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GEOLOGIC REPORT 63

Recent research on Alaskan geology

STATE OF ALASKA

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Cover photo: Active(?) rock glacier, headwaters of Ganes Creek, Iditarod Quadrangle, Alaska.

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METRIC CONVERSION FACTORS

To convert feet to meters, multiply by 0.3048. To convert inches to centimeters, multiply by 2.54.

LEAD ISOTOPE RATIOS FROM THE RED DOG AND DRENCHWATER CREEK LEAD-ZINC DEPOSITS, DE LONG MOUNTAINS, BROOKS RANGE, ALASKA

By Larry Lueck[1]

INTRODUCTION

New lead isotope data have been determined for the Red Dog and Drenchwater Creek base-metal sulfide occurrences in the De Long Mountains of the western Brooks Range, Alaska. To the author's knowledge these are the only two deposits studied by lead isotope methods on the north flank of the Brooks Range.

The Red Dog Prospect (secs. 20 and 29, T. 31 N., R. 18 W., Kateel River Meridian, De Long Mountains Quadrangle), was first described by Tailleur (1970). Mineralization occurs primarily as disseminations of sphalerite, pyrite, and minor galena in country rock; as quartz-rich veins and podiform bodies of zinc, lead, and iron sulfides, some of which contain barite; and as extensive overlying massive barite of unknown configuration. Origin of the mineralization is a matter of considerable debate.

About 160 km east of Red Dog is Drenchwater Creek (T. 10 S., R. 1 E., Umiat Meridian, Howard Pass Quadrangle), the site of an apparently low-grade, disseminated lead-zinc deposit in volcanic, intrusive, and pelagic sedimentary rocks that have been complexly disturbed by folding and low-angle thrusting. Previous investigators believe the mineralization may be volcanogenic-exhalative in origin and therefore contemporaneous with the country rock.

Local geology and structure are presented by Plahuta (1978) and Nokleberg and Winkler (1978) for the Red Dog and Drenchwater Creek deposits, respectively. Soil geochemistry and results of magnetometer and gamma-ray spectrograph surveys of both deposits are reported by Metz and others (1978a).

DATA

Two samples from Red Dog vein material and one sample from Drenchwater Creek disseminated ore were chosen for this study. Two galena splits from each sample were hand separated by the author and analyzed by Teledyne Isotopes, Inc., of New Jersey and Dr. G.L. Cumming of the University of Alberta, Edmonton. Different methods of lead digestion and concentration were used by Teledyne and Cumming, but both sets of analyses were normalized according to the NBS981 lead standard and gave good reproducibility (L.F. Casabona and G.L. Cumming, written comm., 1979).

Results are listed in table 1 and the ratios are plotted in figures 1 and 2, which are graphs of $^{207}Pb/^{204}Pb$ vs $^{206}Pb/^{204}Pb$, and $^{208}Pb/^{204}Pb$ vs $^{206}Pb/^{204}Pb$. These graphs contain: a) a 'lead growth curve' that shows gradual decay of uranium or thorium to stable lead isotopes through time and b) a series of 'isochrons' that radiate from an estimated primordial lead composition and cross the growth curve in increments of years before present. Data points on or near the growth curve are, in theory, dated by the isochron crossing the curve at that point. More thorough discussions of thorium and lead decay are found in Doe (1970), Doe and Stacey (1974), and Faure (1977, p. 227-66). Stacey and Kramers (1975) and Cumming and Richards (1975) describe new growth curves that are based on fewer assumptions and better fit the lead isotope data than do the old single-stage models. Together with Richards (1971), these papers present the most thorough examination of current research in lead isotope theory. On the advice of Cumming (pers. comm., 1980), the Stacey and Kramers (1975) model growth curve is used in this paper because it best approximates relatively young leads such as those of the Red Dog and Drenchwater Creek deposits.

A line of best fit through the Teledyne data points (fig. 2) passes very close to the origin of the graph, which indicates there are probably significant ^{204}Pb errors in these analyses (Kanesewich, 1968, p. 155-60). Because ^{204}Pb constitutes less than 1.65 percent of any natural lead, its analysis is difficult (Cannon and others, 1961, p. 5); nevertheless, these probable errors render the Teledyne points unreliable for dating. The Cumming data points fall close to this same ^{204}Pb error line, but their tight grouping indicates that the ^{204}Pb error is not very significant in these analyses. All three points plot close to the Stacey and Kramers (1975) growth curve. Within the limits of analytical error, they are the same (table 1 and fig. 1) and remain together, even on the larger scale plot of $^{207}Pb/^{204}Pb$ vs $^{206}Pb/^{204}Pb$ (fig. 2). Hence, unless three almost identical samples have accidentally been selected from two otherwise heterogeneous populations (a possibility discussed

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Table 1. *Lead isotope ratios from the Red Dog and Drenchwater Creek deposits*

Sample	206/204	207/204	208/204	References
LL-26-18 (Red Dog)	18.404 ± .013	15.590 ± .014	38.228 ± .046	1
	18.175 ± .000	15.414 ± .037	37.770 ± .091	2
LL-4-14 (Red Dog)	18.413 ± .016	15.604 ± .017	38.197 ± .062	1
	18.325 ± .000	15.535 ± .012	38.026 ± .035	2
78PMO52 (Drenchwater)	18.406 ± .013	15.593 ± .013	38.270 ± .049	1
	18.057 ± .000	15.353 ± .045	37.585 ± .105	2
Stacey and Kramers' average modern lead	18.700	15.628	38.63	3

1. Analyses from G.L. Cumming laboratory, University of Alberta, Edmonton.
2. Analyses from Teledyne Isotopes, Inc.
3. Stacey and Kramers, 1975.

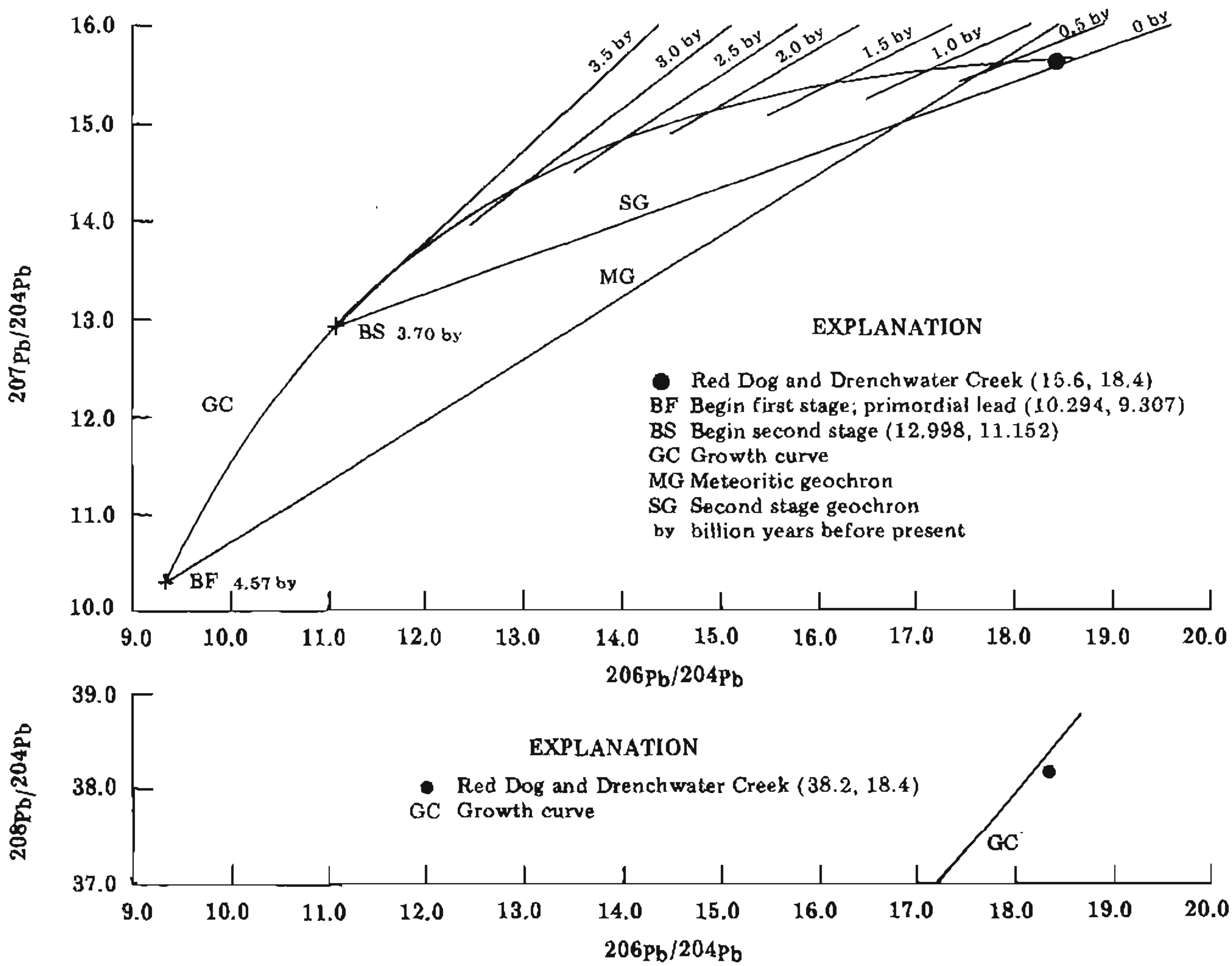

Figure 1. Two-stage model of Stacey and Kramers (1975) plotted with Red Dog and Drenchwater Creek data.

below), the two deposits probably received this lead from the same source or from two genetically related sources.

BACKGROUND

'Ordinary' leads are thought to be derived from a deep, homogeneous source without much crustal contamination. These leads may give meaningful model dates in some cases but are fairly rare. Relatively few deposits contain lead with isotope ratios similar enough to an ideal growth curve to be classed as ordinary and thus datable by Pb/Pb methods (Cumming and Richards, 1975, table 3, p. 165; Stacey and Kramers, 1975, tables 2, 4, and 6, p. 208, 210, and 213). Samples from modern volcanogenic and exhalative deposits do not necessarily plot on the zero isochron, or 'geochron.' Therefore, some investigators believe that even oceanic volcanic rocks have been contaminated by crustal, radiogenic lead (Richards, 1971, fig. 3, p. 429). Thus, even those lead isotope ratios presumed to be ordinary must be cautiously interpreted with regard to other geologic evidence.

All leads that are not ordinary are termed 'anomalous.' Although quite random combinations of lead isotope ratios can be found in anomalous deposits, two special categories have been recognized, 'B-type' and 'J-type.' In the former, lead has a model age greater than that of the host rock, a condition thought to be caused by remobilization and redeposition of older crustal lead in younger country rock. J-type lead deposits are produced by mixing of older crustal lead with juvenile lead from depth and often yield negative, or future, model ages.

Base-metal deposits containing anomalous leads yield such age discrepancies (that is, if age can be estimated at all) and exhibit a wide scatter of isotope ratios (Stanton and Russell, 1959, p. 595). In the Mississippi Valley and Tri-State districts, where anomalous J-type leads are common, the scatter is even more pronounced (Heyl and others, 1974, fig. 2, p. 995). On the other hand, the isotopic similarity of leads from the Red Dog and Drenchwater Creek deposits suggests that these leads are ordinary. Furthermore, as shown above, these analyses fall close to an ideal growth curve for ordinary lead (fig. 2).

ALTERNATIVES IN EVALUATING THE DATA

Leads from only three galena samples from the Red Dog and Drenchwater Creek deposits were analyzed for this study. If these samples are not representative of those deposits, then the data are not significant, the leads could be either ordinary or anomalous, and their datability and other properties are left in question.

If, on the other hand, the samples do represent the deposits, then the leads are probably ordinary and therefore datable. On Stacey and Kramers' (1975) lead-growth curve the sample analyses yield a lead model mineralization age of about 170-190 m.y. (fig. 2). This Early Jurassic age is corroborated by Lange and others (1980), who report an age of 200 m.y. for galena mineralization at Red Dog (although the ratios and model used to determine this age are not reported).

