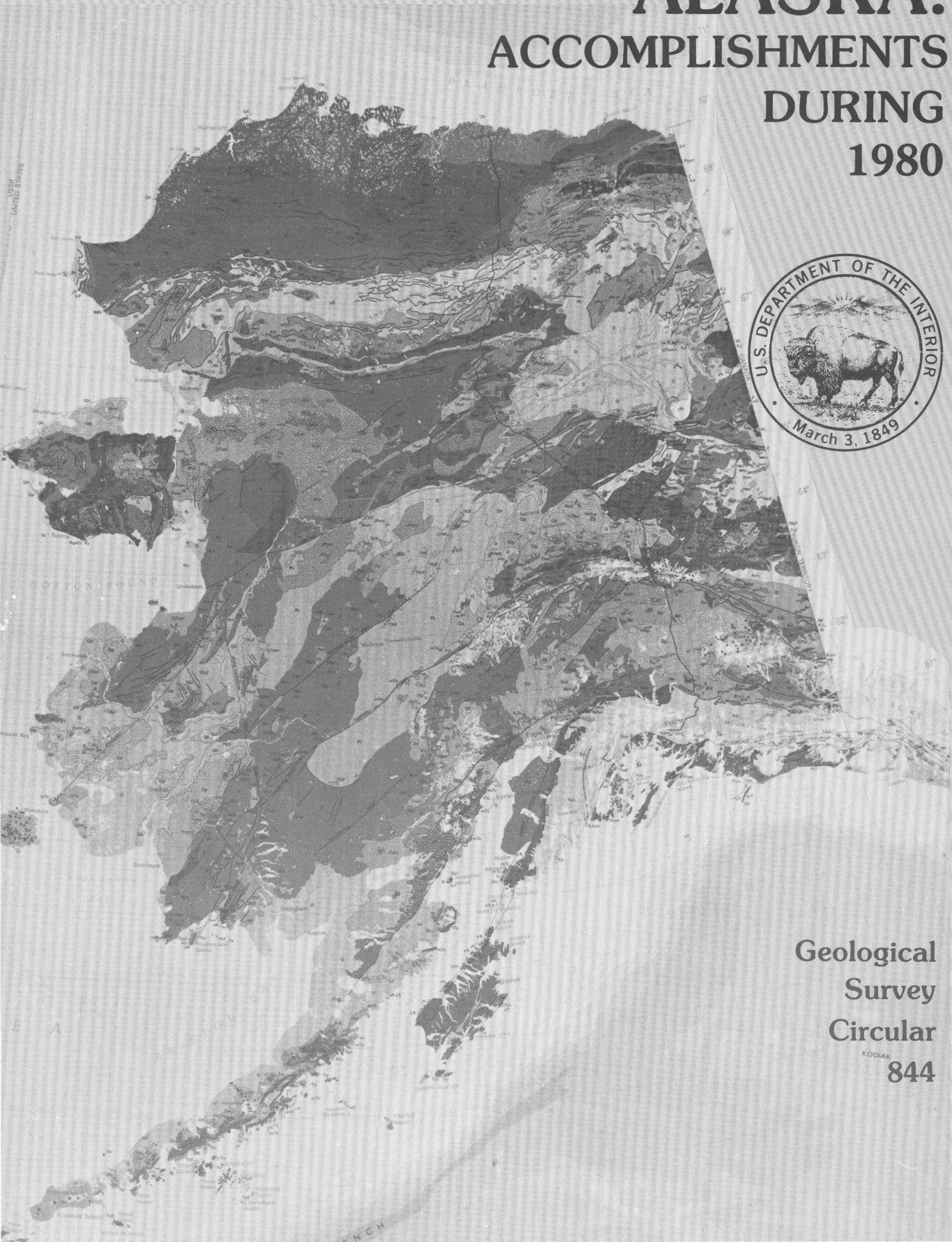
THE UNITED STATES GEOLOGICAL SURVEY IN ALASKA: ACCOMPLISHMENTS DURING 1980



Geological Survey Circular 844

FRONT COVER

The front cover is a reduction of part of the new geologic map of Alaska published in 1980 by the U.S. Geological Survey (Beikman, H. M., 1980, Geologic map of Alaska: U.S. Geological Survey Special Map, scale 1:2,500,000, 2 sheets). This multicolored map portrays the areal distribution of almost 200 rock units and indicates the sources of data used in the compilation.

The United States Geological Survey in Alaska: Accomplishments during 1980

Warren L. Coonrad, Editor

GEOLOGICAL SURVEY CIRCULAR 844

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THE UNITED STATES GEOLOGICAL SURVEY IN ALASKA: ACCOMPLISHMENTS DURING 1980

Warren L. Coonrad, Editor

ABSTRACT

This report of accomplishments of the U.S. Geological Survey in Alaska during 1980 contains summary and topical accounts of results of studies in a wide range of topics of economic and scientific interest. In addition, many more detailed maps and reports are included in the lists of references cited for each article and in the appended compilations of 297 reports on Alaska published by the U.S. Geological Survey and of 177 reports by U.S. Geological Survey authors in various other scientific publications.

