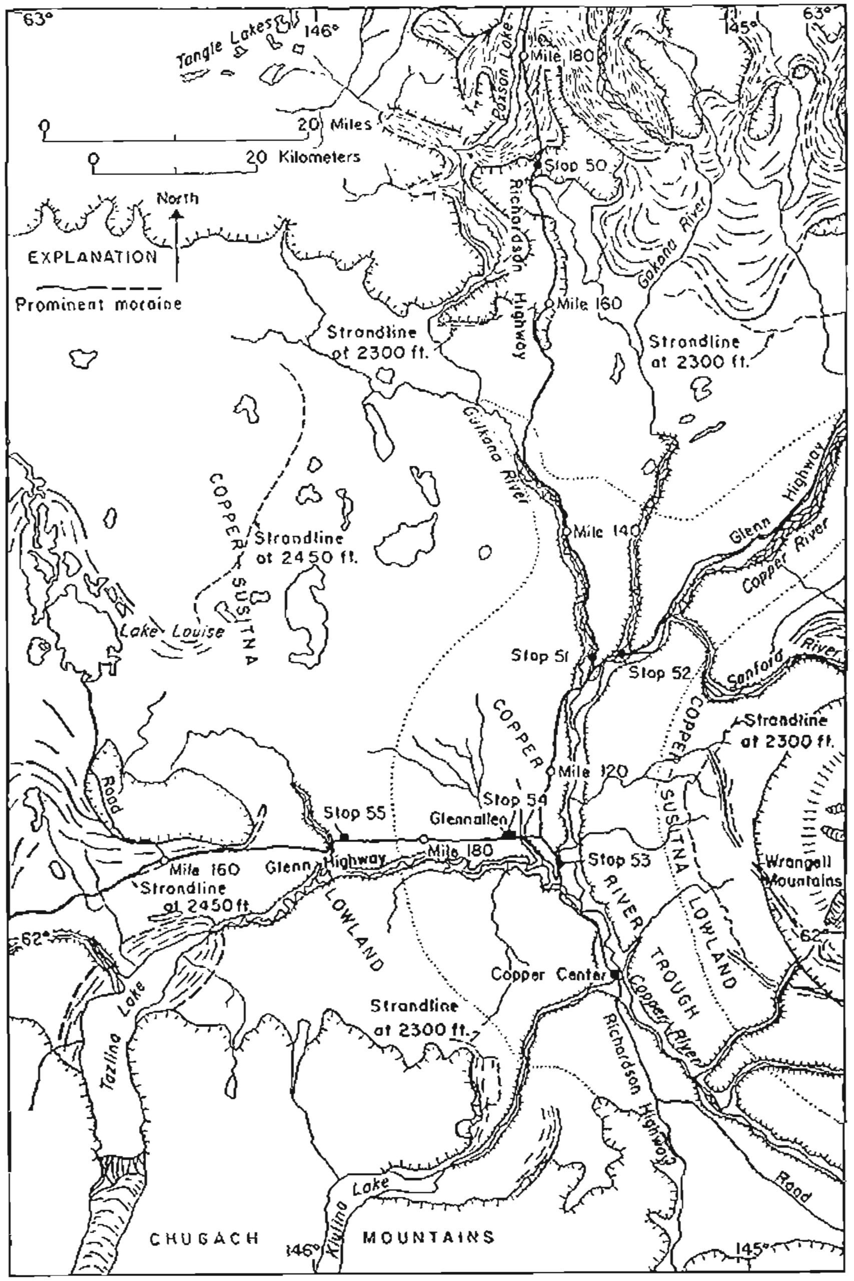


Figure 8-40. Index map of the Copper River Basin, Alaska showing Field Trip Stops.

into the Pleistocene sediments. Between river valleys, muskegs and marshes, which occupy depressions on the old lake floor, are perched on poorly drained, perennially frozen lake sediments.

In the Copper River basin, most of the damage and ground breakage caused by the major earthquake of March 27, 1964, was restricted to the southern half of the basin. Several buildings were shaken from their foundations, and the foundations of several other structures were damaged. Dishes were broken in many dwellings, and, locally, sewer lines and other underground pipes were damaged. Ground cracks commonly occurred in flood plains of major rivers; locally, in low terraces adjacent to flood plains; in deltas; along margins of lakes; along the toes of alluvial fans; in highway fill; along the face of steep slopes of river bluffs and hillsides; and in areas cleared of vegetation. These ground cracks in unconsolidated deposits generally were restricted to areas where one or more of the following conditions existed: (1) permafrost was absent or deep lying, (2) ground-water table was near the surface, (3) bedrock was relatively deep lying, and (4) slopes were steep.

Road Log and Locality Descriptions

183.3.[1] Enter Gulkana D-3 Quadrangle.

182.5. Enter Gulkana D-4 Quadrangle. Straight ahead; Paxson Lake which is approximately 10 miles (16 km) long and ¾ mile (1.2 km) wide.

182.2. Right; delta of Gulkana River which forms forested neck of land in center of lake.

181.5. Roadcut in highly fractured greenstone. Highway parallels eastern shore of Paxson Lake and lies at base of north-south-trending hill to the east. The lower steep slope along the western flank of this hill generally is covered by colluvium derived both from greenstone bedrock, which underlies the hill at relatively shallow depth, and from glacial drift, which generally mantles the bedrock.

180.8. Right; Sportsman Lodge which is built on a spit.

179.0. For the next 6 miles (10 km), highway is underlain by relatively thick deposits of glacial drift, a complex morainal system formed during the last major glaciation (Wisconsin). During this period of glaciation, ice from the Alaska Range flowed southward down the trough now occupied by Paxson Lake (fig. 8-40).

177.8. Enter Gulkana D-3 Quadrangle.

176.7. Roadcut through deposit of till; exposed best on left.

172.0. The rounded hills to the south were not overridden by ice during the last major glaciation; however, they were overridden during one or more earlier glaciations.

171.5. Right; two prominent terraces can be seen at the base of the hill across the drainageway. These terraces were formed during the last major glaciation by an outwash stream emanating from the glacier that occupied the trough now occupied by Paxson Lake. The meltwater stream flowed to the south through the depression now occupied by Meier Lake and formed a large delta where it emptied into a proglacial lake. The highway crosses this delta about 2 miles (3 km) to the south.

171.1. Highly fractured greenstone exposed in roadcut. For the next mile the highway parallels the eastern shore of Meier Lake and for a short distance is underlain by outwash gravel and sand deposited by

[1]Miles from Valdez on the Richardson Highway.

the outwash stream previously mentioned.

170.6. Stop 8-50: Frost-rived granite blocks.

For the next 2,500 feet (760 m) granite bedrock is exposed along the highway. Numerous large blocks of weathered granite are exposed conspicuously on the hillside to the left. The well-jointed character of the granite makes it especially susceptible to frost action. The presence of these large blocks of frost-rived granite indicates that the hill was not only just outside and above the glacier border of the last major glaciation but also above the level of a large lake which existed in the Copper River basin. The hill therefore was exposed to a rigorous periglacial climate, a climate more rigorous than that of today.

170.0. Left; site of Meier Roadhouse. The main structure burned to the ground a few years ago. Wagon in rear was used in the early 1900's to haul supplies from Valdez to interior Alaska.

169.8. Right; roadcut in outwash gravel.

169.3. Right; gravel pit in large delta mentioned at mile 171.5. Good forest beds are present locally in this deposit. A local resident reports that a mammoth tusk was unearthed by construction workers during excavation of this pit. For the next 3 miles (5 km) the highway is underlain by fine, silty, lacustrine sand.

166.0. For the next 5 miles (8 km), highway crosses a north-south-trending hill composed of dioritic bedrock with a thin veneer of glacial drift.

165.0. Enter Gulkana C-3 Quadrangle.

161.0. Highway crosses low terrace of Haggard Creek. To the south, a rock quarry has been developed at the north end of the hill. The dioritic rock has been hauled to Gakona, 30 miles (48 km) to the south, and used to protect a high river bluff from undercutting the Copper River.

160.8. For the next 2.5 miles (4 km) the highway ascends, and then follows, the western flank of this hill which is underlain by a thin veneer of glacial drift overlying dioritic bedrock.

158.2. For the next 3 miles (5 km) the highway ascends and follows the western flank of Hogan Hill. A thin veneer of glacial drift overlies greenstone which crops out conspicuously along the highway at mile 158.1 and mile 155.5. A rock quarry has been developed on the left side of the highway at mile 158.0.

155.4. Straight ahead, on a clear day an excellent view, from left to right, of Mt. Sanford (16,237 ft [4,949 m]), Mt. Wrangell (14,163 ft [4,317 m]), and Mt. Drum (12,010 ft [3,661 m]). All of these peaks are volcanoes. The dome-shaped Mt. Wrangell is still active and occasionally emits steam and ash.

155.2. For the next 5.4 miles (8.2 km) the highway crosses a lacustrine plain which is underlain by stony silt (gravelly, clayey silt). These fine-grained deposits, and other similar fine-grained deposits in the Copper River basin, generally are poor foundation materials. Construction on them commonly results in serious maintenance problems, especially where surface drainage is poor. Construction disrupts the thermal regime of the ground, causing a lowering of the permafrost table and consequently melting of any ground ice which in turn causes settlement. When saturated with water the thawed silt loses its bearing strength with settlement of the road bed.

149.8. For next 1.8 miles (2.9 km) highway crosses near-shore, lacustrine sand and fine gravel deposits; these sediments are underlain by fine-grained lacustrine deposits at relatively shallow depth.

149.5. Enter Gulkana C-4 Quadrangle.

148.0. Highway descends about 50 feet (15 m) over colluvial deposits to the narrow, sand- and silt-mantled terrace of Sourdough Creek. Right; Sourdough Lodge which was established in 1904.

147.7. Sourdough Creek, and a colluvium-covered slope similar to that along the north side of the creek.

147.0. For the next 17 miles (27 km) highway continues on lacustrine plain which is underlain generally by massive pebbly silt.

146.0. Enter Gulkana C-3 Quadrangle.

145.4. Enter Gulkana B-3 Quadrangle.

138.0. Drainageway of small stream which is a tributary to the Gulkana River. For several miles the highway is parallel to the valley of the deeply entrenched Gulkana River which can be seen through the trees to the right.

137.0. Highway underlain by a sandy facies of the lacustrine deposits, and consequently, the area is fairly well drained and supports a relatively tall stand of poplar and white spruce trees.

130.0. Stop 8-51: Permafrost thermal recording station.

The Copper River basin lies in the zone of discontinuous permafrost. Permafrost probably is present everywhere in the basin except beneath large lakes and major streams. It lies 1 to 2 feet (0.3 to 0.6 m) below the surface in some muskegs with thick sphagnum moss, 2 to 5 feet (0.6 to 1.5 m) in lacustrine and fine-grained glacial deposits, and 6 to more than 10 feet (1.8 to more than 3 m) in granular alluvial and glacial deposits. It generally ranges from 100 to 200 feet (30 to 60 m) thick, has a high ice content, and is marginal in temperature (−0.5° to −1.5°C [31.1° to 29.3°F]). Consequently, permafrost here is in a delicate state of equilibrium, and if it is thawed by minor changes in the regimen of the ground-surface temperature, such as brought on by most construction projects, considerable surface subsidence may occur. However, because of its thickness, attempts to thaw the permafrost and stabilize the ground prior to construction generally are impractical.

Because of the problems experienced in constructing highways and buildings on permafrost in the Copper River basin, a cooperative study of the engineering aspects of perennially frozen ground was undertaken in 1954 by the U. S. Geological Survey and predecessor organizations to the Alaska Department of Highways (Alaska Road Commission and Bureau of Public Roads). Six roadway sections and one apartment house were selected for study. All roadway sections were instrumented by a continuously recording thermograph to obtain ground-surface temperature, and four vertical, 20-foot-long (6.1 m) thermistor cables. The cables were installed in the centerline, shoulder, and ditch of the road and one in undisturbed ground beside the road.

A control station was set up at Mile 130 on the Richardson Highway. At the control stations, in addition to the three vertical thermistor cables in the road, a cable was placed horizontally across the road at a depth of 3 feet (0.81 m), and 20-foot (6.10 m) cables were placed vertically in undisturbed ground on both sides of the road and one in the centerline of an adjacent section of an old, abandoned roadway (fig. 8-41). The highway at the control section was built on undisturbed ground in 1951. The roadway was hand cleared and trees were laid normal to the centerline on the undisturbed vegetation mat. In 1952, approximately 3 feet (0.9 m) of rela-

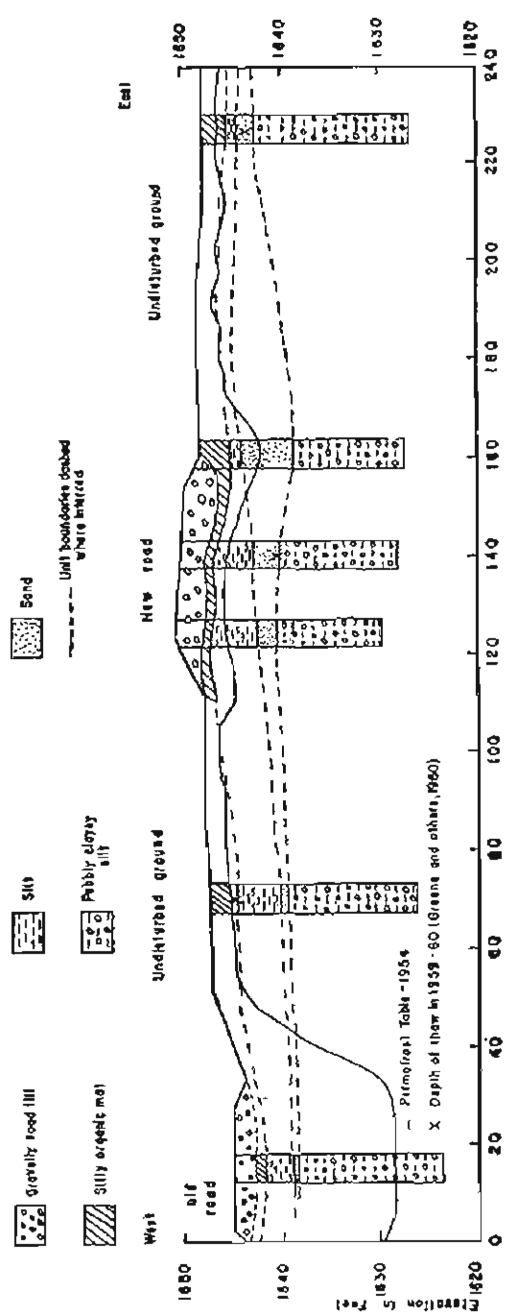


Figure 8-41. Cross section of permafrost table and geology at Mile 130 Richardson Highway as determined from drill holes during installation of thermistor cables.

tively clean, coarse gravel and sand were spread over the section with a minimum of disturbance. However, by 1954, differential subsidence necessitated placing half a foot of gravel through the area to maintain a smooth surface. This section was paved originally in 1956 and has been patched several times and torn up and repaved once in the late 1950's. A good riding surface seldom lasts more than a year after reconstruction.

The adjacent old road was constructed in the pre-World War II era, then widened and additional gravel placed on the surface during wartime construction. When originally constructed, the roadway was cleared and stripped down to the frozen portion of the organic mat. At present, permafrost lies at a depth of a little more than 16 feet (4.9 m), and seasonal freezing of the ground with considerable ice growth may extend to a depth of almost 10 feet (3 m). Under the newer road, permafrost had degraded from 5 feet (1.5 m) in 1954 to about 11 feet (3.4 m) at present, largely the result of an abnormally warm summer in 1957.

129.6. Roadcut showing character of the massive, lacustrine pebbly silt. It is exposed best on the right; however, slope wash somewhat obscures view. Highway descends into old abandoned drainageway connecting the existing valleys of the Gulkana and Copper Rivers and then ascends back to the lacustrine plain.

128.8. Junction of the Richardson Highway and the Tok Cutoff portion of the Glenn Highway.

Glenn Highway

0.6.[1] For next 0.4 miles (0.6 km) highway follows old, abandoned drainageway (mentioned above) to the edge of a high river bluff of the Copper River. Edge of bluff is underlain by windblown sand and silt (cliff-head dunes) interbedded with peat and other organic material.

1.0. Highway descends the steep, 200-foot (60 m) bluff. Along the left side of the highway there is a good exposure of deposits laid down in the extensive proglacial lake which occupied a large part of the Copper River basin during the last major glaciation. These deposits include not only typical laminated lake sediments, but also nonsorted and poorly sorted lake sediments, which have been described as glaciolacustrine diamicton deposits. The nonsorted deposits commonly are till-like in character. Along this section of the highway, considerable slumping has occurred, both upslope from the road and downslope from the road where the Copper River has been undercutting the bluff. Recently a long dike was constructed to protect the bluff from undercutting.

1.4. Sand- and silt-mantled gravel terraces of the Gakona River.

1.7. Gakona River.

1.8. Gakona Roadhouse — one of oldest roadhouses in the Copper River basin.

Stop 8-52: Gakona Section.

During each major glaciation, glaciers advancing in the mountains surrounding the Copper River basin dammed the drainage of the basin, thus impounding an extensive proglacial lake. The lake that existed during the last major glaciation covered more than 2,000 square miles (5,200 km^2) of the basin floor, and numerous glaciers and sediment-laden glacial streams debouched into it. The 300-foot (90 m) high bluff along the west side of the Gakona River offers a good view of the type of sediments that were deposited in this lake.