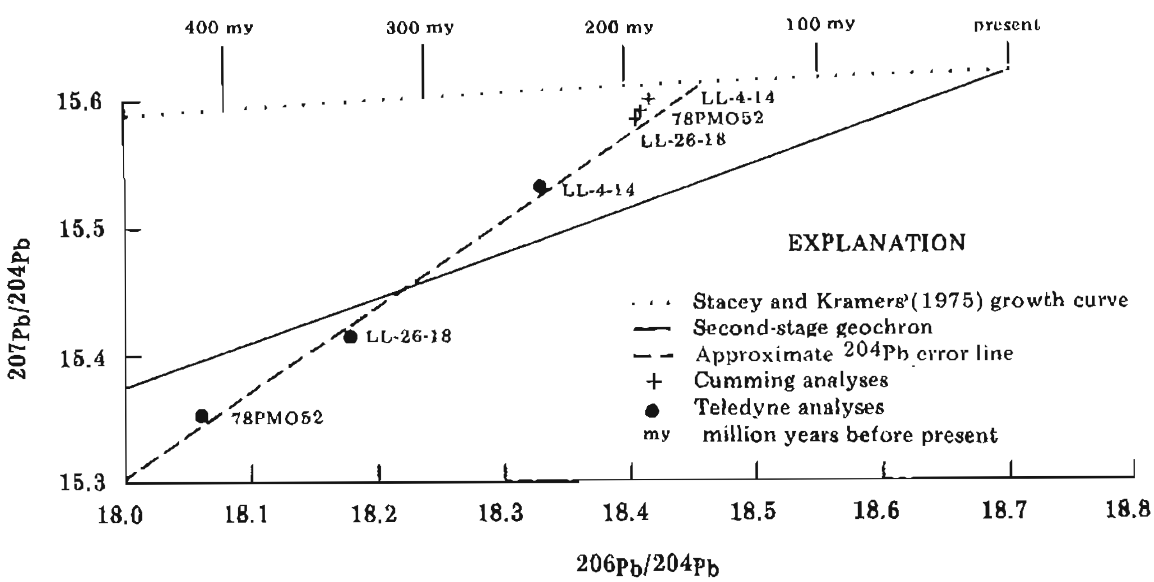

Figure 2. Enlargement of upper end of growth curve shown in figure 1.

DISCUSSION

Though of unproven stratigraphic position, host rock for the Red Dog ore is mainly silicified black mudstone, shale, and chert and contains disseminated sulfides. It has been assigned by Plahuta (1978, fig. 2, p. 4-5), to the Tupik Formation, which is the uppermost formation of the Mississippian Lisburne Group and part of the De Long Sequence of Martin (1970, fig. 2b). The unit has yielded no direct fossil or K-Ar dates at Red Dog, but chert samples collected by the author and coworkers from strata immediately overlying the ore zone contain radiolaria (identified by D.L. Jones, USGS) of Permian age (Nokleberg, pers. comm., 1980). Assuming, therefore, that the country rocks are Permian or older, a Jurassic Pb/Pb age of 170-190 m.y. is reasonable for the discordant vein mineralization, but not for the disseminated ore, which is probably syngenetic. Thus, at least two stages of mineralization, widely separated in time, may be represented in the Red Dog deposit.

Biotites from a keratophyre and a pyroxene andesite from the Drenchwater Creek occurrence yielded K-Ar ages of 319 $\pm$ 10 m.y. and 330 $\pm$ 17 m.y., respectively (Tailleur and others, 1977). Sample 78PMO52 yielded a lead age of 170-190 m.y. and was collected from the unit designated by Nokleberg and Winkler (1978, pl. 1) as the 'black, medium-bedded chert' of the Mississippian Lisburne Group. These rocks include abundant volcanics and intrusives not found at Red Dog, but are thought to be lateral facies equivalents of the black host rock at Red Dog. Nokleberg and Winkler (1978, p. 13) state that galena and sphalerite of the Drenchwater Creek deposit are closely associated with aquagene tuff (the keratophyre and pyroxene andesite) or with dark chert and shale adjacent to tuff. Hence, the mineralization could be largely syngenetic and volcanogenic-exhalative, and therefore Mississippian, as is the host rock. However, a previously unrecognized mineralizing event may have introduced epigenetic galena 170-190 m.y. ago. Pervasive thrusting and folding that continued along the North Slope through Cretaceous time (Nokleberg and Winkler, 1978, p. 10) may have obscured or destroyed evidence of the mode of this lead emplacement. No such evidence was recognized during field work (but there was no suspicion of two generations of lead at that time).

Thus, lead isotope ratios from the Red Dog and Drenchwater Creek deposits, if significant, provide evidence for a Jurassic epigenetic mineralizing event, although other geological evidence points to an earlier Mississippian period of syngenetic ore deposition.

There is mounting evidence that during Mississippian time a large-scale rift, or aulacogen, opened along an east-west axis through the present North Slope of Alaska (Metz and others, 1978b). Volcanic and submarine exhalative activity accompanying the rifting probably produced the Mississippian base-metal sulfide deposits syngenetically in pelagic sediments filling this trough. The later mineralization might have occurred during overthrusting of oceanic crust onto the Brooks Range rocks, which began in Jurassic time, when rifting reversed itself to assume a compressional mode that persisted into Early Cretaceous time, as described by Roeder and Mull (1978, p. 1701).

Several cases of recurrent or continuous mineralization over periods of up to 300 m.y. are discussed in Smirnow (1977). Frequent or continuous ore deposition over a long time would probably produce a set of lead ratios difficult to distinguish from the random scatter of a Mississippi Valley-type deposit. This problem need not be handled here, however, because available geologic and lead isotope data indicate only two periods of mineralization at the Red Dog and Drenchwater Creek deposits, not a continuum.

As Smirnow (1977) emphasizes, the style and mechanism of ore emplacement cannot be expected to remain unchanged over these spans of time. If tectonic activity is renewed or continued along a given trend, for instance, a plate boundary, the style of tectonism—and thus the type of mineralization—will probably vary. Because the North Slope has been the locus of tectonic activity for at least 150 m.y., two or more genetically different types of mineralization may be superimposed at the Red Dog and Drenchwater Creek deposits.

Mitchell and Garson (1976) discuss processes for emplacement of lead-zinc sulfide deposits under both divergent and convergent plate regimes that could produce a twofold mineralization like that postulated here. Type-identified ore deposits have been used in reconstructions of the plate tectonic history of numerous regions. For example, Metz and others (1978b) have reconstructed a history of the North Slope region of Alaska that relies heavily on the occurrence of ore deposits. But to be useful, such deposits must be correctly identified as to type. Incorrect characterization of the deposit type at either Red Dog or Drench-Water Creek would be highly misleading.

CONCLUSIONS

a) Samples selected from the deposits at the Red Dog and Drenchwater Creek occurrences have nearly identical lead isotope ratios. If these samples are not representative of lead in the deposits, the resemblance of ratios is merely a coincidence. But if the samples are typical of the deposits, then the lead is ordinary rather than anomalous, and therefore datable.

b) Available lead isotope ratios yield Jurassic ages of 170-190 m.y. for galena samples from both the Red Dog and Drenchwater Creek occurrences. Other evidence yields a Mississippian age for the host rock and syngenetic sulfides at both locations. This difference in ages is best explained by the occurrence of two mineralizing events, widely separated in time—in which case there may well be two different types of mineralization at the Red Dog and Drenchwater Creek deposits.

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^{40}K-^{40}Ar AGES FROM RHYOLITE OF SUGAR LOAF MOUNTAIN, CENTRAL ALASKA RANGE: IMPLICATIONS FOR OFFSET ALONG THE HINES CREEK STRAND OF THE DENALI FAULT SYSTEM

By Mary D. Albanese[1] and Donald L. Turner[2]

INTRODUCTION

Sugar Loaf Mountain is located 6 km north of the Hines Creek strand of the Denali fault system at lat 63°47' N., long 148°50' W., and consists of rhyolite intruded by andesite. The rhyolite overlies a basement of mica-quartz schist of possible Precambrian to early Paleozoic age. Mineralogically similar masses of rhyolite and andesite intrude the underlying schist up to 4 km south of Sugar Loaf Mountain (fig. 1).

The Hines Creek strand trends east-west and separates the previously mentioned mica-quartz schist from the more southerly Paleocene Cantwell Formation. Wahrhaftig and others (1975) estimate 65 to 315 km of lateral displacement along the Hines Creek strand between Devonian and Cretaceous time. The Cantwell Formation is intruded by mafic dikes feeding the volcanic Teklanika Formation (Bultman, 1972; Wahrhaftig and others, 1975; Gilbert and others, 1976).

Volcanic rocks of the Teklanika Formation (sample localities 40 km west of Sugar Loaf Mountain) yielded ^{40}K-^{40}Ar ages of 41.8 m.y., 57.2 $\pm$ 3.4 m.y., and 60.6 m.y. (Gilbert and others, 1976). Hornblende from basaltic andesite of the Teklanika Formation at Mount Fellows (9 km southeast of Sugar Loaf Mountain) yielded a ^{40}K-^{40}Ar age of 49.5 $\pm$ 2.1 m.y. (Bultman, 1972). This can be interpreted as a minimum age because the dated rock was altered.

Rhyolite and andesite of the Sugar Loaf Mountain area have been interpreted as part of the Teklanika Formation (Gilbert, 1979). The proximity of similar volcanic rocks north and south of the Hines Creek strand led Bultman (1972) to suggest that lateral motion along the strand has been minimal since Paleocene time. However, ^{40}K-^{40}Ar whole-rock ages of the Sugar Loaf Mountain rhyolite reported here do not support this interpretation.

^{40}K-^{40}Ar AGE DETERMINATIONS

Four whole-rock samples of fine-grained rhyolite from the Sugar Loaf Mountain area (fig. 1) were dated by the ^{40}K-^{40}Ar method at the Geochronology Laboratory of the Geophysical Institute, University of Alaska, Fairbanks (table 1). Samples MR78-2 and -39 are from an outcrop on Sugar Loaf Mountain, and samples MR78-50 and -58 are from localities 4.0 and 1.2 km southeast of the mountain, respectively. The petrology of these samples is listed in table 2. Samples MR78-39, -50, and -58 meet petrologic criteria from reliable ages (Mankinen and Dalrymple, 1972). Although sample MR78-2 contains devitrified groundmass glass, its age agrees with those of the other samples. These rhyolites yield middle Oligocene whole-rock ages ranging from 32.4 $\pm$ 1 to 35.2 $\pm$ 1 m.y. (table 1). Field relationships suggest that the associated andesite is contemporaneous or younger (Albanese, 1980).

DISCUSSION AND CONCLUSIONS

The substantial age difference between the dated Teklanika Formation andesite in the Mount Fellows region (Bultman, 1972) and the dated rhyolite in the Sugar Loaf Mountain area indicates that the rocks of the two regions are not genetically related. Thus, they provide no evidence on the dates or amounts of displacement along the Hines Creek strand. The main geologic constraint on the age of the strand is the Buchanan Pluton, located about 70 km east of Sugar Loaf Mountain. This intrusive body has been dated at 95 m.y. and extends northeast across the trace of the Hines Creek strand with no offset, which suggests the absence of lateral movement along the Hines Creek strand in the last 95 m.y.