SUMMARY OF IMPORTANT RESULTS

Introduction

The U.S. Geological Survey is engaged in many scientific investigations on various aspects of the land and water in Alaska. Products of the Survey's investigations are reports describing the physical findings and their significance. This volume includes summary discussions, topical results, and in some instances, narratives of the course of the studies on just a few of the various earth-science subjects being investigated in Alaska. For readers' convenience in perusal of reports covering specific areas of interest, the articles are grouped by areas corresponding to the six regional geographic onshore subdivisions shown on figure 1, offshore coverage, and those articles that cover more than one geographical region or are statewide in scope. Index maps showing the study areas or sites discussed are included near the beginning of each subdivision (figs., 1, 8, 25, 39, 48, 55, 72, and 79). An author index at the back of the volume will assist in indentifying and locating work of specific authors reported herein or in the appended compilations of current reports on Alaska by U.S. Geological Survey authors and their associates.

STATEWIDE ALASKA

(Figure 1 shows study areas discussed)

Tectonostratigraphic terrane map of Alaska

By David L. Jones, Norman J. Silberling, Henry C. Berg, and George Plafker

The concept that Alaska contains separate allochthonous fragments (Hamilton, 1971; Richter and Jones, 1971; Jones and others, 1972; Berg and others, 1972) has slowly gained credibility as increasingly detailed stratigraphic and structural information has become available. It now seems apparent that much of Alaska consists of disjunct, accreted fragments, termed tectonostratigraphic terranes ("terranes," for short), each of which records a distinctive depositional and deformational history (Jones and Silberling, 1979, 1981). Concomitant with the increase in geologic data over the past 10 years has been the application of paleomagnetic techniques that have permitted quantitative measurements of the amount of displacement between cratonal North America and some of the allochthonous terranes (Packer and Stone, 1972, 1974; Stone, 1977, 1979; Stone and Packer, 1977, 1979; Hillhouse, 1977; Van der Voo and others, 1980). Without exception, where discrepancies in paleolatitude are determined for Alaskan rocks, a northward shift is indicated (see summary by Beck, 1980). The amount of displacement for some rocks is very large; for example, minimum northward movement of Wrangellia (Jones and others, 1977), as recorded by paleomagnetic measurements obtained from the Triassic Nikolai Greenstone, may be 9000 km (Hillhouse, 1977; oral commun., 1980). A comparable northern displacement for Jurassic

