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Recent research on Alaskan geology

STATE OF ALASKA

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Cover photo: Flat Top, a summit-type cryoplanation terrace (elev 1,230 m) on upper southwest flank of Indian Mountain. Pocohontas Hills in background.

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HOLOCENE DISPLACEMENTS MEASURED BY TRENCHING THE CASTLE MOUNTAIN FAULT NEAR HOUSTON, ALASKA

By Ronald L. Bruhn[1]

ABSTRACT

During 1977, two trenches spaced 300 m apart were dug across the northeast-trending Castle Mountain fault zone in unconsolidated Pleistocene deposits near Houston, Alaska. The faults exposed in the trench walls vary with respect to geometry and amount of displacement. Most faults dip between vertical and 68° S. and exhibit normal offsets. A total dip-slip displacement of 97 cm was measured in the fluvial deposits of trench 1 and 110 cm of dip-slip displacement was measured in the sand and till deposits exposed in trench 2. A river terrace bank near trench 2 is offset an estimated 2.4 m in the dextral sense. Estimated oblique-slip on the exposed fault zone at the second trench site is 2.6 m.

INTRODUCTION

The Susitna segment of the Castle Mountain fault zone is delineated by a discontinuous series of scarps, banks, and vegetation lineaments that developed from Holocene faulting (Barnes, 1962; Karlstrom, 1964; Detterman and others, 1974). Detterman and others (1974) estimate that the most recent faulting occurred between 222 (tree-ring date) and 1860 ± 250 (C^{14}) years ago.

TRENCHING SITE LOCATION AND GEOLOGY

Two trenches were dug with a small backhoe-equipped tractor across a well-defined Holocene scarp in sec. 28, T. 18 N., R. 3 W. in the Anchorage C-8 quadrangle (fig. 1). The geology of the trench walls was mapped with the aid of a level line and tape measure.

The surficial deposits surrounding the trenching sites consist of Pleistocene glacial, fluvial, and eolian deposits (Detterman and others, 1974). The remnants of the Holocene fault scarp trend between N. 25° E. and N. 85° E. and are up to 2.1 m high. The north side is upthrown relative to the south side and the scarp, which is variably modified by erosion, dips southward.

TRENCH 1

The easternmost trench crosses a 97-cm-high scarp that trends N. 80° E. in fluvial deposits (fig. 1, T-1). Two offset zones occur in the trench walls, one near the top of the scarp and the other near the scarp base (fig. 2).

The offset near the scarp top consists of 44 cm of dip-slip displacement on a fault that dips 78° S. Directly across the trench in the western wall the offset occurs on two normal faults 31 cm apart; one fault dips 64° S. and the other fault dips 77° S. (fig. 2). The total dip-slip displacement across the two faults is 34 cm, 10 cm less than the total offset observed on the single fault directly across the trench. However, the sand and gravel layers in the zone between the two faults were rotated during faulting so that they dip about 20° S. This rotation accounts for most of the difference in displacements measured in the two walls of the trench.

Strata at the base of the scarp are offset on a single, vertically dipping fault that occurs in both walls of the trench. The strata on the south side of the fault are downthrown 46 cm farther than those on the north side. A total dip-slip displacement of 90 cm is estimated at this site by summing the displacements on the faults near the top and base of the scarp (44 cm + 46 cm).

A laterally eroded bank overlain by river channel gravel fill is exposed in the trench and may be the remnants of an old, buried fault-line scarp. The channel buries the truncated, interlayered fluvial sands and gravels along its northern margin, indicating that the river had eroded laterally into a bank that was at least 1 m high. The channel gravel interfingers with the interlayered fluvial sands and gravels along its southern margin. The channel fill is displaced by the fault near the scarp base, proving that the buried bank was eroded and covered by the gravel prior to the most recent faulting event. The old river channel can be traced for at least 100 m on both sides of the trench, where it forms a gentle topographic low about 4 to 5 m wide and up to 30 cm deep parallel to the base of the present fault scarp. Searching beneath the channel fill for evidence of earlier faulting failed because of groundwater seepage.

TRENCH 2

The second trench was dug 300 m west of trench 1. The fault is delineated by a bank about 1 m high trending N. 80° E. (fig. 3). The stratigraphy in the trench consists of wind-blown sand overlying till. A poorly developed, discontinuous weathering horizon

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occurs along the top of the till.

A single fault was intersected in the trench near the center of the bank. In the eastern wall the fault dips 68° S. and the sand-till contact has a normal, dip-slip offset of 110 cm (102 cm throw). In the west wall the fault dips 64° S. in the upper part of the trench but at a depth of 1.9 m the fault plane reverses dip (to 82° N.). The sand-till contact in this wall is offset 102 cm vertically, the same amount of throw measured directly across the trench in the eastern wall. No evidence of splay faulting occurs where the fault plane reverses dip direction, although any such evidence may have been obliterated by postfaulting compaction of the sand. No evidence of a buried fault-line scarp was observed in the trench.

A river terrace bank located about 30 m east of trench 2 was offset dextrally about 2.4 m. This displacement was estimated by measuring the offset of the base of the bank across the fault. The bank morphology was irregular because of erosion and the growth of vegetation; thus, the measurement should be considered an estimate only.

The 2.4-m dextral offset resolved with the 1.1-m dip-slip offset measured in trench 2 yields a total oblique

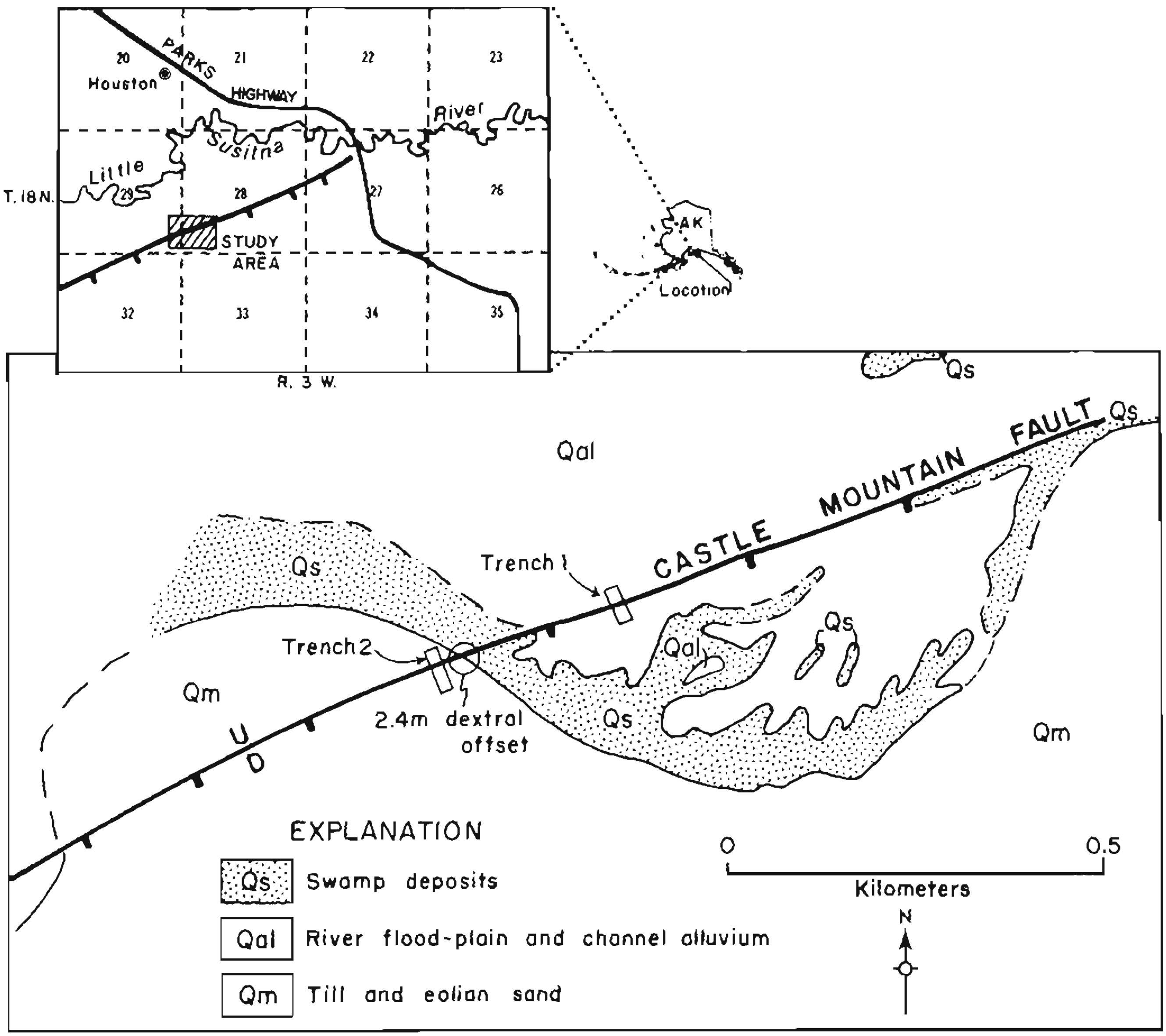

Figure 1. Generalized surficial geologic map of the trenching sites.

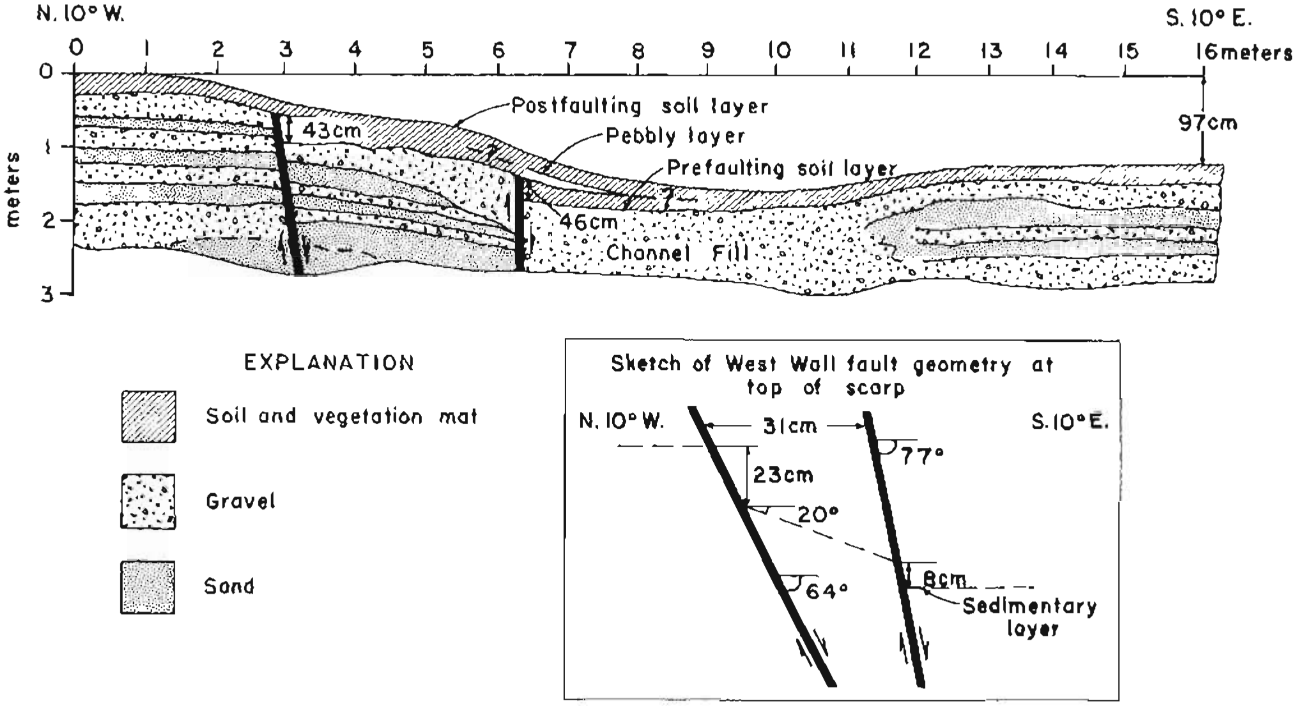

Figure 2. Geologic cross section of the east wall of trench 1.

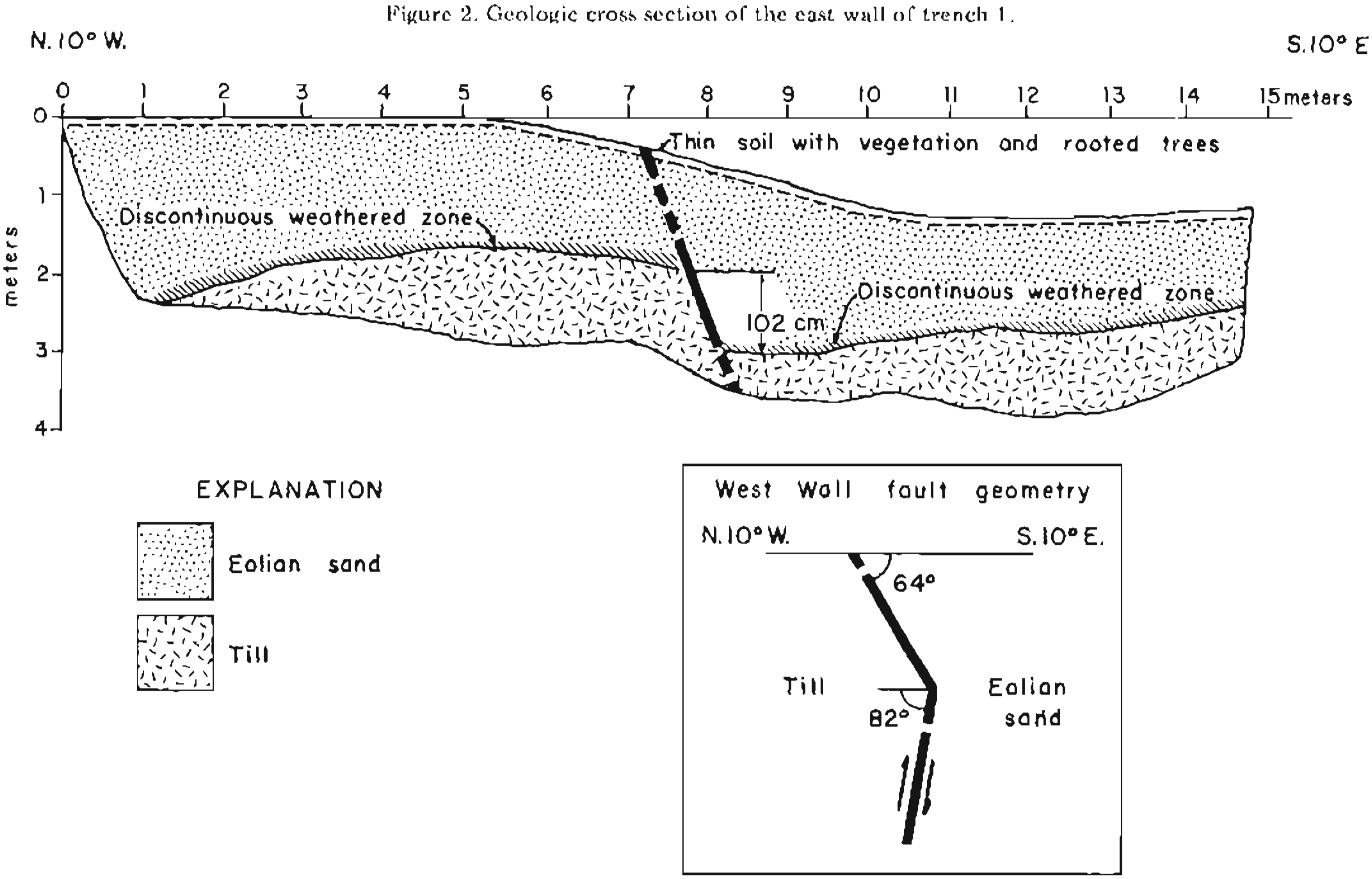

Figure 3. Geologic cross section of the east wall of trench 2.

fault displacement of 2.6 m at this locale.

ACKNOWLEDGMENTS

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BLUFF POINT LANDSLIDE, A MASSIVE ANCIENT ROCK FAILURE NEAR HOMER, ALASKA

By Richard D. Reger[1]

INTRODUCTION

The likelihood of future offshore drilling for petroleum in lower Cook Inlet has increased interest in locations for onshore support facilities along the southwestern shore of the Kenai Peninsula. Because most of this coast consists of near-vertical cliffs, undeveloped locations suitable for onshore facilities near likely drilling sites are rare. The attention of industry has been drawn to one area at the mouth of Kachemak Bay 4.5 km west of Homer, where an extensive, relatively flat area near sea level offers a potential site for development. This area, which was mapped as the Bluff Point moraine by Foster and Karlstrom (1967, pl. 1), is actually an ancient landslide as previously mapped by Barnes and Cobb (1959, pl. 18) and is herein termed the 'Bluff Point landslide' (fig. 1). The purpose of this paper is to briefly describe this feature, propose its origin, and estimate its age.

DESCRIPTION

The Bluff Point landslide is the largest slope failure along the southwestern shore of the Kenai Peninsula (Riehle, Reger, and Carver, 1977). The irregular, 100- to 215-m-high headwall extends for 5.5 km in a northwest-southeast direction roughly parallel to the coastline (fig. 2). Exposed in the headwall is a thin cap of till underlain by rocks of the Kenai Group, consisting of a sequence of moderately to weakly indurated fine to medium sandstone interbedded with siltstone and claystone layers and lenses of lignitic to subbituminous coal (Barnes and Cobb, 1959). These sedimentary rocks of late Miocene age (Triplehorn, Turner, and Naeser, 1977) generally strike northeast to north-northeast and dip about 3° S.E. Steep postlandslide gullies have been eroded as deeply as 160 m into the headwall by intermittent torrents (fig. 3). During the major earthquake of March 27, 1964, fissures as deep as 6 m developed on the crown parallel to the headwall and as far back as 30 m from the edge (Waller, 1966, fig. 5). Linear gullies have been eroded along parallel but older cracks (fig. 2, locality A).