The only known rhyolite and andesite bodies south of the Hines Creek strand that can be correlated with those of the Sugar Loaf Mountain area are the mid-Tertiary felsic volcanic rocks that form a subordinate part of the Mount Galen Volcanics. This formation, located south of the Hines Creek strand and about 60 km southwest of the Sugar Loaf Mountain area, is predominantly andesite and basalt and yields ^{40}K-^{40}Ar ages ranging from 32.3 $\pm$ 1.0 to 43.2 $\pm$ 2.6 m.y. (Decker and Gilbert, 1978). Mid-Tertiary felsic intrusives also occur north and south of the Mount Galen Volcanics (Gilbert, 1979). Although the Mount Galen Volcanics are about the same age as the rhyolite and andesite of the Sugar Loaf Mountain area, the two volcanic se-

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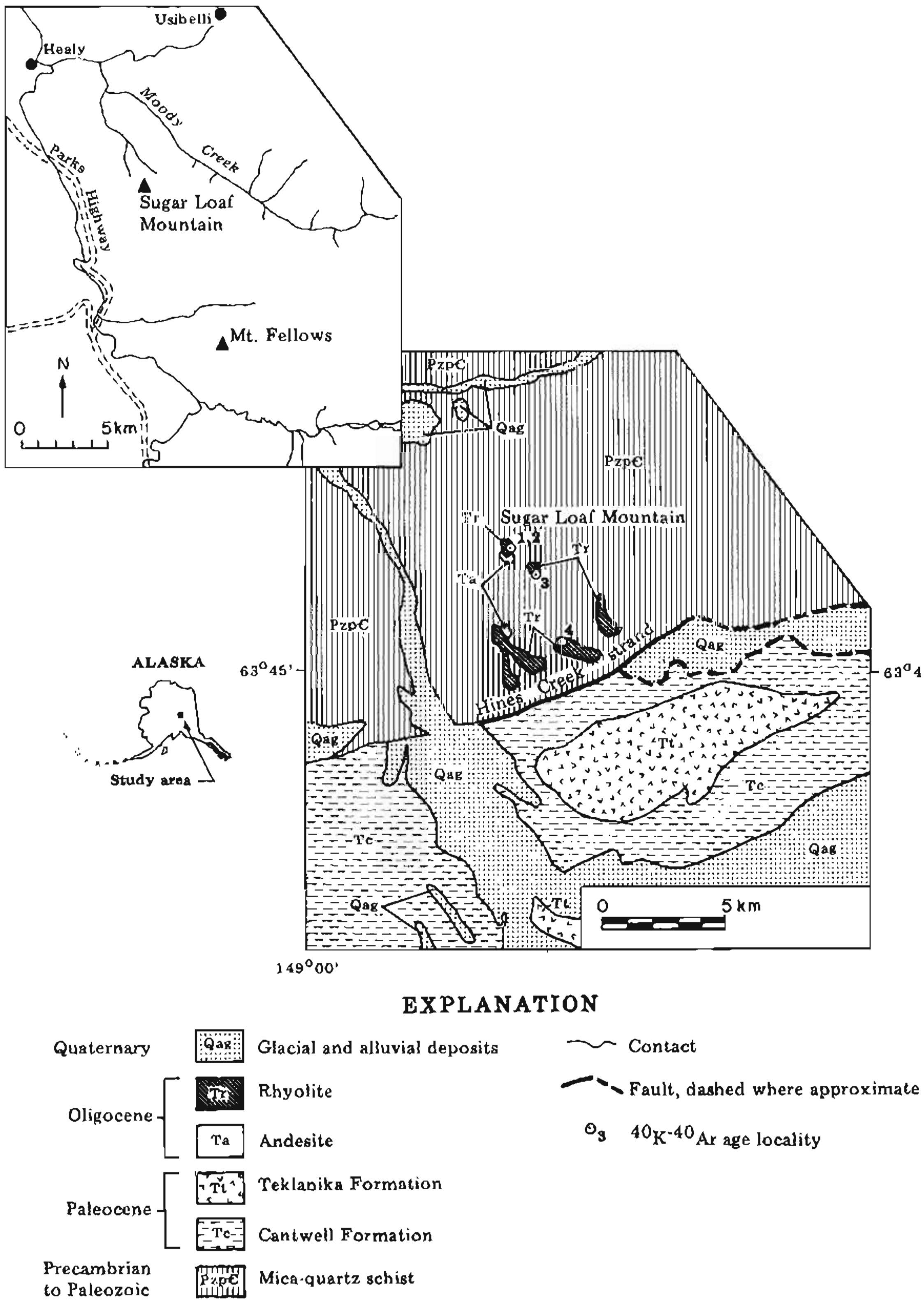

Figure 1. Location and generalized geologic map of Sugar Loaf Mountain area. Geology from Albanese (1980), Bultman (1972), Gilbert (1979), and Wahrhaftig (1970).

Table 1. *Analytical data for whole-rock K-Ar age determinations on rhyolites from Sugar Loaf Mountain area.*

Map No.	Sample	K_2O (wt. %)	^{40}Ar rad (%)	$\frac{^{40}Ar_{rad}}{^{40}K} \times 10^{-3}$	Age $\pm 1\sigma$ (m.y.)
1	MR78-2	$\bar{x}_4 = 4.425$	96.44	2.016	$\geq 34.4 \pm 1$
1	MR78-2 (replicate)	$\bar{x}_4 = 4.425$	93.00	2.001	$\geq 34.4 \pm 1$
2	MR78-39	$\bar{x}_4 = 4.344$	94.28	1.900	32.4 ± 1
2	MR78-39 (replicate)	$\bar{x}_4 = 4.344$	91.87	1.916	32.7 ± 1
3	MR78-58	$\bar{x}_4 = 3.277$	84.67	2.065	35.2 ± 1
4	MR78-50	$\bar{x}_4 = 4.729$	89.19	1.924	32.8 ± 1

Note: rad = radiogenic; σ = standard deviation; $\bar{x}$ = mean, $\lambda_\epsilon + \lambda_\epsilon' = 0.581 \times 10^{-10}\ yr^{-1}$; $\lambda_\beta = 4.962 \times 10^{-10}\ yr^{-1}$; $^{40}K/K$ total = 1.167×10^{-4} mol/mol. Analytical techniques have been described previously (Turner and others, 1973).

quences differ in age composition and may not be genetically related. However, igneous suites from both areas were probably produced during the same mid-Tertiary regional magmatic event.

Table 2. *Petrology of rhyolites from Sugar Loaf Mountain area.*

MR78-2		
	Phenocrysts -	5% quartz (0.5-2 mm dia) 5% alkali feldspar (0.5-1 mm dia) 1% plagioclase (0.5 mm dia)
	Groundmass -	68% quartz 20% devitrified glass 1% hypersthene
MR78-39		
	Phenocrysts -	5% quartz (2.5 mm dia) 5% alkali feldspar (0.5-1 mm dia)
	Groundmass -	80% quartz 9% feldspar 1% clay
MR78-50		
	Phenocrysts -	1% alkali feldspar (1 mm dia)
	Groundmass -	79% quartz 17% feldspar 2% clay 1% sericite
MR78-58		
	Phenocrysts -	5% quartz (2.5 mm dia) 5% alkali feldspar (0.5-1 mm dia)
	Groundmass -	80% quartz 9% feldspar 1% clay

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MULTIPLE GLACIATION IN THE BEAVER MOUNTAINS, WESTERN INTERIOR ALASKA

By T.K. Bundtzen[1]

INTRODUCTION

The Beaver Mountains are a rugged, isolated, igneous-cored massif in the Kuskokwim Mountains of southwestern Alaska, about 40 mi southwest of McGrath, Alaska (fig. 1). Eakin (1913) and Mertie (1936) first recognized evidence of glaciation in the Beaver Mountains and nearby highlands such as the Sunshine Mountains and Cloudy Mountain. During a 1979 mineral resource investigation, the author found evidence of four Quaternary glaciations in the Beaver Mountains and adjacent lowlands. This study is based on 3-1/2 weeks of field work during August 1979 and 6 weeks of photogeologic interpretation of 1:40,000-scale black-and-white aerial photographs.

GEOGRAPHY

The highest parts of the 400-mi^2 study area (fig. 2) have been carved and steepened by glacial erosion, but no glaciers exist today. Elevations in the Beaver Mountains range from 500 ft in the valleys of Windy and Hunter Creeks to a 4,150-ft-high unnamed peak about 2-1/2 mi southwest of Tolstoi Lake. Numerous peaks and ridges that rise above 3,000 ft contrast with the lower, rounded hills of the nearby Kuskokwim Mountains. The Beaver Mountains massif is an elongated north-northwest-trending ridgeline; major streams flow northeast and west-southwest from this hydrographic divide through long, U-shaped valleys.

Vegetation on the lee (east-northeast) side of the divide below the average 2,500-ft timberline consists of large stands of mature white spruce, alder, willow, and minor birch and a ground cover of shrubs and lichens. There are large areas of lichen-covered treeless tundra dominated by *Cladonia* west and north of the divide, where stands of trees are limited to protected valley bottoms.

The present climate combines the continental influence of interior Alaska and the cool maritime weather systems of Bristol Bay and the Bering Sea. Temperatures at McGrath range from -64°F to +89°F, with a mean annual temperature of 25.5°F. According to Fernald (1960, p. 201), mean annual precipitation at McGrath is 19.13 in., with 234 days of annual cloud cover; maximum precipitation occurs in August. Annual precipitation in the Beaver Mountains is unknown but probably higher than at McGrath because of the rainshadow effect of the massif.

SUMMARY OF BEDROCK GEOLOGY

A generalized bedrock geologic map of the Beaver Mountains is shown on figure 1. The oldest recognized rock unit consists of small discontinuous lenses of chert, limestone, and sandstone of Late Paleozoic age (Pzlc, fig. 1) in two outcrops protruding through glacial till 5 mi north-northeast of Crater Mountain (Bundtzen and Laird, 1980). Lithic to sublithic shale, siltstone, sandstone, and conglomerate (Kkss) of the Kuskokwim Group (Cady and others, 1955) of mid- to Late Cretaceous age overlie the older rocks. Intruding and hornfelsing these layered rocks is the 60-mi^2 heterogeneous diorite-to-syenite Beaver Mountains Stock (Kdsm, fig. 2), which has yielded one ^{40}K-^{40}Ar mica crystallization age of 70.1 ± 2.1 m.y. (Bundtzen and Laird, 1980). A 1,600-ft-thick pile of basalt, andesite, crystal tuff, and minor tuffaceous sedimentary rocks (Kbc) overlies the Beaver Mountains Stock in the southern part of the study area. A major northeast-trending, high-angle fault forms the northern topographic boundary of the Beaver Mountains.

The resistant intrusive and extrusive rocks in the Beaver Mountains are very similar to those in other glaciated highlands of southwestern Alaska such as Cloudy Mountain and the Taylor, Horn, and Russian Mountains (Patton and others, 1980; Cady and others, 1955).

BEAVER CREEK GLACIATION[2]

The oldest recognized glaciation in the study area is the Beaver Creek Glaciation, named for the extensive till and glaciofluvial deposits on lower Beaver Creek, well beyond the limits of the Beaver Mountains (Qd, Qdo, fig. 2). During this glaciation, ice occupied at least 78 cirques in the Beaver Mountains, extended down the Ganes, Beaver, Tolstoi, Billy Goat, and Windy Creek drainages, and scoured wide, U-shaped valleys.

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[2]Nomenclature for successive glaciations in this paper is informal and used for comparative and descriptive convenience.

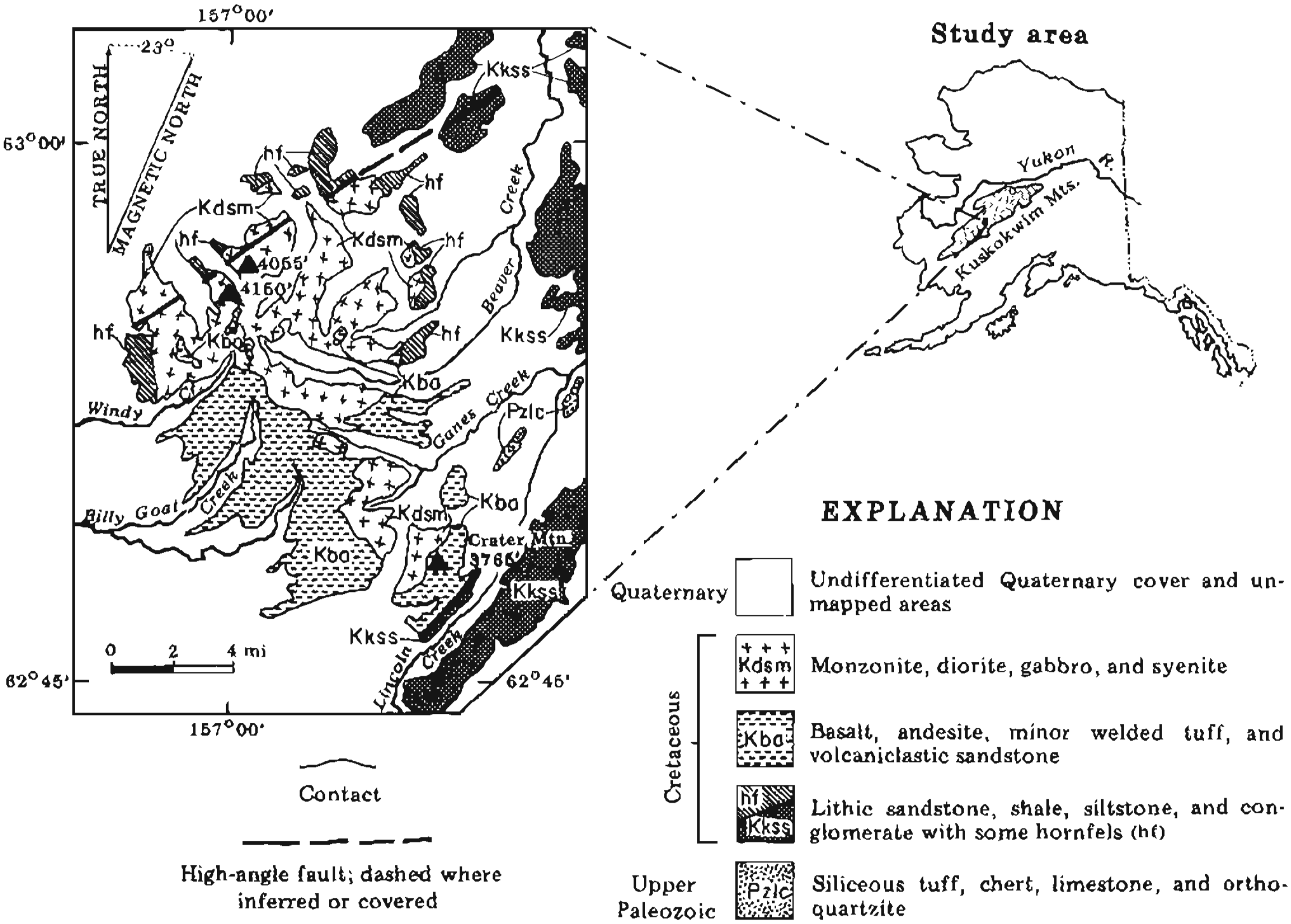


Figure 1. Location map showing generalized bedrock geology of Beaver Mountains, western interior Alaska.