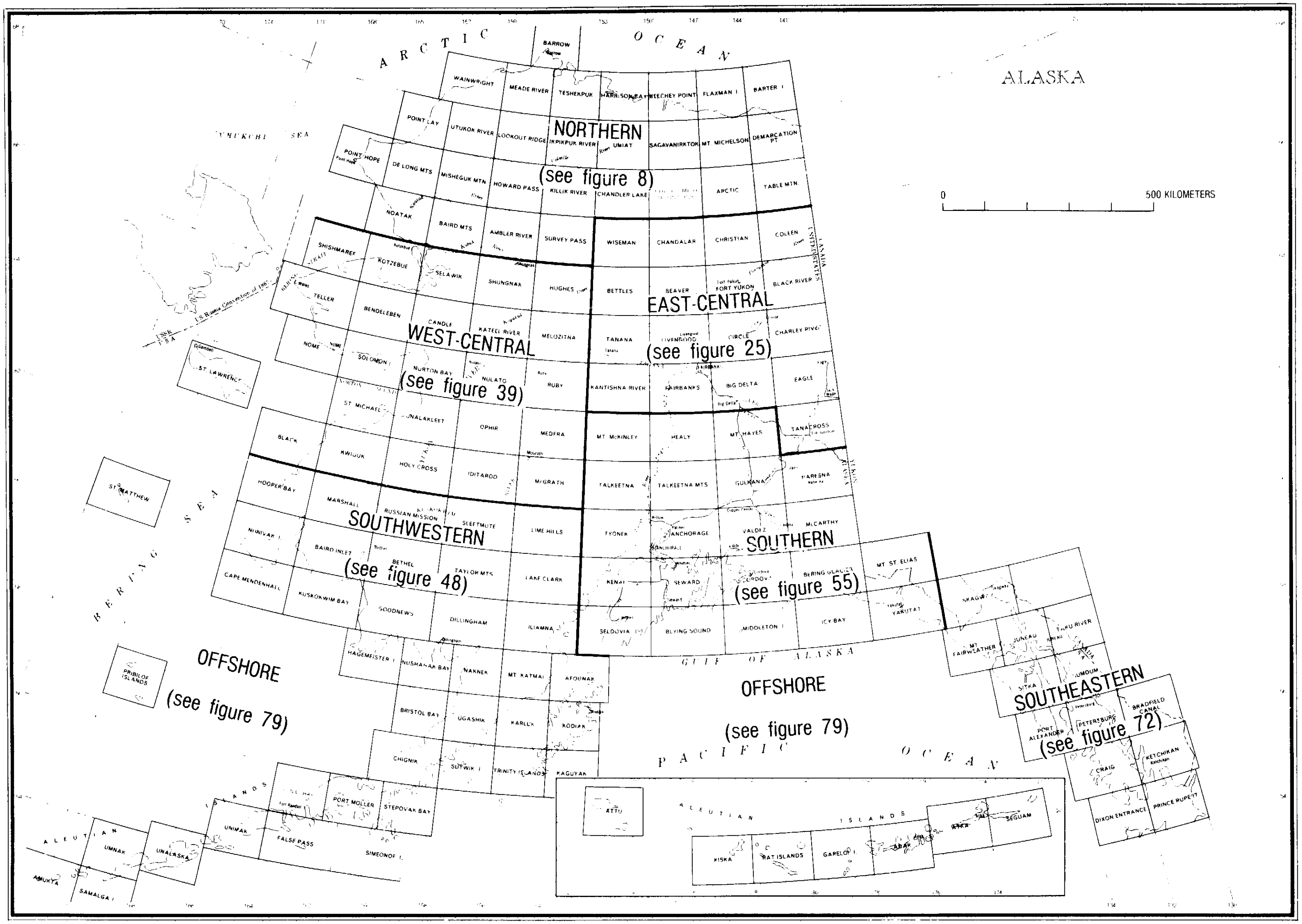

FIGURE 1.—Regions of Alaska used in this report. Referenced figures indicate index maps showing areas discussed in this volume.