Landslide deposits extend as much as 0.9 km southwest from the base of the headwall to the present shore of Kachemak Bay (fig. 2) and continue beneath the waters of the bay at least another 0.4 to 0.5 km as indicated by the outer limit of boulders exposed or barely submerged during lowest tides (fig. 3). Depth soundings on the nautical chart of the area (NOAA NOS, 1976) indicate that the body of the landslide, measured perpendicular to the coastline, may be 1.7 to 2.0 km long. A thickness of at least 30 m of landslide deposits is exposed in sea cliffs. Landslide deposits appear to consist of complexly disrupted blocks of the Kenai Group that have been rotated.[2] The landslide reaches a maximum elevation of about 90 m and consists of subparallel discontinuous ridges with up to 60 m of relief. Shallow ponds occupy some surface depressions (fig. 2). Topographic lows have been partially filled by a mixture of alluvium and colluvium, composed of fine to medium fluvial sand with some silt, scattered to rare angular detrital blocks of coal up to 15 cm across, and rare fragments of buried wood interlayered with 0.1- to 0.4-m-thick tongues of debris-flow diamicton containing few to numerous blocks of coal. At the base of the eroded headwall, talus interfingers with and overlies the mixed alluvium and colluvium (fig. 2). Valley fills reach thicknesses of over 16 m (fig. 4) and undisturbed bedding dips 3° to 5° S. to S.W. Recent slumping has rotated valley fills so that affected strata dip 25° to 30° N. to N.E. (fig. 2). The debris flow near the southeast end of the Bluff Point landslide (fig. 2) occurred in response to seismic shaking during the 1964 earthquake (Waller, 1966, fig. 2).

ORIGIN

The composition of the landslide deposits demonstrates that the Bluff Point landslide resulted from failure of the moderately to weakly consolidated bedrock. The cause of the initial failure is unknown, but it probably was loss of support when the late Wisconsinan glacier that formerly abutted against the slope withdrew from the locality.[3]

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[2]The internal structure has not been systematically examined in detail over the entire landslide. No major surface of failure was directly observed.

[3]Karlstrom (1964, pl. 4) placed the limit of the Killy advance (Naptowne Glaciation) along Diamond Creek and the limit of the Skilak advance (Naptowne Glaciation) along the base of the headwall of the Bluff Point landslide. In my opinion, the abandoned meltwater channels extending from the top of the

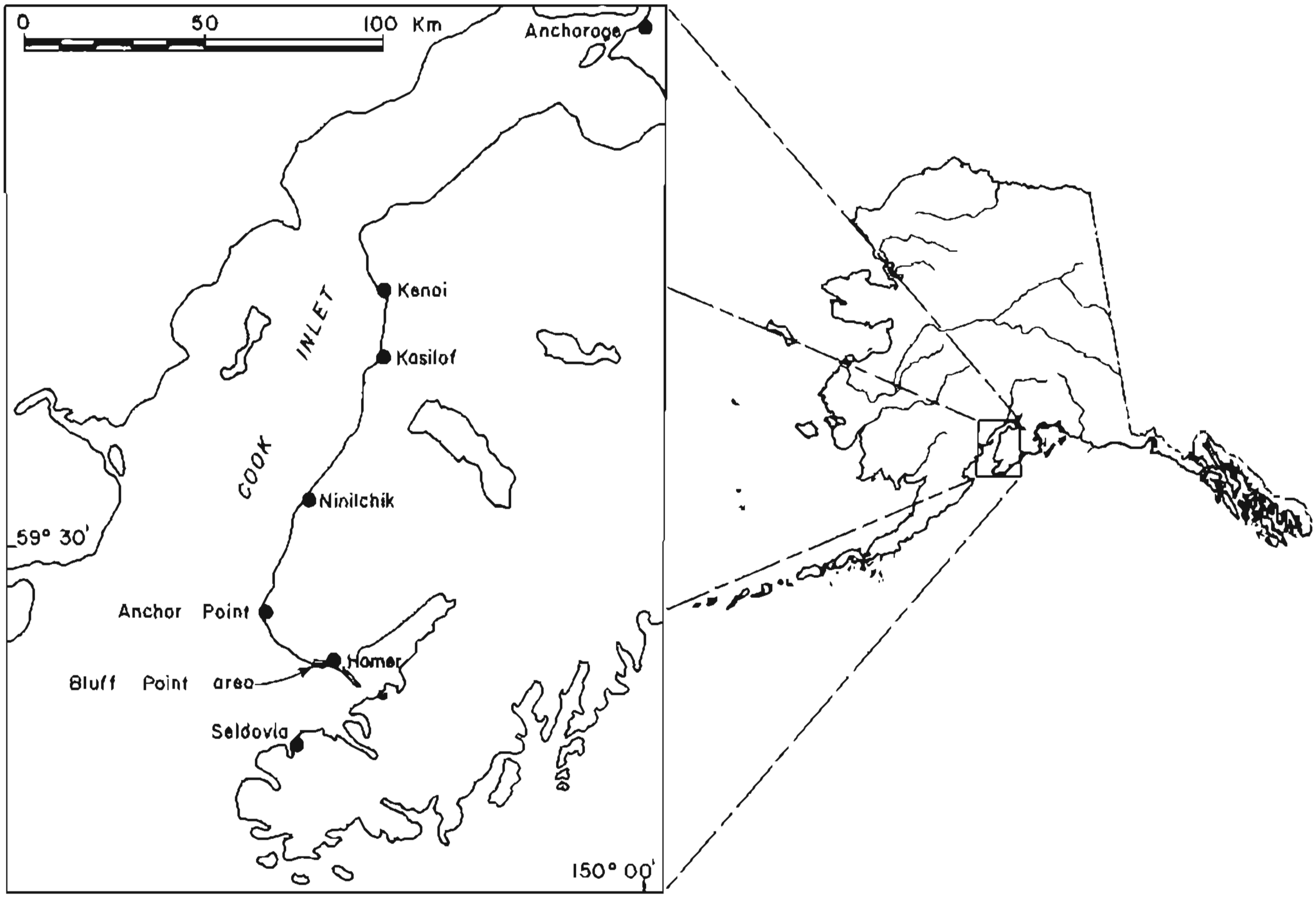


Figure 1. Location of the Bluff Point landslide area.

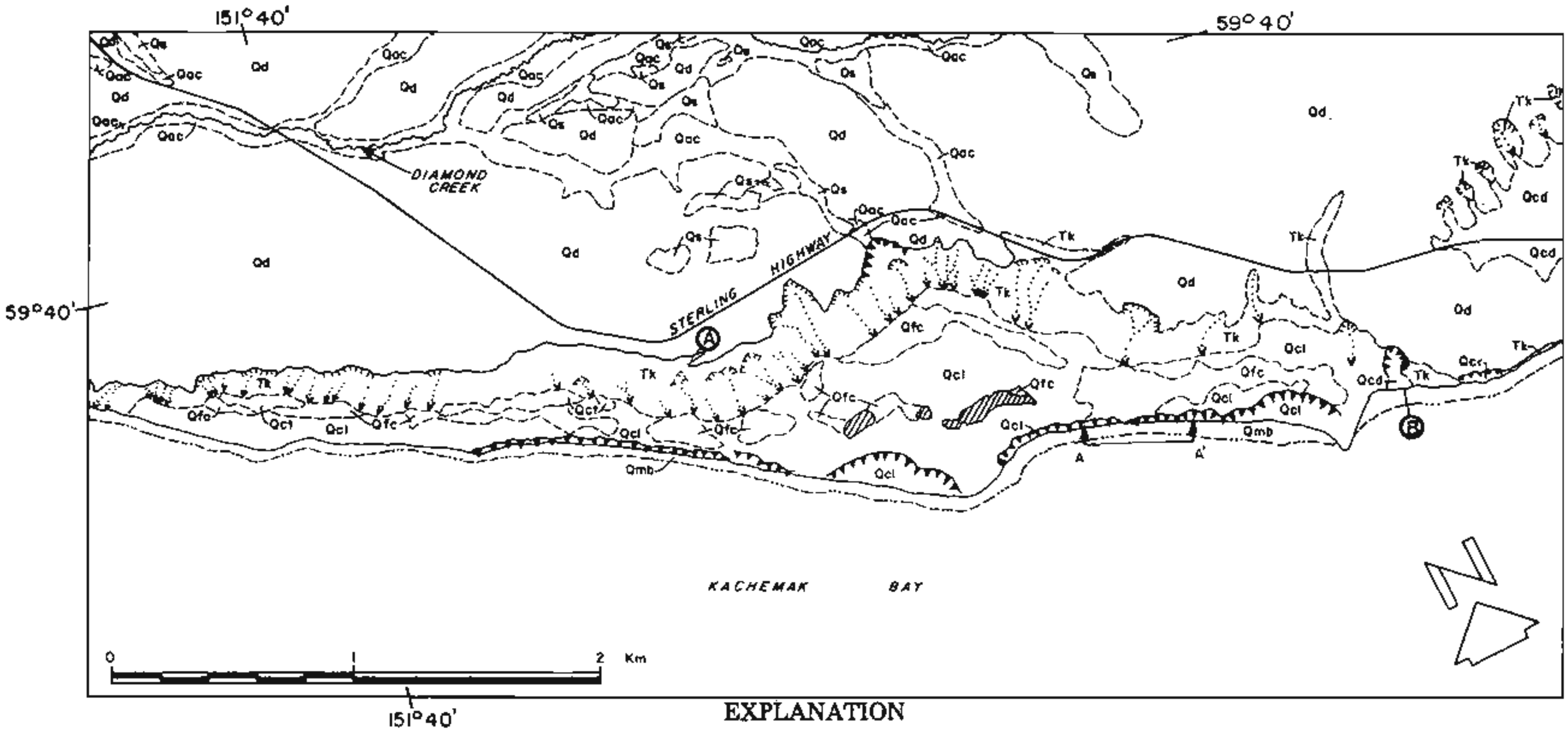
Other possible, although less likely, causes are progressive wave erosion at the base of the high sea cliff and seismic shaking. Certainly the secondary rotational failures along the modern sea cliff transecting the landslide were caused by this basal undercutting. Perhaps secondary slumping was triggered by seismic shaking, but Waller (1966) did not identify any slumps released during the March 27, 1964, Alaska earthquake. According to Foster and Karlstrom (1967) there were no large bedrock failures of the size of the Bluff Point landslide along the western coast of the Kenai Peninsula during the intense shaking of the March 27, 1964, earthquake, although many smaller bedrock failures occurred. The overall width-to-length ratio (0.30 to 0.36), the presence of subparallel discontinuous transverse ridges, and the known structure indicate that the failure was characterized by the falling, tilting, and rotation of discrete blocks and wedges of bedrock. The failure apparently did not flow significantly or move by basal slippage (for example, airborne sliding). The movement may have occurred as a single event, although the irregular form of the headwall suggests that the failure could also have been a series of events in which bedrock wedges or blocks were released soon after the removal of the adjacent mass of bedrock.

headwall north to Diamond Creek (fig. 2) drained the margin of the Skilak-age glacier situated at the *top* of the headwall; this interpretation is more consistent with physiographic and geologic evidence up Kachemak Bay to the northeast (Reger, 1977).

AGE

The antiquity of the Bluff Point landslide is indicated by the deep gullying of the headwall, the development of large talus cones at the base of the headwall, the deposition of considerable mixed alluvium and colluvium on top of the landslide, and the extensive notching of the landslide by wave attack. The considerable degree of postfailure modification and the most likely cause of release (removal of support) suggest that the event happened shortly after glacier ice began to thin and recede from its maximum extent in lower Kachemak Bay following the culmination of the Skilak stadial advance, perhaps as early as 12,000 years ago and certainly by 9,000 years ago (Karlstrom, 1964). Clearly the locality has not been glaciated since the landslide occurred.

Radiocarbon analyses of wood collected from valley



EXPLANATION

Quaternary

- Qac — Abandoned meltwater channel alluvium
- Qfc — Mixed alluvium and colluvium
- Qcd — Debris-flow deposit
- Qcl — Landslide deposit
- Qcr — Rockfall rubble
- Qct — Talus
- Qd — Till
- Qmb — Granular beach deposits
- Qs — Swamp deposits

Tertiary

- Tk — Continental clastic rocks of the Kenai Group

- Ⓐ Locality cited in text
- Steep torrential gully
- Steep, U-shaped headwall of torrential gully
- Relatively fresh landslide scarp
- Pond
- Contact, dashed where approximate
- Tide line
- A A' Cross section shown in figure 4

Figure 2. Geology of the Bluff Point landslide area. Locality B is the same as locality K in Karlstrom (1964, pl. 4). Modified from Reger (1977).

fill deposited on the landslide after the bedrock failure demonstrate that the Bluff Point landslide is at least 1,555 ± 135 radiocarbon years old (fig. 4). The base of the section was obscured in 1977, so I could not determine the thickness of material deposited on the landslide prior to 1,555 ± 135 radiocarbon years ago or if a soil profile was developed on the landslide deposits at this locality before the valley fill was laid down. The younger date in the section is evidence that secondary slumping along the modern sea cliff occurred after 1,160 ± 120 radiocarbon years ago. A date of 2,250 ± 300 radiocarbon years for log fragments collected in 1950 by Daniel B. Krinsley under 1.9 m of slope wash and 'landslide'[4] debris (fig. 2, locality B; Broecker and others, 1956, p. 156) undoubtedly postdates the Bluff Point landslide.

Figure 3. Aerial view toward north of the central part of Bluff Point landslide. Photograph 53-32 taken at low tide on August 15, 1976.

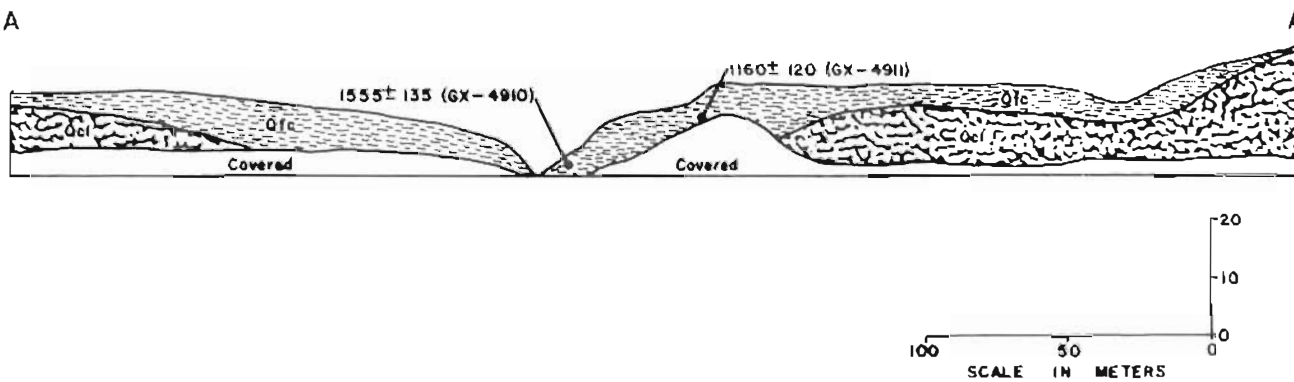


Figure 4. Slumped valley fill (Qfc) and landslide deposits (Qcl) in sea cliff along section A-A', Bluff Point landslide. See Figure 2 for location of section. Radiocarbon ages date buried wood. Numbers in parentheses are Geochron Laboratories' sample numbers.

CONCLUSIONS

The Bluff Point landslide appears to be stabilized, except in the zone of slumping within 10 to 150 m of the modern sea cliff. The main hazard to development on the landslide is rock falls from the headwall. Cracks in the crown pose a serious rock stability problem. These fractures not only weaken the bedrock, but serve as conduits for surface and near-surface water, thereby promoting the development of stress toward the free face and the lubrication of potential failure surfaces. Other processes that potentially threaten future development include debris flows and torrential floods, especially in steep headwall gullies and near the mouths of these features.

[4]Italics are mine. I doubt that the 'landslide' debris was deposited during the massive failure that formed the Bluff Point landslide. During my visit to the site in 1977 I saw only fairly fresh debris-flow deposits overlying older mixed fluvial-colluvial sand on top of disrupted rocks of the Kenai Group. Only the lower unit is part of the Bluff Point landslide proper. The base of the Bluff Point landslide was not exposed.

ACKNOWLEDGMENTS

Invaluable field assistance was provided by Jeffrey T. Kline and Cheri L. Daniels. Many of the ideas presented in this paper evolved during stimulating discussions with them and James R. Riehle. I appreciate thoughtful reviews of the preliminary manuscript by Kline, Riehle, and Randall G. Updike.

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RECURRENT LATE QUATERNARY FAULTING NEAR HEALY, ALASKA

By Robert M. Thorson[1]

INTRODUCTION

A large normal fault that offsets Tertiary and Quaternary strata just south of Healy was discovered during geological mapping of the Nenana River valley in 1978 (fig. 1). The purpose of this report is to document the fault and to describe the inferred episodes of movement that occurred along it during late Quaternary time.

The Healy area is at the northern front of the Alaska Range (fig. 1). The Birch Creek Schist of Precambrian or Paleozoic age is the oldest rock in the area, and generally occurs south of Healy Creek (Wahrhaftig, 1970). Rocks of the foothill belt to the east and north of Healy consist of folded coal-bearing strata of middle Tertiary age. The Nenana Gravel, consisting of oxidized semiconsolidated gravel deposits, is Pliocene(?) and mantles much of the older strata north of Healy.

Four major glaciations of the Nenana River valley were recognized by Wahrhaftig (1958). Deposits of the two oldest glaciations are not present in the Healy area. Deposits of the Healy Glaciation in the study area consist primarily of gravelly hummocky moraines and outwash terraces. They were derived from a glacier that nearly covered the Healy area, and are presently believed to be of early Wisconsinan age (30,000-70,000 years B.P.). During the subsequent Riley Creek Glaciation, glaciers advanced no farther north than about 15 km south of Healy, but their meltwater caused extensive alluviation of the Nenana River north of the mountain front. Deposits of the initial and major phase of the Riley Creek Glaciation (Riley Creek I) are almost certainly correlative to deposits that occur throughout Alaska and northwestern North America, and are believed to date between about 13,000 and 22,000 years B.P. The large outwash terrace west of Healy was built during Riley Creek I time (fig. 1). Riley Creek II outwash was deposited during a significant readvance of the Nenana River valley glacier during late Riley Creek time. The younger Riley Creek II drift predates occupation of the Carlo Creek archeologic site, which occurred at about 8,500 years B.P. in the upper Nenana River valley (Peter Bowers, pers. comm., 1978).

Wahrhaftig (1958) first presented evidence for Quaternary tectonic deformation of the northern Nenana River valley. In his classic study of its glacial deposits, he demonstrated that extensive upwarping of glacial outwash terraces occurred north of the range front. Wahrhaftig (1970) also mapped a large high-angle fault about 4 km north of Healy that offsets late Quaternary terrace deposits.