Ice from several trunk glaciers coalesced on the outer flanks of the Beaver Mountains, as evidenced by the broad, till-covered, planated ridges at the 1,600-ft elevation on the eastern, northern, and western limits of the upland. During the Beaver Creek maximum, ice in Ganes Creek overtopped the 2,950-ft divide with Beaver Creek, and a transfluent ice stream flowed into the Beaver Creek drainage. Similarly, Beaver Creek ice breached the 2,700-ft-high divide on its north valley wall and allowed a diffluent ice stream to enter upper Brown Creek. Ice also breached lower divides between Tolstoi, Ganes, Last Chance, and Lincoln Creeks, which resulted in complex diffluent and transfluent ice flow that changed direction several times, depending on the relative vigor of the ice streams. During the Beaver Creek Glaciation, the central part of the massif was a 150-mi^2 ice sheet; only those peaks above 3,000 ft protruded above the wide valley glaciers. Ice advanced down Tolstoi and Beaver Creeks a minimum of 10 and 15 mi, respectively, and reached elevations as low as 900 ft.

Evidence for the Beaver Creek Glaciation consists of a) isolated patches of till (Qd, fig. 2) on planated summits near the limits of the Beaver Mountains, b) faint breaks in slope as high as 3,000 ft along valley walls that are believed to represent ice limits in the Beaver Mountains and adjacent lowlands, c) distribution of ice-marginal meltwater channels far beyond the limits of younger glaciations, and d) drift (Qdo, fig. 2) in the Beaver, Hunter, and Windy Creek valleys. Mertie (1936) described till of probable Beaver Creek age at the confluence of Last Chance and Ganes Creek near Ganes Creek canyon (fig. 2) but the author did not visit this locality. No terminal or lateral moraines are recognized.

Till and outwash deposits of Beaver Creek age are covered with a 3- to 10-ft-thick loess mantle. Numerous lakes and ponds in the Qdo deposits are probably thermokarst features developed in the ice-rich perennially frozen eolian cover rather than youthful kettles. Breaks in slope denoting former ice limits high in the Beaver Mountains logically project downstream to the distribution of Qdo deposits (fig. 2).

Most cirque headwalls of Beaver Creek age have been rescoured by younger glaciations. The rest are extensively modified by mass wasting, and coalescing talus cones and cirque floors are completely filled with rock rubble and alluvium. Slope angles of the cirques approach those of the surrounding valley walls and only a crude ice–scour morphology is retained. Small, northwest-oriented parabolic tributary valleys on the south side of Lincoln Creek may be cirques incised by later streams or they may be old nivation hollows; in either case, their formation is correlated with the Beaver Creek Glaciation.

BIFURCATION CREEK GLACIATION

The Bifurcation Creek Glaciation is named after till in lower Bifurcation Creek valley. Ice advances were confined to wide, U-shaped valleys and downstream till limits extended to about 1,300 ft. Valley glaciers on the east-northeast side of the Beaver Mountains hydrographic divide averaged 6 mi long, whereas glaciers flowing west-southwest averaged only 3-1/2 to 4 mi in length. During the Bifurcation Creek Glaciation, diffluent ice streams crossed divides below the 2,600-ft elevation between Ganes, Last Chance, and Lincoln Creeks, and between Tolstoi and Bifurcation Creeks. Sixty-eight of 78 cirques in the Beaver Mountains were occupied by ice during the Bifurcation Creek Glaciation.

Two and sometimes three well-preserved terminal and recessional moraines are present in almost all stream valleys. These are particularly well preserved in the Tolstoi and Bifurcation Creek valleys, where they have moderate to steep fronts (10° to 15°), are not extensively dissected, and contain 10 to 30 small kettle lakes per mi^2. Lateral limits 250 to 300 ft above the modern streams are well defined along valley walls. Most Bifurcation Creek till deposits have well-developed soil profiles up to 18 in. thick, are covered with rounded igneous boulders 1 to 6 ft in diameter, and have been partially incised by modern streams.

Cirque headwalls slope to 20° and have been modified by talus cones and stream incision, but retain circular ice-scoured morphology. Cirque floors are partially filled with rock rubble derived from these sources.

TOLSTOI LAKE GLACIATION

The Tolstoi Lake Glaciation, named after a 1.1-mi-long morainal-dam lake on the middle fork of Tolstoi Creek, is defined by several arcuate, steep-fronted terminal moraines in all major stream valleys that drain the Beaver Mountains and by lateral moraines about 150 ft above main valley floors. Ice advanced 2-1/2 to 4 mi from cirque headwalls in Tolstoi, Ganes, and Bifurcation Creeks but only 1/2 to 1 mi in the remaining drainages. No divides were breached and ice was confined to valleys of origin. Most cirques on the north-northeast side of the mountains contained ice, but cirque glaciers developed only in those north-facing basins well protected on three sides (14 of 21 cirques) in the Windy-Billy Goat Creeks drainages.

Several morainal-dam lakes up to 3/4 mi^2 in area occur in north-facing valleys near the termini of maximum Tolstoi Lake advances. Tolstoi Lake till contains large, angular boulders of igneous rock up to 15 ft long and bears a dense cover of *Cladonia*. Modern streams are usually diverted around till borders and only develop mature hydrologic gradients downstream from Tolstoi Lake moraines.

Cirque headwalls slope 30° to 40° and, except for a few talus cones, are largely unmodified. Cirque floors appear relatively unmodified and are composed of crude unsorted rock rubble and lichen-rich soil cover.

CRATER MOUNTAIN GLACIATION

The youngest recognized glaciation in the Beaver Mountains is named after a well developed cirque on the north side of Crater Mountain (fig. 2). Evidence of this advance is confined to cirques above 2,450 ft, particularly those on the east-northeast side of the mountains. During the Crater Mountain Glaciation only 15 of 78 cirques in the Beaver Mountains contained ice; only one cirque southwest of the hydrologic divide shows evidence of ice containment. Steep-fronted terminal moraines covered with rock rubble enclose tarns 300 to 1,500 ft in diameter. The 45° to 70° cirque headwalls are relatively unmodified, except for minor talus-cone development. A minimal soil profile and scattered lichen cover are developed on Crater Mountain till.

ROCK GLACIERS

At least nine rock glaciers are present in the Beaver Mountains. These occur as lobate to spatulate masses of poorly sorted angular boulders and smaller rock fragments derived from cirque headwalls. Seven of the nine have rounded, subdued termini and soil profiles several inches thick; black lichens have developed on stabilized bedrock rubble. The seven rock glaciers are located in northwest- or southwest-oriented cirque basins ranging from 2,000 to 2,500 ft in elevation. Morphology and distribution suggests these rock glaciers are inactive and probably related to the Tolstoi Lake or Crater Mountain Glaciations (Qr, fig. 2).

The two remaining rock glaciers in the Beaver Mountains are probably active—one in a north-facing cirque on upper Ganes Creek at 2,620 ft and the other on the northeast face of Mountain 3957 at 2,650 ft (Qr*, fig. 2). Both are lobate masses with fronts sloping more than 45°; dark lichen covers the upper surfaces. The unvegetated, light-toned, sloping margins result from movement of angular basalt-andesite rubble. Clear water seeping from the terminus of the Ganes Creek rock glacier has filled a 5-acre pond near the cirque headwall; this water may be derived from melting interstitial ice.

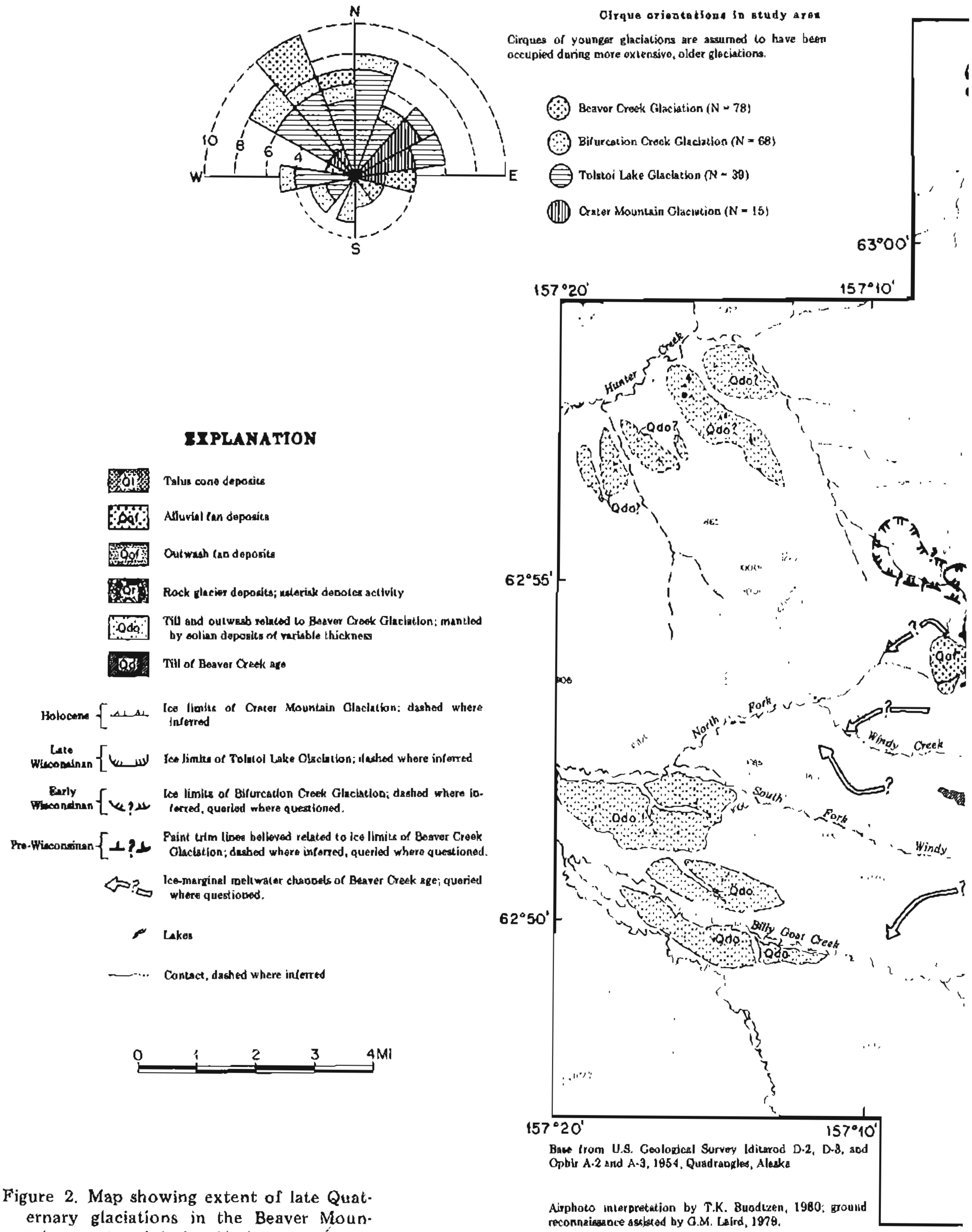
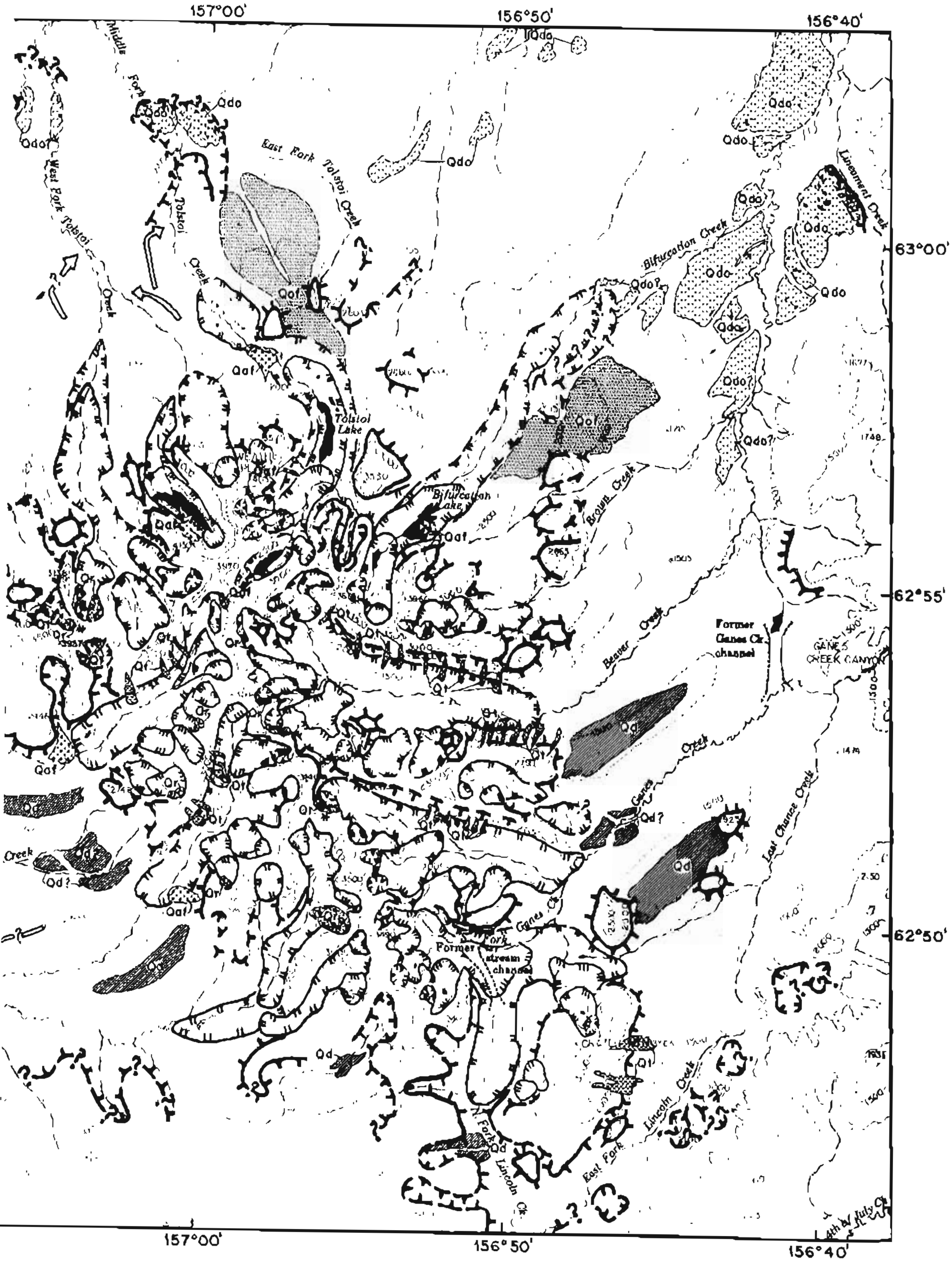