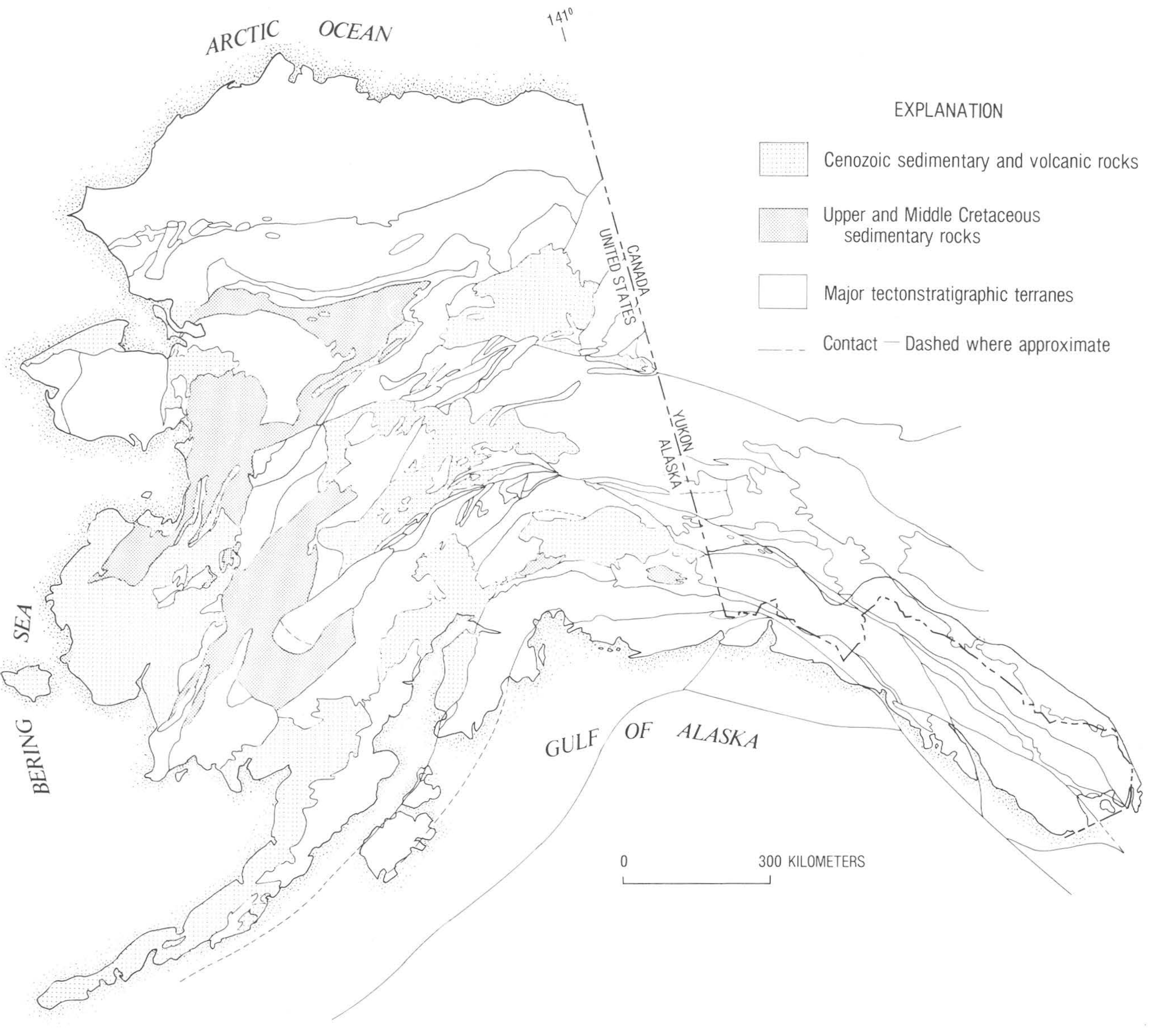

FIGURE 2.—Simplified terrane map of Alaska showing distribution of major tectonostratigraphic terranes and large areas of post-accretionary volcanic and sedimentary rocks. Individual terranes in southeastern Alaska are described by Berg and others (1978); descriptions and columnar sections for the rest of Alaska are given by Jones and others (1981).

rocks of the Alaska Peninsula has been suggested by D. B. Stone (written commun., 1980).

A map with accompanying columnar sections showing the distribution and character of major tectonostratigraphic terranes in Alaska has now been released by the U.S. Geological Survey (Jones and others, 1981). This map, at a scale of 1:2,500,000, incorporates concepts and data from earlier tectonostratigraphic terrane maps (Berg and others, 1978; Jones and Silberling, 1979, 1982) that cover only parts of the State. Figure 2 is a small-scale map showing distribution of terranes tentatively identified in work to date (see Jones and others, 1981). A tectonic map, showing internal structural features and the nature of bounding faults between terranes, is now in preparation at the 1:2,500,000 scale.

The salient points that emerge from examination of this new terrane map are as follows:

1. Alaska is composed of approximately 50 major terranes, each of regional extent, plus an undetermined number of smaller blocks and fragments too small to be shown on the map. The only part of Alaska that may not be allochthonous is near the Canadian border north of the Yukon River, and even that area is suspect.

2. Each terrane is bounded by major faults, the majority of which appear to be thrust faults with very large displacement. Strike-slip faults, although ubiquitous, seem to be younger and to disrupt the nappe-like pattern of terrane distribution.

3. Most terranes in Alaska accreted during the late Mesozoic, although several terranes with disparate points of origin amalgamated before accretion (Berg and others, 1978). By Late Cretaceous time, an overlap sequence composed of sedimentary or volcanic rocks locally covers some of the boundaries and parts of most accreted terranes, except for those along the southern margin that accreted in latest Cretaceous to mid-Cenozoic time.

4. The accreted terranes are highly variable in nature and represent diverse geologic environments. Ophiolitic rocks, thought to have formed as oceanic crust, are the basement for some terranes whereas metamorphosed continentally derived rocks of Precambrian age form the basement for others. Although the terranes containing rocks of oceanic character are obviously out of place in their present continental setting, the terranes of continental character also may be allochthonous. Some geologists (for example, Churkin and others, 1980), however, have speculated that several terranes are para-autochthonous and form the backbone of North America against which the allochthonous oceanic terranes accreted.

5. The pattern of accretion does not fit classic plate-tectonic models that are based mainly on concepts of accretionary growth through off-scraping of soft pelagic sediments. Instead, most of the accreted terranes are large, coherent blocks in which depositional stratigraphic relations are preserved through long periods of time. Melangelike terranes are relatively rare and are abundant only along the southern and southwestern margin of the State. Although large-scale plate movements and consumption of vast tracts of oceanic crust are implicit in the Alaskan data, the mechanism by which these rootless, nappelike thrust sheets are stacked on top of one another to form new continental crust is not yet apparent. Several of the major suture zones between terranes are marked by highly deformed, deep-water chert and distal turbidites of late Mesozoic age. These collapsed sedimentary prisms now constitute a "paste" in which some small terranes are dispersed as rootless nappes (Coney and others, 1980).

6. Gaining an understanding of the basic tectonic history of Alaska—where the terranes originated, where they moved, and how they became assembled into the continental framework—has important ramifications for metallogenesis and mineral resource assessment. Specifically, the origin and distribution of stratabound and certain other syngenetic mineral deposits are related to the depositional and tectonic history of individual terranes (Berg, 1979; Silberman and others, 1980). Specific knowledge of the nature of the terranes thus is an important key to identifying and evaluating geologic tracts favorable for the occurrence of specific types of mineral deposits and, in turn, is useful as a guide to mineral exploration.