The uppermost 15 to 30 feet (4.5 to 9.1 m) of the section consists of

[1]Miles from junction of Glenn Highway and Richardson Highway.

windblown sand and silt interbedded with peat and other organic material. Approximately 200 feet (61 m) of sediments, underlying the windblown material, was deposited in the extensive proglacial lake during the last major glaciation. The age of these sediments is bracketed between a maximum radiocarbon date of greater than 38,000 years (W-531) and a minimum date of 9,400 ± 300 years (W-714).

Radiocarbon age determinations of 28,300 ± 1,000 years (W-1343) and 31,300 ± 1,000 years (W-843) were obtained from organic material collected from a bluff along the Sanford River about 10 miles east of here. These two age determinations date an interval of time when the level of the lake was at an altitude of approximately 2,150 feet (655 m). All these radiocarbon dates indicate that the last major glaciation in the Copper River basin is comparable in age to the last major glaciation (Wisconsin) in central North America. Other exposures less than a mile downstream from here provide stratigraphic evidence for two major glaciations older than the last major glaciation.

Richardson Highway

128.8.[1] Highway on lacustrine plain.

127.6. Highway descends a colluvium-covered bluff of the Gulkana River.

127.5. Low gravel terrace of the Gulkana River. Left: Indian Village of Gulkana.

127.1. Gulkana River is one of the few clear-water streams in the Copper River basin. Most of the major rivers in the basin carry a large load of suspended sediment and consequently are brown.

———

[1]Miles from Valdez on the Richardson Highway.

127.0. Highway ascends a colluvium-covered bluff to the surface of a high gravel terrace.

125.6. Highway on lacustrine plain.

124.8. Enter Gulkana A-3 Quadrangle.

124.8. Underlain largely by pebbly, clayey silt.

121.1. Well at end of access road to right was drilled to a depth of 354 feet (107.9 m) and encountered highly saline water. The minimum temperature recorded over a period of several years in this hole, from below the zone of mean annual temperature fluctuation, was −0.7°C (31.0°F). Permafrost occurs to a depth of 120 feet (36.6 m) below the surface in glacial and glaciolacustrine deposits of Wisconsin age and hence is considered to have formed in the Recent Epoch.

118.2. Poorly drained drainageway on lacustrine plain at west end of Gulkana Federal Aviation Agency airstrip. Note the deterioration of the east-west runway, which is no longer serviceable for aircraft landings. Differential subsidence of the blacktop surface has resulted from degradation of permafrost. Wells drilled at the F.A.A. station have penetrated to a depth of 293 feet (89.4 m) before encountering water. Water is generally high in chlorides, especially at the 433-foot (132 m) level; water from the 293-foot (89.4 m) level is not considered potable without filtration.

117.6. Incised flood plain of Dry Creek.

115.4. Saline water reported from 321-foot (97.8 m) well drilled at Gateway Lodge.

112.5. Stop 8-53: Simpson Hill Roadcut and Copper River Bluff.

Overview of the Copper River and from left to right, Mt. Sanford (16,237 ft [4,949 m]), Mt. Drum (12,010 ft [3,661 m]), Mt. Wrangell (14,163 ft [4,317 m]), and Mt. Blackburn (16,523 ft [5,036 m]). Cones of the Drum group of mud volcanoes also are visible. Shrub mud volcano is 14 miles (23 km) to the east-northeast; Upper Klawasi is 14 miles (23 km) to the east; and Lower Klawasi is 7 miles (11 km) to the east-southeast (fig. 8-42). The Lower Klawasi mud volcano is the largest of the Drum group of mud volcanoes. The cone lines on the lower slopes of the Wrangell Mountains and is composed of clayey silt with small angular rock fragments. The base is approximately 6,000 feet (1,830 m) E-W and 8,200 feet (2,500 m) N-S; the cone is about 150 feet (46 m) high. The pool in the crater is depressed 15 feet (4.6 m) below the crest, is 175 feet (53.4 m) in diameter (fig. 8-43), and discharges carbon dioxide gas through Na-HCO_3-Cl water with up to 28,000 ppm dissolved solids.

Simpson Hill Roadcut. In the spring of 1954, during construction of a new telephone line upslope and parallel to the road, a wide swath of the vegetation cover was stripped through the spruce forest. Early in September of that same year, a maintenance crew made cuts in the bluff immediately uphill and downhill from the fill and dumped the cut material on the outer part of the road to bring it back to grade. Overnight, a 15-foot (4.6 m) wide section of the paved road and shoulder slumped downward through a vertical distance of 10 feet (3 m), apparently along a concave glide plane. To repair this major damage, large amounts of silt were excavated along the cut and dumped into the slump area. Almost a year later the same part of the road again subsided 10 feet (3 m), following attempts to level minor slumps by the addition of a small amount of fill. Subsequently, deeper cuts have been made into the hill in order to obtain the required road width rather than

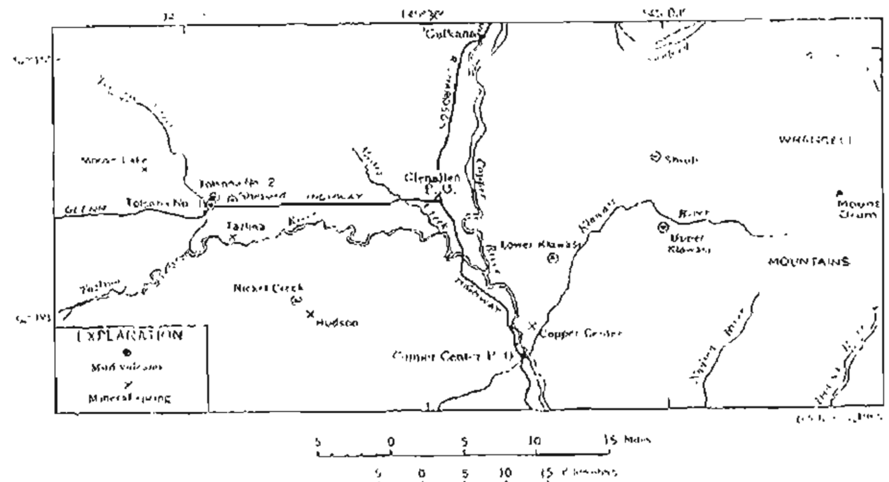


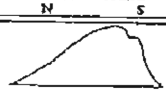
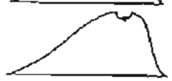
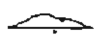
Figure 8-42. Location of mud volcanoes and mineral springs in the southeastern Copper River Basin.

again replacing the slumped material with fill. Nevertheless, sliding has continued on a minor scale and constitutes a never-ending maintenance problem.

The roadcut here exposes several layers of volcanic ash in lacustrine pebbly silt and varved silt and clay.

Copper River Bluff Section. This section exposes the deposits of the last major glaciation in the southeastern Copper River basin.

COMPARISON BETWEEN CERTAIN PHYSICAL CHARACTERISTICS OF MUD VOLCANOES, COPPER RIVER BASIN, ALASKA

	Mud Volcano	Diagrammatic cross-section N S	Approximate dimension * of cone: Base	Hgt.	Alt. * of crest	Approx. * diam. of "crater"	Surf. water temp., F.	Est. water disch., gpm.
Drum Group	Shrub		3600 4200	310	2950	120	54	< ¼
Drum Group	Upper Klawasi		4200 6700	300	3017	150	86.5	2-5
Drum Group	Lower Klawasi		6000 8200	250	1875	175	82	5-10
Tolsona Group	Nickel Creek		800 1000	60	2025	150	cold	< ¼
Tolsona Group	Shepard		1300 1600	25	2172	15	—	—
Tolsona Group	Tolsona No. 1		600 900	25	2045	30	38-55	< ¼
Tolsona Group	Tolsona No. 2		2000 2300	40	2085	150	40-60	< ¼

* in feet, ▾ = active spring, • = inactive spring.

Figure 8-43. Comparison between certain physical characteristics of mud volcanoes, Copper River Basin, Alaska. (From Nichols and Yehle, 1961a, Table 1)

COPPER RIVER BLUFF SECTION

Unit	Thickness in feet	Feet below top of bluff
1. Vegetation mat and soil profile developed on eolian silt and sand.	2.7 (0.9 m)	2.7 (0.9 m)
2. Late-glacial lacustrine deposits: Massive dark-grey clayey silt with lenses of silt and volcanic ash.	6.5 (2.0 m)	9.2 (2.9 m)
——— Gradational contact ———		
3. Late-glacial deposits: Dark-grey silty sand-matrixed till with subrounded pebbles, cobbles, and boulders. Thin sandy lenses in upper 2 ft.	15-25 (4.6-7.6 m)	29.2 (8.9 m)
4. Interstadial lacustrine deposits:		
a. Amorphous masses of blue-grey, phyllitic-appearing clay with a few pebbles. Extreme deformation by shearing locally.	5 (1.5 m)	34.2 (10.4 m)
b. Well-laminated, blue-grey clayey silt with several, 1- to 4-inch (2.5-10 cm) thick volcanic ash beds.	5 (1.5 m)	39.2 (11.9 m)
c. Massive, blue-grey blocky silty clay.	10-15 (3.0-4.5 m)	51.7 (15.7 m)
5. Early-glacial ice deposits:		
a. Massive dark-grey clayey silt with columnar joints and conchoidal fracture and with numerous scattered pebbles and cobbles. Apparent bedding locally.	10-20 (3-6 m)	66.7 (21 m)
b. Gradational zone between (a) and (c).	10 (3 m)	76.7 (23 m)
c. Typical, blocky, fractured till with numerous pebbles, cobbles, and boulders in a clayey, sandy silt matrix. Contorted and faulted varved silt and clay incorporated as large and small masses in till and in shear zones near base.	30-40 (10-12 m)	111.7 (34 m)
6. Early-glacial lacustrine deposits:		
a. Horizontally laminated and varved light-grey and dark-blue-grey silty clay with several thin, highly contorted zones. Includes scattered pebbles and cobbles.	22 (6.7 m)	133.7 (41 m)
b. Medium to coarse dark-grey sand with steep, southward-dipping forest beds. Granules, pebbles, and cobbles scattered throughout.	3 (.9 m)	136.7 (42 m)
7. Preglacial fluvial deposits: (Rapid lateral variations)		
a. Fine pebble and pea gravel in coarse sand matrix.	0.3 (0.1 m)	137 (42 m)
b. Horizontally bedded medium to coarse sand with thin pumiceous pebbly zones and scattered clastic "boulders" of till and varved silt	26 (7.9 m)	163 (50 m)

and clay. 1-foot-thick bed near middle dips steeply northwest.		
——Disconformity——		
c. Horizontally bedded grey to brown silt and fine sand, highly contorted in upper 3 ft.	5 (1.5 m)	168 (51 m)
d. Dark-grey coarse to medium sand with gravelly beds.	4 (1.2 m)	172 (52 m)
e. Well-rounded gravel with a few thin sandy beds. Iron and manganese oxide staining locally and particularly near base.	25 (7.6 m)	197 (60 m)
f. Well-bedded fine to coarse, dark-grey to greenish-grey sand.	0-3 (0-0.9 m)	198.5 (60.5 m)
8. Sanford volcanic mudflow deposits: Silty sand matrix with subrounded to angular pebbles and cobbles, largely of andesite. Sharp color variations laterally and vertically from pink, green, and brick-red to grey. Generally massive but with local lenses of well-sorted sand or gravel.	0-3 (0-0.9 m)	200 (61 m)
9. Pre-Sanford fluvial deposits: Alternating sand and well-rounded coarse gravel with rounded detrital blocks of till and lacustrine sediments, locally iron stained; cross-bedding dips steeply to the south.	60 (18.3 m)	260 (79 m)

Copper River level.

The following reconstruction of events, from the oldest to the youngest, are inferred from the character and relations of the glacial, lacustrine, and fluvial deposits exposed in the Copper River Bluff section.

1) Fluvial deposition with incorporation of blocks of older sediments, probably from nearby riverbanks; 0-60 ft (0-18.3 m) above river level.
2) Downstream terminus of coherent deposits from the Mt. Sanford mudflow. Deposits diluted here; 60-63 ft (18.3-19.2 m)
3) Reworking of Sanford volcanic mudflow deposits and continued alluvial deposition. Disconformity probably represents erosion during one or more glaciations, and/or periods of river downcutting; 63-130 ft (19.2-39.6 m).
4) Rapid change from fluvial to lacustrine deposition—onset of deteriorating climate and glacial conditions; 130-133 ft (39.6-40.5 m).
5) Rising lake level with probable ice-rafting of coarse fragments and periodic slumping or iceberg drag of sediments to cause contortions; 133-185 ft (40.5-56.4 m).
6) Overriding of lacustrine deposits by first ice advance of last major glaciation and incorporation of varves in till; 185-220 ft (56.4-66.1 m).
7) Gradual thinning of ice to increase buoyancy and eventual floating of ice near terminus with rapid deposition from melting at base of ice; 220-245 ft (67-75 m).
8) Retreat of ice front south of this point but with continued rapid lacustrine deposition and at least minor volcanic activity as suggested by thin ash beds; 245-255 ft (75-78 m).
9) Readvance of ice of last major glaciation, strongly deforming lake sediments, and deposition of sandy till; 255-275 ft (78-84 m).
10) Retreat of ice of last major glaciation (final retreat in this area) and resumption of lacustrine deposition; 275-281 ft (84-86 m).
11) Drainage of lake at close of Wisconsin time, downcutting of the Copper River, and inauguration of eolian activity to form the present surface; 281.5-284 ft (86-87 m).

That lacustrine sedimentation was continuous and widespread between deposition of the tills, as postulated in events 6 and 9, indicates that the ice, while it did thin and retreat at least 5 miles (8 km), did not retreat sufficiently to permit complete draining of the lake. Ice retreat in this area may have coincided with evidence by Ferrians for a lowering of lake level 28,000 to 31,000 years ago. Consequently, event 8 is considered to represent interstadial rather than interglacial conditions. Only with complete removal of ice as a barrier to drainage of the Copper River, as indicated by cessation of lacustrine deposition and initiation of subaerial or fluvial conditions, can we assume that an interglacial climate existed.

Glenn Highway

Junction of the Richardson Highway and the Glenn Highway. Much of the land around the junction of the Glenn and Richardson Highways was withdrawn from homesteading to evaluate the area as a townsite. Because of saline water and permafrost problems, plans for the townsite were abandoned. A well drilled to a depth of 323 feet (98 m) at Rosent's Roadhouse in the fall of 1959 encountered water with 2,270 ppm dissolved solids and some gas.

188.1.[1] Enter Gulkana A-4 Quadrangle.

187.6. 100-foot-deep (30 m) well produces slightly hard but potable water from glacial deposits and below permafrost.

186.2. Stop 8-54: Glennallen permafrost problems (fig. 8-44).

Numerous buildings in the Glennallen area have severe structural problems from settlement as the result of differential thawing of permafrost. Most of the buildings at Glennallen are built on colluvial-mantled terrace deposits of Moose Creek. Colluvial deposits, 1 to 15 feet (0.3 to 4.6 m) thick, consist largely of gravelly silty clay; terrace deposits, 10 to 30 feet (3 to 9 m) thick, are largely silty sandy gravel or gravelly sand and overlie a thick sequence of fine-grained ice-rich glaciolacustrine deposits. Permafrost

[1]Miles from Anchorage on the Glenn Highway.

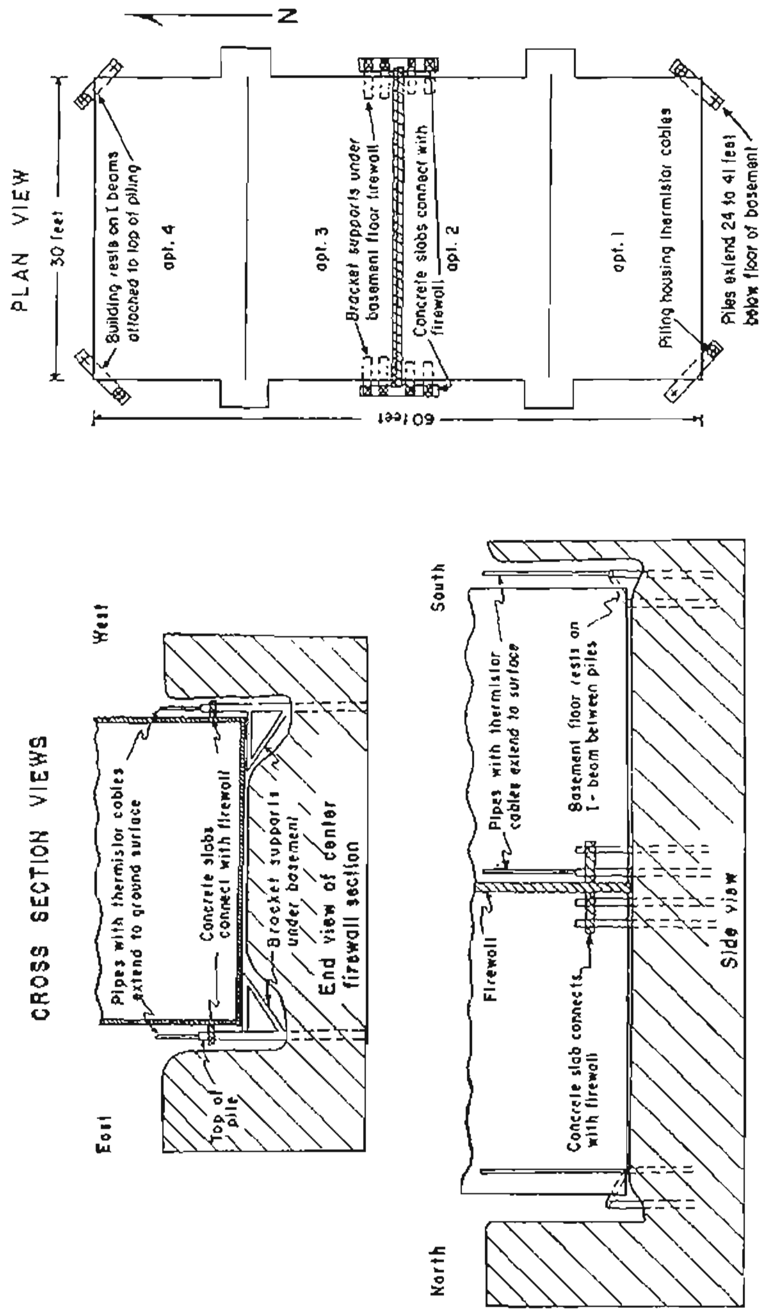


Figure 8-44. Piling and cable installation plan for Alaska Department of Highways apartment house at Glennallen, Alaska.

generally lies 5 to 10 feet (1.5 to 3.0 m) below the surface, deeper in areas of ground scarring. Moisture content is low in unfrozen granular terrace deposits but is sufficient to act as a cementing agent and locally to form ice lenses and stringers. Small amounts of ground water are perched on the permafrost and afford seasonal supplies of potable water. Maintenance of the permafrost level in this area is difficult because of its marginal temperature; artificial destruction of the permafrost before construction is virtually precluded by its considerable depth. A number of methods have been adopted in design and construction of new buildings and in rehabilitating existing structures with varying degrees of success (fig. 8-44). The Glennallen ACS microwave tower constructed in 1960 and utilizing the "Long thermopile" in the foundation, has, almost alone among the structures at Glennallen, remained stable, even through the Alaskan Good Friday earthquake, March 27, 1964.