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EVIDENCE FOR FAULTING

Just south of Healy a recent fault forms a prominent scarp that strikes N. 74° E. for about 2 km and clearly crosscuts the outwash terraces (fig. 1). This fault is hereafter referred to as the 'Healy fault.' The north side of the scarp is upthrown as much as 5.5 m where it crosses the Riley Creek I outwash terrace about 0.7 km south of Healy. Three hundred meters farther east, where the fault crosses a younger Riley Creek I outwash terrace, is only 4 m high. Southeast of Healy the fault crosses a Riley Creek II outwash terrace; where it intersects the Nenana River a gully about 25 m deep and 75 m long has been cut through the Quaternary and Tertiary deposits (figs. 2 and 3; table 1). Railroad maintenance and construction activity has obscured the surface relations of the fault in this area, but it appears to be upthrown on the north. Projection of the fault eastward indicates that it crosses the Nenana River within about 100-200 m south of the existing Golden Valley Electric Association (GVEA) power plant at Healy. The fault trace also extends southwest of Healy but cannot be traced beyond the marginal portion of the Healy moraine.

The fault scarp has been modified along much of its length by colluvial processes and vegetation growth but is very fresh where vegetation is largely absent and where partly stabilized lichen-covered bouldery gravel covers the scarp. The Healy fault scarp slopes as steeply as 32° but is generally less than 20°.

Surface drainage on the terrace surfaces is strongly controlled by the Healy fault. Three large ponds on the downthrown side—one on each Riley Creek I outwash terrace—are dammed against the fault scarp (fig. 1). These ponds have been modified and expanded by beaver-dam construction. Fresh-water springs issue from the terrace scarps near the fault zone and deep, straight gullies occupy the fault trace where it crosscuts terrace edges.

Where the Healy fault intersects the Nenana River a major shear zone 60-90 cm wide is exposed in the north wall of the gully (figs. 2 and 3). Both the coal-bearing strata and the overlying gravel units are clearly offset and foliation is well developed in both units. In the

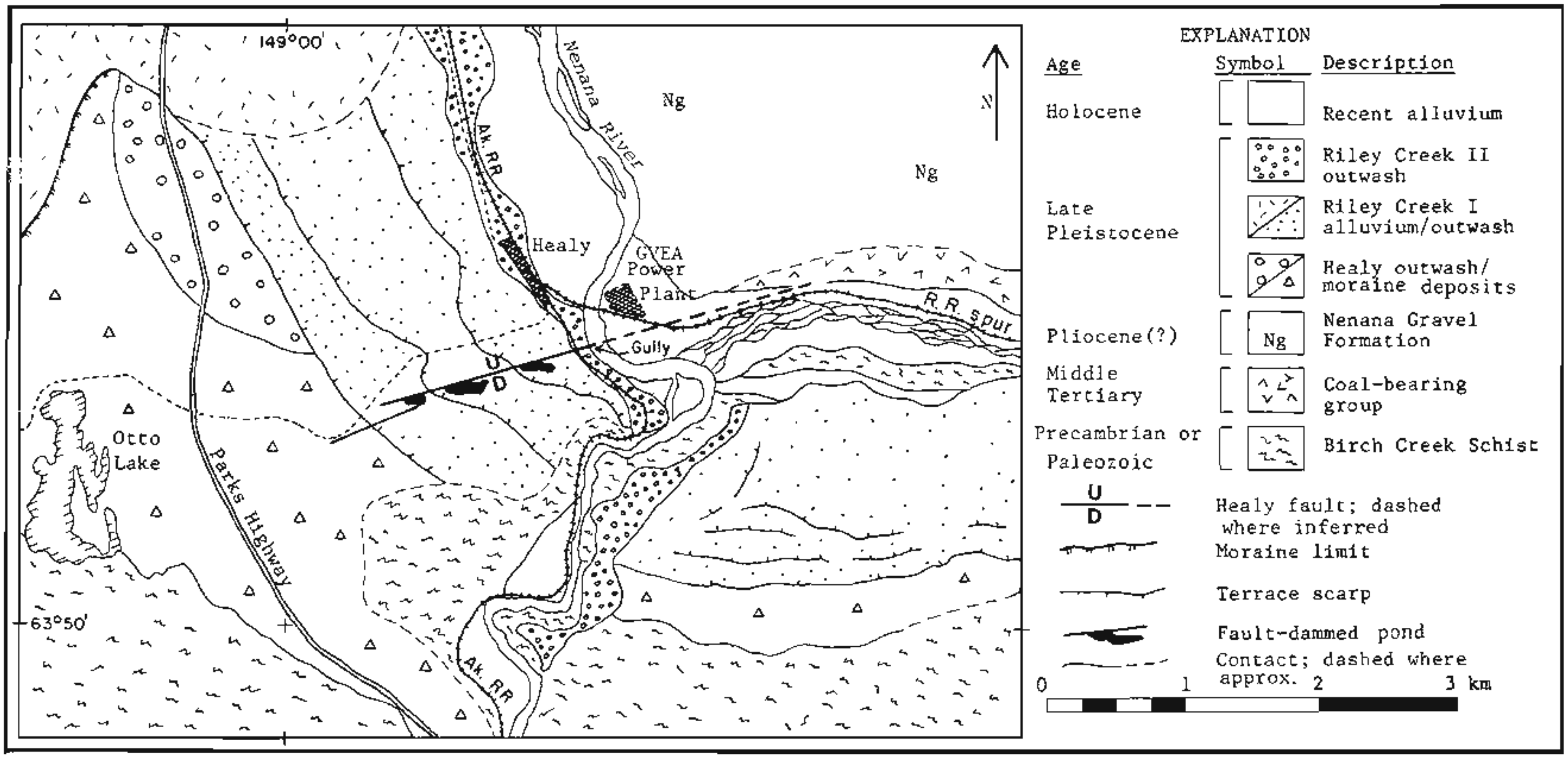

Figure 1. Generalized geologic map of the Healy area showing the Healy fault. Modified from Wahrhaftig (1970).

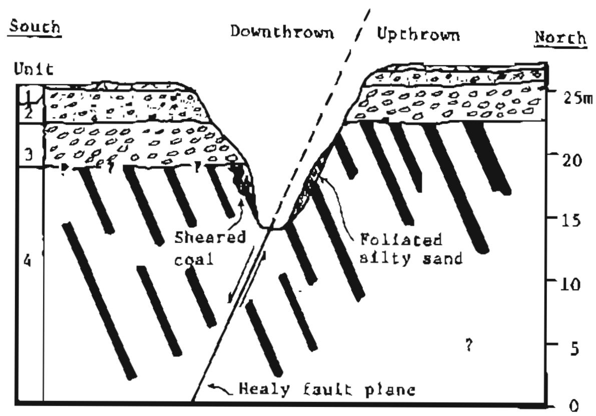

Figure 2. Generalized cross section showing geologic relations exposed in a gully where the Healy fault intersects the Nenana River. The fault plane strikes N. 74° E. and dips 66° S. The horizontal and vertical scales are the same; vertical distances were measured with a hand level and horizontal distances estimated. See table 1 for description of geologic units.

poorly consolidated coal-bearing sandstones, foliation consists of 1-cm-thick parallel sheets that strike N. 74° E. and dip 66° S. In the overlying gravel, foliation consists of discoidal pebbles and cobbles that have been sharply realigned parallel to the fault plane. The attitude of the foliation in the gully wall lies within 10° of the trace of the surface scarp and suggests that the Healy fault is a south-dipping normal fault.

A shear zone in the north gully wall lies several meters west of the major shear zone (fig. 3). The minor shear zone slightly offsets unit 2 and controls oxidation of the gravel. However, the apparent differences in surface altitude of the Riley Creek II terrace on opposite sides of the gully (fig. 2) indicate that faulting of comparable age may have occurred elsewhere in the gully, perhaps along the gully floor. At one locality along the south wall of the gully the coal beds are severely folded and faulted. This deformation may represent a shear zone that crosses the gully floor near its axis. Both lateral erosion by the Nenana River and active faulting are responsible for the steep, irregular, long profile of the gully floor.

INFERRED EPISODES OF FAULT MOVEMENT

The Healy moraine shows less fault displacement than the younger Riley Creek sediments, perhaps suggesting that either fault movement has been confined to the axial portions of the Nenana River valley or that surface displacements on the Healy moraine did not occur because the additional thickness of glacial drift attenuated faulting at depth. Also, large-scale movement on the Healy fault may not have occurred between the Healy and Riley Creek Glaciations west of a point 1 km east of the Parks Highway (fig. 1).

Table 1. *Measured stratigraphic section of units exposed on the south wall of the gully where the Healy fault intersects the Nenana River.*[1]

Unit	Thickness (m)	Description
1	0.5	*Spoil.* Mixed deposit of sand, railroad debris, and gravel. Nearly horizontal profile and uniform thickness. Terrace surface not extensively modified.
2	2.5	*Interbedded sand, gravel.* Rounded to subrounded clasts to medium cobble size (commonly 10-20 cm) in well-sorted sand-granule matrix. Interbedded with 5- to 10-cm-thick beds of finely bedded, well-sorted fine-coarse sand with minor silty-sand beds. Sand beds occasionally exhibit undulose sinusoidal deformation with amplitudes of 1 to 10 cm. Pronounced reddish-brown oxide staining generally follows textural contacts. Lower contact distinct but appears gradational.
3	3.5	*Coarse gravel.* Rounded to subrounded clasts to small boulder size (commonly 20-30 cm) in well-sorted sand-granule matrix. No horizontal bedding, but most clasts are discoidal and exhibit pronounced horizontal fabric.
4	19.0	*Coal-bearing unit.* Interbedded white sandstone, pebbly sandstone, silty claystone, and subbituminous coal. Unit poorly consolidated.

[1]See figs. 2 and 3 for geologic relationships between units.

There is evidence for at least two episodes of movement along the Healy fault during and after the Riley Creek glacial maximum, which probably occurred about 14,000 years ago (table 2). The fault scarp is about 1.5 m higher on the older Riley Creek I terrace than on the younger Riley Creek I terrace, suggesting that at least 1.5 m of displacement occurred during Riley Creek I time. Because both gravel units exposed in the gully appear to have a gradational relationship (fig. 3), they are interpreted to have occurred during Riley Creek II time.

The possible truncation of the major shear zone by the younger gravel (unit 2) suggests that about 3.5 m of displacement occurred during Riley Creek II time, probably between about 8,500 and 13,000 years ago. The possible shear zone developed in unit 2 (fig. 3) and the apparent differences in altitude of the Riley Creek II outwash terrace on opposite sides of the gully (fig. 2) suggest that fault movement also occurred after Riley Creek II time; the freshness of the fault scarp along much of its length suggests that this episode of faulting may have occurred within the past few thousand years.

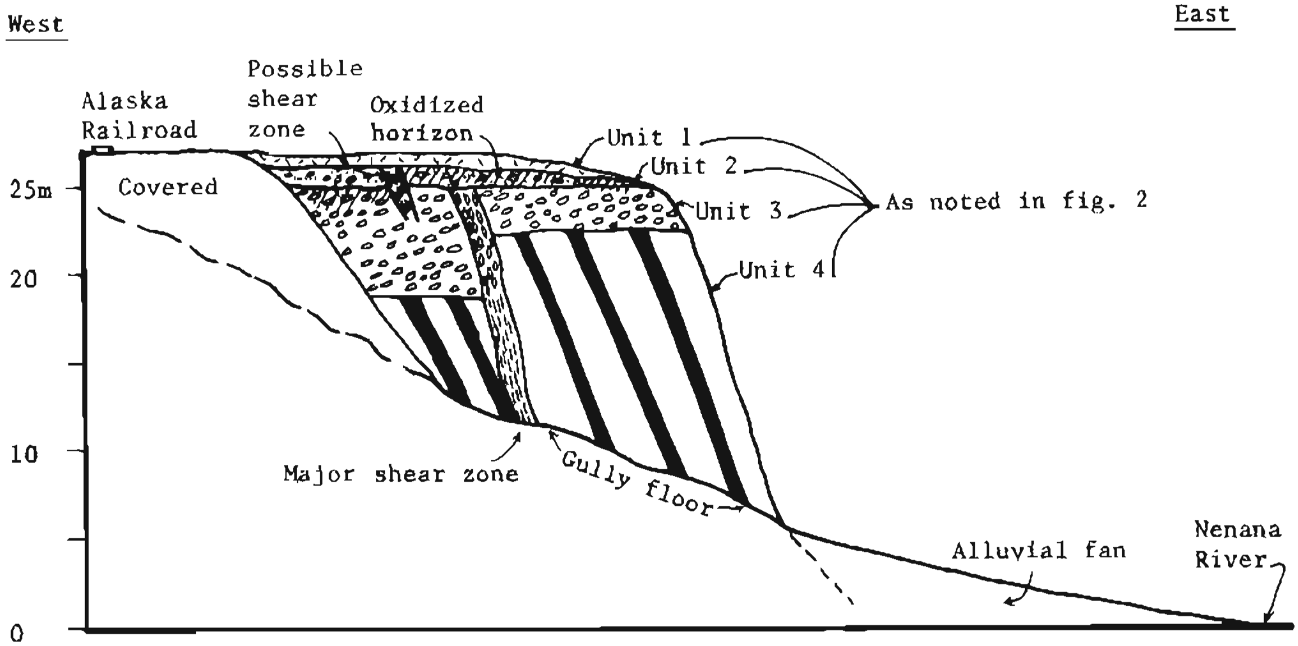

Figure 3. East-west longitudinal cross section showing geologic relations exposed on the north wall of a gully where the Healy fault intersects the Nenana River. The horizontal and vertical scales are the same; vertical distances were measured with a hand level and horizontal distances estimated. See table 1 for description of geologic units.

Table 2. *Inferred episodes of movement along the Healy fault*

Episode	Vertical movement (m)	Relative age	Estimated age (years B.P.)	Evidence
3	1.5(?)	Post-Riley Creek II	100-10,000	Fault apparently causes surface displacement of Riley Creek II terrace. Fault scarp is nearly continuous, gravelly at edge, and as steep as 32°. Surface drainage is controlled by fault. Gullies along fault trace are not deeply eroded.
2	3.5(?)	Riley Creek II time	8,500-13,000	Offset in coal-bearing formation probably postdates deposition of coarse gravel but predates deposition of interbedded sand and gravel. Both units are interpreted to be of Riley Creek II age.
1	1.5(?)	Riley Creek I time	13,000-22,000(?)	Fault scarp appears about 1.5 m higher on older Riley Creek I terrace than on younger Riley Creek I terrace.

CONCLUSIONS

There is convincing evidence that the Healy fault underwent significant movement during late Quaternary time. During each inferred episode of faulting since the beginning of Riley Creek time, at least 1.5 m of vertical displacement may have occurred. The inferred ages and amounts of displacement along the Healy fault are extremely tentative. A more detailed study is required to either support or refute these preliminary conclusions.

ACKNOWLEDGMENTS

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GLACIATION OF INDIAN MOUNTAIN, WEST-CENTRAL ALASKA

By Richard D. Reger[1]

INTRODUCTION

The 1,290-m (4,234-foot) high summit of Indian Mountain is one of a series of generally northeast-trending ridges in the Indian River Upland province of western interior Alaska (fig. 1; Wahrhaftig, 1965, pl. 1). Evidence of at least two glaciations and one postglaciation period of climate colder than today is recognized in the Indian Mountain area.

SLEEPY BEAR GLACIATION

The earliest recognized ice advance, here named the Sleepy Bear Glaciation, left a relatively extensive lobate terminal moraine as far as 6.4 km from the cirque headwall in the drainage of Sleepy Bear Creek (figs. 1 and 2A). Morainal morphology is subdued. The few remaining kettle lakes are nearly filled with organic silt, *Sphagnum* peat, and colluvium or slopewash material.

An olive-gray lake clay is draped over lower portions of the terminal moraine and may interfinger with till of Sleepy Bear age.[2] This lacustrine deposit occurs to an elevation of at least 303 m (1,000 feet) along the eastern flank of Indian Mountain and to 296 m (970 feet) about 8 km east of Utopia (Reger and Reger, 1972). Total clay thickness is unknown; however, the clay overlying the till is not so thick that it completely obscures primary morainal relief. Sections of lake clay up to 61 cm thick were exposed in test pits; frozen ground prevented deeper digging.

During the Sleepy Bear Glaciation the entire compound cirque at the head of Sleepy Bear Creek was full of ice and the basic morphology of the bowl was fashioned. Morainal relief in the cirque is generally from 0.3 to 1 m because of postdepositional modification, primarily by slope processes. Sleepy Bear drift is surficially sorted into well-developed but inactive stone nets, particularly downslope from transverse nivation hollows, and it has been notched by nivation to form scarps 3 to 5.5 m high (Reger, 1975).

Ice of the Sleepy Bear Glaciation extended down the east tributary of Cirque Creek at least 1.8 km from the base of the cirque headwall (fig. 2B). The wide, U-shaped cross profile of Cirque Creek valley relative to the extent of the later Indian Mountain advance indicates that the entire cirque was occupied by ice during Sleepy Bear time. The moraine of Sleepy Bear age on the lower east wall of the cirque has been removed by postdepositional slope processes. On the west side of the cirque the moraine is subdued; yet it retains a distinctive ridge form. Relief of Sleepy Bear drift contrasts sharply with relief of the younger moraine in this drainage. Tills of the two glaciations also differ in lithologic content. Ice of the Sleepy Bear Glaciation plucked granodiorite cropping out of the west wall of the cirque, producing a boulder till relatively rich in granodiorite erratics. The later advance barely reached the granodiorite and produced a till with more greenstone and greenschist clasts. On the west side of Cirque Creek only a thin, dissected blanket of drift, rubble sheets derived from this material, and scattered erratics are preserved to indicate the former ice extent (fig. 2B). Downvalley these deposits are either buried by colluvium and later outwash deposits or have been destroyed by erosion. Where Sleepy Bear moraines are buried, their presence is inferred from a discrete, linear change of slope trending across the colluvial blanket.

The youngest possible age of the Sleepy Bear Glaciation is Illinoian. This assignment is established because the terminal lobe of the Sleepy Bear moraine in the Sleepy Bear drainage is covered by lake clay of Illinoian age. North of the study area Hamilton (1969, p. 186-187; pl. 1) found evidence for an advance of Illinoian ice southward from the Brooks Range into the Koyukuk and Kanuti lowlands. Field relationships indicate that this ice abutted against low ridges at its southern limit, formed an ice barrier to at least 303 m (1,000 feet) elevation, and blocked the lower drainage of Mentanontli River 30 km northeast of Indian Mountain. Study of aerial photographs and maps indicates that an extensive proglacial lake of Illinoian age formed in the Mentanontli drainage as a consequence of this damming. Evidence includes wave-cut cliffs, high-level outwash and deltaic deposits, concentrations of thaw lakes (indicative of ice-rich frozen lacustrine sediments), and anomalous stream drainage patterns. The extent of this ice-dammed lake is only generally known. During Illinoian time lake waters apparently drained across the low divide about 9 km east of Utopia and down lower Indian River into the Koyukuk River about 23 km south southwest of Hughes (fig. 1). Eventual lake drainage,

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[2]Eakin (1916, p. 58) first described this widespread lacustrine deposit.