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Probabilistic estimates of maximum seismic horizontal ground motion on rock in Alaska and the adjacent outer continental shelf

By Paul C. Thenhaus, Joseph I. Ziony, William H. Diment, Margaret G. Hopper, David M. Perkins, Stanley L. Hanson, and S. T. Algermissen

Probabilistic ground-motion hazard maps are being prepared for Alaska and the adjacent continental shelf areas. The maps depict estimates of horizontal ground motions (acceleration and velocity) having a 90 percent probability of not being exceeded in 10, 50, and 250 years, respectively, for return periods of approximately 100, 500, and 2,500 years. The maps currently being prepared are revised versions of those produced by Thenhaus and others (1979).

The general method is that followed by Algermissen and Perkins (1976) for the probabilistic acceleration map of the conterminous United States. Modifications have been made (see Thenhaus and others, 1980) to incorporate broader geologic considerations in defining seismogenic zones and to devise new methods of dealing with zones having only low levels of historic seismicity. General procedures involve (1) delineation of seismic source areas; (2) analysis of the statistical characteristics of the historic seismicity in each area; and (3) calculation of histograms of acceleration exceedances for gridded points in the region.

We partitioned Alaska into 24 seismogenic zones (fig. 3) on the basis of apparent or inferred relations between seismicity and geologic framework. Faults of regional extent that show evidence of Holocene or historic movement are considered seismogenic zones in themselves; such features are confined to southern and southeastern Alaska. Elsewhere, broad zones are used to encompass regions of apparently homogeneous areal rates of earthquake activity and (or) distinctive geologic framework.

Southern and southeastern Alaska.—The Denali, Totschunda, Fairweather, Chatham Strait, Queen Charlotte Island, and Castle Mountain faults were used as individual narrow seismogenic zones. Zone 13 defines the Yakutat

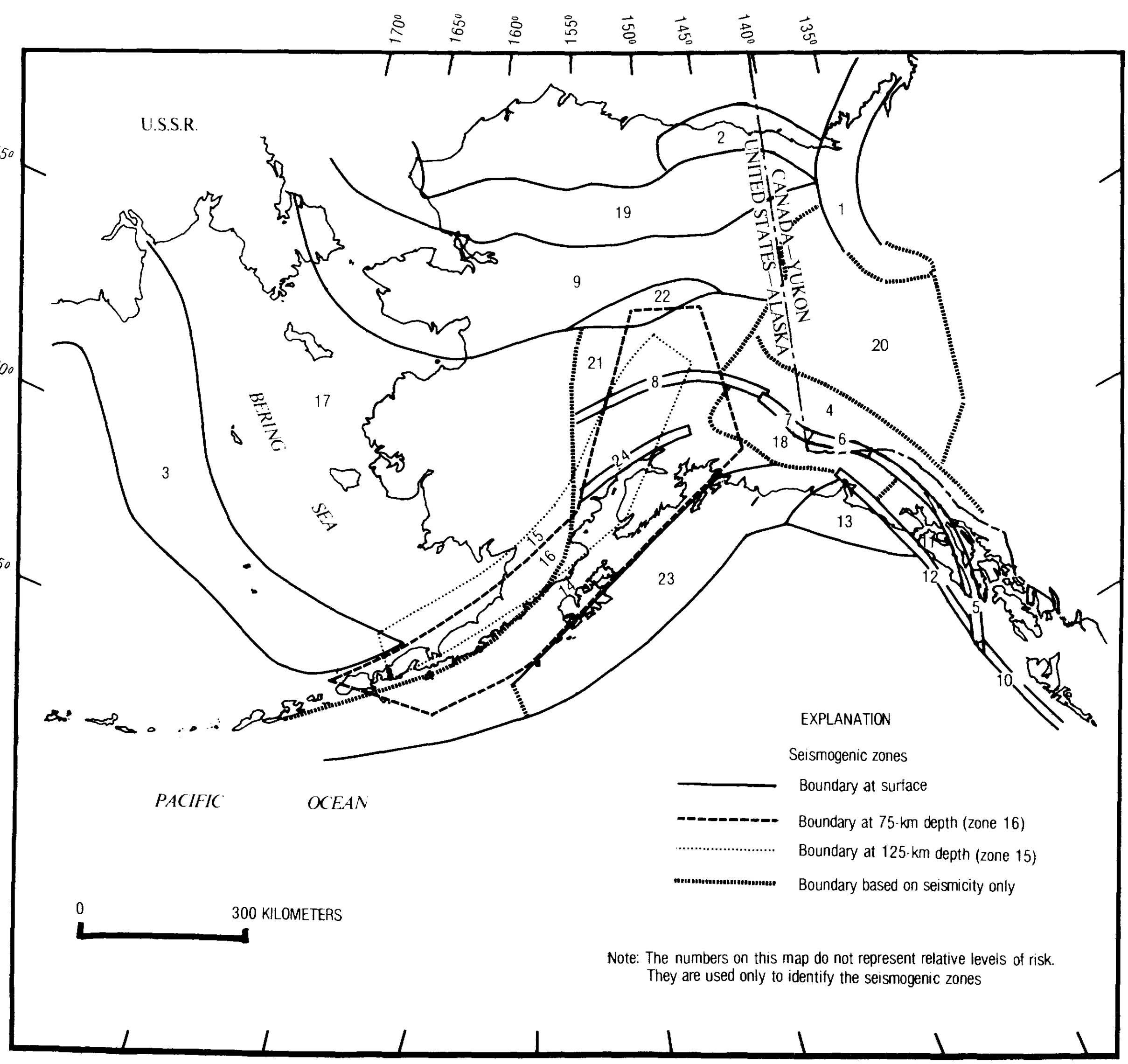

FIGURE 3.—Preliminary map of Alaska seismogenic zones.