Although well water at Glennallen is hard, it is not characteristic of the high salinity of deep wells in the area. Most wells at Glennallen are less than 100 feet deep and do not intersect the saline aquifers 300 to 500 feet (40 to 150 m) deep.

185.9. Swale, crossed by road for next 0.2 miles (.32 km) is probably an early drainageway of Moose Creek.

185.0, 184.6, 184.3. Three minor strandlines occur at about 1,510 (460 m), 1,545 (471 m), and 1,560 (475 m) feet. Other, higher strandlines recognized elsewhere in the southeastern part of the Copper River basin lie at 1,600 (490 m), 1,700 (520 m), 1,800 (550 m), and 1,900 (580 m) feet, but are not apparent along the road, probably because of modification by drainage swales that developed after drainage of the lake at the close of Wisconsin time.

183.3. Road crosses 1.2 miles (1.9 km) of sand and gravel deposits which occur in a broad swath swinging north from the Tazlina River. The swath apparently represents an early course of the Tazlina River that formed about 9,000 years ago, soon after drainage of the lake. Poorly sorted sandy gravel and gravelly sand sediments are exposed in the gravel pit north of the highway at mile 182.7 and south of 182.4.

182.7. A well drilled at Glennallen Lodge to a depth of 502 feet (153 m) encountered Na-Ca-Cl water with 3,240 ppm dissolved solids.

182.0. Construction of the highway for the next 3.3 miles (5.3 km) has impeded the normal flow of drainage in a series of poorly drained swales or muskegs on the lacustrine plain. In these areas water has collected along the shoulders of the highway. Continuous thawing of permafrost in these areas and in culverts, in part by water, causes differential subsidence of the road and calving of the shoulders. Seasonal freezing of the wet ground produced considerable annual frost heaving in wet areas under the road prism.

177.1 to 176.3. Several small, almost imperceptible strandlines lie at altitudes of 2,000 (610 m), 2,050 (625 m), and 2,090 (637 m) feet in the next 0.8 miles (1.3 km). On the strandlines are discontinuous deposits of poorly sorted sand and sandy gravel, generally less than 10 feet (3 m) thick; some of these deposits were used as limited sources of borrow road construction.

173.2. Stop 8-55: Tolsona No. 1 mud volcano (fig. 8-42).

Tolsona No. 1 cone of the Tolsona group of mud volcanoes lies 0.12 miles (0.19 km) north of the highway. This mud volcano is one of four mud volcano cones and two mineral springs that compose the Tolsona group of mud volcanoes.

This group lies largely west of the Copper River and contains much smaller cones than those of the Drum group east of the river (fig. 8-43). Springs discharging from the Tolsona group emit methane gas and cool sodium and calcium chloride water.

The Tolsona No. 1 mud volcano, at an altitude of 2,045 feet, is about 25 feet (7.5 m) high, 600 feet (180 m) wide, and 900 feet (270 m) long. The crater is about 30 feet (9 m) in diameter. The temperature of the water discharged from the vents ranges between 38° and 55°F (3.3°-12.8°C). The cone has gently sloping sides rising to a slightly domed crest on which several areas of activity and of individual gas and water vents have varied from year to year.

Nichols and Yehle (1961) suggested that the water could be derived from meteoric and/or other water circulated with a saline ground water formed by evaporation from glacial lakes and later reconcentrated by permafrost, from highly diluted volcanic water, or from a connate source. They proposed that the gas is probably derived from coal beds similar to the Tertiary or older coal beds exposed near Atlasta House 7.5 miles (12.5 km) to the west, or from marsh gas from buried Pleistocene deposits, or alternatively, from slightly organic, nonpetroliferous connate water associated with Cretaceous deposits.

Grantz and others (1962) proposed that the gas and water are connate in origin and are derived from Cretaceous or older marine rocks that are inferred to underlie the area. In support of this thesis, they report the presence of Foraminifera, *Inoceramus* shell fragments, echinoid spines and plates, and ophiuran ossicles on the surface of the cone. A few *Inoceramus* prisms have been collected from nearby till and proglacial lake deposits but they are strongly abraded. They point out that although the gas resembles marsh gases, and contains little of the higher hydrocarbons, similar gases occur in the oil fields of the United States.

The age of the cone is assumed to be late Pleistocene. Authors of both papers referenced state that the small size of the cones and the lack of included angular rock fragments suggest that the cone formed largely by quiet, gradual accretion of mud rather than by explosive action.

Tolsona No. 2 mud volcano lies 0.4 miles (0.64 km) north-northeast of Tolsona No. 1, but cannot be seen from the highway.

173.1. Road descends bluff composed of glaciolacustrine silt and clay deposits and mantled by colluvium.

172.8. Gravelly terrace deposits of Tolsona Creek.

172.6. Massive lacustrine pebbly silt and varved silt and clay overlying till are exposed in bluffs.

171.9. Lacustrine plain composed largely of pebbly clayey silt.

171.2. Enter Gulkana A-5 Quadrangle.

170.0. Right; side road to Tolsona Lake Resort. Road follows site of former esker now removed for road material. Esker at resort on lakeshore consists of gravel deposits about 80 feet (24 m) thick which rest on glaciolacustrine deposits. Most of the surficial deposits in this area are lacustrine silt, clay, stony silt, and stony clay. Both Moose Lake, located northwest of Tolsona Lake, and Tolsona Lake have small seepages of gas and saline water. For the next 4 miles (6.4 km) highway crosses a series of north-south trending till-cored ridges which are draped with lake sediments.

166.2. Right; Atlasta House. Hill behind lodge is formed of poorly consolidated sandstone, coal, clay, and gravel of Tertiary (Oligocene) age. Coal beds as much as 2 feet (0.6 m) thick occur.

164.0. Right; for the next 6 miles (9.6 km), a moraine of late Wisconsin age occurs as hummocky terrain on the southern slope of Tazlina Hill. This moraine probably is about 11,000 or 12,000 years old. To the left (south) Tazlina Lake and Tazlina Glacier occur in the distance. Because the lake is fed by glacial meltwater it has a considerable amount of gray rock flour in suspension, and its bottom is covered by the same material. The lake is at least 370 feet (110 m) deep. At the outlet of the lake, at its northern end, the Tazlina River crosses conglomerate of the Matanuska Formation (Cretaceous), and then enters a canyon cut into Pleistocene glacial and glaciolacustrine deposits.

156.0. Left; Tazlina Glacier Lodge. Highway is underlain by stony silt or clay, and the lake behind the lodge is perched on this same impermeable material. A well dug beside the lake encountered dry gravel, and water pumped into the well disappeared into the gravel. Many of the lakes in this part of the Copper River basin are perched on impermeable sediments and do not reflect the water table in the underlying materials.

154.8. Right; large pit is at the base of a hill around which a shoreline related to the 2,450-foot (750 m) lake level was formed. The former shoreline is marked by a wave-cut notch with boulders at its base, which have been winnowed from the underlying till. For approximately the next 9 miles (14.5 km) the highway crosses a broad valley which was occupied by a northward-moving glacial lobe during late Wisconsin time. Within this valley the highway crosses several north-south-trending till ridges which are draped with lacustrine sediments. The intervening lowlands generally are underlain by fine-grained lacustrine deposits.

154.1. Enter Gulkana A-6 Quadrangle.

146.0. For the next 4 miles (6.5 km), highway winds through area of kames and eskers composed of sand and fine gravel.

144.5. Enter Valdez D-8 Quadrangle.

142.0. Right; Slide Mountain. For the next 4 miles (6.5 km), highway follows along the southern flank of Slide Mountain (fig 8-45). Large landslides in shale of the Matanuska Formation have moved down this unstable slope in the last 10,000 or more years. Some slides are stabilized and spruce forests are growing on them. Others have moved recently—movement in 1946 formed the large scar that can be seen from near mile 141.0. Minor movement occurred in several places within the slide zone during the earthquake of March 27, 1964. The brown rock at the summit of the mountain is poorly consolidated gravel and sand of Oligocene age.

138.0. Highway begins steep descent into canyon of Little Nelchina River. Shale has been used as fill; consequently, road has slumped in several places.

137.4. Little Nelchina River. Shale of the Matanuska Formation, capped with terrace gravel, occurs along the east side of the river near the bridge. The age of the terrace gravel and of the canyon is slightly older than 10,250±250 years (W-767); therefore the glacier had to have retreated from this point before that date. Highway begins ascent to till-covered surface bordering the canyon of the Little Nelchina River.

135.5. Enter Anchorage D-1 Quadrangle.

133.0. Stop 8-56: Nelchina Glacier view.

To the south there is a good view of the Nelchina Glacier, the broad outwash plain in front of the glacier,

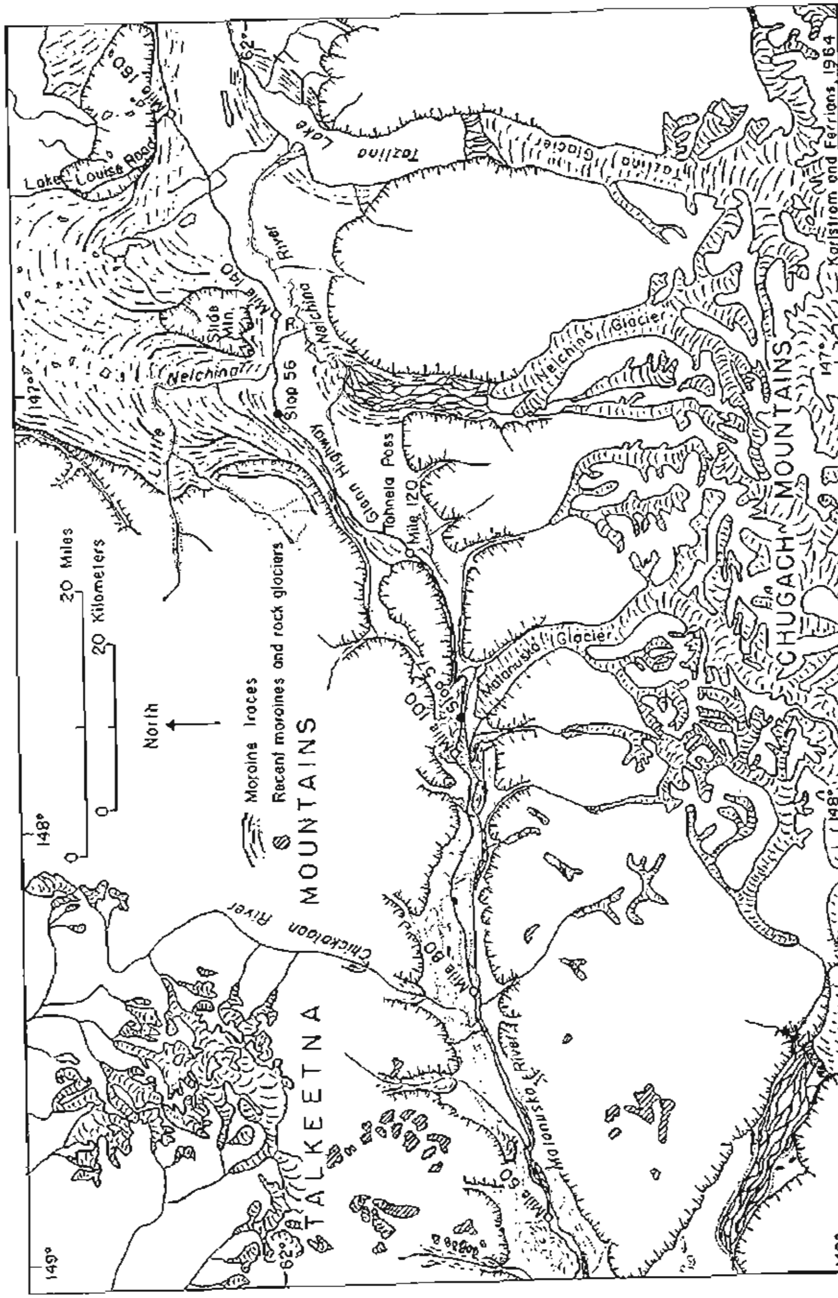


Figure 8-45. Index map of the upper Matanuska River Valley and Nelchina River area, Alaska showing Field Trip Stops.

and the Nelchina River canyon which is cut into glaciolacustrine deposits and other glacial drift. The western part of the end moraine of a major glacial advance during late Wisconsin time can be seen to the northwest.

During Wisconsin time the Talkeetna Mountains, remote from moist maritime air, were not heavily glaciated. Glaciers from the Chugach Mountains, because of their proximity to these air masses, advanced more than 50 miles (80 km) north of their present position and coalesced to form a great piedmont glacier which fronted in a glacial lake. Evidence of pre-Wisconsin glaciation is obscure and lies either at high elevations or in areas north of the highway; therefore it is not discussed here.

In late Wisconsin time the expanded glaciers from the Nelchina and Tazlina valleys in the Chugach Mountains were separated by low bedrock hills into three lobes; a lobe moved northward through the Little Nelchina valley, a lobe moved northward through the valley now occupied by Old Man Lake east of Slide Mountain and west of the Lake Louise Road, and a lobe moved eastward down the Tazlina valley below Tazlina Lake. Outwash from the Little Nelchina lobe was graded to deltas at about 2,650 feet (808 m) above sea level. Successively younger deltas, which can be traced back along outwash channels to recessional moraines, were formed at vertical intervals of about 50 feet (16 m) as the lake level lowered. One of these outwash deposits, which is graded to a delta at about 2,500 feet (760 m), is dated radiometrically as slightly older than 13,280±400 years (W-583). The lake level had lowered to about 2,450 feet (750 m) by the time the ice strongly readvanced, the last of the major readvances of Wisconsin age. As the ice retreated from this moraine, lake level held steady at 2,450 feet (750 m), forming weakly developed beaches. Apparently the lake at this level was held in by a bedrock threshold, perhaps in the Susitna canyon. The age of this advance and of the lake at this level is between 10,250 and 13,280 years old, and it is probably about 11,000 to 12,000 years old.

As the glacier retreated it split into two individual valley glaciers, that from the Nelchina Valley, and that from the Tazlina Valley. About this time, probably about 10,250 years ago, the lake level dropped from 2,450 feet (750 m) to a level of about 2,300 or 2,350 feet (700 or 715 m), which persisted until the glaciers retreated back to within 8 or 10 miles (13 to 16 km) of their present positions. This drop in lake level marks the beginning of the disintegration of the ice dam blocking drainage through the Chugach Mountains. Although there were halts in the lowering, and possibly some rises in lake level, the lowering of the lake was rapid, and according to evidence in the east-central part of the basin, the lake level was below 1,700 feet (520 m) before 9,440±300 years (W-714) ago.

Since withdrawal of the glaciers, minor advances have taken place, and the major rivers have cut deep canyons through the glacial and glaciolacustrine deposits. Also, a minor amount of loess and some cliff-head dune sediments have been deposited along the northern edge of the river canyons.

133.0. From here to Eureka Lodge, at mile 128, the highway follows a series of low till ridges that form a summit more than 3,300 feet (1,000 m) above sea level.

130.0. Right; an unsuccessful oil well 8,546 feet (2,600 m) deep (Eureka No. 2) was drilled here. Shale of the Matanuska Formation is overlain by glacial deposits.

128.0. Eureka Lodge. For next 7 miles (11.2 km) highway crosses Tahneta Pass over an undulating surface that has been scoured by overriding glaciers. Shale of the Matanuska Formation is very close to the surface, and the road fill is largely shale obtained from pits along the southern flank of Sheep Mountain and from side borrow. The shale is a poor aquifer, and attempts to obtain fresh water from it have failed.

125.0. Right; site of Alaska Oil and Gas Development Company well, Eureka No. 1. Well was drilled in 1953-54 and reached a depth of 4,818 feet (1,460 m) without obtaining production of oil or gas.

120.0. For the next 3 miles (4.8 km), highway winds along the southern shoulder of Sheep Mountain. Roadcuts are alternately in till, in gravel, or in sandstone of the Matanuska Formation. Fossil mollusks and some leaves can be found locally in some of the exposures of sandstone.