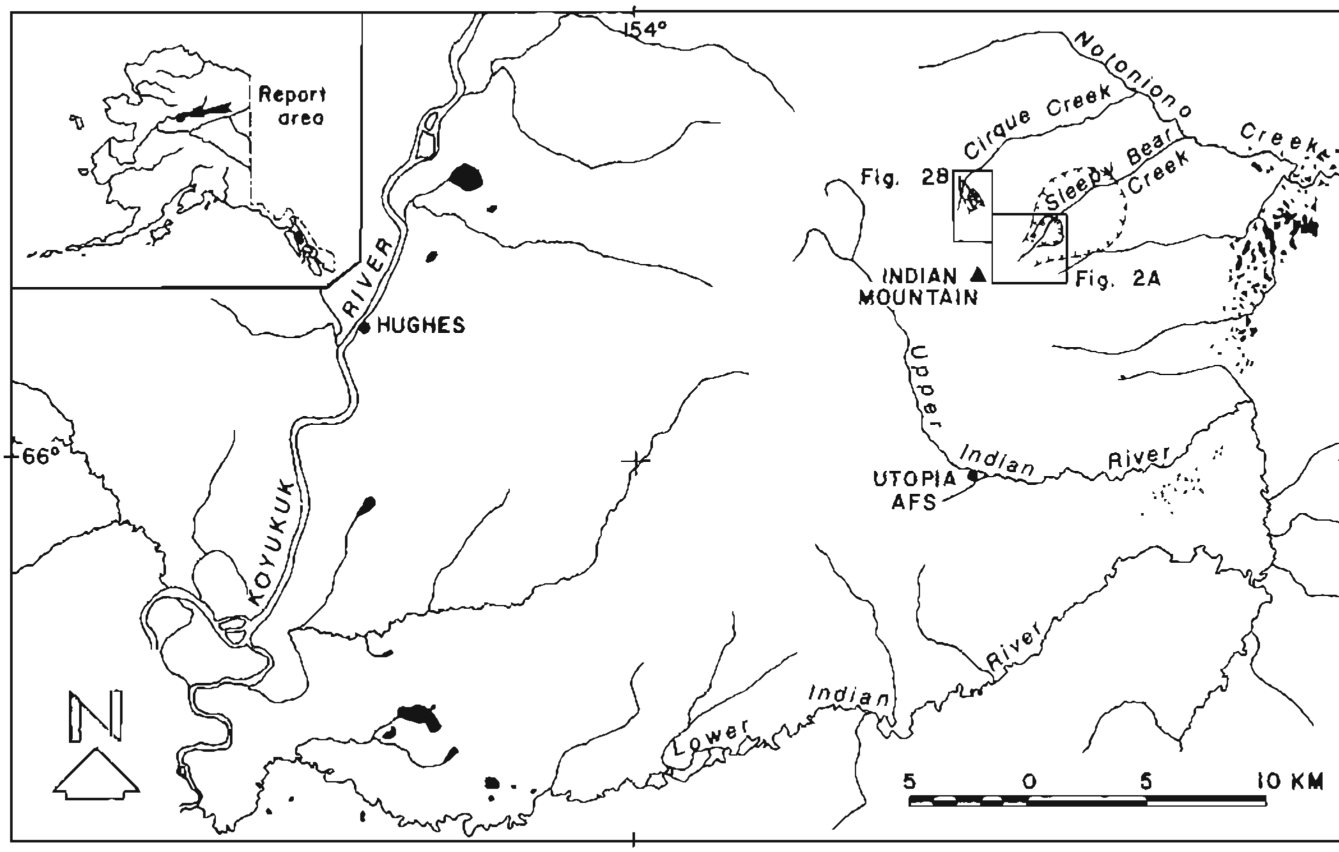

Figure 1. General location of Indian Mountain and geographic features in the vicinity. Single-hachured dashed line - limit of Sleepy Bear Glaciation; double-hachured solid line - limit of Indian Mountain Glaciation.

probably due to the northward retreat of Brooks Range ice, left the overflow channel as the deeply incised, anomalously wide, and underfit valley of lower Indian River.

INDIAN MOUNTAIN GLACIATION

A later advance of ice in the drainage of Sleepy Bear Creek, here named the Indian Mountain Glaciation, is defined by an arcuate, steep-fronted terminal moraine at the mouth of a glacial trough 2.4 km downvalley from the cirque headwall. This morainal ridge has higher and sharper relief than the Sleepy Bear terminal moraine 4 km farther downvalley; maximum relief on the Indian Mountain moraine is 6 to 12 m. Well-developed subarctic brown forest and podzolic soils 33 to 39 cm thick occur along the crest of the terminal moraine.

Morainal distribution indicates that the valley of Sleepy Bear Creek was not completely filled by Indian Mountain ice. The former limit of the Indian Mountain glacier on the Illinoian cirque floor is indicated by a distinctive topographic change that is emphasized by a demarcation of vegetation cover. *Cladonia* (reindeer lichen) is dominant on Indian Mountain drift and Sleepy Bear drift has more *Betula nana* (dwarf birch), *Salix* (willow), and *Cassiope tetragona* (mountain heather). Indian Mountain moraine on the cirque floor has a maximum relief of 1 to 2 m. The till surface is crudely sorted into stone nets. Indian Mountain drift also remains as lumpy accumulations along the lower side slopes and upper floor of Sleepy Bear Creek valley.

During Indian Mountain time ice also advanced down the east tributary of Cirque Creek to about 1.3 km from the cirque headwall and deposited a conspicuous, tongue-shaped moraine with a maximum local relief of about 18 m (fig. 2B). Most drift of Indian Mountain age consists of greenstone and greenschist clasts in a silty matrix, although the western quarter of the till sheet contains granitic components. A subarctic brown forest soil 25 cm thick occurs on the crest of this moraine. Till surfaces bear crudely sorted stone nets. During and following retreat of Sleepy Bear ice in Cirque Creek, a gully at least 30 m deep was cut through Sleepy Bear drift just beyond the cirque mouth. An outwash fan bearing a subarctic brown forest soil of Indian Mountain age was deposited in this gully 2.9 km downvalley from the cirque headwall.

The storied cross profile of Sleepy Bear Creek valley indicates that considerable time elapsed between the recession of Sleepy Bear ice and the expansion of the

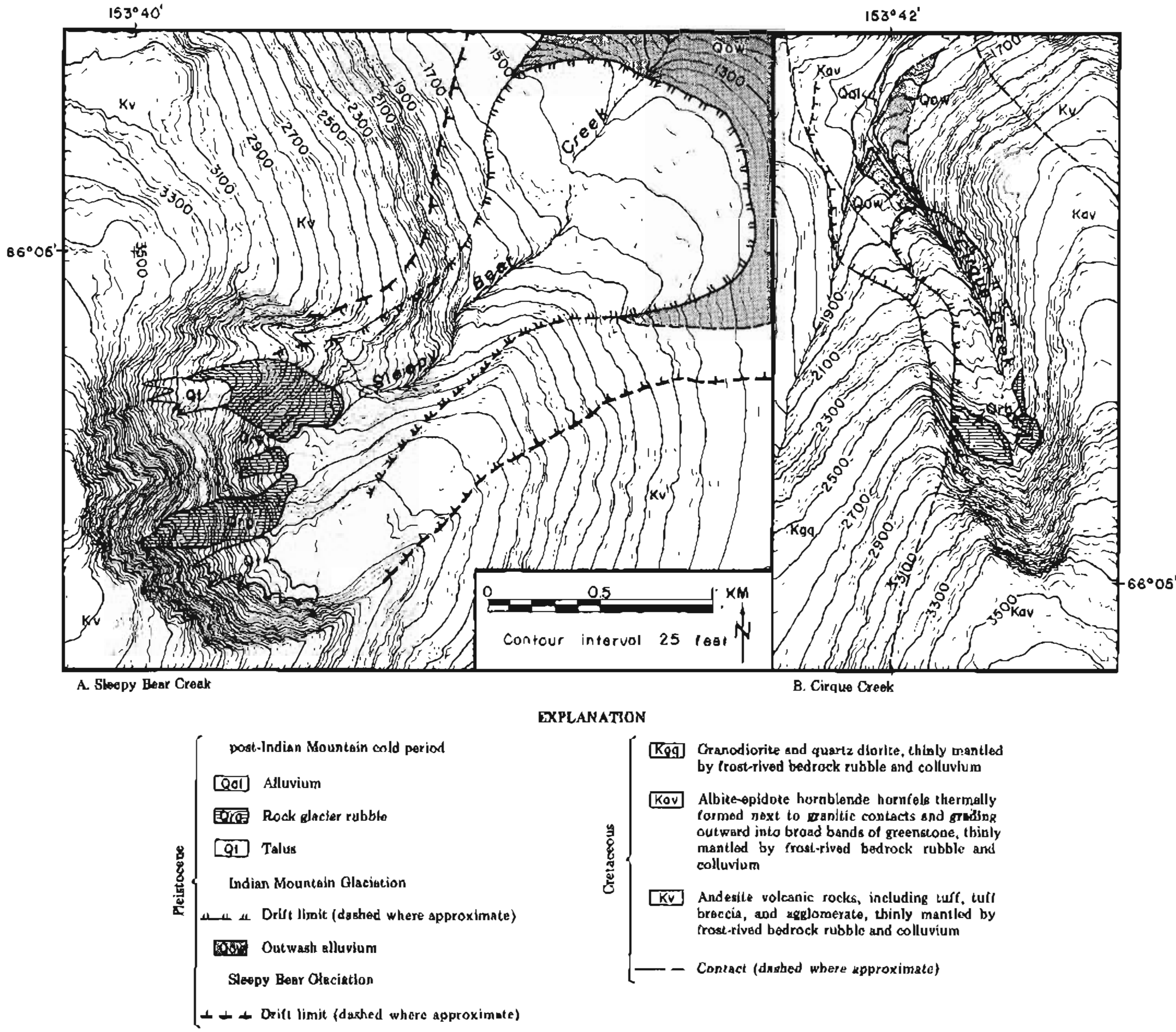

Figure 2. Geology of the headwaters of Sleepy Bear and Cirque Creeks. See figure 1 for location. Bedrock geology modified from Patton and Miller (1966), Miller and Ferrians (1968), and Miller (1971).

Indian Mountain glacier. The glacier floor of Sleepy Bear age was incised as deep as 100 to 150 m before the advance of Indian Mountain ice, which changed the V-shaped stream valley in the cirque floor to an inset glacial trough. Comparisons of the glacial chronology and soil development at Indian Mountain with those of the Ray Mountains 125 km to the east (Yeend, 1971) and the Alatna River valley about 50 km to the north (Hamilton, 1969) indicate that the Indian Mountain Glaciation is probably early Wisconsinan. However, until these moraines are more precisely dated this assignment must be considered tenuous.

Assuming that the accumulation area ratio of the former glacier in Cirque Creek at the culmination of the Indian Mountain Glaciation was 0.6 (Porter, 1975), the equilibrium-line altitude was about 756 m (2,495 feet) elevation.

POST-INDIAN MOUNTAIN COLD PERIOD

There is no evidence of glaciation on Indian Mountain after early Wisconsinan time, although substantial proof is found for a cold period after the Indian Mountain Glaciation when the climate was more severe than now. Several well-developed rock glaciers occur on the lower headwalls of the cirques in Sleepy Bear Creek and Cirque Creek valleys (figs. 2A and 2B). The following evidence indicates that all of these debris bodies are now inactive: 1) stabilized blocks of surface rubble are scattered and wedged apart 3 to 10 cm, yet remain essentially in place; 2) heavy coats of large lichens and moss cover most rock surfaces; 3) soil profiles 15 to 24 cm thick have developed on the surface of the rock glaciers; 4) the rounded profiles of the margins clearly contrast with the steep, unvegetated terminal flanks of active modern rock glaciers elsewhere in Alaska; and 5) very weak flows of clear water emerging from the toes of the features suggest that little interstitial ice remains.

These stable rock glaciers must have been active in a past climate colder than now and they must postdate early Wisconsinan time, when most of their positions were occupied by glacial ice. Several generations of lichens on rubble and significant soil development indicate that these rock glaciers have been stable for a considerable time. Comparisons with weathering characteristics and late Pleistocene chronologies in the Ray Mountains and Alatna River valley indicate that the rock glaciers were probably active in late Wisconsinan time.

ACKNOWLEDGMENTS

Field work was accomplished during my investigation of the cryoplanation terraces on the upper slopes of Indian Mountain under the supervision of Troy L. Péwé, Department of Geology, Arizona State University. Funding for that study was provided by United States Army Cold Regions Research and Engineering Laboratory. Colonel Richard R. Coursey, Jr., former Chief of Staff to Headquarters, Alaska Air Command, facilitated logistical arrangements at Utopia AFS. Albert E. Adams ably served as field assistant. Early drafts of this paper were reviewed by Troy L. Péwé, Don L. McGee (DGGS), and David M. Hopkins.

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THE CANTWELL ASH BED, A HOLOCENE TEPHRA IN THE CENTRAL ALASKA RANGE

By Peter M. Bowers[1,2]

INTRODUCTION

Recent geoarcheological research has led to the recognition, correlation, and dating of a Holocene volcanic ash layer in the central Alaska Range. The importance of tephrochronology in Quaternary studies such as archeology, geomorphology, palynology, and pedology lies primarily as a time-stratigraphic marker horizon through which intersite and intrasite stratigraphic comparisons can be made. This paper briefly describes one unreported tephra from the Cantwell vicinity, including its stratigraphic setting, distribution, chemical and petrographic characteristics, and possible correlation with other known Alaskan ash beds or volcanic vent sources.

STRATIGRAPHIC SETTING

The author observed a volcanic ash layer of apparent Holocene age in six exposures in the upper Nenana River valley in 1976 and 1977 (Bowers, 1978a,b). This tephra was found to extend from the Carlo Creek archeological site southward as far as mile 103.0 on the Denali Highway (figs. 1 and 2). The tephra was not observed farther north than the Carlo Creek locality, nor has it been reported by other workers in the Nenana River valley (Thorson and Hamilton, 1977; Wahrhaftig, 1958). The six localities reported here represent one area of the probably widespread geographic distribution of these pyroclastic deposits.

Tephra samples were collected from four locations: 1) Carlo Creek archeological site, mile 223.5 Parks Highway, 2) roadcut, mile 218.3 Parks Highway (type locality; table 2), 3) roadcut, mile 130.8 Denali Highway, and 4) 'the ash' archeological site, mile 103.0 Denali Highway (figs. 1 and 2). Where observed, the ash ranges in field-moist color from yellowish brown (10YR5/4) to very pale brown (10YR8/4). Ash depth ranges from 5 to 73 cm (table 1, fig. 2). Tephra thickness appears to increase to the east; this is probably a function of local deposition and preservation conditions. On the basis of field observations, these pyroclastic deposits were laid down during a variety of localized episodes of eolian sedimentation.

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DESCRIPTION

In an attempt to correlate the tephra deposits in the Cantwell area, four ash samples were examined under a petrographic microscope. Sample pretreatment (removal of simple iron compounds and organic matter) followed previously established procedures (Smith and others, 1968, 1975). On the basis of preliminary comparisons of the refractive index of glass shards, phenocryst suite, and glass-shard morphologies, this tephra set evidently represents one eruptive event of a single volcanic source.

Visual comparison of tephra grains indicates a textural range of about 0.25 to 0.1 mm (fine-sand class). Because of an apparent fining of tephra grains in an eastern direction, a southwestern or western source is suggested (R. Okazaki, pers. comm., 1977).

The refractive index of the samples ranges from 1.500 to 1.504. Phenocrysts include an abundance of plagioclase and hornblende; hypersthene, magnetite, and ilmenite are minor constituents. Trace amounts of augite and detrital fragments of mica schist are present (Okazaki, pers. comm., 1977).

The glass-shard morphologies indicate a dominance of medium vesicular forms, containing thin cell walls (terminology based on Smith and others, 1968). Tubular forms are also present, but are less common (Okazaki, pers. comm., 1977).

One sample of ash from the type locality was submitted for elemental analysis by electron microprobe. The technique used was identical with that reported by Smith and Westgate (1969), as modified by Smith and others (1975, 1977a, 1977b). Data were recorded for Al, Mg, Ca, Fe, and K by using a set-beam ARL-EMX microprobe at the Idaho Bureau of Mines and Geology. Analysis of the Cantwell tephra indicates the following proportions (in percent) of major elements: Al = 7.82, K = 1.85, Ca = 1.78, Fe = 1.48, and Mg = 0.24 (U. Moody, pers. comm.). Percentages of Ca, Fe, and K particularly have proven to be key elements for describing glass components of like tephra (Smith and Westgate, 1969; Smith and others, 1975, 1977a, 1977b).