block, an area of major thrusting with significant left-lateral strike-slip movement (Plafker and others, 1978). Northwest of the Aleutian trench, the megathrust between the North American and Pacific plates and the shallow thrust faults in the overriding plate constitute zone 23. The area is characterized by a very high rate of seismic activity.

The remaining zones of southern and southeastern Alaska have been defined on the spatial distribution of historic seismicity. The deeper events of the Benioff zone are delineated in such a way that zone 15 encompasses seismicity from 100 km and below, and zone 16, from 50 to 100 km. Earthquakes in zones 15 and 16 are modeled as ruptures up dip from planar, horizontal areas at depths of 125 and 75 km, respectively.

Southwestern Alaska.—The inner Bering Sea and southwestern Alaska (zone 17) constitute a distinctive broad zone with low rates of seismicity and inferred low potential for large earthquakes. Although several major fault systems such as the Denali (which to the east is seismogenic) extend into this region, none is considered active within the zone. The outer shelf of the Bering Sea has been separately zoned (zone 3) on the basis of its geologic framework (fault-bounded northwest-trending basins of Tertiary and Quaternary age) and the sparse occurrence of moderate-size earthquakes.

Central Alaska.—Zone 9 is a distinctive east-trending seismogenic province that extends across interior Alaska, the Seward Peninsula and parts of Norton Sound, and the Hope Basin and has significant potential for large earthquakes. The zone is characterized by a relatively thin crust, scattered Quaternary volcanism, Neogene basins, Quaternary faulting, relatively high heat flow (as evidenced by widespread hot springs), and low average elevation. It is inferred to represent a regime of extensional tectonics that has been established in late Cenozoic time across diverse older tectonic elements of central Alaska. The boundaries of this zone offshore are uncertain.

Northern Alaska.—In northeastern Alaska, zone 2 encompasses seismicity apparently associated with late Cenozoic faulted folds that flank the Romanzof Mountains. This zone may merge with the Richardson-Mackenzie Mountains seismic belt of northwestern Canada (zone 1). The Brooks Range constitutes a distinctive zone (zone 19) on the basis of its east-trending folds and thrusts of chiefly late Mesozoic age and low diffuse seismicity that cannot be currently associated with specific geologic features.

Methods and product.—Data on the historic seismicity of Alaska and its continental shelves were derived from the data base available from the National Geophysical and Solar-Terrestrial Data Center (Myers, 1976). Extensive work was done on this data file in order to obtain a single uniform magnitude, Ms, for all earthquakes of interest. Maximum magnitudes assigned to each of the zones were estimated from the largest observed magnitude in the historic past or, for individual faults of regional extent, by applying Mark and Bonilla's (1977) magnitude versus fault-length regressions. In zones of low seismic activity, maximum magnitudes were adopted from more active zones that have a comparable seismotectonic setting.

Earthquake-magnitude/frequency relations were derived from each zone from the available data set after the data set had been corrected for incompleteness according to the method of Stepp (1973) and after obvious aftershock sequences were deleted. For zones that had too few events to calculate a statistically valid b-value (the rate of change of occurrence frequency with respect to magnitude), zones were combined in order to obtain a more reliable estimate. The b-values were then applied to each of the individual zones. Where possible, zones were combined that have similar seismotectonic characteristics (for example, the various regional strike-slip faults of southern and southeastern Alaska). Where this was not possible because of an incomplete understanding of the seismotectonics, zones that are contiguous or have similarly low seismic activity were grouped.

An example of the type of product of this study is figure 4; a preliminary map of the 50-year exposure time (return period of 500 years) for estimated accelerations. Ground motions with one chance in 500 of being exceeded in one year are 500-year return-period ground motions. Gridded map values are contoured to produce the 500-year return-period map. The map shows that the highest expected ground accelerations in the region are along the Fairweather fault and in the Gulf of Alaska. Because of their low rate of activity in historic time, the Totschunda, Chatham Strait, and Castle Mountain faults

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Europium-rich dark monazite in Alaska

By Sam Rosenblum and Elwin L. Mosier

The investigation of europium-rich monazite in Alaska developed out of a study of the minor elements in 1072 alluvial concentrates from Alaska (Overstreet and others, 1975, 1978). Mineralogic examination of nine concentrates that showed anomalous spectrographic values for Eu led us to recognize this unusual dark pelletlike variety of monazite. Subsequent spectrochemical analysis of hand-picked dark monazite grains revealed the high Eu and low Th contents. The mineral resembles rounded fragments of shale and is difficult to recognize, but careful examination of the concentrates disclosed dark monazite present at 63 sites in 14 different regions between the northwest border of Canada and the west tip of Seward Peninsula (fig. 5).