116.0. Left; a good view of northern part of Chugach Mountains and of the divide separating the Matanuska Valley drainage basin from the Copper River drainage basin. Glacial ice probably moved westward from the Copper River basin to the Matanuska Valley during Wisconsin time, but during the later part of Wisconsin time the pass probably was free of ice and was the site of outwash streams emanating from glaciers in the Copper River basin.

115.0. For the next 3 miles (4.8 km), highway crosses large alluvial fans along the southern side of Sheep Mountain. The rusty color of the gravel is caused by oxidation of iron by large amounts of sulfuric acid in the water. The acid comes from oxidation of pyrite in rocks of the Talkeetna Formation which is of Early Jurassic age. Within some gullies of the mountains, talus and alluvial deposits of Pleistocene age are cemented into conglomerate by iron oxide. During early Pleistocene time Sheep Mountain probably was completely covered by glacier ice from sources in the Chugach Mountains; erratic boulders from that source are found as high as 200 feet (60 m) below the summit.

112.0. For the next 2 miles (3.2 km), on the right, the angular accumulations of rock in talus cones and in small inactive rock glaciers formed from frost-shattered rock of the Talkeetna Formation. The contact between the Talkeetna Formation, chiefly volcanic rocks altered to greenstone, and the Matanuska Formation is a fault that follows the southern slope of the mountain just north of the highway.

109.1. Right; major roadcut in toe of an inactive rock glacier consisting of boulders of Talkeetna Formation in a matrix of fine-grained material.

108.8. Right; roadcut. Radiometric date of organic material collected at contact between outwash and overlying talus indicates that the outwash deposit is older than 3,620±250 years (W-573).

108.0. Begin steep descent into canyon of Caribou Creek.

107.1. Caribou Creek. After crossing bridge, highway ascends steep bedrock slope. Bedrock is Talkeetna Formation.

106.1. Right; sand deposits. These deposits and others at similar elevations across the valley of Caribou Creek are ice-contact deposits that bordered an advanced position of the Matanuska Glacier during late glacial time, sometime prior to 4,000 years ago.

105.9. To the south, Glacier Point, a steep glacier-scoured hill of porphyritic intrusive rock of Tertiary age. The south face overlooking Matanuska Glacier is nearly vertical and is 1,400 feet (430 m) above the Matanuska River.

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UPPER COOK INLET AREA AND MATANUSKA RIVER VALLEY

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RESUME OF THE QUATERNARY GEOLOGY OF THE UPPER COOK INLET AREA AND MATANUSKA RIVER VALLEY

Upper Cook Inlet is the most densely populated region in Alaska, having in the metropolitan area of Anchorage and in the adjoining rural areas of Palmer and the Kenai Lowland nearly half of the total population of the state. The economy of the region is principally based on federal support of large military installations, on expanding exploitation of oil and gas reserves found on the Kenai and adjoining lowlands, on salmon fishing, and on agriculture. The region has a modified marine climate characterized by a moderate range between summer and winter temperatures and by relatively low precipitation. At Anchorage the mean annual temperature is 36°F (2.2°C) and mean annual precipitation is 14.5 inches (36.8 cm).

Below treeline (between 2,000 and 3,000 feet [600 and 900 m] above sea level) the better drained surfaces of the region are covered by interior type forest including white spruce, birch, aspen, cottonwood, willow, and alder. Vanguard coastal species of mountain hemlock and Sitka spruce occur locally along the eastern and southern margins of the region. A significant part of the lowlands is poorly drained and characterized by extensive areas of muskeg and marsh vegetation of sedge, grass, moss, low brush, and black spruce and birch.

The extensive lowlands at the head of Cook Inlet occur at the intersection of three main valley systems: The Susitna River valley that lies between the Alaska Range and the Talkeetna Mountains to the north; the Matanuska Valley that separates the Talkeetna Mountains from the coastal Chugach Mountains to the east; and the Turnagain Arm Fiord that separates the Chugach Mountains from the Kenai Mountains to the south and southeast. Peak elevations of the coastal mountains range in general from 3,000 feet (900 m) to more than 13,000 feet (4,000 m), those of the interior Talkeetnas from 3,000 feet (900 m) to more than 8,000 feet (2,400 m), and those of the Alaska Range from 4,000 feet (1,200 m) to more than 20,000 feet (6,000 m).

The region has been repeatedly glaciated. The surrounding rugged mountains are alpine in character and retain on their higher portions numerous valley glaciers and extensive ice fields. The existing glaciers are larger and occur at lower elevations near the coast, reflecting a modern climatic snow line that is

steeply inclined coastward from more than 6,500 feet (2,000 m) in the northern part of the Alaska Range to less than 3,000 feet (900 m) along the coastal margin of the Chugach Mountains in the Prince William Sound region. Climatic snow line in the intermediate Talkeetna Mountains lies as high as 6,000 feet (1,800 m) elevation. The lowland areas are all underlain by thick sections of glacial deposits and are characterized in most areas by hummocky morainal topography.

Five Pleistocene Glaciations

During the earliest recognized and most extensive glaciations (the Mount Susitna [oldest] and the Caribou Hills) expanding ice caps on the Chugach and Talkeetna Mountains completely filled the Matanuska Valley; largely, if not completely, covered the bordering mountains; and pushed out to join other major glaciers to fill upper Cook Inlet to elevations of about 3,000 to more than 4,000 feet (900 to 1,200 m). Evidence for these glaciations is fragmentary, and is restricted to remnants of highly weathered glacial drift and erratic material found at high elevations only at great distances from the main alpine nourishment centers in the surrounding mountains. These deposits have been mapped where they lie above less modified moraines of the younger Eklutna Glaciation, on the upper slopes of Mount Susitna, along the southwest flank of the Talkeetnas, and locally along the Chugach and Kenai Mountain fronts south of Anchorage. On protected slopes, deposits of Caribou Hills age still retain crude morainal form; on higher slopes the evidence for the Mount Susitna Glaciation is in the form of highly modified drift remnants or scattered erratic material. During these and all succeeding glaciations, ice in the Matanuska Valley was coextensive with ice that spilled into the Copper River basin in the vicinity of Tahneta Pass at the head of the valley.

During the next glaciation (the Eklutna) the upper part of Matanuska Valley was again buried beneath a vast ice cap except perhaps for scattered nunataks along the valley margins. Below the confluence of the Matanuska and Knik Valleys, however, many of the divides between tributary glaciers flowing out of the Chugach Mountains stood above the ice surface as well as above a large upland area on the southwest flank of the Talkeetna Mountains. As marked by lateral moraines on both the Talkeetna and Chugach Valley walls, ice filled the valley up to elevations of about 3,000 feet (400 m) near Palmer and about 1,000 feet (300 m) near Anchorage. West of Anchorage, the compound Matanuska-Knik ice lobe coalesced with ice flowing out of the Susitna River Valley, Turnagain Arm, and other trunk valleys draining the Kenai Mountains and Alaska Range to completely cover the floor of Cook Inlet for the last time in the Pleistocene.

During the last two glaciations (the Knik, older, and the Naptowne), the Matanuska-Knik ice lobe did not coalesce with ice flowing down the Susitna River and the Turnagain Arm Valleys. End moraines were deposited in the lowlands surrounding Anchorage. The Elmendorf moraine lies just north of Anchorage and represents the maximum Naptowne extent of the Matanuska-Knik ice lobe. As marked by remnants of end moraines partially buried in the gravel plain bordering the Elmendorf moraine, and by buried drift exposed in sea bluffs at the lower end of Knik Arm, the ice lobe of Knik age was from 3 to 6 miles (4.8 to 9.6 km) more extensive at its maximum than the ice lobe of Naptowne age. Proglacial lake deposits associated with these and correlative moraines along with vague strandline features and compound hanging deltas locally preserved throughout the upper part of Cook Inlet indicate that the glaciers of Knik and Naptowne age advanced into and retreated from regional proglacial lakes. The regional distribution of Naptowne and Knik moraines in the southwestern, more constricted part of Cook Inlet indicates that during both glaciations ice from the Kenai Mountains coalesced with ice from the Alaska Range to form a high ice dam that created the interior lake environments. Because of the instability of the ice dam, lake levels fluctuated drastically. Instead of one or several well-defined simple strandlines, broad zones of poorly to fairly well defined terracettes, were characteristically developed in Cook Inlet region.

The inference of major intervals of retreat and weathering between the named glaciations suggested by geomorphic relations of moraines in the region is supported by the following stratigraphic data. Sea bluffs at the lower end of Knik Arm and near Boulder Point on the Kenai Lowland transect moraines of both Naptowne and Knik age. In both localities the Naptowne Till is separated from the Knik Till by a sequence of weathered deposits including peat layers and locally a marine horizon recording a higher sea level stand, the Woronzofian transgression, during the Knik-Naptowne interval. In both localities Knik Till unconformably overlies deeply weathered drift of Eklutna age. Measured oxidation profiles on Naptowne drift range from 2 to 10 feet (0.6 to 3.0 m), those on buried Knik drift from 5 to 20 feet (1.5 to 6.1 m), and those on Eklutna drift are greater than 40 feet (12 m). Deposits throughout Cook Inlet contain little or no carbonate material, and it has not been possible to determine depths of leaching. Statistical studies of surface boulders on moraines of Naptowne, Knik, Eklutna, and Caribou Hills age indicate progressive and distinctive differences in granite to graywacke-argillite boulder ratios. The boulder ratios suggest approximately equal intervals of time between the Naptowne, Knik, Eklutna glaciations and about twice as long an interval between the Eklutna and Caribou Hills glaciation. Data bearing on the duration of the Mount Susitna/Caribou Hills interval are scanty, but it is believed that it was at least as long as the shortest of the preceding interglacial intervals.

Naptowne advances and associated proglacial lake phases

The Naptowne moraines throughout Cook Inlet region include two or three belts of recessional moraines, suggesting intervals of important stillstands or readvances following retreat from the outermost belt of moraines. On the Kenai Lowland these have been named from type localities the Moosehorn (oldest), the Killey, the Skilak, and the Tanya moraines. From the level of strandlines, abandoned meltwater channels terminating in hanging deltas, and of lake deposits associated with the type Naptowne morainal belts, the following high lake phases of Glacial Lake Cook have been reconstructed: *ca.* 750 feet (230 m) elevation or higher during the Moosehorn maximum; 500 to 600 feet (150 to 180 m) during the Killey maximum; *ca.* 250 to 300 feet (75 to 90 m) during the Skilak maximum; and *ca.* 100 to 150 feet (30 to 45 m) and *ca.* 50 feet (15 m) during Skilak recession and prior to final drainage of the lake. Partial or complete lake drainages between these high lake phases are recorded by diastems, buried soils, and peat layers in the lake bottom sediments. The intervals of lake drainage reflect failures of the ice-dam in lower Cook Inlet during the periods of general glacial recession; the high lake phases reconstitution of effective ice-dam conditions during glacial advances.

The Matanuska-Knik ice-lobe retreated from the same regional lake, and experienced the same sequence of Naptowne recessional events. Correlations with the type deposits on the Kenai lowland are based on comparable morainal sequences and on the evidence that during the advance marked by the third belt of moraines meltwater streams flowing from the ice front emptied into the same 250 to 300 foot (75 to 90 m) high lake phase of the Skilak maximum. Well-defined, meltwater channels graded to these moraines west of Palmer terminate downvalley at this critical regional strandline level.

Differences in the type of deposits laid down during Naptowne recession of the Matanuska-Knik ice lobe strikingly reflect its compound character. Long crevasse-fill ridges confining oriented lake basins resulted from stagnation of ice of the Matanuska sublobe following retreat from the Killey end moraines. These stagnation deposits are separated by an interlobate complex of eskers from typical recessional moraines deposited by the Knik sublobe that apparently remained active because of shorter distance to nourishment center. This dynamic difference in regimen is further recorded in the outer boundary of the Skilak moraines that indicates that the Knik sublobe advanced farther downvalley than did the ice derived from the more distant Matanuska Valley sources. Similarly, whereas during Skilak recession, the Matanuska sublobe ice began to stagnate in place as recorded by the spectacular development of ice-contact deposits about Palmer, the Knik sublobe remained active and on retreat determined a uniform series of progressively lower marginal terraces deposited by meltwaters diverted along the retreating ice front. The more active Knik sublobe eroded a deep trough that is now occupied by Knik Arm, whereas bedrock lies at much shallower depth beneath the deposits laid down by the less active Matanuska sublobe.

Recession in Matanuska River valley

Because of the numerous tributary glaciers that supplied ice to the Matanuska sublobe and because of the complications resulting from local stagnation of glacial ice, the Naptowne recessional history up the Matanuska River valley is somewhat obscure. The late recessional history does not involve progressive withdrawal of a single glacier but involves the independent movements of tributary glaciers that receded from the trunk valley and into their own valleys at different times. Details of Naptowne recession from the Matanuska River valley must therefore await detailed mapping in the tributary valleys as well. Some of the broader features of recession within the main valley, however, may be derived from available data. High-level gravel terrace deposits and quiescent-water deposits along the axis of the valley indicate that the middle part of the valley became ice free before the stagnating ice of Skilak age had melted out of the mouth of the valley. This stagnating ice plug acted as a dam to drainage out of the valley and determined aggradation in the middle ice-free part. Subsequent downcutting by meltwater stream through the ice plug resulted in a series of kettle-pocked terrace deposits. According to radiocarbon dating of moraines in the upper part of the valley, ice had already retreated within 1 mile (1.6 km) of the present front of Matanuska Glacier sometime prior to 8,000 years ago. The moraines deposited by the Matanuska Glacier

	COOK INLET GLACIAL CHRONOLOGY	ASSOCIATED EVENTS IN COOK INLET		DATED MARINE TRANS-GRESSIONS		CORRELATIONS WITH CONTINENTAL CHRONOLOGIES: N. AMERICA	N. EUROPE	SIBERIA
ALASKAN GLAC.	TUNNEL ADVS. A.D. 500 TUSTUMENA ADVS.	Minor sea level fluctuations	KRUZENSTERNIAN	GIRDWOODIAN	RECENT	"Little Ice Age"	Sub-Atlantic* Sub-Boreal*	Postglacial*
	3500 B.C.	High Sea Level Stand		KASILOFIAN 5,000-6,000		ALTITHERMAL*	ATLANTIC*	
NAPTOWNE GLACIATION	TANYA ADVS. 7000 B.C. SKILAK ADVS. 10,500 B.C. KILLEY ADVS. 13,500 B.C. MOOSEHORN ADV. 17,000 B.C. Pre-MOOSEHORN ADVS.	Glacial Lake Cook: final drainage 275' high lake drainage 500-600' high lake drainage 750' high lake drainage? Advance lake phases			CLASSICAL WISCONSIN	COCHRANE* MANKATO OR VALDERS* CARY* TAZEWELL* Morton, Farmdale, Roxana Loesses*	Boreal* Pre-Boreal* O & Y Dryas* Bolling* OLDER WURM II DEPOSITS	NORILSK (SARTAN) GLACIATION*
	45,000 B.C.	High Sea Level Stand		WORONZOFIAN 38,000-48,000		PORT TALBOT*	GOTTWEIG*	KARGINIAN* TRANSGRES.
	KNIK GLACIATION	Proglacial Lake				PRE-CLASSICAL WISCONSIN	WURM I	ZYRYANKA
	85,000 B.C. EKLUTNA GLACIATION	Deep Weathering		PELUKIAN 78,000-100,000		SANGAMON	R/W	KAZANTSEVIAN TRANSGRES.
		Cook Inlet Icefilled				ILLINOIAN	RISS	YENESEY
		Major Weathering		KOTZEBUAN 170,000-175,000		YARMOUTH	M/R	OPLYUNY-SAMBURG BEDS
	CARIBOU HILLS GL.	Cook Inlet Icefilled				KANSAN	MINDEL	BAKHTA
	210,000 B.C.	Deep Weathering		MIDDLETONIAN 210,000-224,000		AFTONIAN	G/M	PANTELEYEV BEDS
	MOUNT SUSITNA GL.	Cook Inlet Icefilled				NEBRASKAN	GUNZ	BAIKHA
	Beginning of Quaternary uncertain			Isotope dates in years B.P.		*Isotope-dated deposits		

Table 2. Cook Inlet glacial record, chronology and continental correlations (summarized from Karlstrom, 1956, 1961, and 1964; and for Siberia from V. A. Zubakov, pers. comm. 1963). Deposits defining the informally named marine transgressions *Kotzebuan*, *Pelukian*, *Woronzofian*, and *Kruzensternian* are described by Hopkins (1965). The Cook Inlet deposits defining the *Woronzofian*, *Kasilofian* and *Girdwoodian* transgressions are described in Karlstrom, 1964; the *Woronzofian* is defined from the middle marine horizon of the Bootlegger Cove Clay near Pt. Woronzof, the *Kasilofian* from the elevated tidal flats near the mouth of the Kasilof River, and the *Girdwoodian* from the thick middle tidal silt section in the Girdwood Bog. The *Middletonian* transgression is defined by bedded sand deposits in unit 20 of the thick conglomeratic glaciomarine section described from Middleton Island by Miller (1953) and reexamined by the author in 1962. The isotope dates applied to these inferred marine transgressions are discussed in Blanchard (1963); Hopkins (1965) and Karlstrom (1964). The Middleton Island isotope dates are incompatible with Hopkins' (1965) designation of the section as Pliocene in age but not with the interpretation that the fossils on Middleton Island indicate that the upper part of the section is definitely Pleistocene, and probably pre-Yarmouth, in age (MacNeil and others, pers. comm. 1964). The *Kasilofian* and *Girdwoodian* are events of the *Kruzensternian* transgression as this is broadly datdd as postglacial by Hopkins (1965).
Note: The stratigraphic nomenclature used here is that of the published sources and does not necessarily conform to that of the U. S. Geological Survey.

just prior to this late recession are associated with abandoned glacial spillways graded approximately to the present gradient of the postglacial Matanuska River canyon. Establishment of present grade this far up the valley at this early time emphasizes the general rapidity of ice recession out of the Matanuska Valley and of the drainage adjustment down through the stagnating ice plug following retreat from the Skilak maximum.