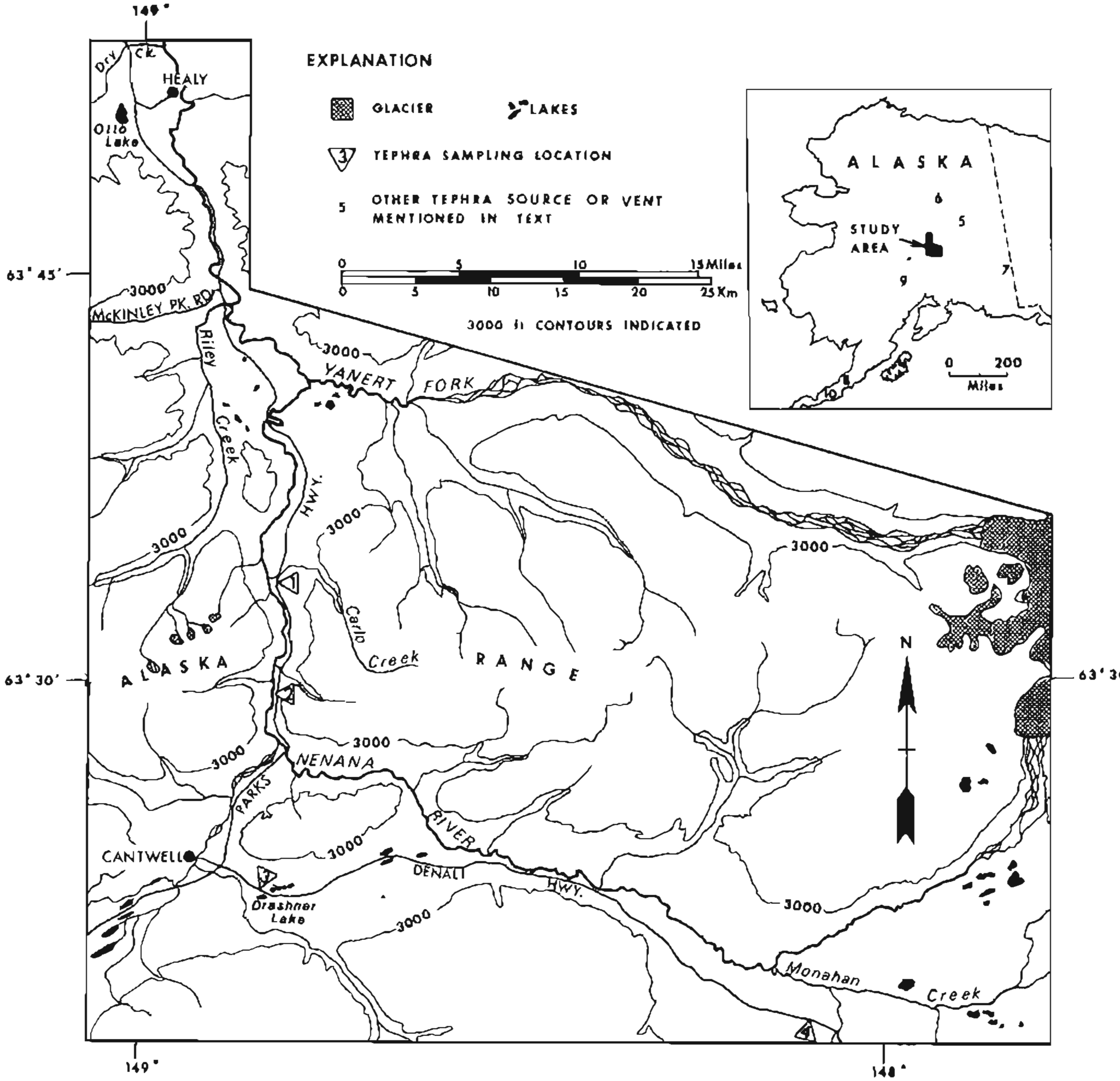

Figure 1. Location map showing tephra sampling localities and possible source vents mentioned in text. 1 - Carlo Creek Site, 2 - Cantwell Ash type locality, 3 - mile 130.8 Denali Highway, 4 - Ash site, mile 103.0 Denali Highway, 5 - Jarvis Creek type locality, 6 - Wilber Creek type locality, 7 - White River Ash source, 8 - Aniakchak caldera, 9 - Hayes volcanic vent, and 10 - Veniaminof Caldera.

AGE

A single radiocarbon date, obtained from sampling locality 2, provides a maximum age limit for this ash horizon of 3,780 ± 80 B.P. (WSU 1747). This date was determined from a 20-gm sample of wood collected from between 0.5 and 1.5 cm below the tephra layer. Within the 2-m-thick loess section at the type locality, the ash is 72 to 73 cm below ground surface and is fairly continuous.

Allowing for the date's standard deviation, maximum estimates of wood growth and an ensuing 0.5 cm of eolian deposition at locality 2, sample WSU 1747 conservatively dates the Cantwell area ashfall to within a 100- to 200-year span between about 3,600 and 3,800 years ago.

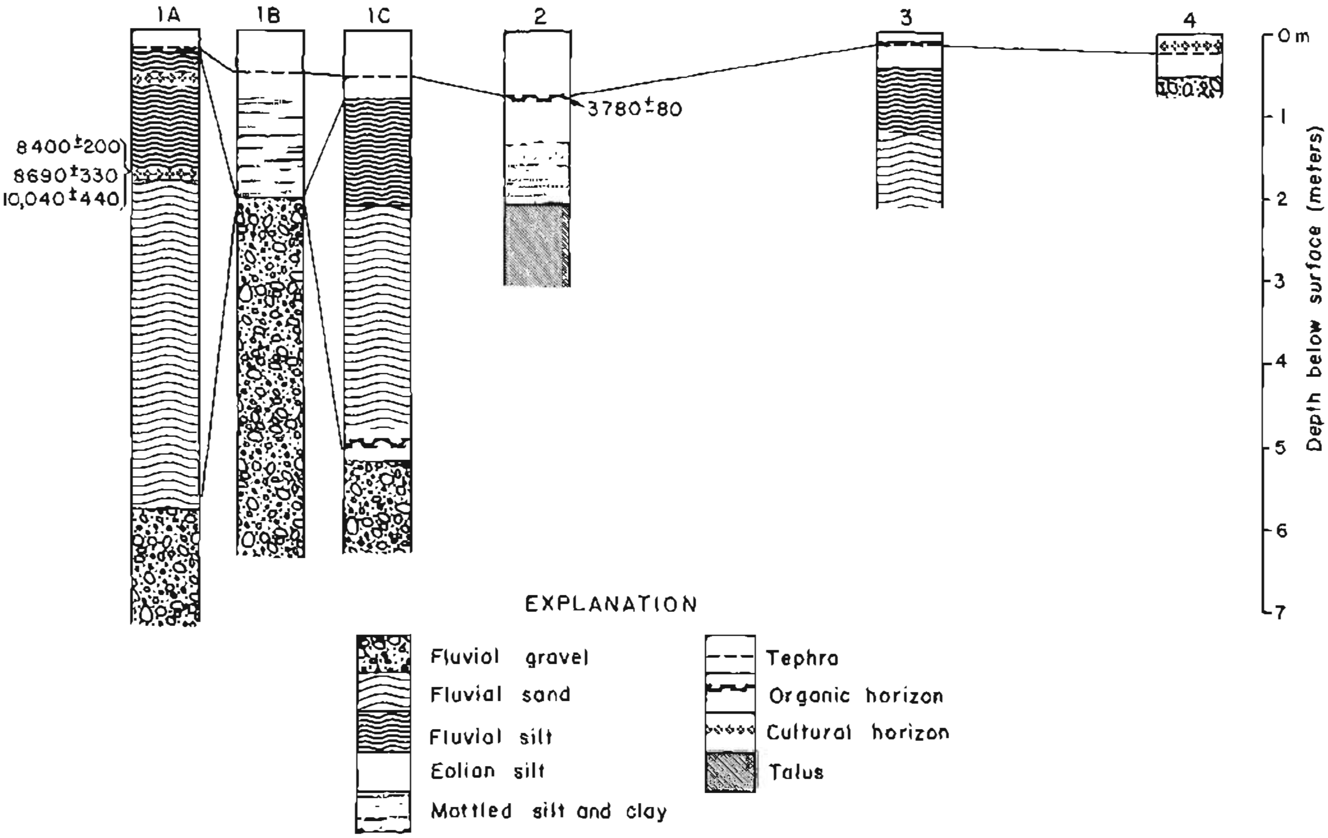

Figure 2. Correlation of six generalized stratigraphic sections, upper Nenana River valley. Refer to locations in fig. 1. Dates indicated in radiocarbon years B.P. Relative elevations above sea level not indicated.

POSSIBLE CORRELATIONS WITH OTHER KNOWN ALASKAN ASH BEDS

Although the samples collected from the Cantwell region appear to represent a single ashfall, it has not been possible to trace this tephra to its source or to definitely correlate it with other recognized ash horizons in interior Alaska.

JARVIS CREEK ASH BED

A temporal similarity is indicated with the Jarvis Creek Ash Bed (Reger and others, 1964; Péwé, 1975), which has been radiocarbon dated at 3,500 years B.P. (Reger, oral comm.). However, the refractive indexes do not compare favorably. The average refractive index of the Jarvis Creek Ash is 1.515, with a major mode of 1.506 to 1.520 and a minor mode of 1.555 (Reger, oral comm.). In addition to different refractive indexes, a difference was also noted in general shard morphologies and phenocrysts (Okazaki, pers. comm.).

The type locality for the Jarvis Ash Bed (fig. 1, loc. 5) is about 175 km northeast of Cantwell. There it occurs as a 6-mm-thick layer in Engineer Loess near the junction of the Delta and Tanana Rivers (Péwé, 1975). Péwé (1975, p. 19) suggests that the Wrangell Mountains may be a possible source for this tephra.

WILBER CREEK ASH BED

The poorly known Wilber Creek Ash Bed, formally named by Péwé (1975) for the type locality 10 km southwest of Livengood, 240 km north of Cantwell, may also represent a correlative tephra horizon with the volcanic ash from the Nenana River valley. At the type locality (loc. 6, fig. 1) the Wilber Ash is 5 to 25 mm thick (Péwé, 1975). Péwé (1975) suggests that the Jarvis Creek and Wilber Creek volcanic ashes may be correlative. However, little is presently known about the Wilber Creek Ash Bed except that it is of Holocene age and may be less than 4,200 years old (Péwé, 1975, p. 18).

WHITE RIVER ASH BED

A less likely correlation is the well-documented source of the White River Ash. The source of this widespread tephra is located on the Alaska-Yukon border, near Mt. Bona, 435 km southeast of Cantwell (loc. 7, fig. 1). White River Ash has been described (Lerbekmo and Campbell, 1969, p. 109) as a "...rhyodacite composed of glass (n = 1.502), andesine, hornblende, hypersthene, and magnetite. The average chemical composition is: SiO_2 = 67.4%, Al_2O_3 = 15.1%, TiO_2 = 0.5%, MgO = 2.0%, FeO = 2.0%, Fe_2O_3 = 2.2%, Na_2O = 4.1%, K_2O = 2.5%, CaO = 4.1%." The northern lobe of the bilobate White River Ash has been fairly precisely dated by more than a dozen radiocarbon dates at 1,890 years B.P. (Lerbekmo and Campbell, 1969, p. 110; Lerbekmo and others, 1975, p. 203; Denton and Karlèn, 1977, p. 72).

Although some similarities are noted in comparisons of refractive indexes and phenocrysts, correlation of the Cantwell and White River tephras can probably be ruled out because of differences in geographical distribution, age, and sorting of the former. The White River Ash has not been reported farther west than the Delta area, located about 175 km northeast of the upper Nenana River sampling localities. Even if a western lobe of pyroclastic ejecta from the White River source did reach as far west as the Nenana River valley, there still exists an apparent temporal difference of about 1,900 years between the two tephra units.

POSSIBLE SOUTHWESTERN ALASKA SOURCES

On the basis of radiocarbon dates, other possible sources for the Cantwell Ash are the Aniakchak and Veniaminof Calderas (loc. 8 and 10, fig. 1), located about 885 and 995 km southwest of Cantwell, respectively, and the Hayes volcanic vent (loc. 9, fig. 1), which is about 45 km northwest of Mt. Spurr and 275 km southwest of Cantwell (T. Miller, pers. comm). Aniakchak apparently underwent a major eruption about 3,500 years ago (Miller and Smith, 1977, p. 174; Miller, pers. comm.) whereas Veniaminof has been tentatively dated at 3,700 years ago (Miller and Smith, 1975, p. 1201). A dated organic layer beneath a tephra horizon near the vent indicates that the Hayes volcanic vent also erupted about 3,700 years B.P. (T. Miller, pers. comm.).

CONCLUSIONS

The Cantwell volcanic ash cannot be unequivocally correlated with the Wilber Creek, Jarvis Creek, or White River Ash Beds. Comparisons of phenocryst morphologies and refractive indexes suggest that Jarvis Creek and Cantwell Ashes are not the same. Correlation with the White River Ash source can probably be discounted because of the differences in apparent age and geographical distribution of the two ash horizons and the apparent decrease in grain size of Cantwell Ash in the direction of the White River source. On the basis of the sparse available data, the Cantwell Ash Bed is most likely the result of an eruption on the Alaska Peninsula or southwestern Alaska Range between about 3,600 and 3,800 years ago.

Table 1. *Summary of stratigraphic data for six Cantwell Ash localities (figs. 1 and 2).*

Locality	Site	Location[1]	Ash depth below surface (cm)	Enclosing sediment	Ash color (moist)	Ash thickness (cm)
1A	Carlo Creek archeological site (HEA 031), mile 223.5, Parks Highway	NW¼ NE¼, sec. 1, T. 16 S. R. 7 W.	19-20	Eolian silt loam	10YR5/4	0.5-1.0
1B	Terrace 6 m below Carlo Creek site		42-43	Eolian silt loam	10YR6/4	0.5-1.0
1C	Roadcut, mile 223.0, Parks Highway		44-45	Eolian silt loam	10YR6/4	1.0
2	Roadcut, mile 218.3, Parks Highway (Type locality)	SW¼ SE¼, sec. 25, T. 16 S. R. 7 W.	72-73	Eolian silt clay	10YR7/4	1.0
3	Roadcut, mile 130.8, Denali Highway	SE¼ SW¼, sec. 2, T. 18 S. R. 7 W.	5-10	Eolian silt	10YR7/4	2.5-5.0
4	Roadcut, 'the ash' archeological site (HEA 100), mile 103.0, Denali Highway	SE¼ NW¼, sec. 3, T. 19 S. R. 3 W.	20-27	Eolian silt	10YR8/4	3.0-7.0

[1]Source: USGS Healy Quadrangle 1:250,000 series.

ACKNOWLEDGMENTS

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Table 2. *Pedologic description of Cantwell Ash type locality, mile 218.3 Parks Highway*[1]

Soil horizon	Depth (cm)	Description
H+Ap	25-0	Modern organic mat, decomposed vegetation; minor highway disturbance and overburden.
Ah	0-5	Dark-reddish-brown (5YR2.5/2, moist) silt loam; massive; very friable, slightly sticky, slightly plastic; many fine roots; abrupt, wavy boundary.
Bh	5-15	Dark-brown (10YR3/3, moist) silt loam; massive; friable, slightly sticky, slightly plastic; many fine roots; abrupt, wavy boundary.
2Bhir	15-16	Dark-yellowish-brown (10YR3/6, moist) silt loam; massive; friable, slightly sticky, slightly plastic; many fine roots; abrupt, wavy boundary.
2Bir	16-32	Dark-brown (10YR3/3, moist) silt loam; massive; friable, slightly sticky, slightly plastic; few very fine interstitial pores; few small roots; abrupt, wavy boundary.
2Bg	32-72	Weak, red (2.5YR4/2, moist) silt clay; massive; firm, sticky, plastic. Mottles: distinct, medium common, dark-yellowish brown (10YR4/6, moist), and distinct, medium common, weak red (10YR4/2, moist). A few discontinuous organic lenses (wood?) present just above lower contact; abrupt, wavy boundary.
3C	72-73	Very pale-brown (10YR7/4, moist) volcanic ash; structureless; abrupt, wavy boundary (tephra sample PB-76-1).
4Bhb	73-78	Dark-yellowish-brown (10YR3/4, moist) silt loam; massive; slightly sticky, slightly plastic; few roots. Wood fragments and peat(?) are present from 0.5 to 1.5 cm below upper contact (radiocarbon sample: WSU 1747). Clear, distinct boundary.
4Cg	78-138	Dusky-red (2.5YR3/2, moist) clay loam; massive; firm, sticky, very plastic; few roots. Unit contains a number of diffuse buried soil horizons that are very dark brown (10YR2/2, moist). Frequency of angular rock fragments (up to 25 by 26 cm) increases with depth; abrupt, wavy boundary.
4Cgr	138-200+	Dark-reddish-gray (5YR4/2, moist) silty clay; massive; firm, very sticky, very plastic. Strong gleying and mottling. Contains many angular rock fragments.
R	200+	Angular decomposed bedrock and talus.

[1] Convention: 1975 USDA.

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GEOCHRONOLOGY OF METAMORPHIC AND IGNEOUS ROCKS IN THE KANTISHNA HILLS, MOUNT MC KINLEY QUADRANGLE, ALASKA

By T.K. Bundtzen[1] and D.L. Turner[2]

INTRODUCTION

This paper summarizes the results of a radiometric dating reconnaissance of the Kantishna Hills, a low, rugged group of hills on the north flank of the Alaska Range about 150 km southwest of Fairbanks (fig. 1).

Thirteen total-fusion ^{40}K-^{40}Ar mineral and whole-rock ages were determined for seven metamorphic and four igneous rocks from the study area (fig. 2, table 1). Analytical work was done at the Geochronology Laboratory of the Geophysical Institute, University of Alaska, Fairbanks. Analytical techniques have been described previously (Turner and others, 1973).

SUMMARY OF GEOLOGY

The geology of the area has been studied by Wells (1933), White (1942), Morrison (1964), Reed (1961), Bundtzen and others (1976), and Gilbert and Bundtzen (1978). The oldest layered units consist of four mappable sequences of regionally metamorphosed rocks, ranging in age from late Precambrian(?) to Mississippian.

The oldest rocks (pCs on fig. 2) are an undifferentiated sequence of massive quartzite, chlorite-quartz-muscovite-biotite-garnet-schist, feldspathic ortho(?)schist, graphite-rich chlorite schist, and amphibole-rich metabasite. The protoliths of these rocks were metamorphosed to lower amphibolite facies and subsequently retrograded to a lower greenschist facies (Morrison, 1964); they commonly exhibit 1) relict oligoclase, biotite, and muscovite replaced by albite and chlorite, 2) garnet extensively replaced by magnetite and chlorite, and 3) hornblende replaced by biotite. Kyanite(?) is found in the feldspathic ortho(?) schist. On the basis of apparent stratigraphic relations and comparative grade of metamorphism with younger(?) units, the pCs unit is believed to be either of late Precambrian or earliest Paleozoic age. Folded and discontinuous metabasite lenses and sills are included in the same pCs unit because they have undergone the same polymetamorphic history. However, they often show discordant structural relationships with enclosing pCs lithologies. Contacts of pCs with other metamorphic rocks (fig. 2) are probably, in part, tectonic.

Porphyroblastic chlorite-rich phyllite, graphitic schist, marble, and minor metafelsite (Ps in fig. 1) and dark-gray slate, phyllite, limestone, and stretched pebble conglomerate (Ppcl on fig. 2) have lower greenschist-facies metamorphic mineral assemblages. Pks and Pkpcl are correlated with the Keevy Peak Formation, defined by Wahrhaftig (1968). On the basis of fossil and stratigraphic evidence, Gilbert and Bundtzen (1978) assigned the Keevy Peak Formation an Ordovician to Middle Devonian age.