Figure 2. Map showing extent of late Quaternary glaciations in the Beaver Mountains, western interior Alaska.


157°00'
156°50'
156°40'
63°00'
62°55'
62°50'
Middle Fork
West Fork Tolstoi
Tolstoi Creek
East Fork Tolstoi Creek
Lineament Creek
Bifurcation Creek
Tolstoi Lake
Bifurcation Lake
Brown Creek
Beaver Creek
Former Ganes Cr. channel
GANES CREEK CANYON
Ganes Creek
Last Chance Creek
Ganes Cr.
Former stream channel
Lincoln Creek
N. Fork Lincoln Cr.
East Fork
4th of July Cr.
Qdo
Qaf
Qd
Qr

AGE AND CORRELATION

No radiometric or paleontological age control is available for glaciations in the study area, and reconstruction of a Beaver Mountains glacial chronology relies on relative-age-dating techniques presented by Birkland and others (1979). Table 1 correlates glaciations of the Beaver Mountains with those described by other workers in selected regions of southwestern and western Alaska.

The absence of morainal landforms, the extensively modified character of till, outwash, and cirque headwalls, and the deep stream incision in valleys occupied by Beaver Creek ice are features similar to those of pre-Selatna (Fernald, 1960) and Clara Creek (Porter, 1967) Glaciations. Both are regarded as Illinoian.[3] Characteristics of the Bifurcation Creek Glaciation include morainal landforms rounded by mass wasting and incised by stream erosion, soil profiles up to 18 in. thick, kettles filled with *Sphagnum* peat, and subangular to rounded till boulders with minor oxidation. These features are similar to those of the Selatna Glaciation[3] near Farewell (Fernald, 1960), Porter's (1967) Chagvan Bay Glaciation, the Indian Mountain Glaciation near Hughes (Reger, 1978), and the Mak Hill Glaciation in the Iliamna Quadrangle (Detterman and Reed, 1973); all are regarded as early Wisconsinan.

Advances of the Tolstoi Lake Glaciation are believed to be late Wisconsinan for several reasons: a) well-preserved glacial landforms, b) poor development of soil cover, c) incomplete incision by modern streams, d) presence of water-filled kettles and larger morainal-dam lakes, e) existence of large, angular till boulders, and f) minor modifications to cirque headwalls. The Tolstoi Lake advances are correlated with stades of the Farewell Glaciation (Fernald, 1960), the Brooks Lake Glaciation (Detterman and Reed, 1973), or the Unaluk Glaciation near Goodnews Bay (Porter, 1967).

The unmodified moraine morphology, lack of mature soil development, presence of fresh, angular till boulders, and general absence of cirque headwall modification are characteristic of the Crater Mountain Glaciation. Similar features are described by Karlstrom (1964) and Detterman and Reed (1973) for the Alaskan Glaciation of post-Wisconsinan to Holocene age. Active(?) rock glaciers in two Beaver Mountains cirques suggest near-threshold conditions for reinception of glaciation in the study area above 2,600 ft.

[3] Fernald (1960) concludes that the age of his Selatna Glaciation is problematical—either Illinoian or early Wisconsinan. He regards the two Farewell advances as late Wisconsinan. Péwé (1975) has shown the Selatna Glaciation as Illinoian on his statewide correlation chart and the two advances of Farewell Glaciation as early and late Wisconsinan. However, J.T. Kline (oral comm., 1980) believes the Selatna Glaciation may be synchronous with the Delta Glaciation in the eastern Alaska Range. Cumulative evidence suggests that the Delta Glaciation is early Wisconsinan (R.D. Reger, oral comm., 1980).

CIRQUE ANALYSIS

Orientation data of cirques of the Beaver Creek, Bifurcation Creek, Tolstoi Lake, and Crater Mountain Glaciations (table 2, fig. 2) demonstrate that cirques of progressively younger glaciations assume a more northerly orientation with dominant mean direction of each glaciation varying from 13° to 37°. Statistical data on cirque orientations from the Bifurcation Creek and Beaver Creek Glaciations indicate polymodal distributions in the northwest, northeast, and southwest directions, whereas cirque orientations of the Tolstoi Lake and Crater Mountain advances show bimodal distributions to the northeast and northwest. Thus, the mean vectors for the different glaciations are the average of several distinct cirque orientations from each glaciation, as shown by the R values and mean angular deviations (table 2). Cirque orientation data imply that orographic influences had less effect during the older, more extensive glaciations than during the younger, more confined ice advances. By Crater Mountain time, almost all active cirques were confined to protected northerly basins.

Beaver Mountains cirque levels range from 2,150 to 2,900 ft and average 2,495 ft in elevation (fig. 2, table 2), a level almost 1,500 ft lower than cirques described in the Ray Mountains (Yeend, 1971) and the Yukon-Tanana Upland (Péwé, Burbank, and Mayo, 1967). Low cirque levels in the Beaver Mountains during Pleistocene time can be attributed to higher precipitation rates or a cooler climate or both. Cady and others (1955) describe cirque basins of unknown age at elevations as low as 1,500 ft in the Taylor Mountains 150 mi southwest of the Beaver Mountains.

POSTGLACIAL READJUSTMENTS

After ice receded from the Beaver Mountains, talus cones, outwash fans, and alluvial fans began to modify slope morphology. Subsequent to recession of Tolstoi Lake ice, small tributary streams deeply incised valley walls in Tolstoi and Bifurcation Creeks and deposited alluvial fans over valley floor moraines. This process resulted in the division of the formerly 1-mi-long 'Bifurcation' Lake (fig. 2). Talus cones modified valley slopes and cirque headwalls subsequent to all major glaciations.

Deposition of till and outwash in several stream drainages resulted in stream piracy and sharp drainage deflections. Mertie (1936) first described the stream capture of the headward reaches of the preglacial south fork of Beaver Creek by Ganes Creek. During the Beaver Creek and Bifurcation Creek Glaciations, drift from upper Beaver Creek built up along the valley floor next to the low, rounded hills 10 mi east of the Beaver Mountains. Beaver Creek diverted around this material and incised a steep-walled canyon in the upper water-

Table 1. *Correlation chart of local Quaternary glacial sequences in the Beaver Mountains and selected glacial chronologies in western and southwestern Alaska.*

Age of glaciation	Farewell area, southern Alaska Range (Fernald, 1960)	Chagvan-Goodnews Bay area (Porter, 1967)	Iliamna Quadrangle (Detterman & Reed, 1973)	Beaver Mountains (this study)	Indian Mountain (Reger, 1978)
Holocene			Alaskan Glaciation (Tunnel Stade and Tustumena Stade)	Crater Mountain Glaciation rock glaciers	
Late Wisconsinan	Farewell II Glaciation Farewell I Glaciation	Unaluk Glaciation	Brooks Lake Glaciation (Kuichak Stade, Iliamna Stade, Newhalen Stade, Iniuk Stade)	Tolstoi Lake Glaciation (3 unnamed advances) rock glaciers	rock glaciers
Early Wisconsinan	Selatna Glaciation	Chagvan Bay Glaciation	Mak Hill Glaciation (Kukuklek Stade)	Bifurcation Creek Glaciation (2 unnamed advances)	Indian Mountain Glaciation
Illinoian	glaciation(?)	Clara Creek Glaciation		Beaver Creek Glaciation (1 unnamed advance)	Sleepy Bear Glaciation
Pre-Illinoian		Kemuk Glaciation			

Table 2. *Summary of cirque measurements from Beaver Mountains, Iditarod Quadrangle. (Orientation statistics after Reyment, 1971)*

Glaciation	Number of cirques measured (N)	Mean orientation vector (°)	Distance from origin (R value)[a]	Mean angular deviation (°)	Average elevation (ft)
Crater Mountain	15	39	0.6444	48	2,650
Tolstoi Lake	39	15	0.6212	49	2,583
Bifurcation Creek	68	13	0.4193	61	2,516
Beaver Creek	78	22[b]	0.3100[b]	68[b]	2,495[b]

[a]R value, or correlation coefficient, may vary from a value of 1.000 (unimodal population) to 0.000 (random population).
[b]Average for all cirques in study area.

shed of preglacial Ganes Creek to produce the present Ganes Creek valley. Drift from a Bifurcation Creek advance blocked the upper north fork of Lincoln Creek and caused the stream to divert along the north side of the moraine into Ganes Creek.

ECONOMIC IMPLICATIONS

Gold placers on lower Ganes Creek immediately east of the study area may have been derived from a localized bedrock source or from copper-silver-tourmaline-bearing vein faults that cut alkaline intrusive rocks of the Beaver Mountains Stock on Ganes and upper Beaver Creeks (Bundtzen and Laird, 1980). If the mature auriferous bench and stream placers on Ganes Creek are derived from the latter, they must have formed since Beaver Creek time, inasmuch as Ganes Creek captured an upper fork of Beaver Creek after the Beaver Creek till and outwash were deposited.

It is more plausible that the auriferous benches on Ganes Creek formed before the Beaver Creek Glaciation and that their source is a structurally controlled, mineralized rhyolite-basalt dike swarm trending northeast through upper Ganes and Yankee Creek mapped by Bundtzen and Laird (1980).