Post-Naptowne sea level changes and the Alaskan Glaciation

Following Naptowne recession and final drainage of Glacial Lake Cook around 9,000 years ago, rising sea levels caused progressive marine invasion of Cook Inlet trough. Highest postglacial sea levels, contemporaneous with maximum contraction of glaciers in the region, attained an elevation of from 5 to 10 feet (1.5 to 3.0 m) above present datum. This high sea level stand is recorded by elevated tidal flat deposits found near the head of Knik Arm and elsewhere through Cook Inlet. It is informally named the Kasilofian transgression from elevated tidal flats at the mouth of the Kasilof River on the Kenai lowland.

Radiocarbon dating of coastal bogs places the culmination of this high sea level phase at between 5,000 and 6,000 years old or Altithermal in age. Low terraces just above the present flood plains of the Matanuska and Knik Rivers are graded to these elevated tidal flats, and thus appear to record aggradation of the Knik and Matanuska Rivers during the Altithermal and post-Altithermal periods. The subsequent falls in sea level were contemporaneous with general advances of glaciers throughout the region. Two sets of moraines characterized by slight differences in vegetation cover and soil development lie close to present glacier fronts. Radiocarbon dating of these moraines places the outermost set (the Tustumena moraines) between 3,500 B.C. and 500 A.D. and the innermost set (the Tunnel moraines) between 500 A.D. and the present. As cross-dated with tidal bog records in Cook Inlet, these post-Altithermal Tustumena and Tunnel advances were contemporaneous with a series of minor sea level regressions. The last minor sea level regression culminated about 1850, since that time the tidal bog record is of progressive but oscillatory transgression.

The Cook Inlet chronology and continental correlations

The glacial and marine deposits of Cook Inlet are dated in part by the radiocarbon and ionium/uranium ratio dating methods, and in part by roughly quantitative relative geologic data. The internally dated Cook Inlet chronology and provisional correlations with the North American and Eurasian continental glacial sequences are summarized in the accompanying chart (Table 2).

Road Log and Locality Descriptions

101.7.[1] Anchorage D-3 Quadrangle.

Stop 9-57: Scenic overlook of Matanuska Glacier.

The Matanuska Glacier is one of the largest glaciers draining the interior side of the central ice fields of the Chugach Mountains. It drains 250 square miles (647 km²) of the fiords at the head of Prince William Sound.

During the maximum of the Naptowne Glaciation (around 20,000 years ago) expanded ice fields almost completely covered the bordering mountains, with a few higher peaks rising above the ice surface as nunataks. Sheep Mountain may have been completely covered by

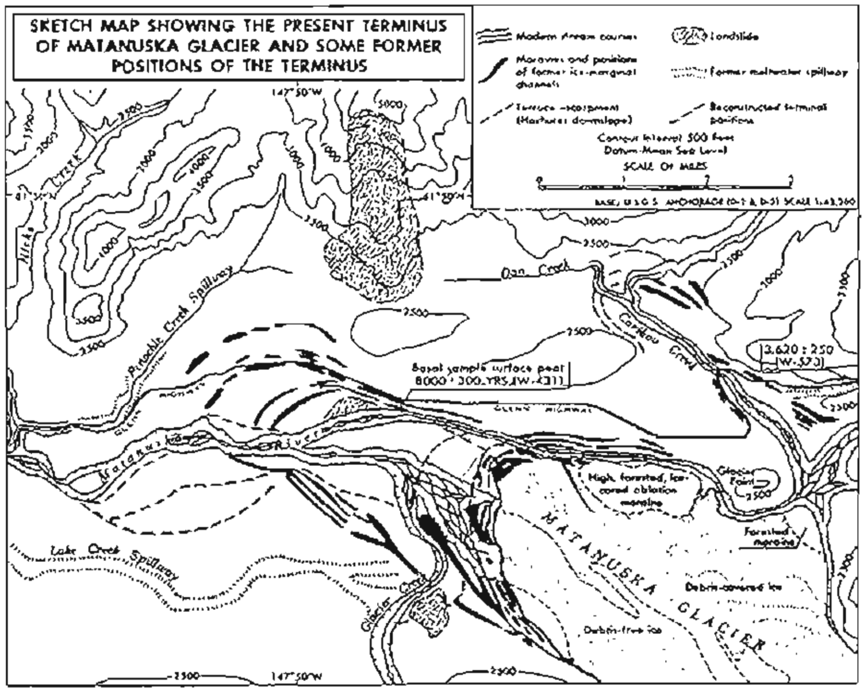
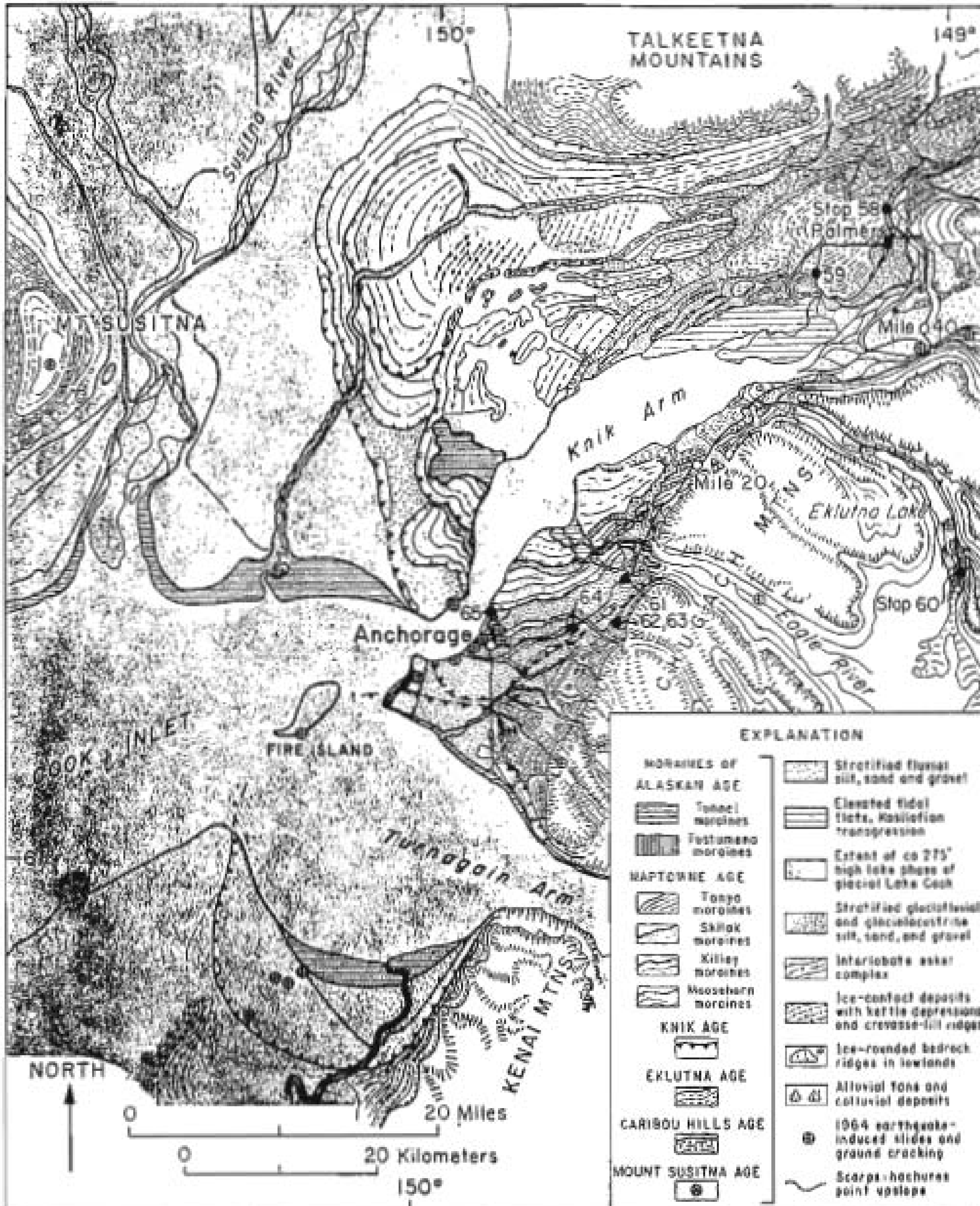


Figure 9-46. Sketch map showing the present terminus of Matanuska Glacier and some former positions of the terminus. (From Williams and Ferrians, 1961, fig. 1)

highest part of the mountains between Mt. Witherspoon (12,023 ft; 3,665 m) and Mt. Marcus Baker (13,176 ft; 4,016 m). Glaciers draining the coastal side of the same high ice fields reach tidewater in deep ice, or may have jutted above the ice surface as suggested by remnant lateral moraines and lateral channel features concentrated between elevations of 4,000 to 5,000 feet (1,219 to 1,524 m). More work needs to be done to determine whether this ice boundary represents the Naptowne maximum or a recessional ice position. The virtually continuous ice

[1]Miles from Anchorage on the Glenn Highway.



Compiled by Thor Karlstrom, USGS, 1964

Figure 9-47. Glacial geology map of the Upper Cook Inlet area, Alaska, showing Field Trip Stops.

cap that formed at the head of Matanuska Valley and drained into Cook Inlet was coextensive with ice that flowed into the Copper River basin from an ice divide in the vicinity of Tahneta Pass.

Williams and Ferrians have delineated the following late Naptowne and post-Naptowne history of Matanuska Glacier (fig. 9-46).

(1) Rapid retreat from extended positions into the upper reaches of the valley accompanied by canyon cutting by meltwater streams.

(2) Readvance 2.5 to 5 miles (4 to 8 km) from the present glacier front blocking the Matanuska River and diverting Caribou and Glacier Creeks as recorded by moraines on lower valley slopes and by lake deposits in the floor of the upper part of the Matanuska Valley east of the glacier. During recession from this advance Pinochle Creek spillway was cut and graded approximately to or below present level of the Matanuska River flood plain. Radiocarbon dating of basal peat in a bog overlying moraines deposited during this late Naptowne recession indicates that the readvance to Pinochle Creek spillway predates 8,000 years ago.

(3) Continued recession to an ice-front position perhaps upvalley from the present terminus prior to the past several thousand years. Culmination of this recession is correlated with the buried soil (Altithermal?) overlain by talus and dated 3,620 ± 250 years (W-573).

(4) Readvance into the canyon cut during deglaciation and formation of outwash and associated moraines that are located 1 mile (1.6 km), ¼ mile (.4 km), and less than a quarter of a mile from the glacier front. From analysis of soils, vegetation cover, and historical records the recent moraines are dated as follows: the oldest set a few thousand years, but probably less than 4,000 years; the intermediate set probably greater than several hundred years; and the youngest set prior to 1898 when the first photographs of the glacier were taken. Since 1899 the glacier front has retreated little, but considerable thinning has occurred.

The history of Matanuska Glacier appears to be comparable in intensity of advance and sequence to that recorded by the morainal and coastal deposits in Cook Inlet.

From this point westward to Palmer the road crosses the drift-mantled, irregular bedrock floor of Matanuska Valley and, in places, descends into the proglacial Matanuska River and tributary canyons incised 200 to 500 feet (60 to 152 km) below the glaciated floor. The glaciated valley floor is underlain by relatively soft sandstone and shale of the Matanuska and Chickaloon Formations of Cretaceous and early Tertiary age, respectively. Both of these formations have been intruded by diabasic rocks of Tertiary age. Owing to superior hardness, the intrusive rocks rise as ridges and cliffs above the general floor level. The glacial topography of the valley floor is only in small part constructional. Most of the local relief reflects glacial erosional forms scoured into the thinly mantled bedrock. Modification of these glacial forms has been slight except where they are transected by canyons and gullies or buried beneath colluvial and alluvial deposits on steeper slopes. Reflecting their relative youthfulness, the surface drift deposits of Naptowne age throughout the valley have shallow soil profiles with depths of oxidation generally between 1 and 3 feet (0.3 and 1 m). Younger deposits such as recent end moraines, rock glaciers, and alluvial and colluvial deposits are correspondingly less weathered, and have oxidation profiles generally less than 6 to 12 inches (15.2 to 30.4 cm).

101 to 98. Moraines deposited during retreat of Matanuska Glacier from its stand near the Pinochle Creek. Roadcuts are largely in till and locally in lacustrine deposits. Underlying bedrock is shale and sandstone of the Cretaceous Matanuska Formation.

100. Hundred Mile Lake to right occupies a basin dammed by recessional lateral moraines on which the road is built. Ahead; rock glacier of Alaskan age on steep mountain slope.

98. Top of Pinochle Creek grade. Pinochle Creek is an underfit stream that occupies the glacial spillway (fig. 9-46) formed during the last recognized Naptowne readvance or stillstand position of the Matanuska Glacier. Bedrock in the valley walls of Pinochle Creek and in the roadcuts in shale of the Matanuska Formation. The shale is very susceptible to landsliding.

96.8 to 95.1. Hicks Creek Inn. On low terraces just above the flood plain of the Matanuska River canyon, and at approximately the same level as the Pinochle Creek spillway floor. Ascend wall of Matanuska River canyon cut in the Matanuska Formation. Roadcuts of thin drift and colluvial deposits on bedrock. Left; good view of Matanuska River canyon and of the glaciated valley floor flanking the Chugach Mountains.

95.3. Location of buried contact between the Matanuska and the Chickaloon Formations.

95.1. Packsaddle Creek. Creek occupies a broad-floored spillway cut into Tertiary shale and marks a stillstand of readvance position of the Matanuska Glacier. The Pinochle Creek spillway upvalley may represent a recessional phase during retreat from Packsaddle Creek or a distinctly separate advance. If both represent recessionals then the morainal drift between Pinochle and Packsaddle Creeks may be of Tanya age; if Pinochle Creek represents a separate advance, then the drift between Pinochle and Packsaddle Creeks is probably of late Skilak age. That the spillways are closely related in time is suggested by the fact that both are graded close to the present gradient of the Matanuska River flood plain. To right, large double-pronged rock glacier of Alaskan age on upper and middle slope of Strelshia Mountain.

94.5. On upper slope of large alluvial fan deposited by Muddy Creek in the Packsaddle Creek spillway. Roadcuts are in stratified alluvial deposits over glacial drift of late Naptowne age.

94.5 to 89.2. Road traverses broad, thinly drift-mantled valley floor underlain by sandstone and shale of the Tertiary Chickaloon Formation. More resistant diabasic skills and dikes cut the Chickaloon and form partly talus-covered cliffs on the slopes of Anthracite Ridge to right. Vegetation is mainly grass and brush (alder and willow) with scattered stands of white spruce and birch.

89.2. Enter narrow glacially scoured trough cut in Tertiary shale between intrusive rocks and occupied by Purinton Creek.

89.6. Enter Anchorage D-4 Quadrangle.

88.5 to 85.8. Along margin of large, irregularly floored glacial trough cut in Tertiary shale and sandstone between intrusive rock on right and the ridge of Matanuska sandstone and shale on left.

88. Weiner Lake and beautiful Long Lake at 87-86 occupy elongate depressions scoured in the trough floor.

85.3. Road passes through erosional gap in ridge of Matanuska Formation onto the Matanuska River canyon side.

85 to 82. On drift-covered bench along margin of Matanuska River canyon. Left; braided gravel flood plain and low forested terraces of the Matanuska River. Across the river large alluvial fans are developed along the lower courses of streams draining the steep Chugach Mountain front which is underlain by resistant rocks of Jurassic and Cretaceous age intruded by granite of Mesozoic or Tertiary age.

79 to 78.3. Descend into mouth of Chickaloon River canyon. Chickaloon River is one of the major tributary streams draining the Talkeetna Mountains to the right. Roadcuts expose 2 to 10 feet (.61 to 3 m) of laminated silt containing 3 discontinuous ash layers; over 0 to 20 feet (0 to 6 m) of sand and gravel; over 0 to 15 feet (0 to 4.5 m) of Naptowne Till on sandstone of the Chickaloon Formation of early Tertiary age. The Chickaloon River flows in a broad terraced valley bottom incised 200 to 500 feet (60 to 152 m) below the level of the glaciated valley floor.

78.3. Chickaloon River enters the Matanuska River just above a restriction in the Matanuska River canyon, the King Mountain Gorge. The narrow gorge is confined to the left by a large intrusive body of diabase and to the right by the massive granite intrusive underlying King Mountain.

78.3 to 74.4. Margin of Matanuska River flood plain below high terrace gravels and steep rock walls.

78.3 to 76.4. Canyon walls expose contorted ice-contact sand and gravel underlying a pitted, channeled terrace surface about 200 feet (60 m) above flood plain and at an elevation between 900 and 1,000 feet (274 to 304 m). The ice-contact deposits overlie a thick section of Naptowne Till on sandstone and shale of the Chickaloon Formation.

77 to 76.5. Low terrace, 5 to 15 feet (1.5 to 4.5 m) above flood plain, on which is located King Mountain Lodge and public campground.

76.5. Enter Anchorage D-5 Quadrangle.

76.5 to 74.4. Base of bedrock ridge underlain by Chickaloon and Matanuska Formations intruded by diabase and other types of intrusive rocks.