The predominant source rocks for this dark monazite are clayey and silty sedimentary rocks that have undergone low-grade dynamothermal

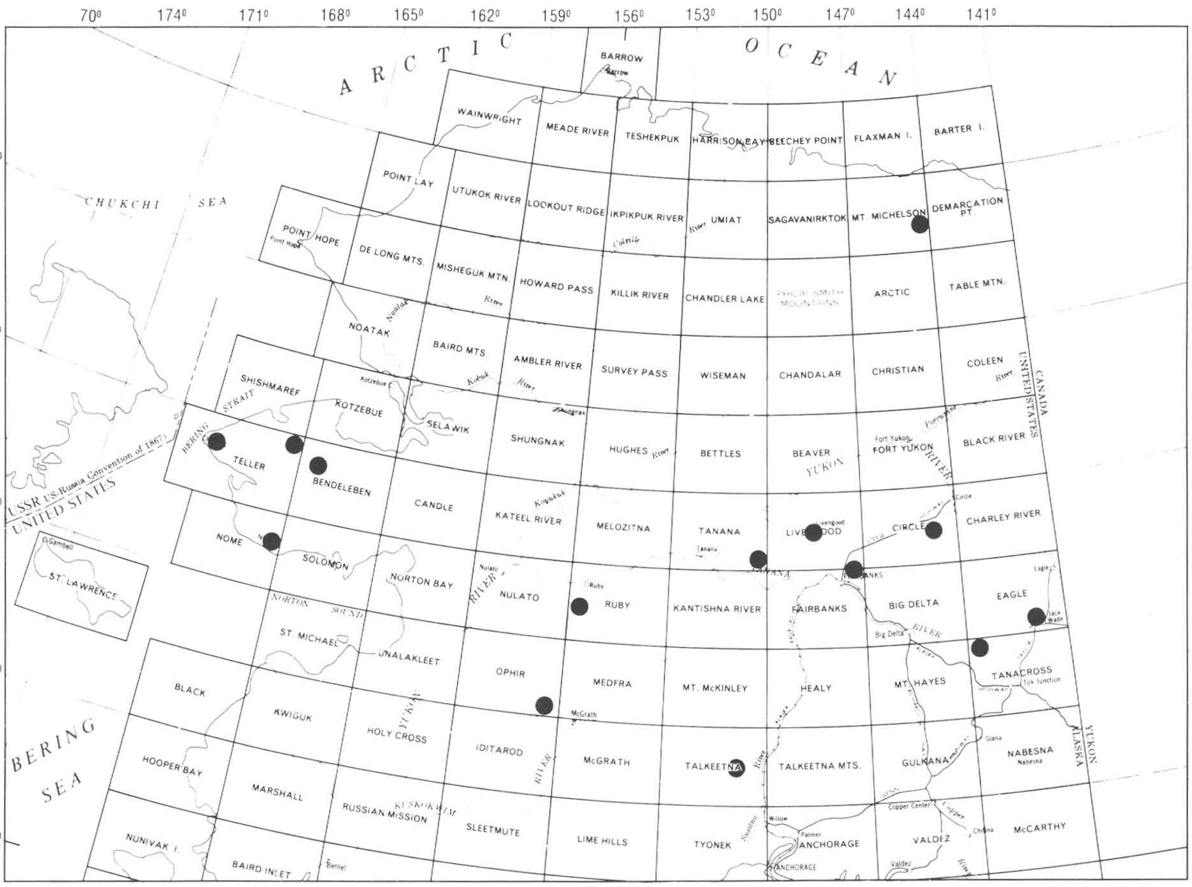

FIGURE 5.—Map of part of Alaska showing general regions (dots) where europium-rich dark monazite has been recognized in alluvial concentrates.