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FOSSIL ALGAE IN LOWER DEVONIAN LIMESTONES, EAST-CENTRAL ALASKA

By James G. Clough[1]

INTRODUCTION

Two genera of calcareous fossil algae, *Girvanella* and *Renalcis*, have been identified in thin sections of rocks from the Ogilvie Formation located in the Charley River Quadrangle, east-central Alaska and at one locality in the Yukon Territory (fig. 1). These fossil algae are important paleoenvironmental indicators of shallow-shelf lagoonal (*Girvanella*) and reef or bank-edge (*Renalcis*) deposition within a carbonate platform. Strata of the Ogilvie Formation were deposited during Early Devonian time on the southern part of the Yukon Stable Block (Lenz, 1972), an area of shallow-water carbonate sedimentation throughout much of the early Paleozoic.

CHARACTERISTICS AND PALEONTOLOGY

GIRVANELLA

Girvanella consists of tubular filaments 7 to 30 microns in diameter entwined in loose irregular masses (fig. 2) and encrusting or perforating fossil fragments (Machielse, 1972). The filament walls are fine grained and dark in thin section. Differentiation of species within this genus is based on wall thickness and tube diameter. The *Girvanella* filaments shown in figure 2 have a tube diameter of 22.8 to 26.6 microns and a wall thickness of 5 to 7 microns.

The genus *Girvanella*, established by Nicholson and Etheridge (1880), is considered a blue-green (Schizophyta) alga referred to the Family Porostromata on the basis of similar morphologic characteristics (Machielse, 1972).

It commonly occurs in a lagoonal back-reef setting (Wray, 1972) in limestones "representing a quiet, slightly restricted subtidal environment" (Machielse, 1972, p. 214). *Girvanella* has also been observed in voids between stromatoporoid skeletons in Devonian reefs (Konishi, 1958). It is found in rocks of Cambrian to Cretaceous age.

RENALCIS

Renalcis consists of chambered algae ranging in size from 30 to 300 microns in diameter that occur in grape-like clusters (fig. 3) in a fan-shaped pattern (Machielse, 1972).

The genus lacks diagnostic characteristics readily related to living algae, and *Renalcis* has been referred to the red algae (Rhodophyta) by Vologdin (1962) and to the blue-green algae (Schizophyta) by Johnson (1964) and Wray (1967), and is more recently considered a problematical blue-green algae (Machielse, 1972; Wray, 1977).

Renalcis commonly grew in "voids beneath laminar stromatoporoids, and separately in lime-mudstone intervals" (Cheshire and Keith, 1977, p. 30), where they often formed the core of Devonian reefs. Wray (1972) reports that their presence indicates deposition in reef and bank-edge environments. The genus is reported from Cambrian- through Devonian-age limestones.

DISCUSSION

Fossil *Girvanella* is present in some intervals of limestone from all three localities shown in figure 1. At localities 1 and 2, *Girvanella* occurs in association with abundant crinoidal debris and less commonly with ostracod, brachiopod, and trilobite fragments. At locality 1 this alga occurs with the ostracod *Moellerita canadensis* (identified by R.B. Blodgett, 1978), which has been described in restricted paleoenvironments within the Ogilvie Formation by Lenz (1972). At locality 3, *Girvanella* is found in interskeletal voids between colonial corals and stromatoporoids. This occurrence is similar to that reported by Konishi (1958), and suggests a shelf-margin biohermal paleoenvironment.

The fossil alga *Renalcis* is present in one interval of limestone from locality 2, and a single intraclast containing this genus was observed in a sample of limestone from locality 3. The genus occurs in a 1.8-m interval of *Renalcis* lime-mudstone capped by the tabular stromatoporoid *Trupetostroma* (identified by R.B. Blodgett, 1978) at locality 2, indicating a biohermal environment (Cheshire and Keith, 1977). There, the limestones form a part of the 30-m-thick massive reef complex observed 0.6 km to the southwest by Blodgett (1978).

These genera of fossil algae are the westernmost recognized in upper Emsian strata of the region. Their presence and the associated biota that indicate a shelf-margin and shallow-shelf environment agree with the findings of Blodgett (1978) on the Ogilvie Formation in this area.

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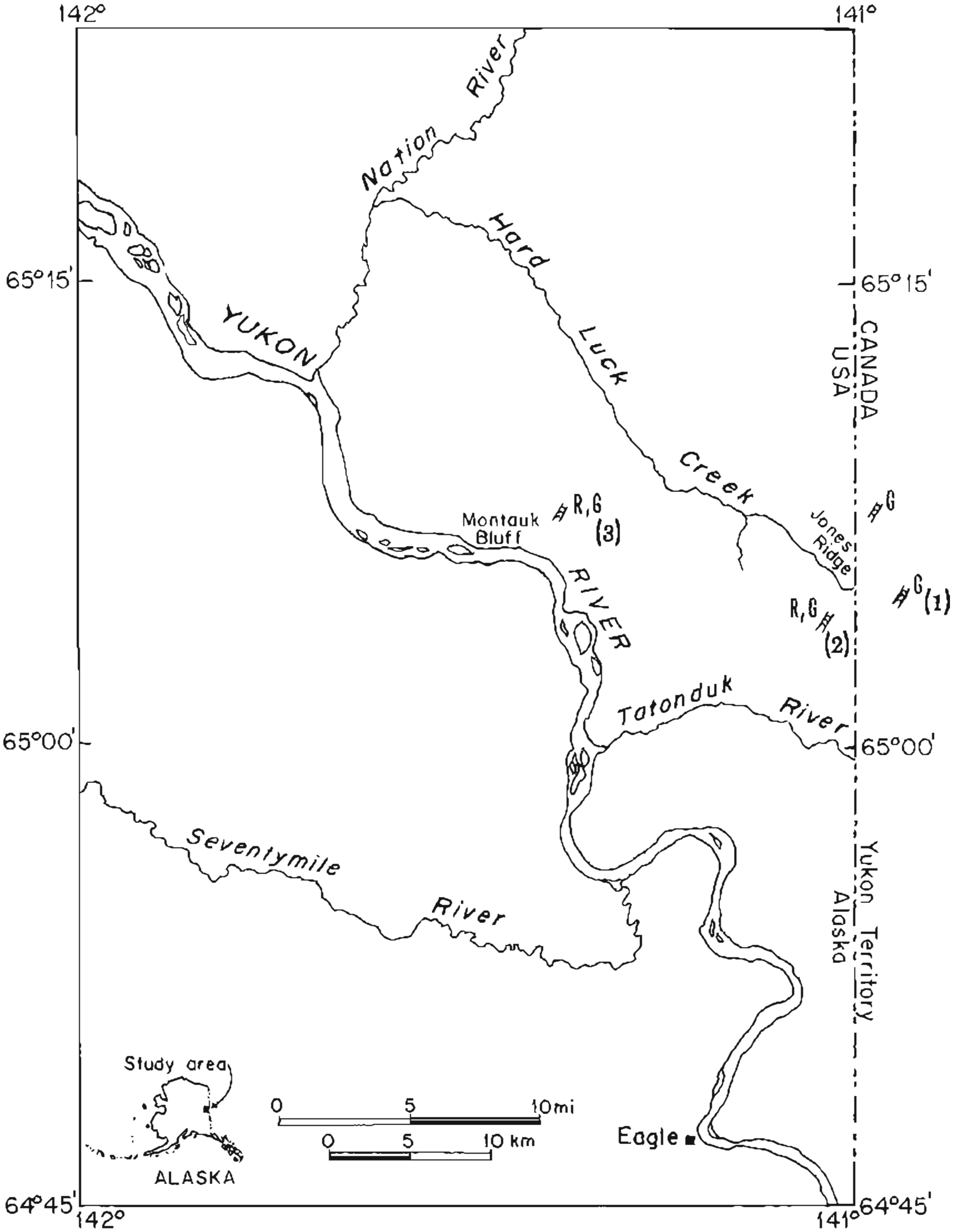

Figure 1. Map showing microfossil localities in east-central Alaska (G - *Girvanella*, R - *Renalcis*).
Locality 1 - 1.4 km east of International Boundary in Yukon Territory, Canada.
Locality 2 - NE 1/4 of NW 1/4 sec. 22, T. 3 N., R. 33 E., Charley River A-1 Quadrangle.
Locality 3 - Center of sec. 34, T. 4 N., R. 31 E., Charley River A-1 Quadrangle.

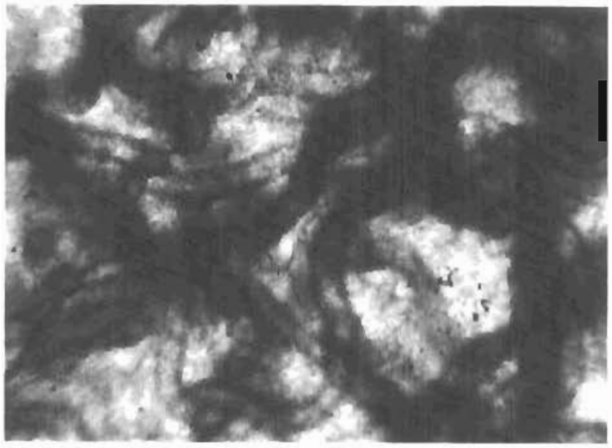

Figure 2. Photomicrograph of entwined *Girvanella* filaments in micrite (X160)

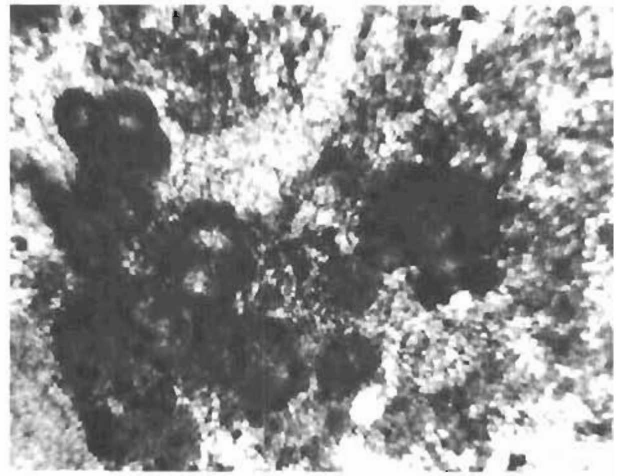

Figure 3. Photomicrograph of grapelike clusters of *Renalcis* surrounded by micrite and microspar. Dark grains are hematite (X75).

ACKNOWLEDGMENTS

The author gratefully thanks C.W. Allison of the University of Alaska Museum and W.G. Gilbert of DGGS for reviewing the manuscript. The fossils shown in figures 2 and 3 are on deposit in the University of Alaska Museum, Fairbanks.

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TERTIARY TILLITES(?) ON THE NORTHEAST FLANK OF GRANITE MOUNTAIN, CENTRAL ALASKA RANGE

By L. David Carter[1]

INTRODUCTION

Tilted beds of moderately consolidated pebble, cobble, and boulder conglomerate compose three ridges west of the Gerstle River on the northeast flank of Granite Mountain, central Alaska Range (figs. 1 and 2a). These strata are especially interesting because some of the beds contain boulders as large as 2 m in diameter and may be Tertiary glacial deposits, and some boulders are of lithologies not present in modern drainage basins adjacent to the ridges. Knowledge of the depositional and postdepositional history of these strata is important to understanding a) the fault movement along the northeast side of Granite Mountain, b) the uplift of Granite Mountain and the Alaska Range, c) regional drainage development, and d) the age of initial late Cenozoic alpine glaciation in the Alaska Range.

The conglomerate beds were first noted by Moffit (1942, 1954), who correlated them with the Nenana Gravel, but he did not mention the presence of large boulders. Other deposits in the vicinity that have been correlated with the Nenana Gravel are on Independent Ridge (Holmes and Foster, 1968), at the northernmost corner of Granite Mountain (Holmes and Péwé, 1965), and southwest of Granite Mountain near McCumber and Jarvis Creeks (Moffit, 1942, 1954). The term 'Nenana Gravel' was first applied by Capps (1912) to a series of beds that is exposed along part of the Nenana River and that is widespread on the north flank of the Alaska Range. These beds are generally considered to predate glaciation in the Alaska Range and to have been deposited by streams rejuvenated during initial uplift of the range (Capps, 1940). The most recent estimate of the age of the Nenana Gravel is late Miocene and early Pliocene (Wolfe and Tanai, 1980, p. 9).

DESCRIPTION

The conglomerate ridges extend about 2 km north from the steep margin of Granite Mountain. The maximum elevation on the ridge crests is about 880 m, whereas elevations of as much as 1,800 m occur on the gently rolling but deeply dissected erosion surface at the top of Granite Mountain. The contact between the tilted beds and the quartz and quartz-mica schist that make up this part of Granite Mountain is not exposed. However, an extensive crushed zone occurs near the base of the range-front scarp about 9 km to the northwest, and the contact between the sedimentary beds and the schist is inferred to be a fault (fig. 1).