74.3 to 72.6. Rise from low terraces onto hummocky terrain with deep kettle depressions and eskerlike ridges. These coarse gravelly ice-contact deposits form a plug in the mouth of a largely drift-filled channel that probably represents a pre-Naptowne segment of the Chickaloon River Valley.

72.6 to 69.3. Low terrace remnants at base of steep bedrock canyon walls.

71.2. Enter Anchorage C-5 Quadrangle.

69.3 to 67. Edge of Matanuska River flood plain at base of bedrock ridge underlain by Matanuska Formation. The ridge forms a northeast-trending divide between the Matanuska River and the King River, a major tributary stream of the Talkeetna Mountains that enters the Matanuska River several miles downvalley. The drift-covered ridge is flanked by remnants of gravel terraces up to elevations of 1,100 feet (335 m). It is transected by a gravel-filled channel recording a previous course of the Matanuska or King Mountain River.

68.8. Gravel pit in channel gravels.

67.3. Sharp bend in the Matanuska River where it passes through a narrow gorge. The gorge transects structure and resulted from postglacial superposition of the Matanuska River down from the aggradation levels marked by the high level terrace gravels. Here such gravels occur at an elevation of *ca.* 1,000 feet (304 m) on both sides of the canyon.

67.0. Good exposures of folded and faulted shale and sandstone of the

Matanuska Formation in the gorge walls.

66.9. Low terraces at the mouth of King River. King River flows in a broad, terraced, flat-floored canyon excavated along the contact of the Matanuska and Chickaloon Formations. Remnants of high terrace gravels present on both sides of the canyon are graded to an end moraine deposited at the mountain front 5 miles (8 km) upvalley and to the high terrace gravels present in the Matanuska River canyon behind us. Ahead; good view of high level terrace remnant (*ca.* 1,000 feet [309 m] elevation) cut below till-covered ridge top.

66.9 to 63.8. Low terraces built below the confluence of the Matanuska and King Rivers. Right; steep 400-foot (122 m) scarp cut in drift-covered bedrock ridge underlain by Chickaloon Formation. End moraines in Little Granite Creek valley behind the ridge record a recessional position of the Matanuska Glacier in Skilak time just prior to retreat of the trunk glacier upvalley and of tributary glaciers back into the bordering mountains. The moraines in Little Granite Creek valley extend downvalley to about mile 65. Beyond this point the deposits mantling the ridge in front of the moraines are finely laminated sand and silt. Deposition of such fine-grained material adjacent to an ice front strongly suggests that during this phase of valley deglaciation the receding glaciers were fronted by a lake or an outwash plain of extremely low gradient. These relations virtually require that an ice dam or other drainage obstruction persisted downvalley after this part of the valley became partially ice-free.

63.8 to 60. Toe of coarse, cobble coalescent gravel fans deposited in the lower reaches of Granite and Eska Creeks that drain broad, U-shaped valleys in the Talkeetna Mountains to the right. The alluvial fans record a sequence of depositional intervals, and are cut below hummocky, kettle-pitted ice-contact deposits of Skilak age locally characterized by long, sinuous crevasse-fill gravel ridges. These deposits have extensive distribution in the lower part of the Matanuska Valley between here and beyond Palmer for a distance of more than 10 miles (16 km). They occur up to elevations of 1,000 feet (304 m).

The field relations of these ice-contact deposits to the high terrace gravels and the high-level quiescent-water deposits upvalley indicate that a plug of stagnating ice persisted after the middle part of the valley had become partly ice-free. This ice plug acted as a dam causing aggradation and development of the high-level terraces near Chickaloon River (1,200 feet [365 m] elevation) and near King River (*ca.* 1,000 feet [300 m] elevation). Progressive downcutting through the stagnating ice plug by meltwaters trapped upvalley is recorded by a series of kettle-pitted terraces that will be crossed ahead.

61. Enter Anchorage C-6 Quadrangle. New milepost numbers begin.

60.58. Rise from Eska Creek valley onto terraced ice-contact deposits. Six crude terrace surfaces are crossed. The higher terraces are more pitted and irregular in topographic relief than the lower ones, recording progressive melting of buried ice blocks. The ice-contact deposits and associated pitted terraces lie upvalley from the Skilak maximum moraines dated *ca.* 12,000 years old, and above stream gradients developed in the valley prior to 8,000 years ago and associated with the Pinochle Creek and Packsaddle spillways. The recorded sequence of ice-stagnation, aggrada-

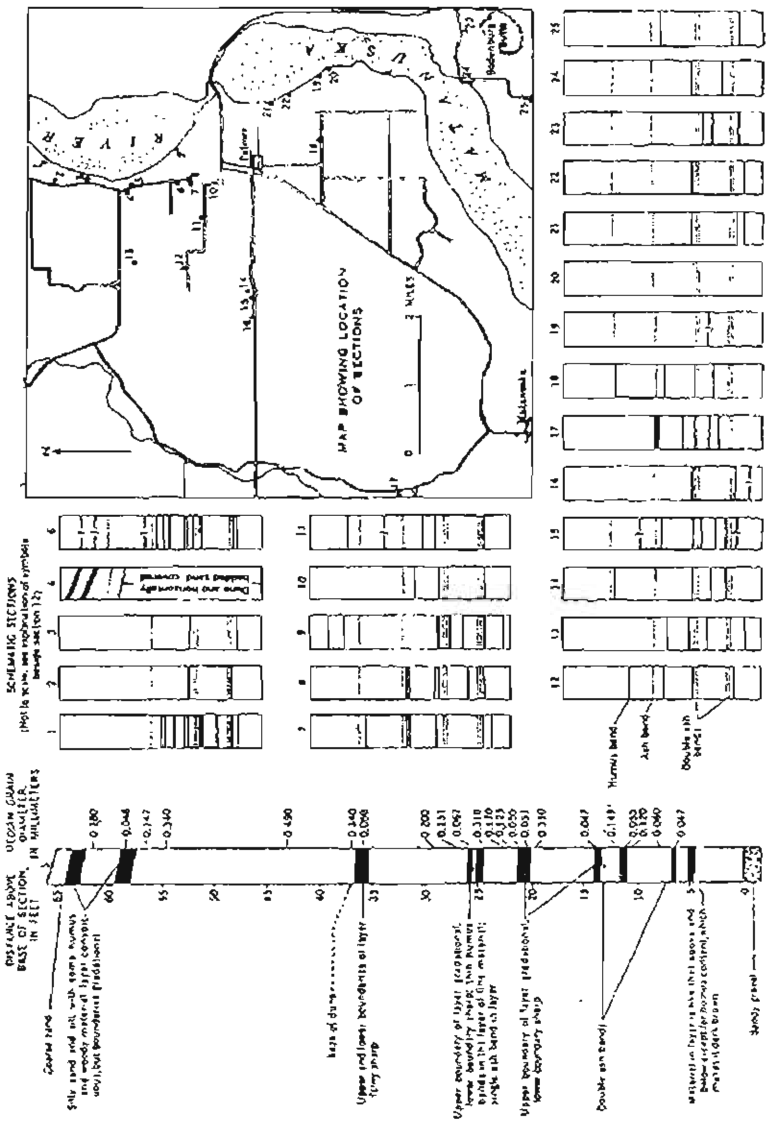


Figure 9-48. Stratigraphic features of the eolian deposits near Palmer. Schematic sections 1 to 4 and 6 to 25 (not to scale) show number and relative positions of humus bands and layers of volcanic ash in sections through the eolian mantle. Section 5 shows details at Stop 9-58. (From Trainer, 1961, fig. 6)

tion, and dissection must therefore have taken place in less than, and probably much less than, 4,000 years.

58 to 55. On hummocky kettle-pitted ice-contact deposits characterized by numerous oval to kettle-hole depressions and eskerlike ridges of contorted coarse gravel.

55 to 54. On low terraces and coarse cobble flood plain of Moose Creek.

54 to 49.5. Rise to terraced and pitted surface of ice-contact deposits surrounding Palmer (fig. 9-47). The deposits are predominantly coarse cobbly gravel overlain by from 4 to 10 feet (1.2 to 3 m) of loess and locally by thick sections of dune sand. This ubiquitous mantle of windblown silt provides the agricultural soils in the earliest and most successful farming community in Alaska.

49.2. Stop 9-58: Cliff-head dunes and loess stratigraphy in the Palmer area.

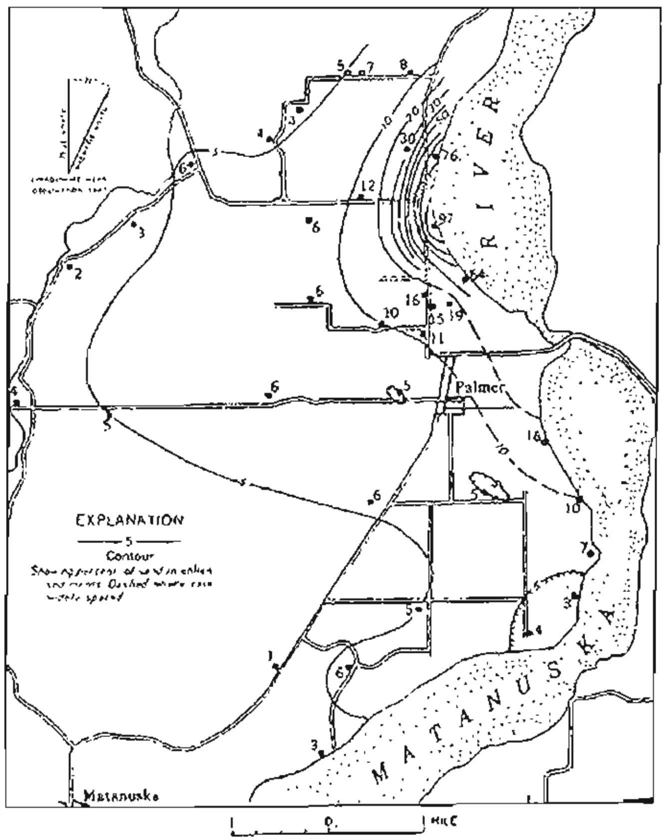


Figure 9-49. Percentage of sand (particles coarser than 0.074 mm diameter) in eolian sediment near Palmer, Alaska. (From Trainer, 1961, fig. 3)

The cliff-head dunes and associated loess formed at the top of the Matanuska River canyon walls represent the maximum thickness of windblown materials measured in the Palmer area (fig. 9-48). More than 30 feet (9.1 m) of dune sand overlies 36 feet (10.9 m) of loess. Nine buried soils are recorded by weathered, fine-grained silt layers containing organic material. These and comparable buried soils present in the windblown deposits throughout the area record multiple cycles of accelerated loess deposition interrupted by periods of diminished deposition in which vegetation stabilized the surface, and weathering profiles were developed. The thickness and the coarseness of the windblown deposits decrease away from the bluffs in the direction of dominant winds blowing downvalley and across the bare flood plains of both the Matanuska and Knik Rivers (figs. 9-49, 9-50). These relations combined with the observations that present-day storms generate dust clouds of fine-grained materials picked up from the nearby flood plains led Trainer (1961) to conclude (1) that the glacial stream flood plains were the dominant source of wind-transported materials in the past; and (2) that changes in flood plain regimen, due either to changes affected by glacial advances and retreats in the valley head or to sea-level changes at the river mouths, or to both, probably explain the cyclical sequence recorded by the buried soils in the deposits. According to this interpretation dune and loess deposition was accelerated during periods of flood plain aggradation, and diminished during periods of degradation.

Proximity of the Palmer area to tidewater at the head of Knik Arm, and the known record of past glacioeustatic sea level oscillations, favor sea level changes as the controlling process on flood plain aggradation and degradation. If true this would mean that periods of accelerated loess and sand deposition probably took place during periods of rising sea levels and glacial recession, and that the buried soils mark periods of marine regressions and glacial advance. Wood samples collected from these buried soils are pending C-14 analysis. Their dating should provide a direct test of this thesis. From present regional relations it is believed that most of the loess in the Palmer area was deposited during and following Altithermal time because only in this time interval after ice retreat were sea levels high enough to determine appreciable aggradation in the lower courses of the Knik and the Matanuska Rivers.

49.2 to 48. Terraced ice-contact deposits. As exposed in roadcuts the thick loess mantle maintains approximately constant thickness on all terrace levels and on intervening terrace scarps. This suggests that the dominant period of loess deposition began only after the lower terraces had been constructed by meltwater streams cutting down through the stagnating ice block. What appear to be good terraces here, when traced to the right off the road, disappear into the spectacular eskerlike complex southwest of Palmer, indicating terrace formation at a time when large buried masses of glacial ice still persisted in the area.

48.3. Enter Palmer. The town of Palmer is at the apex of the lowest terrace level of late Naptowne age. The Palmer terrace is in the form of an alluvial fan deposited by meltwater flowing out of the Matanuska Valley. At the time of its construction, buried glacial ice blocks had melted out except for a few scattered blocks that determine the kettle lakes and drained pits in its distal part to the south. The terrace gradient is greater than that of present

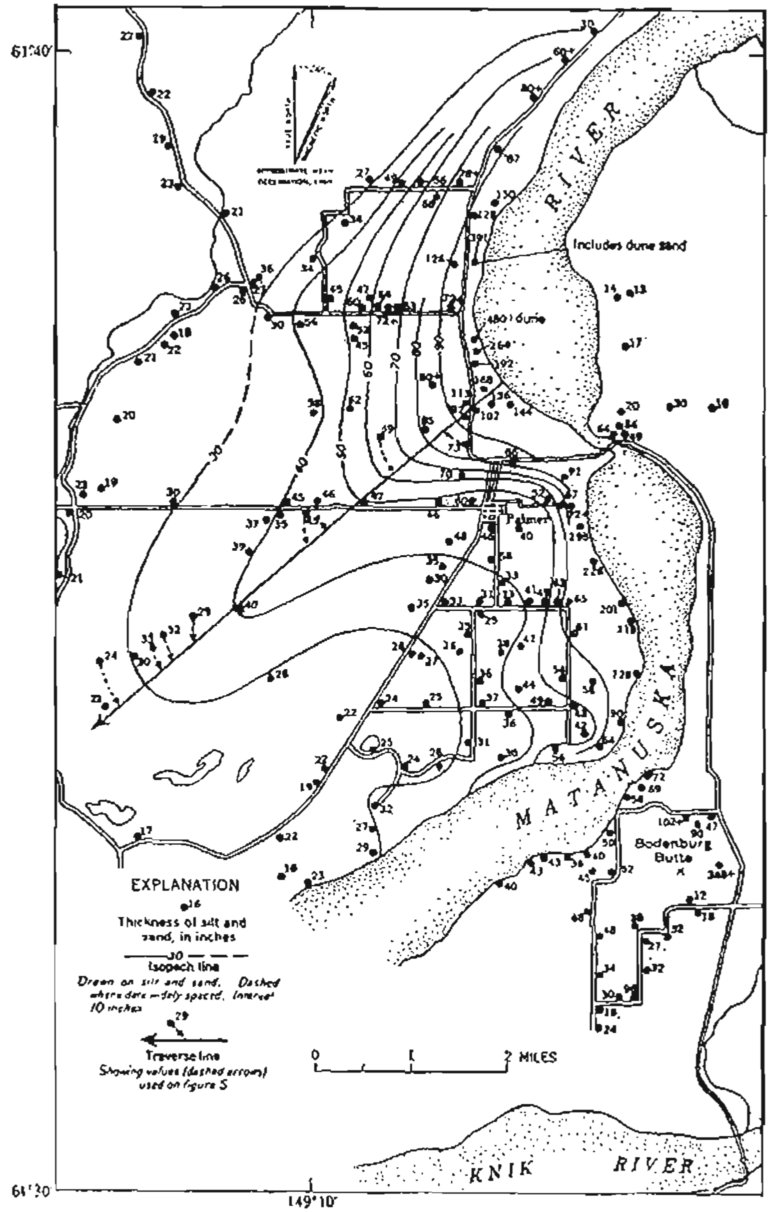


Figure 9-50. Contours showing the thickness, in inches, of the eolian silt and sand near Palmer, Alaska. (From Trainer, 1961, fig. 4)

drainage. The terrace lies about 100 feet (30 m) above the Matanuska River flood plain near here and about 50 feet (15 m) above the flood plain 5 miles (8 km) to the south. A few bedrock hills project above the fan surface, and with the ice-block depressions, provide local relief of a few feet to a few tens of feet. The terrace is underlain by coarse gravel and sand averaging 30 to 50 feet (9 to 15 m) thick overlying a thick sand section. Some wells in the area penetrate what appears to be an underlying till, others terminate in bedrock directly overlain by sand. The loess mantle ranges in thickness from about 2 feet (.6 m) to more than 10 feet (3 m) with the thickest sections present near the Matanuska River flood plain.

Stop 9-59: Experimental farm loop road.