Metabasite, metafelsite, and metasedimentary rocks (tsf, tsb, tss) of the Totatlanika Schist occur in fault contact with the Keevy Peak Formation. Fossils collected from the schist by Wahrhaftig (1968) and Gilbert and Redman (1977) are of Upper Devonian and Lower Mississippian age. The Totatlanika Schist is predominantly a metamorphosed bimodal volcanic sequence with mineral assemblages that indicate up to a lowest greenschist-facies metamorphic grade locally.

Intruding the metamorphic units is the 'Stony Creek' granodiorite (gd), which shows petrologic similarities to the Mt. Eielson pluton in nearby Mt. McKinley Park. The Mt. Eielson pluton is Eocene in age (Decker and Gilbert, 1978). A distinctive bimodal dike system of quartz-K-spar porphyries (Tf) and augite-olivine gabbros (Tb) and associated small plugs intrudes metamorphic rocks roughly parallel to northeast-trending fold structures. Poorly consolidated Miocene sand and gravel and Quaternary glaciofluvial deposits overlie all older lithologies.

The metamorphic basement underwent an early deformation that produced northwest-trending isoclinal folds; a later deformation event resulted in the development of broad northeast-trending folds, thrust faults, and high-angle longitudinal faults.

The Kantishna mining district, centered in the Quigley Ridge and Stampede areas (fig. 1), is known for polysulfide vein mineralization, metalliferous skarns, and placer gold. The polysulfide veins in the former area have been classified as 1) quartz-arsenopyrite-(gold)-(scheelite)-pyrite veins, 2) galena-sphalerite-pyrite-siderite-tetrahedrite veins, and 3) massive stibnite-pyrite-quartz veins (Wells, 1933). The veins were mined sporadically for lead, zinc, and silver in the 1920s, for gold in the 1930s and 1940s, and for antimony since World War I. Most of the Kantishna vein deposits are

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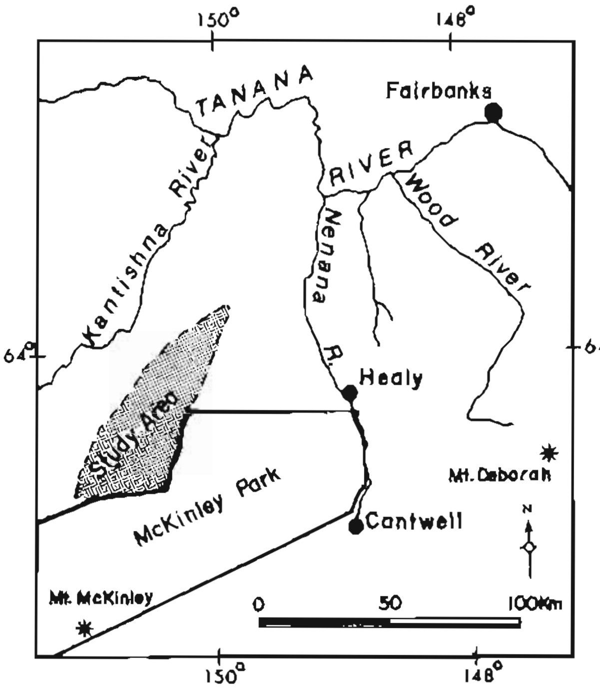

Figure 1. Location of study area.

fracture controlled and parallel the northeast-trending assymetrical 'Kantishna Anticline' (fig. 2). Previous workers such as Brooks (1911) and Wells (1933) recognized a possible genetic relationship between the mineralized veins and the quartz porphyry-gabbro dikes and plugs (Tf, Tb). Stratiform sulfide mineralization is hosted in metamorphic rocks of the Totatlanika schist in the northern Kantishna Hills and within units of the Keevy Peak Formation (Gilbert and Bundtzen, 1978; Hawley, 1978).

RADIOMETRIC DATING

Samples of regionally metamorphosed rocks of the pCs unit, Totatlanika Schist metarhyolite (tsf), and quartz porphyry-gabbro dikes (Tb, Tf), were dated radiometrically to investigate the time of regional metamorphism, igneous activity, and possible age of sulfide vein mineralization in the study area.

Four mica separates from two samples of biotite-garnet-muscovite-quartz-feldspar schist in the pCs unit (Nos. 2, 3 on table 1; fig. 2) yield ^{40}K-^{40}Ar ages ranging from 100.2 to 85.7 m.y. Muscovite ages were concordant and significantly older ($\bar{x}$ = 100.0 m.y.) than the biotite ages ($\bar{x}$ = 89.7 m.y.). Four hornblende separates from garnet amphibolites within the same rock sequence (Nos. 1, 4, 5, 6 on table 1; fig. 2) yield ages ranging from 195 to 104 m.y.

The observed order of apparent argon retention in the dated minerals (hornblende $\geq$ muscovite $\geq$ biotite) is identical with that generally observed for these minerals (Dalrymple and Lanphere, 1969). These data probably represent varying amounts of radiogenic argon loss from older metamorphic events during a Middle or Late Cretaceous thermal event. This is consistent with petrographic evidence, which shows a retrograde greenschist-facies event overprinting an earlier lower amphibolite-facies metamorphism (Morrison, 1964). The oldest hornblende apparent age of 195.4 m.y., probably represents a minimum age for the earlier metamorphism, whereas the mid-Cretaceous micas could represent either a total resetting of the K-Ar clock during the retrograde lower greenschist-facies event or cooling ages caused by uplift and removal of overburden.

Mica and amphibole ages in the polymetamorphic basement of the Big Delta Quadrangle of the Yukon-Tanana Upland to the northeast have apparent age patterns similar to those reported here, which suggests that both areas underwent similar thermal histories (Foster and others, 1979). Plutonic rocks intruding polymetamorphic terrane in the Big Delta Quadrangle yield hornblende and biotite ages of 105 to 88 m.y. (Foster and others, 1979). Concordant mineral-age pairs from these intrusions and from nearly coeval volcanics indicate that a major thermal event affected the Yukon-Tanana Upland during mid-Cretaceous time.

One whole-rock analysis from sparsely porphyritic metarhyodacite of the Totatlanika Schist (No. 7 on table 1; fig. 2) yields an age of 108.0 $\pm$ 3.2 m.y. Because fossils within the Totatlanika Schist to the east are Devonian-Mississippian (Gilbert and Redman, 1977; Wahrhaftig, 1968) the age probably reflects the mid-Cretaceous thermal event described above and does not indicate the magmatic crystallization age of the volcanic rock.

Four samples (No. 8-11 on table 1; fig. 2) of dikes and small plugs (Tf, Tb) that intrude the metamorphic rocks yield ages of 81.3 to 46.7 m.y. Three of the ages—two from feldspar separates and one whole rock—average 49.8 m.y., but an amphibole separate from a fourth yields an age of 81.3 m.y.; all four are considered minimum ages because of alteration. The 49.8-m.y. average age of three samples suggests that they may be related to late-stage Teklanika Formation volcanism that took place in nearby McKinley Park (Gilbert and others, 1976). Sample 11 is from a sericitized porphyritic K-spar-quartz porphyry plug at the Bunnell zinc-lead-silver-antimony deposit (Morrison, 1964). Here, sulfide vein mineralization is found both localized within fracture systems in the quartz porphyry plug and disseminated throughout the intrusion. The age for sample 11 (48.3 $\pm$ 1.4 m.y.) should be regarded as a minimum date for sulfide mineralization there because the rhyolite-gabbro dikes in the area are believed to have been intruded along longitudinal fractures adjacent to the Kantishna Anticline about the same time sulfide-vein mineralization occurred (Bundtzen and others, 1976).

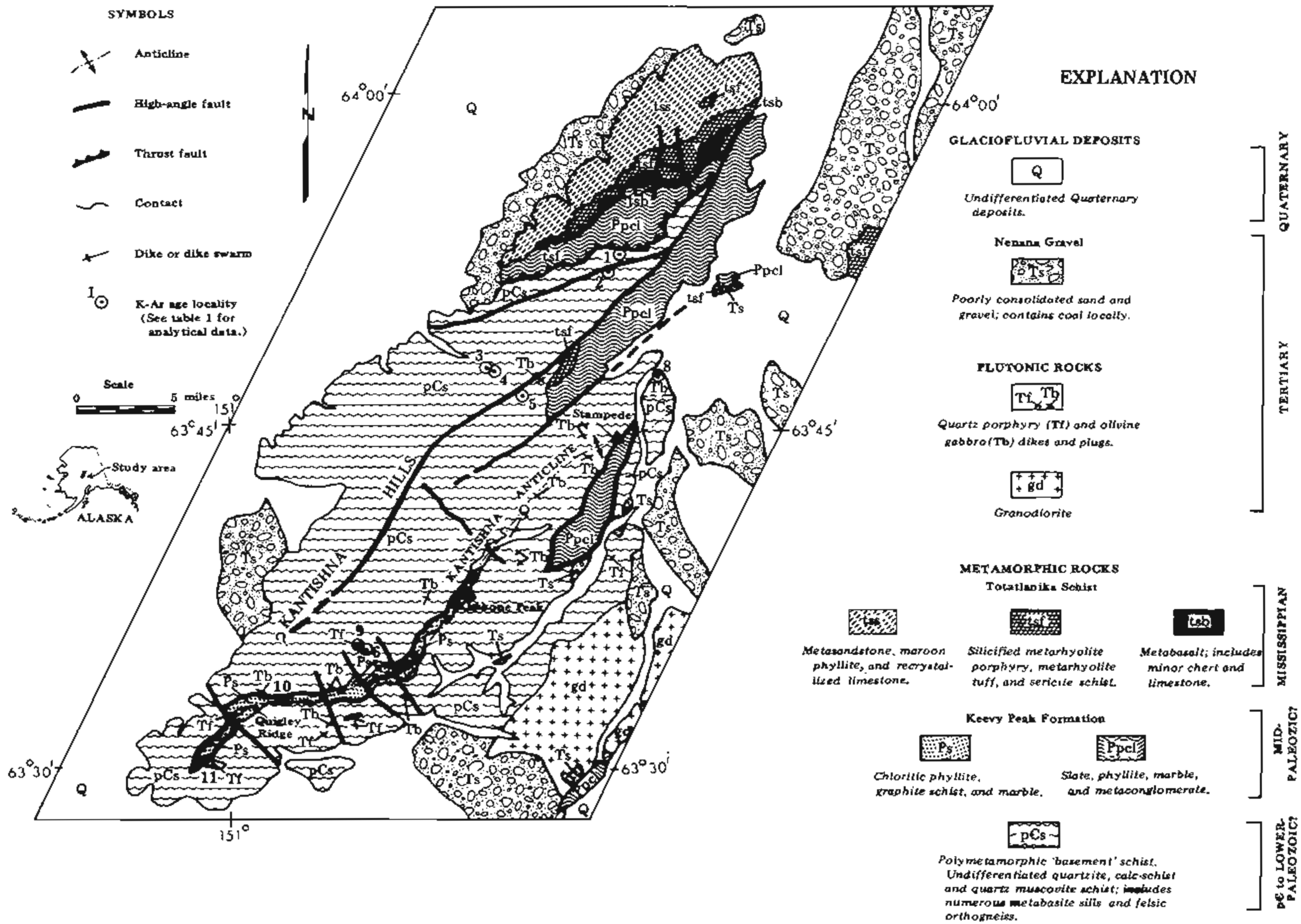

Figure 2. Generalized geologic map of the Kantishna Hills, from Bundtzen and others (1976), Gilbert and Bundtzen (1978), and additional field data collected by Bundtzen in 1976 and 1978.

Table 1. *Analytical data for ^{40}K-^{40}Ar age determinations*[1]

Map No.	Sample (Lab. No.)	Rock type	Mineral dated	K_2O (Wt. %)	Sample weight (g)	$^{40}Ar_{rad}$ (moles/g) $\times 10^{-11}$	$\frac{^{40}Ar_{rad}}{^{40}K}$ $\times 10^{-3}$	$\frac{^{40}Ar_{rad}}{^{40}Ar_{total}}$	Age $\pm 1\sigma$ (m.y.)
			Polymetamorphic 'basement' terrane (pCs on fig. 2)						
1	DT75-25 (76132)	Amphibolite dike	Hornblende	0.320 0.320 0.320 0.320 $\bar{x}$ = 0.320	1.8713	6.008	7.43	0.611	122.9 ± 3.7
2	DT75-28 (76196)	Biotite, muscovite, quartzo-feldspathic schist	Biotite	8.430 8.443 $\bar{x}$ = 8.437	0.3026	109.33	5.130	0.663	85.7 ± 2.6
2	DT75-28 (76125)	Biotite, muscovite, quartzo-feldspathic schist	Muscovite	9.718 9.718 $\bar{x}$ = 9.718	0.2742	150.511	6.13	0.812	101.9 ± 3.1
3	DT75-31 (76134)	Biotite, muscovite, quartzo-feldspathic schist	Biotite	8.830 8.830 $\bar{x}$ = 8.830	0.1541	125.104	5.61	0.877	93.5 ± 2.8
3	DT75-31 (76133)	Biotite, muscovite, quartzo-feldspathic schist	Muscovite	10.280 10.240 $\bar{x}$ = 10.260	0.442	156.049	6.02	0.839	100.2 ± 3.0
4	DT75-30 (76135)	Garnet amphibolite	Hornblende	0.330 0.330 0.330 0.330 $\bar{x}$ = 0.330	1.1340	10.040	12.050	0.372	195.4 ± 5.9
5	DT75-29 (76129)	Amphibolite	Hornblende	0.387 0.380 0.390 0.390 $\bar{x}$ = 0.387	1.6278	8.532	8.73	0.805	143.7 ± 4.3

6	DT75-34 (76109)	Garnet amphibolite	Hornblende	0.450 0.450 0.450 0.450 $\bar{x}$ = 0.450	0.9016	7.124	6.27	0.734	104.2 ± 3.1
				Totatlanika Schist (tsf on fig. 2)					
7	DT75-26 (76094)	Metarhyo-lite	Whole rock	4.897 4.870 4.880 4.880 $\bar{x}$ = 4.882	0.6382	80.188	6.50	0.939	108.0 ± 3.2
				Bimodal dike swarm (Tf, Tb on fig. 2)					
8	DT75-32 (26130)	Olivine gabbro plug	Plagioclase	0.670 0.660 0.660 0.660 $\bar{x}$ = 0.663	2.7715	5.415	3.24	0.476	54.5 ± 1.6 minimum age
9	DT75-33 (76107)	Altered hornblende-rhyodacite dike	Hornblende	1.380 1.380 1.380 $\bar{x}$ = 1.380	0.8658	16.952	4.86	0.754	81.3 ± 2.4 minimum age
10	76BT270 (7020)	Olivine gabbro dike	Whole rock	0.640 0.640 0.630 $\bar{x}$ = 0.637	0.2783	4.447	2.765	0.145	46.7 ± 1.4 minimum age
11	75AST-1987 (76200)	Sericitized porphyritic quartz porphyry plug	K-feldspar	3.249 3.235 3.260 3.247 $\bar{x}$ = 3.248	1.4479	23.482	2.863	0.838	48.3 ± 1.4 minimum age

[1]Constants used in age calculations: $\lambda_{\epsilon} = 0.585 \times 10^{-10}\ yr^{-1}$; $\lambda_{\beta} = 4.72 \times 10^{-10}\ yr^{-1}$; $^{40}K/K$ total $= 1.19 \times 10^{-4}$ mol/mol.

CONCLUSIONS

Available radiometric evidence indicates that the early Paleozoic(?) 'basement' schist pCs complex in the Kantishna Hills was affected by a pre-Jurassic regional metamorphism that was later overprinted by a mid-Cretaceous thermal event that may represent the lower greenschist-facies regional metamorphism petrologically recognized in the rock units. The Totatlanika Schist was also affected by the same mid-Cretaceous event. Complex sulfide vein mineralization in the Quigley Hill area was emplaced synchronously with a bimodal dike system that may represent late-stage Teklanika Formation volcanic activity.

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THE CHILIKADROTNA GREENSTONE, AN UPPER SILURIAN METAVOLCANIC SEQUENCE IN THE CENTRAL LAKE CLARK QUADRANGLE, ALASKA

By T.K. Bundtzen[1], W.G. Gilbert[1], and R.B. Blodgett[2]

INTRODUCTION

This paper describes a sequence of weakly metamorphosed basalt, andesite, chert, limestone, and tuffaceous sedimentary rocks of Upper Silurian age, hereby named the 'Chilikadrotna Greenstone,' recently mapped by Eakins and others (1978) in the central Lake Clark Quadrangle of southern Alaska (fig. 1). The Chilikadrotna Greenstone trends northeast and dips steeply northwest to vertical. Contact relationships with younger rock units are obscured by extensive glacial drift and vegetation (fig. 2).

The Chilikadrotna Greenstone is in probable fault contact on all sides with a northeast-trending, northwest-dipping sequence of interbedded marine volcaniclastic sandstone, andesitic tuff, siltstone, and shale of Upper Jurassic age. Quartz monzonite, granodiorite, diorite, dacite porphyry, basalt, and rhyolite that both overlie and intrude the older bedded sequences yield amphibole and mica ages ranging from 71-56 m.y. (Eakins and others, 1978). A prominent conjugate system of northwest- and northeast-trending high-angle faults cuts all bedded lithologies; one fault near Ptarmigan Creek (fig. 2) shows evidence of displacement during Holocene time.

CHILIKADROTNA GREENSTONE

About 60 percent of the known exposure of the Chilikadrotna Greenstone forms massive, resistant ridges of olive-gray to dark-green metabasalt that are commonly broken into angular blocks by northwest-trending joints. The metabasalt (Sg, fig. 2) is characteristically very fine grained to aphanitic, equigranular to porphyritic, and has undergone low-grade metamorphism. Flows display spheroidal weathering suggestive of pillow structure. Amygdaloidal zones are less resistant and form rubble in depressions and saddles. Epidote and quartz vein systems cut the flows and locally make up to 40 percent of the rock.

Hematite-rich metabasalt and minor meta-andesite (Spb, fig. 2) form brown to maroon eroded outcrops in depressions or saddles. Textures are usually subporphyritic and amygdaloidal zones are common. The maroon color is due to fine-grained hematite or limonite evenly distributed throughout a very fine grained groundmass. These rocks appear more altered than normal gray-green metavolcanic units, and epidote-chlorite alteration commonly gives hand specimens a distinctive greenish-maroon splotchy appearance.