Bedding in the conglomerate is emphasized by differential erosion. Beds richest in boulders are more resistant than adjacent strata and stand in relief on the flanks and crests of the ridges (fig. 2). The beds strike northwest, parallel to the range front, and dip northeast at 40° to 60°. In general, the beds could not be traced from ridge to ridge and the presence of unconformities or faults within the section could not be determined. However, one massive bouldery bed that occurs at about the middle of the section is traceable across all three ridges and apparently thins to the northwest.

Measurement of the strata of the central ridge demonstrated that the total thickness exceeds 1,000 m (fig. 3). In the lower 325 m of the section only one boulder-bearing bed occurs. It is 20 m thick and occurs 145 m above the base of the section. Boulders are present in the upper 675 m of the section except in two zones: a) a 60-m-thick zone that is 580 m above the base of the section and b) a 90-m-thick zone 710 m above the base. None of the beds are well exposed and sedimentary structures, aside from the gross differentiation of beds due to differential erosion, were not observed.

Boulder-free beds consist of pebble to cobble conglomerate. Clasts are predominantly hard, fine-grained siliceous rocks, including quartzite, quartz, chert, and chert conglomerate, but volcanic and metamorphic rocks and felsic to mafic intrusive rocks are also present. Clasts are well rounded and generally have orange-colored iron oxide coatings. Weathering rinds are as thick as 2 to 3 mm.

In boulder-bearing beds, clasts are as much as 2 m in longest dimension, and a length of 1 m is common (fig. 2b). The largest boulders are generally concentrated in units from 10 to 20 m thick, except for one massive 140-m-thick zone in the center of the section in which boulders are especially abundant. Boulder shapes range from angular to rounded; subrounded clasts are most common. Boulder lithologies include schist, quartz, felsic to mafic intrusive rocks, gneiss, and a distinctive porphyritic granitic rock in which phenocrysts are as

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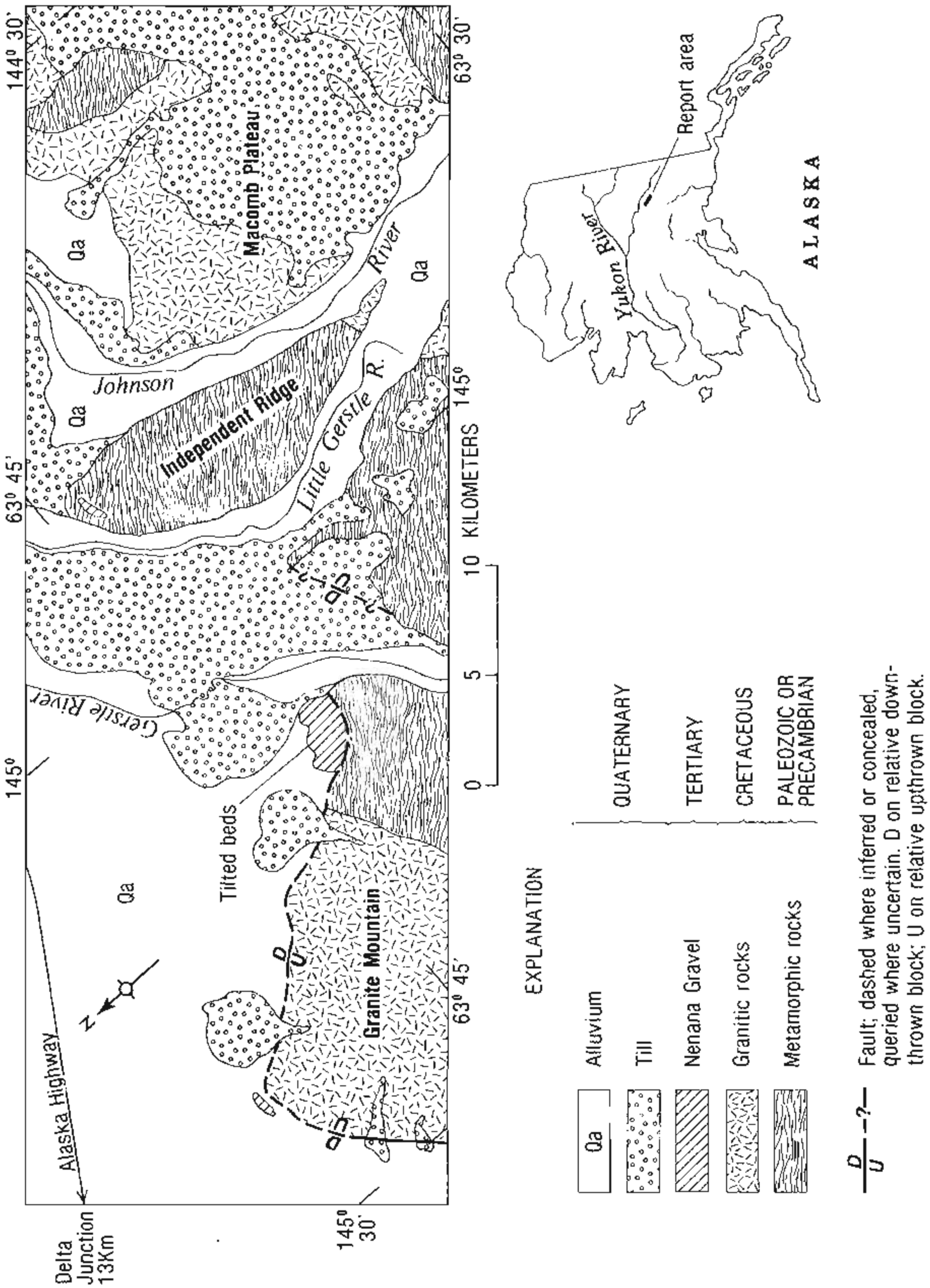

Figure 1. General geology of Granite Mountain area (modified from Holmes and Péwé, 1965; Holmes and Foster, 1968; and Moffit, 1954) and location of tilted conglomerate beds.

Figure 2a. View toward east from Granite Mountain showing tilted beds in center, Gerstle River beyond, and Independent Ridge in background.

Figure 2b. Boulders exposed on slope of central ridge. Shovel near center of photograph is about 1 m long.

long as 3 cm. Rounded pebbles and cobbles like those described for the boulder-free beds constitute much of the pebble- to cobble-size material in the lower two-thirds of the section; in the upper one-third these clasts are less common, and angular to subrounded clasts of quartz schist and mica schist are abundant. The sandy matrix contains abundant angular grains.

The state of weathering of the boulders ranges from relatively fresh clasts that are difficult to break with a hammer to thoroughly rotted ones that disintegrate easily under a few hammer blows. Boulders on the surface appear relatively fresh, whereas those recently exposed in a soil failure are highly weathered, and some of the granites are grusified. Apparently, boulders in the soil zone at this locality weather more rapidly than those on the surface, perhaps because of prolonged contact with soil moisture. Striations were not observed on clasts of any size. However, no original surfaces may remain.

SIGNIFICANCE

I agree with Moffit (1942) in correlating the boulder-free beds with the Nenana Gravel and agree that they were deposited under nonglacial, fluvial conditions. However, the presence of large boulders in the rest of the beds implies either a powerful transporting mechanism (glaciers, debris flows, torrential streams) or a nearby source terrane of high relief, or both. The variety of boulder lithologies suggests either long-distance transport or a nearby heterogeneous source terrane. Most of the rock types composing the boulders occur as dikes and small intrusive bodies in the metamorphic terrane south of the ridges, but a notable exception is the porphyritic granitic rock. The nearest exposures of granitic rocks are 5 km northwest on Granite Mountain and 16 km southeast on Independent Ridge and Macomb Plateau (fig. 1). The only granitic pluton known to occur in the Gerstle River drainage is about 25 km upstream from the valley mouth. Porphyritic granite is present in parts of the pluton, but the maximum phenocryst size observed is about 1.2 cm, and this pluton is therefore not believed to be the source of the porphyritic granitic rock in the tilted beds. The problem of the source of these granitic boulders is intimately related to the mode of deposition, relation to faulting, and regional drainage development. Unfortunately, the source cannot be determined with the data at hand, but speculations serve to emphasize the importance of the tilted beds to an understanding of regional geologic history.

If Granite Mountain were the source of the granitic boulders and if the beds were deposited during uplift of the range and initiation of the present drainage system, the boulder-bearing beds must have been deposited at the mouth of a valley that is 5 km northwest of the ridges, because this is the only valley draining the northeast side of Granite Mountain in which both granitic and metamorphic rocks are exposed. If this valley were the source of the granitic clasts, 5 km of right-lateral movement must have occurred along a fault at the northeast margin of the range after deposition of at least 1,000 m of gravel to produce the present separation between the deposits and their source. However, movement along the Granite Mountain fault, which occurs along the northwest side of Granite Mountian, has been dominantly normal (Holmes and Péwé, 1965; Hudson and Weber, 1977). The fault along the northeastern edge of the range is presumably part of the same fault system, and 5 km of right-slip is considered here to be improbable because of the style of normal offset.

Derivation of the boulders from Granite Mountain before uplift of the range is also not considered likely. The subdued erosion surface on Granite Mountain indicates that this was an area of low relief before uplift of the range, and if the boulders had been transported across this surface from the northwest, residual

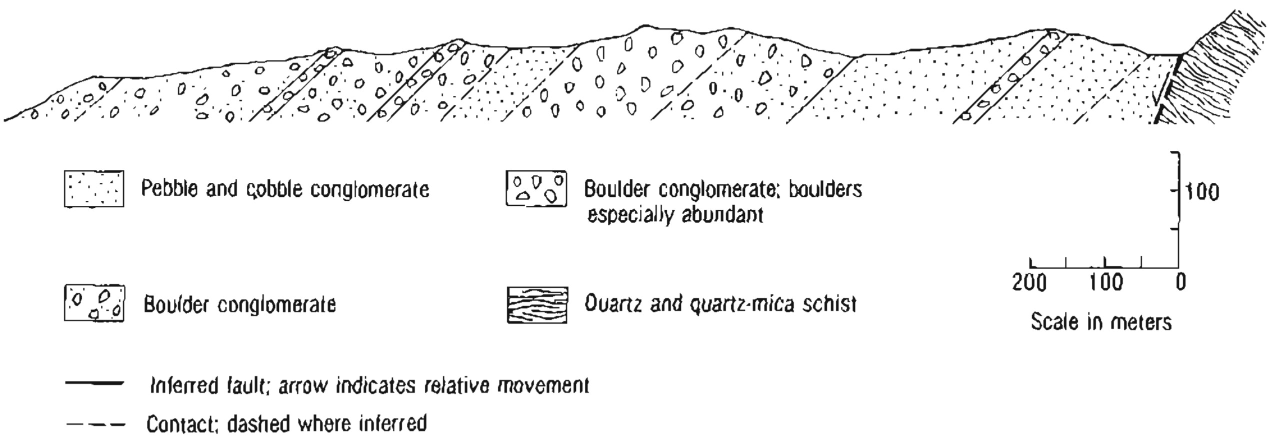

Figure 3. Longitudinal section through central ridge.

boulders should occur on this ancient surface adjacent to the tilted beds. However, neither the Nenana Gravel nor granitic boulders have been observed there.

A more likely hypothesis is that the source for the porphyritic granitic boulders was to the southeast, near Independent Ridge and Macomb Plateau (fig. 1). This hypothesis requires deposition of the boulder-bearing beds before development of the modern drainage system. The minimum distance of 16 km between the 2-m-diameter boulders and their potential source argues against either torrential streams or debris flows as the mechanism of transport, unless an extremely high range-front scarp was present in the granitic terrane. Therefore, ice most likely played a role in the emplacement of the boulders, either in the form of ice rafting or glacial ice.

Ancient glaciations previously recognized in the Alaska Range (Péwé and others, 1953; Wahrhaftig, 1958) were considered to be younger than the Nenana Gravel and are probably Pleistocene, but Tertiary glacial deposits have been identified elsewhere in Alaska (Plafker and Addicott, 1976; Hamilton, 1979). If the boulder-bearing beds are tills, the Nenana Gravel includes glacial deposits at this locality and the tilted strata may record the inception of late(?) Tertiary glaciation in the Alaska Range and document a change from generally nonglacial conditions to a glaciated Alaska Range.

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EVIDENCE FOR SUPRAPERMAFROST GROUND-WATER BLOCKAGE, PRUDHOE BAY OIL FIELD, ALASKA

By Gail D. March[1]

INTRODUCTION

Permafrost in gravel pads and roads at the Prudhoe Bay oil field blocks the flow of ground water. The resultant dammed water forms ponds, creating a potential threat to the facilities.