Through Palmer and south on the terrace level of the City of Palmer. On right, the scarp along the inner edge of the terrace is cut below pitted terrace deposits associated with an extensive crevasse-fill complex. Near McLean Lake, a kettle lake, enter transition zone between terrace and hummocky ridge topography of crevasse-fill complex. Near Kepler Lake enter crevasse-fill complex. Kepler Lake and associated lakes occupy long linear kettle depressions, 50 to 100 feet (15 to 30 m) deep between long, narrow, sinuous gravel ridges. Ahead and on left; forested low terraces and old flood plain level of Matanuska River which are graded to elevated tidal flats bordering Knik Arm. These and comparable elevated tidal flats throughout the upper Cook Inlet area record a post-Naptowne sea level stand 5 to 10 feet (1.5 to 3 m) higher than present sea level, and are radiocarbon dated between 5,000 and 6,000 years in age. This maximum post-glacial transgression is informally named the Kasilofian transgression (Table 4). The Farm is along the edge of the crevasse-fill complex of sinuous gravel ridges and intervening linear kettle depressions. Loess here ranges in thickness from less than 1 foot (0.3 m) to more than 3 feet (.9 m). The cultivated fields to the right are on the flatter crestlines of the crevasse-fill ridges; those to the left are on pitted gravel terraces developed along the margins of the retreating Knik ice sublobe after the stagnating ice of the Matanuska sublobe had been buried by coarse gravels deposited by marginal drainage. Between Experimental Farm and Four Corners, cross edge of crevasse-fill complex to right and onto terraces developed along the Wasilla Creek drainage line. Between Four Corners and Palmer cross hummocky topography of the crudely terraced portion of the crevasse-fill complex. The topography of the central part of the crevasse-fill complex to the right is so rugged that it has been impossible to cultivate, and thus retains a virtually undisturbed forest cover of mature white spruce, white birch, with associated aspen, alder, willow, and some cottonwood. Loess thickness ranges from less than 3 feet (.9 m) near Four Corners to more than 5 feet (1.5 m) near Palmer (fig. 9-50).

47.8 to **46.6.** Cross apex of Palmer Terrace. On right, the loess-covered hills are underlain by bedrock of probably Jurassic age. Ahead; Lazy Mountain. The lateral moraines around 2,000 feet elevation (610 m) on the mountain slopes mark the maximum height of Naptowne ice in the area.

46.6. Matanuska River bridge. Constriction of the river here to a narrow gorge resulted from postglacial superposition from the terrace level at the City of Palmer through a buried bedrock ridge. Subsurface data indicate a buried bedrock channel just north of Palmer that probably marked the pre-Naptowne course of the Matanuska River into Knik Arm.

46.6 to 38.5. Bodenburg Terrace. Area underlain by stream-deposited silt, sand, and sand gravel of post-Naptowne age. Bodenburg Butte and smaller hills of bedrock and a few hills of till protrude through the stream deposits. The terraced surface lies just above present flood plain levels and records a sequence of aggradational and degradational intervals accompanying Altithermal and post-Altithermal sea level fluctuation. The general gradient of the low-lying surface is to the south and southwest indicating that supply was predominantly from the Matanuska Valley. The forest cover consists of white spruce, white birch, aspen, alder, willow, and cottonwood.

45.2. Left; remnant of late Naptowne moraines.

44.5. Ahead; good view of Bodenburg Butte, an ice-rounded loess-covered bedrock hill of Jurassic greenstone.

43.7. To right and ahead, clearwater road ditch used by spawning salmon migrating up spring-fed Palmer Creek. During salmon runs numerous King, Red, Humpy, and Dog Salmon may be observed in the deeper pools along the road.

39.6. Ahead; view of steep slopes of Pioneer Peak, elevation 6,398 feet (1,950.1 m).

38.5. Knik River bridge. The Knik River floods annually in June, July, or August when Lake George, impounded upvalley by the front of Knik Glacier, overtops the ice and erodes a gorge along the valley wall (fig. 9-51). The discharge of the Knik River prior to flooding is on the order of 5,000 to 6,000 cfs (142-170 $m^3\ sec^{-1}$). Since the beginning of U.S. Geological Survey gage recording at the bridge in 1949, peak discharges of flooding have ranged from 41,500 cfs (1,180 $m^3\ sec^{-1}$), when no lake formed in the Lake George basin in 1963, to 359,000 cfs (10,200 $m^3\ sec^{-1}$) in 1958. Lake discharge has begun as early as June 26, 1962, and as late as August 13 in 1949. The duration of augmented discharge from the lake has ranged from 8 to 18 days.

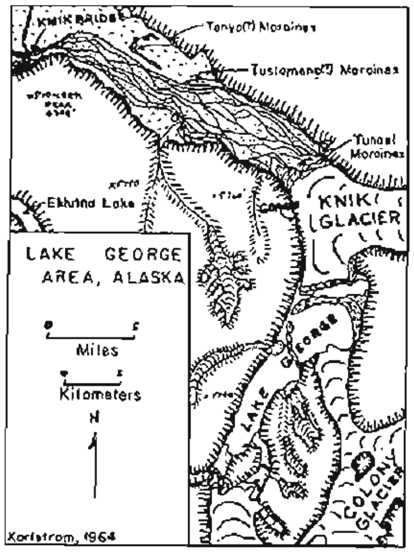


Figure 9-51. Lake George area, Alaska.

The existence and the capacity of the ice-dammed lake basin depend on a very delicate balance between the position of the Knik Glacier front and those of the other glaciers extending into the basin. The lake basin could not have come into existence until late Tanya time when the lake basin glaciers had retreated out of the valley prior to retreat of the master Knik Glacier from a position athwart the valley mouth. In all probability during the ensuing Altithermal time, centered around 3,500 years B.C., Knik Glacier retreated upvalley from the valley mouth and no lake existed. During the extended phases of the Tustumena advances dated between 2,500 B.C. and A.D. 500, the tributary glacier appears to have coalesced with Knik Glacier, and the lake basin was completely filled with ice except possibly during the recessional periods between extended advance phases.

The last cycle of Lake George began with the Tunnel I advance dated *ca.* A.D. 1000. During Tunnel I maximum, Colony Glacier coalesced with Knik Glacier, forming a high ice dam for the upper part of the valley that remained ice-free. Whether or not water rose high enough in this upper lake to periodically overtop the ice dam is not certain, but likely. Retreat from the Tunnel I moraines reestablished, at least for a short time, the same conditions existing today in the basin. During the Tunnel II maximum dated A.D. 1600 to 1700 the Colony and Knik Glaciers did not coalesce and a lower lake and an upper lake were created in the ice-free portions of the valley. The middle lake portion was created during retreat of the Colony Glacier from its Tunnel II moraines that today form an arcuate island chain during all but the highest lake phases.

The future of the lake is uncertain. If Knik Glacier continues to retreat, the ice dam will be destroyed and the lake will go out of existence such as happened in 1963. If, however, climatic changes cause Knik Glacier to advance, a stronger lake environment will be generated that will persist until the time the Lake George valley is completely filled with glacial ice or until Knik Glacier again retreats upvalley from the tributary valley mouth.

The Recent moraines in the Lake George basin are instructive in showing the nature of moraine modification within a changing proglacial lake environment. The moraine slopes are characterized by a series of closely spaced but inconspicuously developed terracettes reflecting repeated changes in lake levels resulting from unstable ice-dam threshold conditions. Despite repeated partial or complete submergence of the moraines during high lake phases no lake deposits mantle morainal crests or slopes. Instead the predominant process involved is one of erosion, as recorded by a discontinuous lag veneer of pebbles, cobbles, and boulders concentrated by removal of the finer-grained components of the till matrix. This veneer apparently serves as an armor retarding further erosion. The vast quantities of silt deposited in the lake are virtually restricted to the deeper parts of the lake basin. Silt that may be temporarily deposited on the moraine slopes during higher lake phases is apparently swept away along with the finer-grained constituents of the underlying till by vigorous wave action or current action during periods of catastrophic drainage.

37.5. Enter Anchorage B-6 Quadrangle.

38.0 to 35.7. Along margin of Knik River flood plain at base of steep Chugach Mountain front. Road maintenance problems in this sector are caused by avalanches off the steep mountain face, and by periodic flooding of the Knik River. During the 1964 earthquake a snow and debris avalanche swept away a house near mile 37.7 and deposited debris half way across the flood plain. The sand and gravel bars in the flood plain were extensively fractured. During late spring and early summer floods, low points of the road, such as at Goat Creek are inundated for short periods of time.

35.7 to 34.4. Low, vegetated Knik River terrace surface covered by alluvial fan deposits at the mouths of gullies along the mountain front.

31.0. Right; bedrock knob of Jurassic greenstone.

30.1. Enter Anchorage B-7 Quadrangle.

29.5 to **27.0.** On elevated tidal flat underlain by thick section of stratified estuarine silt and sand. Muskeg and marsh vegetation reflects poor surface drainage conditions. Grass, reed, and sedge cover the wetter parts of the flat; heath moss with scattered black spruce, willow, and birch cover the drier parts. During the 1964 earthquake the tidal flat surface differentially subsided between 2 and 4 feet (.6 and 1.2 m). As a result during periods of high tides, parts of the road were inundated by 1 to 3 feet (0.3 to 0.9 m) of water. Some of the subsidence resulted from differential compaction of the fine-grained silt. Measurements on bedrock in the area, however, indicate that about 2 feet (0.6 m) of the total subsidence resulted from crustal depression of this part of the Chugach Mountains, whereas the axial part of the mountains near Portage was depressed as much as 6 feet (1.8 m).

27.0 to **26.6.** Terraced alluvial fan gravel deposited at the mouth of the Eklutna River accompanying and following Naptowne recession. The coarse gravel is mantled by 1 to 3 feet (0.3 to 0.9 m) of loess and provides a well-drained stable surface for settlement. The nearby native village of Eklutna is one of the larger native villages remaining in Cook Inlet, and has a history that extends back to pre-Russian days.

26.6. Stop 9-60: Eklutna Valley.

Eklutna Valley was occupied by a hanging tributary glacier during Naptowne time, and reconnaissance observations indicate that the valley fill includes deposits recording at least the last three major glaciations. Moraines deposited by Eklutna Glacier during the Naptowne recession have in part been mantled by deposits laid down in a proglacial lake dammed by persisting ice in the lowland. Subdued ridges near the valley mouth and at the outlet of Eklutna Lake are provisionally correlated with the Killey and Skilak moraines, pending more detailed studies in the valley. At the head of the valley there are three sets of moraines, which on the basis of position, vegetation cover, and weathering profiles are correlated with the Tanya moraines of late Naptowne age and with the Tustumena and Tunnel moraines of Alaska age (fig. 9-47). The remnants of the Tanya moraines lie above outwash graded to the two younger moraine sets that lie upvalley within 2 miles (3 km) of the present front of Eklutna Glacier. A mature climax forest of white spruce and birch covers the moraines that are oxidized through a vertical interval of 2 to 5 feet (.6 to 1.5 m). In contrast the Tustumena moraines are covered largely by a mature cottonwood forest and are weathered through a depth generally less than 1 foot (0.3 m). The Tunnel moraines show incipient soil development beneath an immature forest cover.

The type section of Eklutna drift is exposed in the Eklutna River canyon below the mouth of Eklutna Lake. The section is as follows: (1) at the base, 40 to 60 feet (12 to 18 m) of yellow-buff, weathered, cobbly gravel and sand of Eklutna age unconformably overlain by (2) 30 to 50 feet (9 to 15 m) of blue-gray laminated silt and sand (proglacial lake deposits) under buff-weathered till of Knik age unconformably overlain by (3) 10 to 30 feet (3 to 9 m) of gray stratified silt, sand, and gravel under 2 to 25 feet (.6 to 7.6 m) of olive drab till of Naptowne age. The Naptowne Till is overlain by a variable thickness of lake, outwash, terrace and colluvial deposits capped generally by 2 to 4 feet (.6 to 1.2 m) of loess.

During the 1964 earthquake, delta deposits at the head of the lake slumped and fractured. Near the head of the valley, a zone of fracturing developed in the valley fill. The fracture pattern transects outwash and the moraine topography, and is aligned with shear zones in the adjacent valley walls. Unusual avalanche activity continued throughout the entire summer within these shear zones that were apparently activated during and following the main earthquake.

Bedrock in the valley below the head of Eklutna Lake is of Jurassic greenstone. The rocks exposed in the valley head are brecciated volcanic tuffs and lavas mapped as Tertiary or Quaternary in age.

25.4 to **14.0.** On hummocky, dissected, and, in part, alluvial-fan-covered lateral moraines of Naptowne age. The upper boundary of Naptowne ice in this sector is marked by moraine remnants and marginal channels that lie as high as 1,500 feet (457 m) and 1,000 feet (305 m) in elevation near Eklutna and Eagle River, respectively. Roadcuts expose gravelly Naptowne Till overlain by 1 to 3 feet (0.3 to 0.9 m) loess and locally by stratified alluvial fan deposits.

22.2. Pass from lateral moraines of Skilak age onto moraines of Killey age. The Killey-Skilak boundary is apparently marked by the broad lateral channel occupied by Edmonds and Mirror Lakes and is crossed between mile 22.2 and Peters Creek. Accompanying and following retreat of Naptowne ice from the lowland large alluvial fan complexes were deposited at the mouth of Peters Creek and other tributary valleys along the mountain front. The most extensively developed complex of fans, outwash channels, and terraces formed during the Naptowne recession is at the mouth of Eagle River valley at Milepost 13.2.

13.2 to **11.5.** Ascend Naptowne end moraines deposited by Eagle River Glacier when it was in confluence with the lowland ice lobe. Roadcuts expose little weathered gravelly till overlain by from 1 to 4 feet (0.3 to 1.2 m) of loess and in places by extensive deposits of stratified recessional gravel. Two morainal belts, attributed to the Moosehorn (outermost) and Killey advances, are separated by a terraced and channeled gravel plain graded to the inner Killey moraines. Fossil Creek, an abandoned spillway, originates within the Killey moraines and apparently was formed during recession from the Killey maximum (fig. 9-47).

11.2. Stop 9-61: Apex of gravel plain.

Behind and to the right is the forested Elmendorf end moraine that represents the maximum extension of the Matanuska-Knik ice lobe during the Naptowne Glaciation (fig. 9-47). To the left on the Chugach Mountain front are older lateral moraines of Knik and Eklutna age. The Knik boundary lies at about 1,500 feet (457 m) elevation here at the mouth of Eagle River valley and falls rapidly in elevation in line with a discontinuous series of subdued, gravel-capped moraine ridges partially buried in the gravel plain. These Knik end moraines can be traced in an arc towards Knik Arm south of Anchorage (fig. 9-47). The upper boundary of the Eklutna moraines descends in elevation from about 2,000 feet (610 m) at the mouth of Eagle River valley to less than 1,000 feet (305 m) south of Anchorage, and then rises to about 1,500 feet (457 m) near the mouth of Turnagain Arm. This reversed gradient marks the zone of coalescence of the Matanuska-Knik ice lobe with the Turnagain Arm trunk glacier during Eklutna time. The Eklutna moraine boundary is one of the most distinctive boundaries present in the area and can be traced nearly continuously along the flanks of the Talkeetna Mountains to the north, around the flanks of Mt. Susitna, and along the front of the Kenai Mountains to the south. Eklutna moraines form reentrant embankments in the mouths of tributary valleys and gullies cut in drift of older age, suggesting a major period of erosion and weathering prior to Eklutna time. In contrast the Knik boundary is not so clearly expressed topographically.

The northern boundary of the Anchorage gravel plain is along the flanks of the Naptowne end moraines (the Elmendorf moraine). The gravel is in erosional contact with the Naptowne Till and the surface gradient of the plain in general parallels the till contact. The boundary of the plain along the Chugach Mountain front is much more irregular and characterized by a complex pattern of partly buried and isolated moraine remnants and by terraced alluvial fan-shaped deposits that are in part coextensive with and in part overlap the margins of the plain. Although gross spatial relations to the Elmendorf end moraine suggest a simple subaerial outwash origin for the plain, and it is so described in the literature, regional relations require formation and modification within a rapidly changing proglacial lake environment.

11.0 to 8.3. On Anchorage gravel plain. This upper part of the plain has a subdued alluvial fan shape and is highest where axially drained by the north fork of Ship Creek. The gravel is extremely coarse, reflecting proximity to the source area near the mouth of Eagle River valley, and is mantled by 1 to 3 feet (0.3 to 0.9 m) of loess.

10.2. Enter Anchorage B-7 Quadrangle.

8.3 to 6.9. On subdued gravel-capped remnant of Knik end moraine partly buried in the gravel plain. The moraine remnant is made up of two parallel ridges separated and crossed by broad-floored, gravel-filled channels. At mile 8.3 the moraine remnant is completely buried by the alluvial fan-shaped deposit drained by the north fork of Ship Creek, the other end at mile 6.9 stands with steep erosional slopes 50 to 120 feet (15 to 37 m) above a fan-shaped gravel deposit drained by Ship Creek. Stratigraphic and geomorphic relations suggest the following multiple sequence of events: (1) deposition of Knik Till; (2) deposition of gravel cap after a period of weathering and erosion as recorded by a truncated soil profile beneath the gravel cap; (3) dissection and deposition accompanying development of the broad channel between ridges; (4) burial beneath the north fork alluvial fan deposits with damming and muskeg development in the buried end of the channel (item 3); and (5) dissection and partial burial during development of the Ship Creek alluvial fan-shaped deposit, either contemporaneous with item 4 or later.

6.9 to 6.6. On Ship Creek fan-shaped gravel deposit. The gradient of the fan is up toward the Chugach Mountains to the left, where the fan apexes in a canyon cut below the lip of the hanging mouth of Ship Creek glacial valley. Remnants of older fan gravels occur above and within the canyon. To the right the distal edge of the fan form terminates at about an elevation of 240 feet (73 m) where Ship Creek channel abruptly widens into a broad flat-floored valley and where an abandoned flat-floored channel cut into the Anchorage gravel plain is beheaded. These geomorphic relations strongly suggest burial of the heads of preexisting channels by fan development. Insofar as the regional evidence is for a high lake phase at an

elevation of 250 to 300 feet (76 to 91 m) during the Skilak maximum, it is believed that this fan-shaped deposit, as it terminates just below this critical elevation, is actually a fan delta deposit laid down by Ship Creek along the margins of this lake. If this is true it should be expected that similar deposits were laid down at about the same elevation by the other streams flowing out of the Chugach Mountains. Similar fan-shaped deposits, also associated with beheaded channels, do occur at the proper elevations along the middle courses of these tributary streams. One along the south fork of Campbell Creek is particularly well developed, and lies above a younger fan-shaped deposit graded to a lower lake phase, at about an elevation of 100 to 150 feet (30 to 46 m).