Thin sections from the southeastern portion of the Chilikadrotna Greenstone show clinopyroxene(?) largely altered to chlorite and opaques, plagioclase invaribly albitized or altered to calcite, pumpellyite, and prehnite(?), and minor hornblende rimmed with chlorite. The groundmass is largely secondary chlorite, clinozoisite, and magnetite. Epidote-group minerals locally constitute as much as 25 percent of the rock. Some specimens contain stretched amygdules of calcite in a groundmass of nondirectional dusty plagioclase laths and opaques. Veinlets of pumpellyite and calcite cut the most altered samples.

Associated with the hematite-rich volcanic rocks (Spb) are minor intervals of banded hematitic radiolarian chert (not shown in fig. 2) that form nonresistant rubble. The chert consists of dark-gray silica laminated with 1-cm-thick reddish hematite stringers that have a botryoidal texture. The chert has undergone enough thermal alteration to recrystallize the quartz into radiating spheres within a granoblastic matrix of magnetite, limonite, and quartz. Faint outlines of rounded organic features seen in plain light may be recrystallized radiolaria.

Light-gray fossiliferous limestone lenses (Sl, fig. 2) up to 25 m thick discontinuous crop out along strike for about 1 km in several areas. Partial to complete recrystallization of calcite and minor epidote minerals give the rock a coarse-grained texture. The limestone is commonly folded and appears to be structurally discordant within the enclosing metabasalt.

The northern and western exposures of metabasite (Sg?, fig. 2) are less altered than the Sg and Spb units immediately to the south and can only be tentatively correlated with the Chilikadrotna Greenstone. This metabasite consists of dark-greenish-gray mafic and intermediate amygdaloidal flows, tuff, breccia, and intrusive(?) greenstone. Calcite and epidote stringers cutting amygdaloidal units cause a blocky, crackled, weathered appearance in outcrop. Some of the greenstone has a high specific gravity and is composed of up to 35 percent megascopic magnetite and clinopyroxene.

In thin section Sg(?) units are composed of albitized

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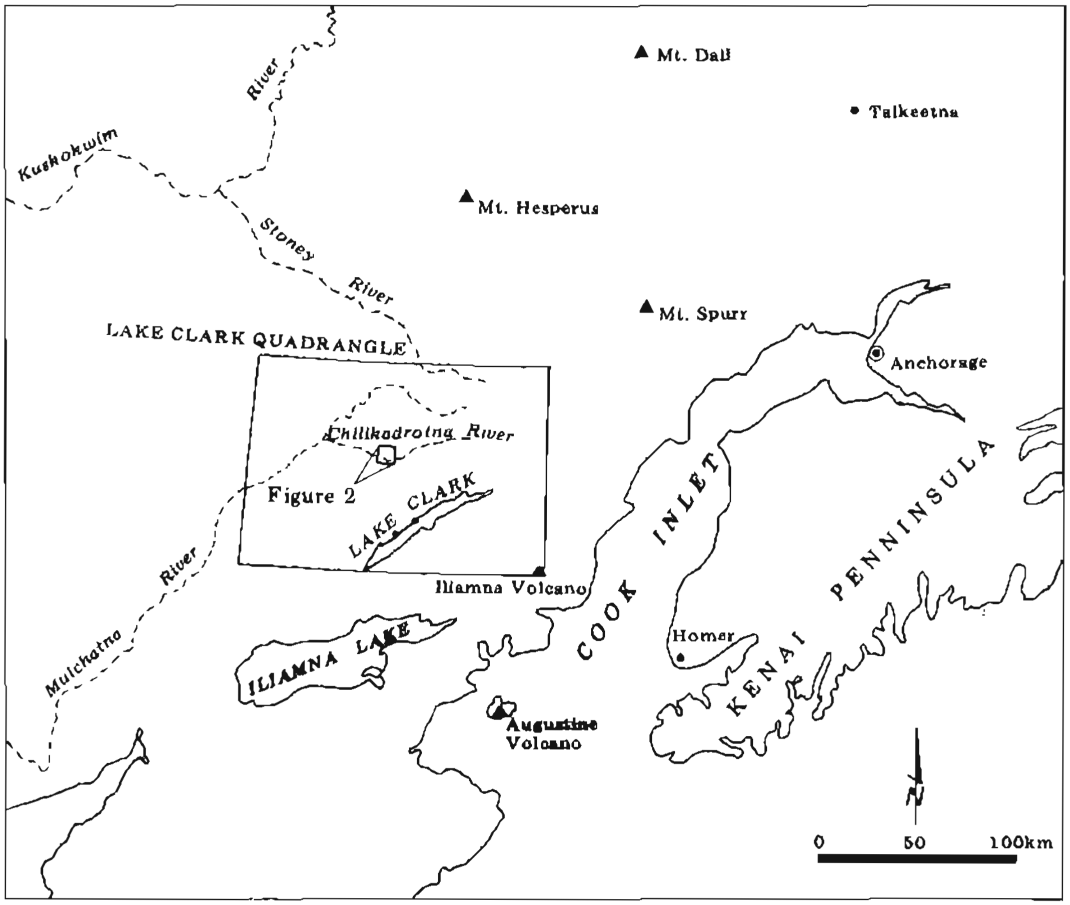

Figure 1. Location of study area, central Lake Clark Quadrangle, Alaska.

plagioclase, augite, hornblende, chlorite, traces of olivine, and secondary zeolite-epidote masses. Augite occurs as euhedral phenocrysts up to 5 mm long and as anhedral glomeroporphyritic aggregates with plagioclase and magnetite. Rims of augite are commonly replaced by penninite and magnetite. Dusty plagioclase laths, rarely fresh enough to determine original composition (An 35-50), are usually completely albitized. Uncommon anhedral olivine crystals have been largely altered to antigorite(?). The groundmass of most of the metabasalt consists of abundant magnetite, epidote-group minerals, and chlorite masses cored with zeolite(?).

Red shale, medium- to dark-gray slate, and volcaniclastic sandstone (Ssh, fig. 2) are interbedded with the greenstone units. They include medium-grained sandstone that is composed of angular grains of dusty plagioclase (An 20-35), rounded clasts of quartz (< 0.5 mm) that show embayed contacts, and andesite lapilli in a groundmass of limonite, undetermined anhedral to rounded feldspar laths, and quartz. The volcaniclastic sandstones are almost identical with nearby Jurassic rocks but are tentatively mapped as part of the Chilikadrotna Greenstone because of their apparent interbedded relationship with the greenstone, limestone, and chert.

Whole-rock chemical analyses were performed on five basalts and two andesites (table 1). Most samples have undergone low-grade metamorphism and lack original mineralogy, but their Na_2O, K_2O, and TiO_2 contents are similar to those of spilitic tholeiites found in mid-ocean ridge environments (Irvine and Baragar, 1971; Hyndman, 1972, p. 99-100; Sun and Nesbitt, 1978). The association of mafic volcanic rocks and fine-grained clastic units and chert would also suggest an ocean-floor assemblage.

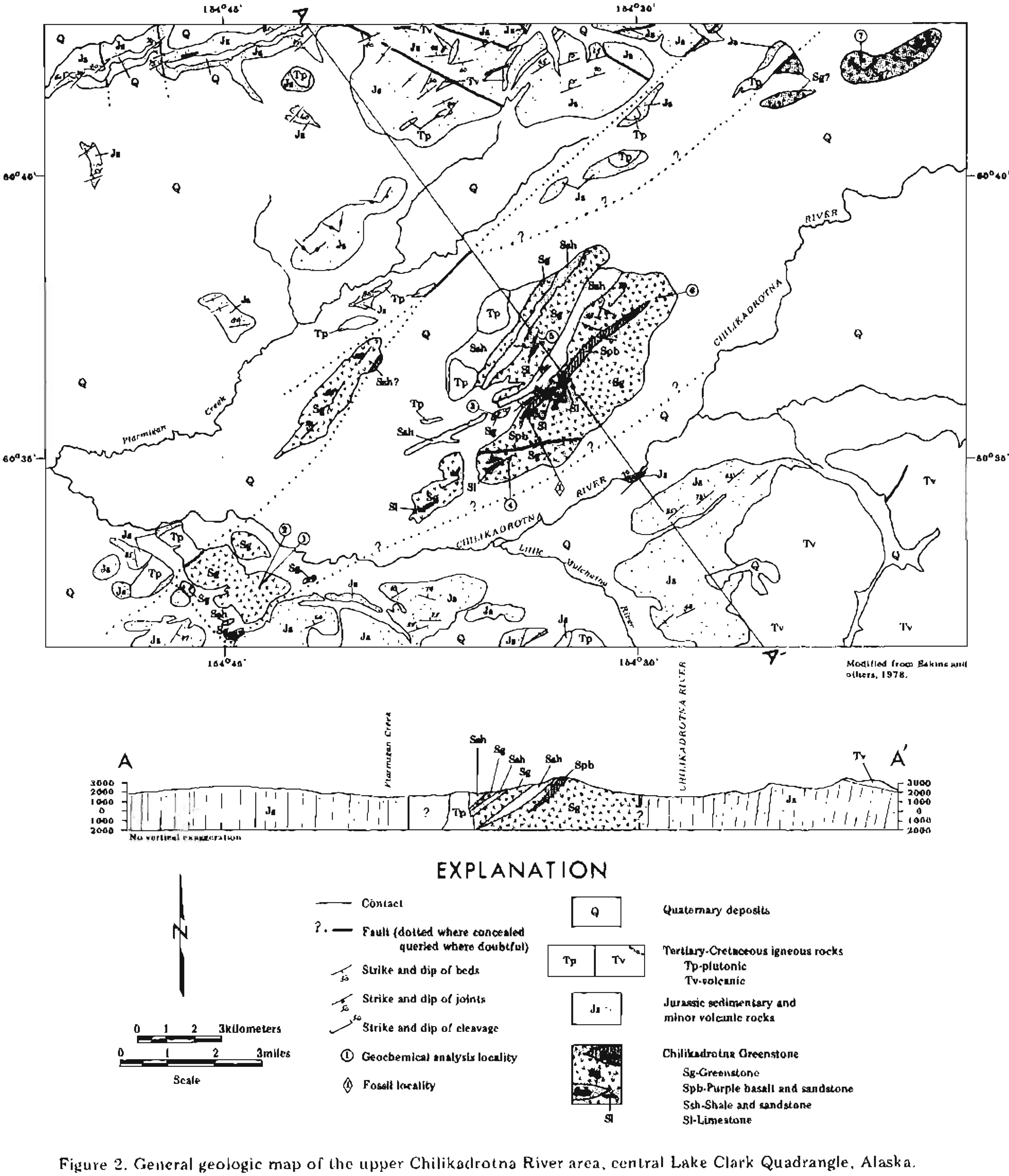

Figure 2. General geologic map of the upper Chilikadrotna River area, central Lake Clark Quadrangle, Alaska.

AGE

Fossils were collected from a lens of light-gray, partially recrystallized limestone (fig. 1). The lens is about 10 m thick and can be traced for about 200 m along strike. The limestone displays vague sedimentary laminations, is locally folded, contains greenstone detritus, and appears to be interbedded with the surrounding greenstone. The following fossils were identified by Blodgett:

-Stromatoporoids
-Undetermined solitary rugose coral
-*Pentamerus* or *Pentameroides* sp.
-*Palaeopecten* sp.
-Undetermined rynchonelloid brachiopods.

The brachiopods *Pentamerus* or *Pentameroides* sp. were identified on the basis of external form; these two closely related genera are distinguished on the basis of internal morphology. The stratigraphic range of these genera is given by Amsden (Williams and others, 1965) as between upper Llandoverian (upper Lower Silurian) and Wenlockian (lower Upper Silurian). Both genera have been reported in similar age strata of the Tolovana Limestone of the White Mountains, central Alaska (Dutro, *in* Oliver and others, 1975, p. 26). The range of the bivalve *Palaeopecten* is Upper Silurian-Lower Devonian (Newell, *in* Cox and others, 1969, p. N334). The composite range of the two identified taxa suggests a Wenlockian (early Late Silurian) age for the locality.

DISCUSSION

The Chilikadrotna Greenstone represents the first discovery of mid-Paleozoic rocks in an area thought to be underlain by rocks of Mesozoic and Tertiary age (Beikman, 1978). Capps (1935) originally classified extensive greenstone terranes in the southern Alaska Range as Triassic, and the Sg(?) units described in this report may be the same age. The nearest mid-Paleozoic rocks that have been dated are the Silurian-Devonian limestones of the Holitna Group, 80 km to the northeast (Cady and others, 1955). Hoare and Coonrad (1978) have mapped mid-Paleozoic tuffaceous limestone with associated mafic volcanic rocks that are overlain by Jurassic(?) ophiolite near Cape Newanham, 450 km to the west. R.L. Detterman (USGS, pers. comm.) reports Upper Silurian-Lower Devonian limestone near the north shore of Becharof Lake, 390 km to the south. The Chilikadrotna Greenstone may be a low-grade part of the undated pre-Jurassic pelitic and mafic metavolcanic schist and gneiss terrane that crops out in the core of the southern Alaska Range (Capps, 1935; Eakins and others, 1978); however, such a correlation would require both a rapid increase in metamorphic grade to the southeast and the presence of as-yet-undocumented Upper Silurian units in the metamorphic complex. Although further studies on the tectonic setting of Paleozoic rocks of southwest Alaska are needed, the Chilikadrotna Greenstone may mark the remnant of a Paleozoic ocean floor.

Table 1. Geochemical analyses (wt %) of seven metavolcanic rocks from Chilikadrotna Greenstone. Rapid rock technique by Skyline Labs., Wheatridge, CO.

	Map No. (single digit) and Field No.						
	1	2	3	4	5	6	7
	77WG162	78WG5a	78BT9	78BT8	78BT1	77BT141	77BT200
SiO_2	50.9	54.3	62.9	47.2	48.4	49.0	49.2
Al_2O_3	12.8	13.8	16.8	16.6	14.9	16.4	14.5
FeO	9.6	6.9	2.9	6.7	9.8	6.0	4.9
Fe_2O_3	3.2	5.1	1.1	3.4	2.1	3.7	5.2
MgO	3.9	3.7	3.0	8.3	5.6	7.1	7.3
CaO	8.7	6.6	5.0	9.8	6.9	10.6	10.8
Na_2O	3.4	5.1	4.4	2.7	3.4	2.7	4.3
K_2O	0.18	0.18	0.55	1.1	0.24	0.57	0.45
TiO_2	2.7	2.0	<0.02	1.2	3.2	1.2	1.1
P_2O_5	0.17	0.11	0.02	0.05	0.17	0.05	0.34
MnO	0.23	0.19	0.036	0.18	0.15	0.17	0.17
H_2O+	2.8	2.5	3.1	3.1	4.1	2.4	2.5
H_2O-	0.5	0.5	0.2	0.5	0.5	0.4	0.4
CO_2	0.3	0.2	0.2	0.2	0.2	<0.1	<0.1
Total	99.38	101.18	100.23	101.03	100.06	100.29	101.16

Sample 1. - Sheared greenstone
2. - Sheared greenstone
3. - Medium green, with maroon splotches, aphanitic meta-andesite
4. - Medium- to dark-greenish-gray metabasalt
5. - Dark-green, fine-grained pyroxene-bearing(?) metabasalt
6. - Dark-greenish-gray, aphanitic to fine-grained pillow(?) basalt
7. - Dark-green, fine-grained pyroxene-rich metabasalt.

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TECTONIC AND ECONOMIC SIGNIFICANCE OF LATE DEVONIAN AND LATE PROTEROZOIC U-PB ZIRCON AGES FROM THE BROOKS RANGE, ALASKA

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INTRODUCTION

Preliminary interpretation of new zircon U-Pb dates from the central Brooks Range yields reliable Devonian and possible Late Proterozoic magmatic crystallization ages for metaigneous rocks that were previously thought to have crystallized from Cretaceous magmas. These dates are the result of geologic mapping and sampling by Pessel and Dillon, mineral separations and petrography by Veach and Dillon, and isotopic analyses by Chen.

METHOD

Two to 20 kg of unweathered rock were collected from each of eight outcrops determined to be free of inclusions, country rock, or crosscutting igneous rocks (fig. 1). The samples were slabbed to ensure the absence of inclusions and then ground to pass 200 mesh. Heavy minerals, including the zircons, were concentrated on a Wilfley table and with bromoform and methylene iodide. The densest minerals were sonicated and magnetic minerals were removed from the separates with vertical-drop and Franz magnetic separators. The zircon-rich fraction was hand picked to about 98 percent purity and examined under binocular and petrographic microscopes for morphologic details. Some zircon separates were bathed in nitric acid. The zircons were hand picked again under ultraviolet light to ensure virtual 100 percent purity. Isotopic analyses were performed on 1- to 9-mg zircon fractions by using an extraction method similar to that described by Krogh (1973). A mixed 205Pb-235U spike was used and isotopic measurements were done on a 35-cm-radius, 90°-sector, single-focusing, solid-source Avco mass spectrometer at UC-Santa Barbara. The standard silica gel-H_3PO_4 emitter technique was used for lead isotopic determination; the uranium isotopes were measured by using the graphite-H_3PO_4 emitter technique.

RESULTS

All the zircons are euhedral and contain tiny euhedral zircon inclusions. Only zircons from sample 77 Dn 217 have rounded cores (in 2 percent of the separate), which may be traces of inherited zircons. Except for this sample, replicate analyses of samples from the same outcrop area collected by different geologists at different times, of magnetic zircon separates from the same rock, and of hand-picked zircon separates from the same rock yield U-Pb and Pb-Pb ages that deviate less than 1 and 5 percent, respectively (table 1). The analytical uncertainty demonstrated by replicate analyses is similar to the calculated precision of each U-Pb and Pb-Pb isotopic analysis (1 and 2 percent, respectively).

The difference between replicate analyses of sample 77 Dn 217 could be due to analytical error, multiple episodes of lead loss, or contamination by inclusions or inherited zircons. Consistent agreement of duplicate analyses of the other samples makes analytical error unlikely and the deviations between samples 217a and 217b are probably due to multiple episodes of lead loss and inherited zircons. If both of the metamorphic events discussed below affected sample 77 Dn 217, they may have caused multiple episodes of lead loss. Microscopic examination of 100 zircons from sample 77 Dn 217 (table 1) reveals that 5 percent have dark inclusions and 2 percent have rounded zircon cores that appear to be relicts of older zircons. If the dark inclusions contain uranium or lead, differences in the numbers of compositions of inclusions in samples 217a and 217b may have caused the deviation (although similar dark inclusions in zircons from duplicate samples 77 Dn 76 and 77 Mw 76 did not cause their ages to differ). The rounded zircon cores probably contain old lead and uranium of inherited zircons; differences in the number of cored zircons the separates also could have caused the deviation, although no reverse discordance is present (Higgins and others, 1977, p. 25).