PHYSIOGRAPHIC DESCRIPTION

The Prudhoe Bay oil field is located on the Arctic Coastal Plain of Alaska (fig. 1), a physiographic province characterized by a low-relief tundra landscape cut by rivers flowing from the Brooks Range north to the Beaufort Sea. The tundra surface is underlain by complex ice-rich sediments of Pleistocene age (Sellman and others, 1975). Lakes and drained lake basins make up from less than 25 percent to more than 90 percent of the tundra surface (Brewer, 1958a). Much of the tundra surface is characterized by high- and low-center polygons ranging from 2 to 30 m in diameter (Péwè, 1966).

AIRPHOTO COMPARISONS

Black-and-white aerial photographs (scale 1:18,000) of the Prudhoe Bay area taken July 20, 1949 were examined to determine the extent of lakes and flooded polygons. A representative sample of the area covered is shown in figure 2. The discovery well at Prudhoe Bay was completed in 1968 (Morgridge and Smith, 1972). Within the next 5 years, two airstrips, more than 25 gravel pads for wells and support facilities, and many miles of connecting roads were built. A photomosaic (scale 1:24,000) taken in July 1973 was examined for changes in the extent of lakes and flooded polygons since 1949. The low quality of the mosaic made precise spatial determinations difficult, but it was possible to see that some formerly drained lake basins next to gravel pads and roads had flooded since 1949.

By July 1979, more than 30 gravel pads existed in the eastern half of the area alone, and there was a corresponding increase in connecting roads. Color airphotos (scale 1:18,000) taken July 13, 1979 were examined for further evidence of refilled lake beds and flooded polygons. Figure 3 shows the extent of ponded water in the same representative area as figure 2.

A comparison of figures 2 and 3 shows that in 30 years, many changes occurred in the area which now contains the main support facilities for production wells in the Prudhoe Bay oil field. A former lake basin next to the road southwest of Lake Colleen has refilled and polygons near the road south and east of the lake have flooded. An extensive area of ponded water has formed west of the Sagavanirktok River, surrounding numerous gravel pads. Smaller areas of flooded polygons in the vicinity of gravel pads and roads are also visible in figure 3.

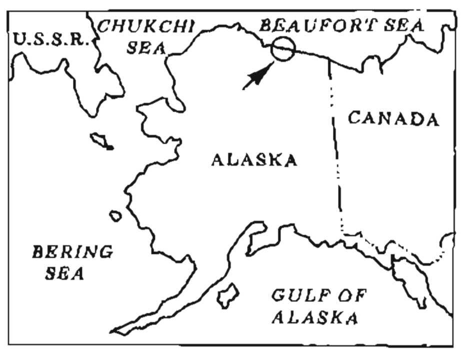

Figure 1. Location of Prudhoe Bay oil field, Alaska.

DISCUSSION

Refilled lake basins and greatly increased areas of flooded polygons appear to have a direct correlation with the presence of gravel fill. A hypothesis that ponded water next to the fill could be a result of drifted snow or blown dust was rejected. Benson and others (1975) state that storm winds responsible for drifted snow are from the west, causing snow accumulation on the east side of gravel roads, and that summer dust is transported by prevailing northeasterly winds, causing deposition on the west side. Ponded water appears to bear no relation to storm or prevailing winds. Instead, its orientation with respect to gravel fill appears to depend on topography, with water ponding on the upslope side of the fill.

In the Prudhoe Bay area, gravel fill 1.5 m thick is used for roads dynamically loaded by vehicles, whereas fill 1 m thick is used on statically loaded well and support facility pads to prevent thaw of underlying perma-

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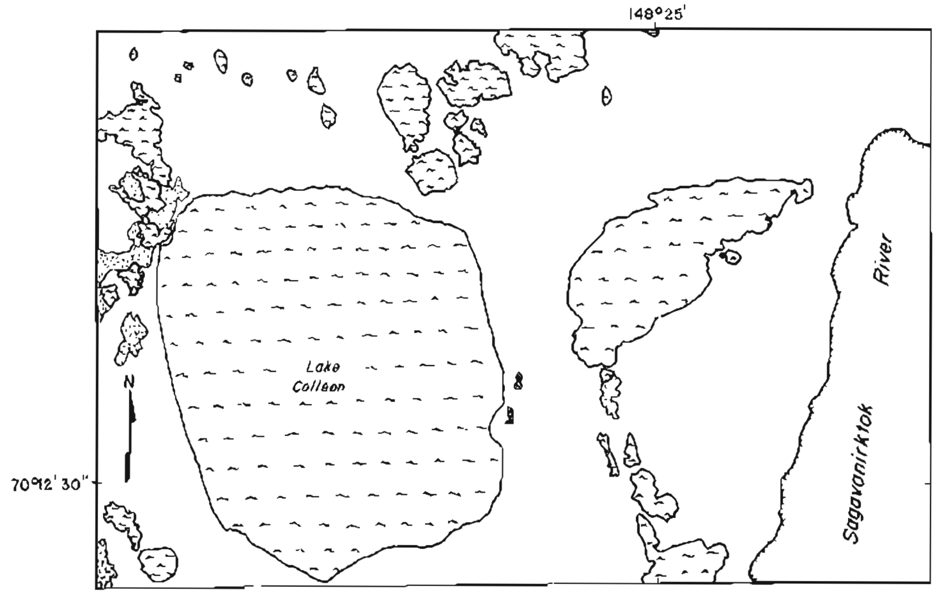

Figure 2. Extent of inundated polygons in part of Prudhoe Bay oil field, Alaska, July 20, 1949.

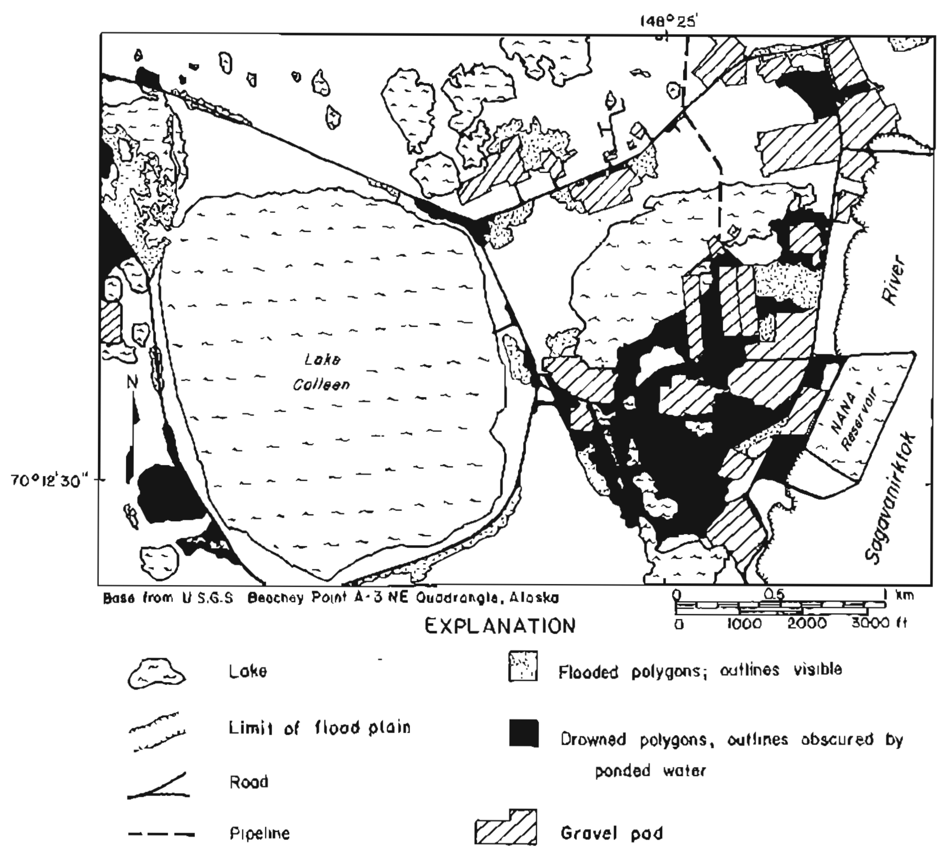

Figure 3. Extent of inundated polygons in part of Prudhoe Bay oil field, Alaska, July 13, 1979.

frost (R.G. Updike, pers. comm., 1980). Because the fill has a greater insulating value than the original active layer, the permafrost table rises (fig. 4). An impermeable barrier of permafrost is then created within the fill (Muller, 1945), which obstructs the flow of water, forcing it to the surface (Federal Energy Regulatory Commission, 1979; Curran and Etter, 1976). This is the probable mechanism for ponding of water adjacent to gravel pads and roads in the Prudhoe Bay area.

EFFECTS OF OBSTRUCTION OF GROUND-WATER FLOW

Curran and Etter (1976) suggest that damming of suprapermafrost ground water by the frozen impermeable barrier beneath the roadway can cause ground icings. When seasonal frost penetrates the saturated active layer, water expelled to the ground surface freezes in layers. They suggest various measures for mitigating the icing problems, including deliberately creating icings away from the road.

Ponded water warms the underlying permafrost. Lakes less than 2 m deep freeze to the bottom each year and are underlain by a seasonally thawed zone 0.3 to 0.7 m thick. Permafrost temperatures at depth under these shallow lakes are up to 3° C warmer than temperatures under the surrounding tundra (Brewer, 1958b). Lakes deeper than 2 m do not freeze to the bottom and develop a perennial thaw bulb (Sellman and others, 1975). As ponds expand, warming progresses downward and outward. Melting of ice-rich permafrost can ultimately cause subsidence of gravel fill adjacent to the ponds. If ponded water begins to flow laterally along gravel fill, both mechanical and thermal erosion can result (Ferrians and others, 1969).

CONCLUSIONS

Permafrost in gravel fill at the Prudhoe Bay oil field is damming suprapermafrost ground-water movement, as evidenced by ponded water adjacent to gravel roads and pads. The increased ponding coincides with the extent of fill deposited between 1973 and 1979. As more roads and pads are built, ground-water blockage will increase.

Ground-water damming and subsequent ponding may cause both ground icings and mechanical and thermal erosion of adjacent fill. Existing fill should be monitored for evidence of these problems. In new construction, elimination of ground-water blockage should be attempted before damming becomes a problem. Surface drainage has been facilitated by use of culverts, but ground-water drainage patterns are more difficult to determine. Potential damming problems require careful study in the early stages of development planning.

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James R. Riehle and Randall G. Updike provided helpful discussions and were very helpful in reviewing the manuscript. Constructive criticism was provided by Richard D. Reger. The 1973 photo mosaic was obtained by Updike from Sohio-BP, and the 1979 airphotos were obtained from ARCO.

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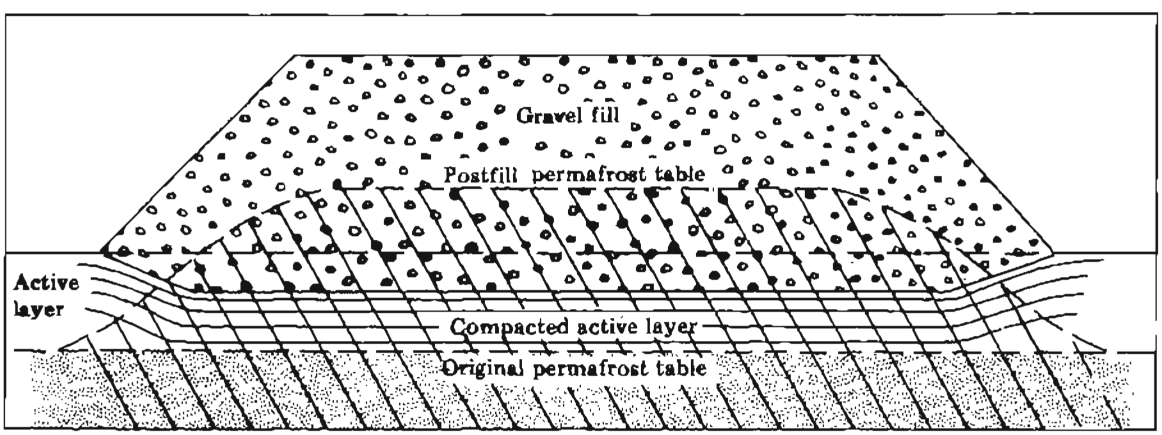

Figure 4. Effect on permafrost table of fill with insulating value greater than that of original active layer (from Ferrians and others, 1969).

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