Older gravel deposits perched along the mountain front above Ship Creek and the other tributary streams are concentrated at about the same elevations as hanging deltas mapped to the south on the Kenai Peninsula, and therefore would appear to record the same high proglacial lake phases established for the Moosehorn maximum (*ca.* 750 feet [229 m] or higher elevation) and the Killey maximum (*ca.* 500 to 600 feet [152 to 183 m] elevation).

6.9. Left; turnoff to Ski Bowl Road. Rise from Ship Creek fan delta onto lateral moraines of Eklutna age.

Stop 9-62: Roadcut in Eklutna lateral moraines. (Ski Bowl Road, 2.7 miles from Glenn Highway junction. Anchorage A-8 Quadrangle.)

The section exposed is as follows (from bottom to top): (1) 15 feet (5 m) plus of buff to brownish-buff, fractured and jointed, laminated silt-proglacial lake deposits laid down during Eklutna advance; (2) 15 feet (5 m) of buff to brownish, coarse, bouldery Eklutna Till—a lag concentration of cobbles, pebbles, and boulders suggests erosion and removal of finegrained materials prior to deposition of overlying laminated silt unit; (3) 2 to 6 feet (.5 to 2 m) of blue-gray finely laminated silt—proglacial lake deposits of Naptowne age; and (4) 2 to 3 feet (.5 to .9 m) of loess with podzolic soil profile extending locally below base into underlying sediments.

The deep weathering extending through the Eklutna Till into the underlying proglacial lake sediments is typical of the Eklutna deposits where they are exposed elsewhere in the region. Oxidation profiles of greater than 40 feet (12 m) are common. This may be contrasted with maximum oxidation profiles of 10 to 20 feet (3 to 6 m) found on buried Knik drift, and of generally less than 10 feet (3 m) found on surficial Naptowne drift.

The roadcut is exposed between an elevation of 700 and 760 feet (213 to 232 m), indicating that the surficial proglacial lake sediments of Naptowne age were probably deposited during the Moosehorn maximum, or around 18,000 to 20,000 years ago.

Stop 9-63: Panoramic view. (Ski Bowl Road, 3.9 miles from Glenn Highway junction. Anchorage A-7 Quadrangle.)

Anchorage in the near distance is located on the distal end of the Anchorage gravel plain between Turnagain Arm and Knik Arm (fig. 9-47). The tidal range in Turnagain Arm is more than 30 feet (9 m) and is second only to that of the Bay of Fundy. During low tides, extensive flats of tidal silt are exposed in both Knik and Turnagain Arms, and Fire Island is nearly joined to the mainland.

The forested Elmendorf end moraine of Naptowne age lies just north of Anchorage and can be traced as a part of a series of morainal belts across the Susitna lowlands on the other side of Knik Arm (fig. 9-47). Although presumably the same age as the Donnelly moraines observed in the Delta River area, the Naptowne moraines in the lowlands differ appreciably in topographic aspect. Sharp secondary ridges and circular kettle depressions typical of the Donnelly moraines are rare or absent on the Naptowne moraines. Only the gross moraine ridge forms remain, and the associated lakes are largely restricted to the interridge depressions that are commonly floored with fine-grained proglacial lake deposits. Insofar as the Naptowne moraines that were deposited above proglacial lake levels are comparable in topographic aspect to the Donnelly moraines, it is concluded that the differing topographic form of the Naptowne lowland moraines is not a function of age or lithologic differences, but principally reflects deposition and subsequent reworking within the proglacial lake environment.

Mount Susitna in the middle ground rises from the Susitna River lowland to a peak altitude of 4,349 feet (1,325.5 m) (fig. 9-47). The rugged mountain range in the distance is the Alaska Range. On clear days, Mt. McKinley, the highest mountain on the North American Continent, can be seen from Anchorage.

Glacial erratic material has been found within 100 feet (30 m) of the ice-rounded summit of Mt. Susitna, which is underlain by coarse-grained quartz diorite. The erratic material includes rock types that could only have been derived from formations exposed in the Alaska Range to the north. Mt. Susitna is the type locality of the Mount Susitna glaciation. The presence of ancient drift on the mountain crestline indicates that during Mount Susitna time the entire upper Cook Inlet region was covered by ice whose surface stood well above the altitude of 4,000 feet (1,220 m). Highly modified remnant moraines of Caribou Hills age and less modified lateral moraines of Eklutna age occur on the middle and lower slopes of the Mountain. Below an elevation of 1,000 feet (305 m), the Eklutna moraines are cut by a series of proglacial lake terraces. The *ca.* 500- to 600-foot (152 to 183 m) and 250- to 300-foot (76 to 91 m) terrace levels are particularly well developed.

6.5 to **5.6.** Rise from surface on Ship Creek fan delta deposit onto gravel-capped moraine of Knik age.

5.6 to **4.2.** On gravel-capped moraine of Knik age.

5.5. Enter Anchorage A-8 Quadrangle.

Stop 9-64: Gravel pit in Knik moraine.

The gravel pit section exposes (from bottom to top): (1) 10 to 15 feet (3 to 5 m) of buff to buff-gray gravelly till of Knik age with a truncated weathering profile; (2) 5 to 10 feet (1 to 3 m) of cobbly sand and gravel of Naptowne age; and (3) 2 to 3 feet (.6 to .9 m) of loess with a podzolic soil profile.

The section is typical of the Knik moraine ridges throughout the lowland in illustrating burial of the Knik moraines beneath glaciolacustrine and glaciofluvial deposits of Naptowne age after an interval of subaerial weathering of the Knik drift. The mantling deposits grade from coarse gravel here to finer-grained gravel and stratified sand and silt to the southwest and at lower elevations. Likewise over the same transect there is a comparable progressive change from gravel to sand and silt in the materials underlying the lower Anchorage gravel plain surface, suggesting similar environments of deposition. Thus, the gravel cap on the Knik moraines is believed to record an older and higher depositional level in Glacial Lake Cook. This older surface was largely destroyed during development of the younger surface, except for remnants preserved on the crests of exhumed moraines, and on other higher ground.

4.2 to **2.9.** Channeled part of the Anchorage gravel plain downvalley from the Ship Creek fan delta. The shallow, flat-floored abandoned channel crossed between 3.4 and 2.9 is 1,000 to 1,500 feet (305 to 457 m) wide and 10 to 25 feet (3 to 8 m) deep. If the interpretation of the fan delta deposits is correct these channels must have formed during or prior to the proglacial lake drainage that preceded regeneration of the proglacial lake during the Skilak advance. Peat intercalated in proglacial lake sediments recording this late Killey lake drainage of the Kenai lowland dates between 11,000 and 10,000 B.C. The preservation of the abandoned channel as a topographic form despite resubmergence beneath proglacial lake waters requires that either it occupied a zone of little deposition within the lake, or, more likely, that the fine-grained deposits that were deposited in the channel were largely flushed out during or following the subsequent lake drainages.

2.9 to **0.** Undissected part of Anchorage gravel plain confined by Ship Creek valley to the right and by Campbell Creek valley to the left. This part of the plain is underlain by 50 to 70 feet (15 to 21 m) of sand and gravel and as much as 200 feet (61 m) of the Bootlegger Cove Clay, which consists of finely laminated to massive silt and clay with minor amounts of sand and gravel. The Bootlegger Cove Clay overlies the Knik Till unit that is now traceable, through subsurface data, from near sea level in the sea bluffs near Point McKenzie (the type section of the Knik Glaciation), beneath the Anchorage area at greater depths, to the surface in the Knik moraines we crossed upvalley. The subsurface data from wells indicate a 400- to 700-foot (122 to 213 m) thick Quaternary section near Anchorage deposited in a deep depositional basin and including at least three tills older than the Knik. These older tills are also associated with quiescent-water deposits and are separated by weathering profiles recording interglacial subaerial intervals. Toward the mountain front the Quaternary section thins and coarsens, with the associated quiescent-water deposits grading into sand or gravel, or replaced by unconformities separating till units. These lateral changes record shallower water or subaerial deposition and erosion along the upper margins of the depositional basin. This subsurface analysis is particularly pertinent in providing stratigraphic evidence for at least four major pre-Naptowne glaciations in the Anchorage area, and in emphasizing the repetition of the same proglacial lake environmental conditions during all the glaciations. The two youngest drift units are the equivalent of the Knik and Eklutna deposits, as these are defined from surficial exposures near Anchorage. The two oldest drift units may or may not be the precise equivalents of the Caribou Hills and Mount Susitna Glaciations, as defined by deposits elsewhere in Cook Inlet, depending on the completeness of the two types of records involved.

0.0. Downtown Anchorage. Ship Creek valley is incised 60 to 80 feet (18 to 24 m) below the Anchorage gravel plain. Ship Creek valley was trenched as much as 9 feet (3 m) below present sea level after final drainage of the Naptowne proglacial lake, but before rising postglacial sea levels reversed the downcutting trend. During the Kasilofian transgression, culminating about 5,500 years ago, sea level rose about 5 to 10 feet (1.5 to 3 m) higher than present datum, entered the mouth of the valley, and deposited a relatively thin clay upon the trenched older clay and alluvial deposits in the valley bottom.

Stop 9-65: Cairn Point.

The channeled surface bordering Elmendorf end moraine and above the level of the Anchorage gravel plain is underlain by stratified clay, silt, and sand (with minor amounts of gravel) deposited in Glacial Lake Cook during and following the advance of Naptowne ice to the Elmendorf moraine. Along the sea bluffs these deposits can be traced discontinuously into the type section of the Bootlegger Cove Clay south of Anchorage, and beneath the Naptowne deposits of the Elmendorf moraine near Cairn Point. Near and beneath the moraine the proglacial lake deposits are deformed and folded due to plowing and overriding by the advancing ice front. Exposures are available near here that show undeformed silt, sand, and gravel unconformably overlying the folded deposits, indicating continued lake deposition after deformation of the underlying sediments. The surface lies below the 250- to 300-foot-high (76 to 91 m) lake phase level; therefore, it was submerged intermittently during the Skilak as well as the older Killey and Moosehorn high lake phases. During the Moosehorn maximum, the ice that deposited the moraine fronted a fluctuating lake that rose, for short intervals of time, to elevations of 750 feet (229 m) or higher. Therefore, since the moraine crests below elevations of 400 feet (129 m), it was deposited at least in part below lake level. During the succeeding Killey maximum when maximum lake levels attained elevations between 500 and 600 feet (152 to 183 m), the Elmendorf moraine was completely submerged and subject to additional subaqueous modification. During the Skilak maximum high lake phase, only the lower parts of the moraine were submerged, while the higher parts stood as islands in the lake. This partial submergence is apparently recorded in the change in slope characteristics above and below an elevation of 250 to 300 feet (76 to 91 m). The slopes below this elevation affected by the last episode of subaqueous modification are much more subdued and rounded than those above.

Although the Anchorage gravel plain has been mapped as outwash laid down during the time ice was depositing the Elmendorf moraine, it has peculiarities atypical of outwash deposits subaerially deposited by meltwater streams marginal to a glacier front. The gravel plain is virtually restricted to the Anchorage area and is absent along the front of the Naptowne end moraines throughout most of the Upper Cook Inlet region. Elsewhere, as here, these moraines are bordered by finely stratified, fine-grained proglacial lake sediments recording contemporaneous sublacustrine rather than fluvial deposition. Further, the Anchorage gravel plain is not radially graded to the Elmendorf moraine, as is characteristic of outwash aprons, but slopes parallel to the moraines and coarsens in texture toward marginal source areas at the mouths of tributary valleys in the Chugach Mountains. These relations combined with the evidence that the gravel plain lies below the level of lacustrine deposition during and following maximum extension of Naptowne ice require that the Anchorage gravel plain was formed after ice had retreated from the Elmendorf end moraine and proglacial lake levels had fallen. Within the framework of the regional evidence of Glacial Lake Cook, it is believed that the gravel plain commenced as fan deltas deposited at the mouths of Eagle River and Ship Creek during thus to exhume preexisting topographic features such as moraines and gravel channels constructed of, or in, coarse, less readily transported materials.

Drifts of Eklutna, Knik and Naptowne age are exposed in the sea bluffs of Knik Arm (fig. 9-52). Upvalley from the Naptowne end moraines and near Goose Bay, the Naptowne

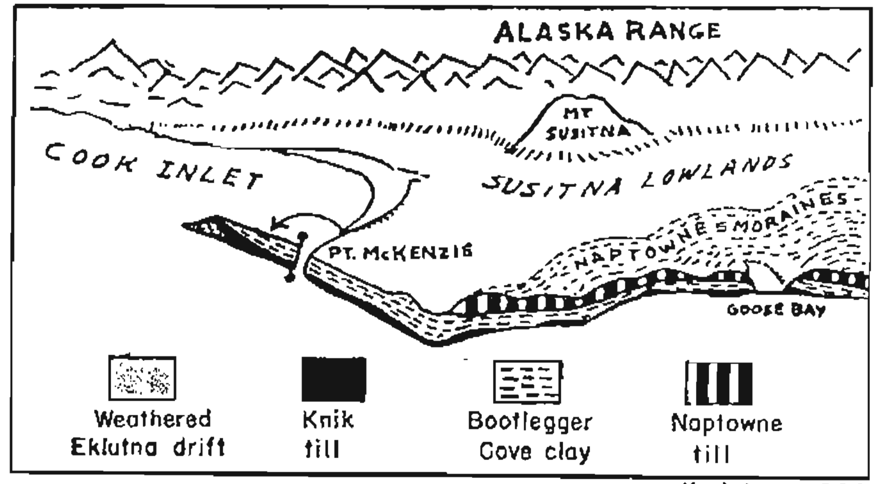


Figure 9-52. Sea bluff stratigraphy along west shore Knik Arm, Cook Inlet region, Alaska.

the highest lake level stands and then extended southwestward into the Anchorage area as fluctuating lake levels progressively dropped. Gravels deposited at the mouths of tributary valleys during the higher lake phases were dissected and reworked by extending streams during lower lake phases to form the distal parts of the plain. The repeated catastrophic drainages of the lake and the wave and current turbulence in the fluctuating littoral zone of the lake could be expected to flush out the finer-grained sediments deposited during higher lake phases and

Till at the top of the bluffs is overlain by lacustrine silt and sand, and is separated from the lower Knik Till by a sequence of weathered silt, sand, and gravel deposits containing peat layers, indicating a long warm interval between the two glaciations. Downvalley from the Naptowne end moraines, the basal Knik Till is overlain by the Bootlegger Cove Clay that contains a middle marine zone recording a high sea level stand during the Knik-Naptowne Interglaciation and just prior to the proglacial lake deposition that accompanied and followed the Naptowne maximum advance. The buried peat near Goose Bay has been dated by radiocarbon as greater than 32,000 to 40,000 years old. Peat from the same stratigraphic horizon on this side of the Arm exposed along Eagle River dates within the same time range. Shells from the marine beds of the

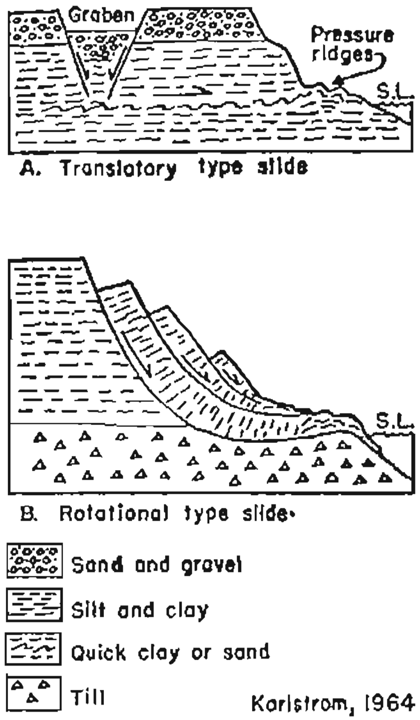


Figure 9-53. Types of landslides formed in the Anchorage area, Alaska, caused by the Alaskan earthquake of March 27, 1964.

Bootlegger Cove Clay date 33,000 to 48,000 years old by the ionium/uranium ratio dating method. Based on these dates and other data from Cook Inlet the culmination of the Knik-Naptowne high sea level stand (the Woronzofian transgression) is placed *ca.* 45,000 B.C.

Stop 9-66: Earthquake landslides.

All the devastating slumps that occurred during the 1964 earthquake in the Anchorage area are restricted to bluff areas underlain by thick sections of the Bootlegger Cove Clay. Mapping by U. S. Geological Survey geologists Ernest Dobrovolny, Wallace Hansen, Clifford Kay, and Robert Miller indicates two main mechanisms of failure. The Turnagain Heights, L-Street, 4th Avenue, and Government Hill slides resulted principally from lateral translation of bluff blocks with horizontal slippage at depth through one or more weak zones in the underlying clay (either quick clay beds or sand beds or both). Development of inner graben zones resulted from this lateral displacement (fig. 9-53A). The slide that can be seen along the bluffs across Knik Arm near Point McKenzie (fig. 9-52) is associated with the thickest exposed section of the Bootlegger Cove Clay that failed by rotational slippage on concave fracture planes (fig. 9-53B).

Stabilization of the slide areas has largely been accomplished, but some of the main features of failure still remain.

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