U-Pb ages for sample 77 Dn 74D are concordant (table 1). U-Pb ages for all other samples display normal discordance (206Pb*-238U ages $<$ 207Pb*-235U ages $<$ 207Pb*-206Pb* ages), demonstrating that the zircons have either lost lead or gained uranium since crystallization (Davis and others, 1968). Lead loss is by far the most common type of disturbance of the isotopic system (York and Farquar, 1972, p. 87) and uranium gain is unlikely because the samples were not exposed to a uraniferous metasomatic environment such as the contact zones of younger uranium-rich intrusions (Davis and others, 1968; Grauert and others, 1974). The Pb-

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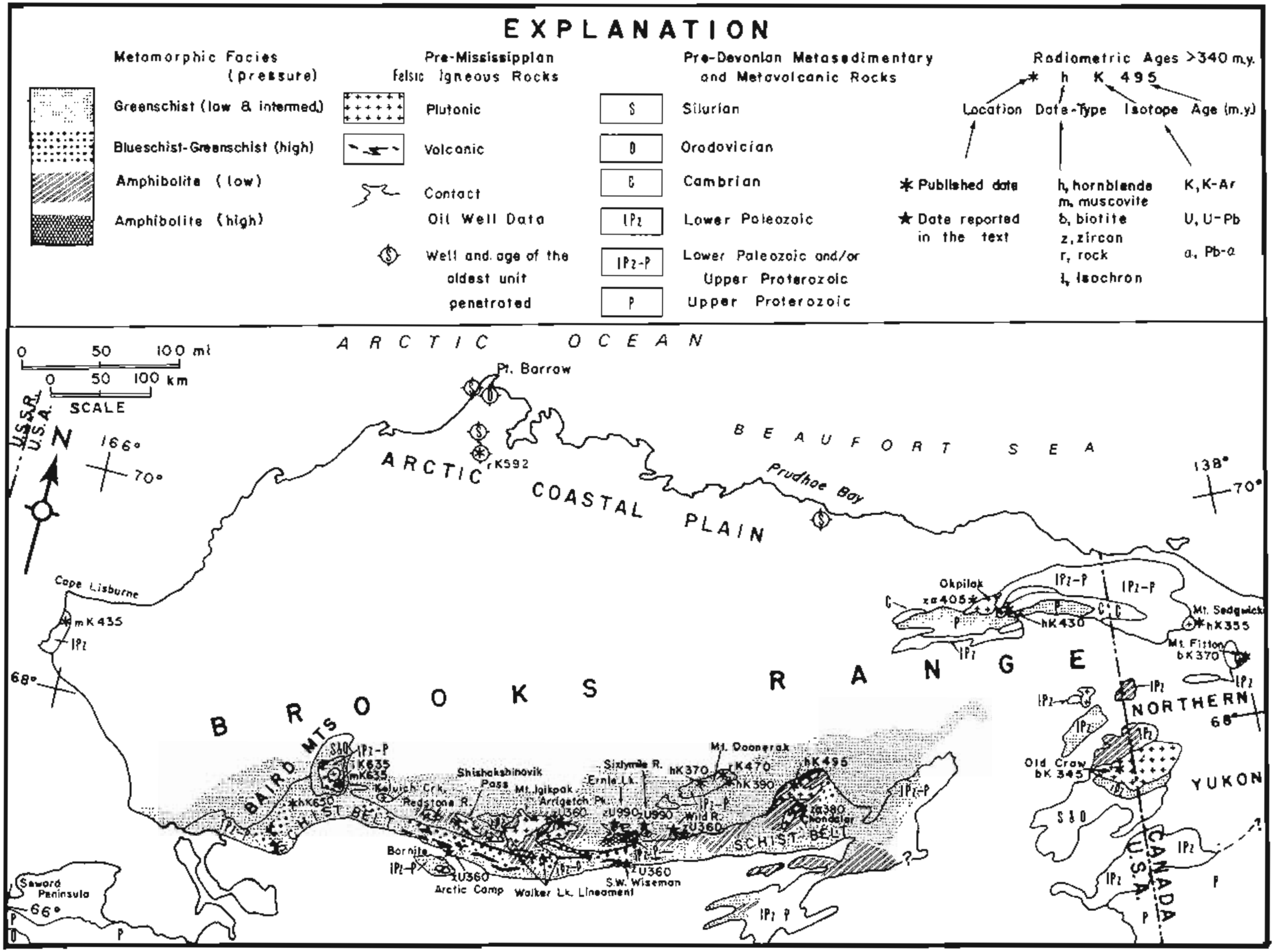

Figure 1. Map of the Brooks Range showing pre-Mississippian radiometric dates, felsic igneous rocks, pre-Devonian metasedimentary rocks, and metamorphic facies. Radiometric dates are from Sable (1977); Turner and others (1978); Grybeck and others (1977); M. Silberman, oral comm. (1978); and this report; rock units are redrawn from Beikman (1978), Grybeck and others (1977), and Churkin (1973); metamorphic facies are shown only within the Brooks Range and are from a map by Mayfield, in Grybeck and others (1977).

Pb age of normally discordant zircons is closest to the original crystallization age (table 1) because 206Pb usually is not fractionated from 207Pb during lead loss (York and Farquar, 1972).

The original crystallization age of discordant zircons must be determined from a concordia diagram. When discordance is due to episodic lead loss, the upper intercept of a chord through the data points with the concordia curve is the age of original crystallization and the lower intercept is the age of lead loss. If point 217b is disregarded, the discordant U-Pb dates appear to lie along two chords on the concordia diagram (fig. 2). The younger (chord 1) is defined by U-Pb isotope ratios in zircons from the Arrigetch and Wild River metaplutonic rocks and the Wiseman and Arctic Camp metavolcanic rocks. It intersects concordia at 360 ± 10 m.y. and at 80 ± 25 m.y. The seven points at the upper end of chord 1 (inset, fig. 2) closely control its upper intercept with the concordia, thereby minimizing the error resulting from construction. The lower intercept (80 ± 25 m.y.) is less well determined because no data points occur near the intercept. However, Late Cretaceous lead-

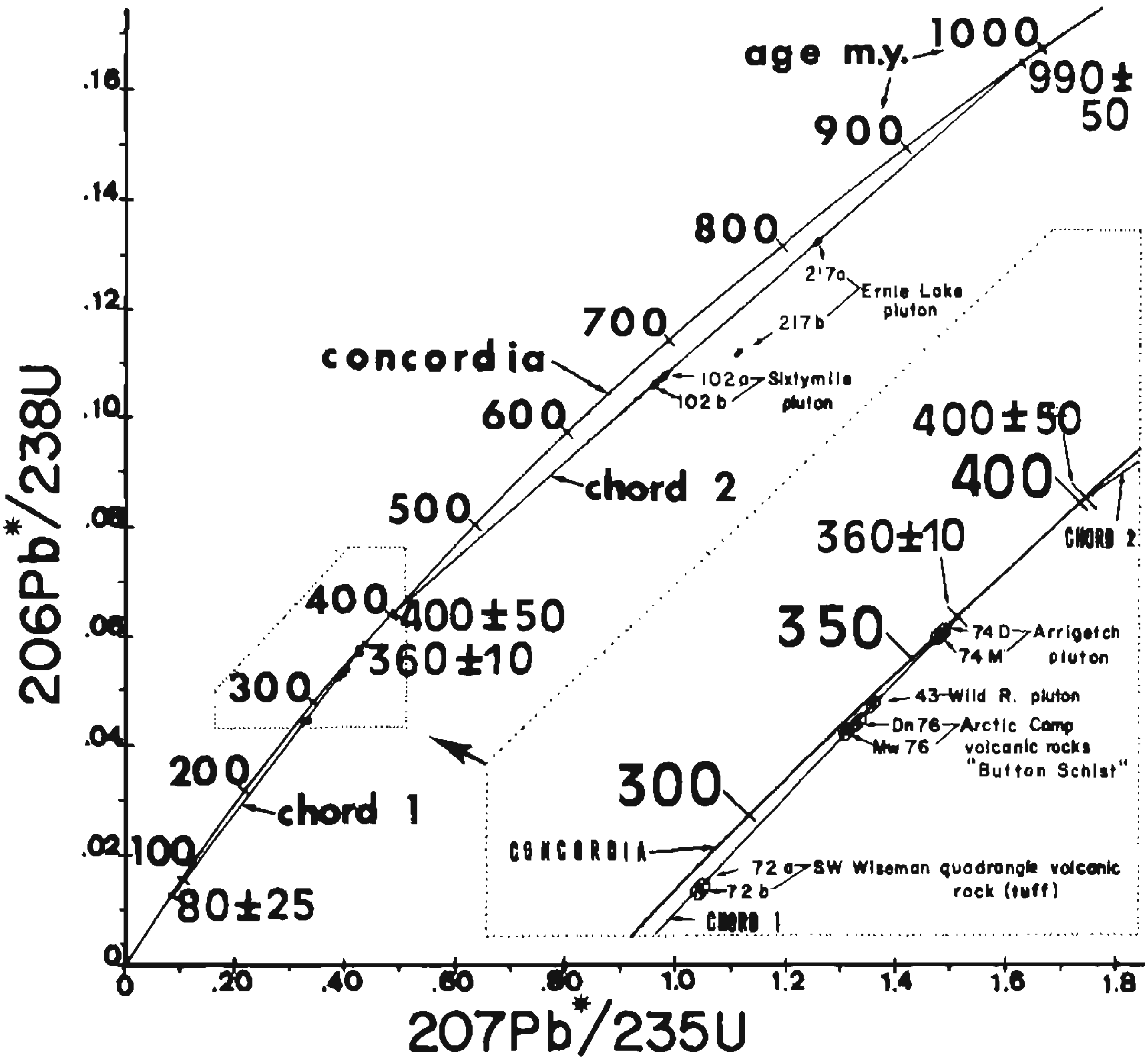

Figure 2. Concordia diagram (Stacey and Stern, 1973) of U-Pb isotopic ratios from the Brooks Range, Alaska. Analytical uncertainty lies within the small polygonal area surrounding each data point. Inset shows detail in central portion of concordia. (Sample numbers are abbreviated; see table 1.) *Radiogenic Pb, corrected for common Pb.

loss age is supported by four 90 ± 2 m.y. K-Ar dates on metamorphic muscovite and biotite from the Arrigetch pluton (Turner and others, 1978). Lead loss at 80 ± 25 m.y. from zircons first crystallized in intrusive and extrusive magmas about 360 m.y. ago is suggested.

Chord 2 on figure 2 is defined by U-Pb ratios in three of the four zircon separates taken from plutons located in the western Wiseman Quadrangle. This chord intersects concordia at 990 ± 50 m.y. and 400 ± 50 m.y. Although the Late Proterozoic age of the upper intercept is supported by the U-Pb and Pb-Pb ages from the same zircons, the time of crystallization of the magma cannot be precisely determined from the concordia diagram because one of the four data points has been disregarded. The lower intercept is close to the time of the magmatic event at 360 ± 10 m.y. Metamorphism accompanying Devonian magmatism may have caused the lead loss from zircons first crystallized during the Late Proterozoic.

GEOLOGIC INTERPRETATION

If lead loss is assumed, a seven-point U-Pb (zircon) chord from plutonic and volcanic rocks of the central Range metaigneous rocks record a Middle to Late Devonian magmatic event and a Late Cretaceous metamorphic event. The presence of Late Proterozoic plutons metamorphosed during Devonian time is not firmly established but can be inferred from the chord defined by most of the samples from the Ernie Lake and Sixtymile River plutons; this is supported by the U-Pb and

Table 1. Zircon U-Pb data for samples from Brooks Range, Alaska

Sample[1]	Ages (m.y.)[2] $^{206}Pb^*/^{238}U$	$^{207}Pb^*/^{235}U$	$^{207}Pb^*/^{206}Pb^*$
77 Dn 76	329	334	373
77 Mw 76	327	332	364
77 Dn 72a	277	284	350
77 Dn 72b	276	285	367
77 Dn 74D	357	357	357
77 Dn 74M	356	359	376
77 Dn 43	336	338	357
77 Dn 102a	652	688	808
77 Dn 102b	651	688	808
77 Dn 217a	798	829	912
77 Dn 217b	681	761	1002

[1]M = magnetic and D = diamagnetic separates; a, b = duplicate separates; 77 Dn 76 and 77 Mw 76 are duplicate samples.

[2]Constants used in age calculation: $\lambda^{238}U = 1.5513 \times 10^{-10}$, $\lambda^{235}U = 9.848 \times 10^{-10}$, $^{238}U/^{235}U = 137.88$. Analytical uncertainty of U/Pb ages estimated less than 1% of the age; in Pb-Pb ages, less than 2%.

*Indicates radiogenic Pb, corrected for common Pb.

Pb-Pb ages on all of these samples. We combine these findings with other geologic data (below) and suggest that Cretaceous deformation and metamorphism affected a mid-Paleozoic ensialic island arc developed on lower Paleozoic and Late Proterozoic(?) continental basement that now underlies the southern Brooks Range (Dillon and Pessel, 1979).

Mid-Cretaceous K-Ar dates on biotite and muscovite from the Arrigetch and Igikpak plutons formerly were taken as the most likely age for intrusion of the plutons (Brosge and Reiser, 1971; Turner and others, 1978; Beikman, 1978; Grybeck and others, 1977). Nelson and Grybeck (1978) and Dillon (unpub data) discovered that the dated minerals define a metamorphic fabric within the plutons and that the Cretaceous dates may reflect postmetamorphic cooling. This metamorphism may have caused the Cretaceous lead loss suggested above. Cretaceous regional greenschist-facies metamorphism of Brooks Range rocks can also be inferred from the 47 Cretaceous K-Ar dates obtained from the 76 pre-Mesozoic mineral separations (Turner and others, 1978). Metamorphism may have accompanied northward overthrusting of oceanic or quasi-oceanic crust onto the Brooks Range rocks during Jurassic and Early Cretaceous time (Newman and others, 1977; Patton and others, 1977; and Roeder and Mull, 1978).

Widespread Late Devonian plutonic and volcanic rocks of the central Brooks Range provide compelling evidence for a major mid-Paleozoic magmatic event. The 360 ± 10 m.y. age obtained from the Arrigetch and Wild River plutons correlate well with a 370 ± 40 m.y. Rb-Sr whole-rock isochron date from the Arrigetch and Igikpak plutons (Silberman, oral comm., 1978). Petrologically similar plutons occur farther west at Shishakshinovik Pass, on the Redstone River, and on Kaluich Creek (Pessel and Brosge, 1977). Other plutons in the eastern and northeastern Brooks Range and the northern Yukon Territory that yield middle Paleozoic radiometric ages are those of the Chandalar area, the Okpilak and Old Crow batholiths, and Mounts Fitton and Sedgwick (fig. 1). Within the contact aureoles of all these mid-Paleozoic plutons the metasedimentary country rocks are prograded to the amphibolite facies.

Felsic metavolcanic and hypabyssal intrusive rocks interlayered with metasedimentary rocks and strata-bound ore deposits crop out 15 to 30 km south of the mid-Paleozoic plutonic belt along the Brooks Range schist belt for over 250 km. U-Pb ages of 360 ± 10 m.y. for volcanic rocks within this package of ore-bearing strata at Arctic Camp and the southwest Wiseman Quadrangle are in general agreement with mid- to late-Paleozoic model lead ages from metavolcanic rocks at Arctic Camp (Smith and others, 1978). Interlayers of marble with poorly preserved Devonian(?) fossils occur within the volcanic units at Arctic Camp (Smith and others, 1978) and 30 km farther east (Brosge and Pessel, 1977). The U-Pb ages are also contemporaneous with Devonian fossils from marbles at the Bornite stratabound deposit, 30 km to the south (Fritts, 1970).

The Igikpak pluton intruded and metamorphosed schist-belt rock units near Walker Lake (Brosge and Pessel, 1977), where it is exposed less than 10 km north of coeval felsic metavolcanic rocks (fig. 1). The metamorphic aureole of Igikpak pluton affects rocks on both sides of the Walker Lake fault (Fritts and others, 1971). Petrologic similarity of adjacent Late Devonian volcanic and plutonic rocks in the central Brooks Range leads us to conclude that they are genetically related and may well be comagmatic. The similarity of the magmatic and metamorphic history of rocks on both sides of the Walker Lake fault makes major post-Devonian displacements along this lineament unlikely; we suggest that it represents a deformed and locally faulted unconformity.

During mid-Paleozoic magmatism, shallow marine sedimentary rocks (Brosge and Tailleur, 1970) and stratabound ore deposits were deposited throughout the Brooks Range. The carbonate rocks, adjacent plutons, and nearby volcanic rocks probably formed in an island-arc or on a submerged continental margin environment while contemporaneous clastic rocks to the north were being deposited in a foreland basin or marginal sea. Ore deposition within this setting formed at least 10 large volcanogenic Cu-Pb-Zn-Ag deposits (Wiltse, 1975, 1977; Grybeck and DeYoung, 1978; Smith and others, 1978). Near the plutons other syngenetic Pb-Zn-Ba-Ag-Cu deposits occur in Middle and Upper Devonian metasedimentary rocks that are coeval with, but not close to Devonian felsic metavolcanic rocks (Grybeck and DeYoung, 1978, areas 4, 11, and 19; Dutro, 1978). A nearby source for these Devonian stratabound ore deposits is the hydrothermal systems of the mid-Paleozoic magmas. These systems may have also emplaced as-yet-undiscovered porphyry, vein, and contact ore deposits near the plutons.

Pre-Middle Devonian basement rocks in the central Brooks Range are exposed around the Old Crow and Okpilak batholiths, at Mt. Doonerak, in the Baird Mountains, and possibly near Ernie Lake and within the Brooks Range schist belt (fig. 1) (Grybeck and others, 1977). Lower Paleozoic and Proterozoic(?) basement rocks of the mid-Paleozoic magmatic arc were affected by deformation and amphibolite-facies contact-metamorphism accompanying emplacement of the mid-Paleozoic plutons (fig. 1). Distinctive kyanite-amphibolite facies rocks that crop out just south of Ernie Lake (Grybeck and others, 1977) may be part of a pre-middle Paleozoic high-pressure metamorphic terrain within the schist belt (Gilbert and others, 1978) that strikes westerly along the south side of the Igikpak pluton to the Upper Proterozoic(?) schists in the Baird Mountains (Turner and others, 1978). The Baird Mountain schists trend southward into Proterozoic metamorphic rocks on the Seward Peninsula.

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