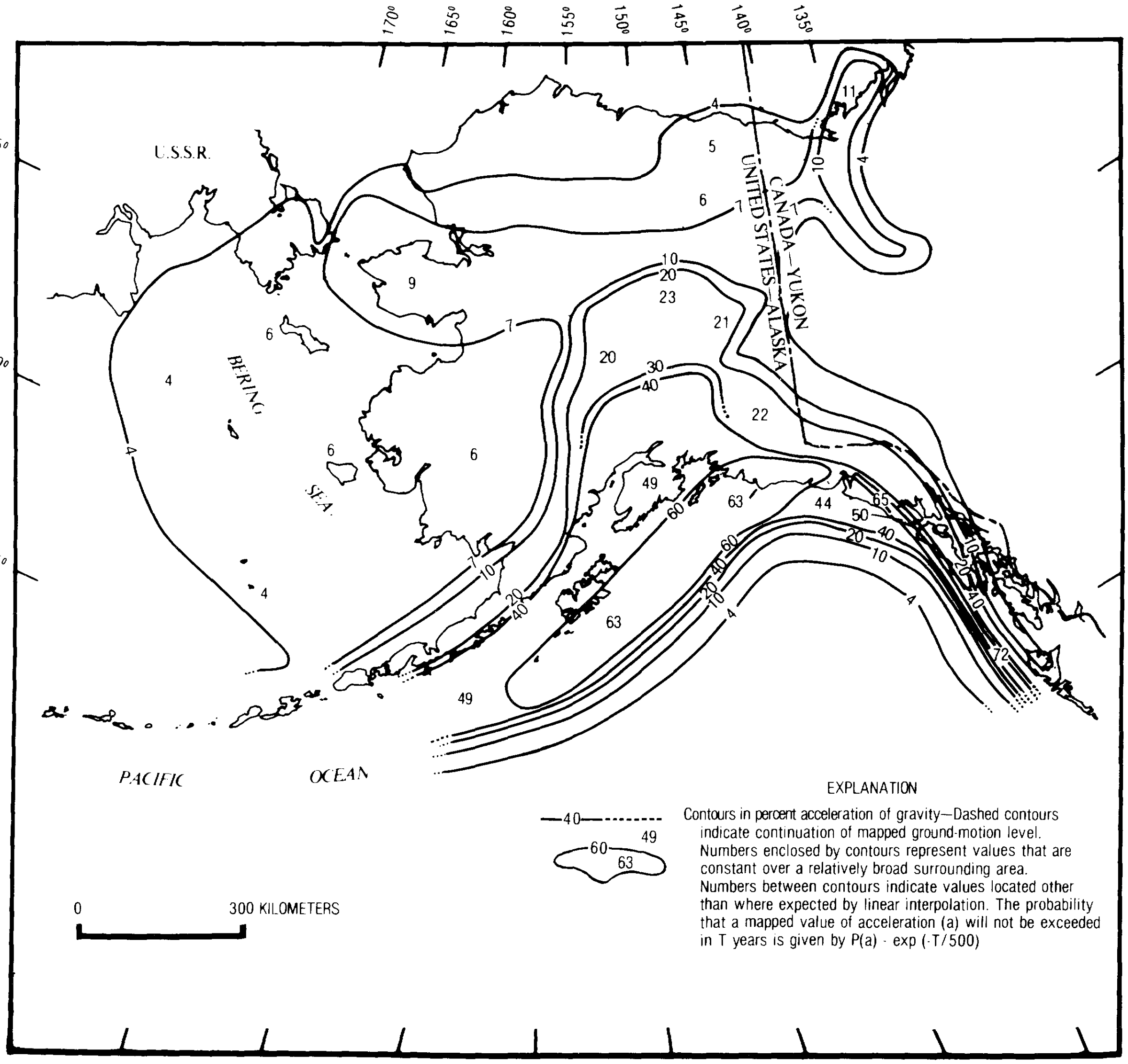

FIGURE 4.—Preliminary map showing estimated horizontal ground motion (acceleration) on bedrock. There is a 90 percent probability that estimated values will not be exceeded in 50 years (or one chance in 500 that estimated values will be exceeded in one year).

contribute values of ground motion commensurate with their neighboring, more regional zones.

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metamorphism or low-grade thermal (<300°C) metamorphism caused by intrusive masses of felsic, mafic, or ultramafic rocks. Favorable rocks that have been subject to temperatures higher than about 300°C contain allanite, sphene, or other rare-earth silicates rather than the dark monazite.

Grains of dark monazite from Alaska are usually gray to black and ellipsoidal, as shown in figure 6*A*. Sericite and chlorite fringes commonly form "dogtooth" projections at the ends of the long axes of the dark monazite grains, as shown in figure 6*B*. This illustration indicates the sequential release of the grains of dark monazite from phyllitic host rocks. Grains may be monocrystalline, microcrystalline, or cryptocrystalline, as determined by optical extinction properties. Fine-grained inclusions of host-rock minerals (phyllosilicates, oxides, carbon) are common. The dark color is caused by carbon. Sizes of the grains range from 0.1 to 1 mm and average 0.5 mm in Alaska. Worldwide, the colors, shapes, and sizes of this variety of monazite are rather similar.

Data from X-ray diffraction and optical and magnetic properties of dark monazite from Alaska seem to be identical to those of yellow monazite, but hardness is less (3 to 4 for dark monazite as compared to 5 or more for yellow monazite), and density is considerably less (4.27 to 4.72 as compared to 5.15 for yellow monazite). These lower values for hardness and density are

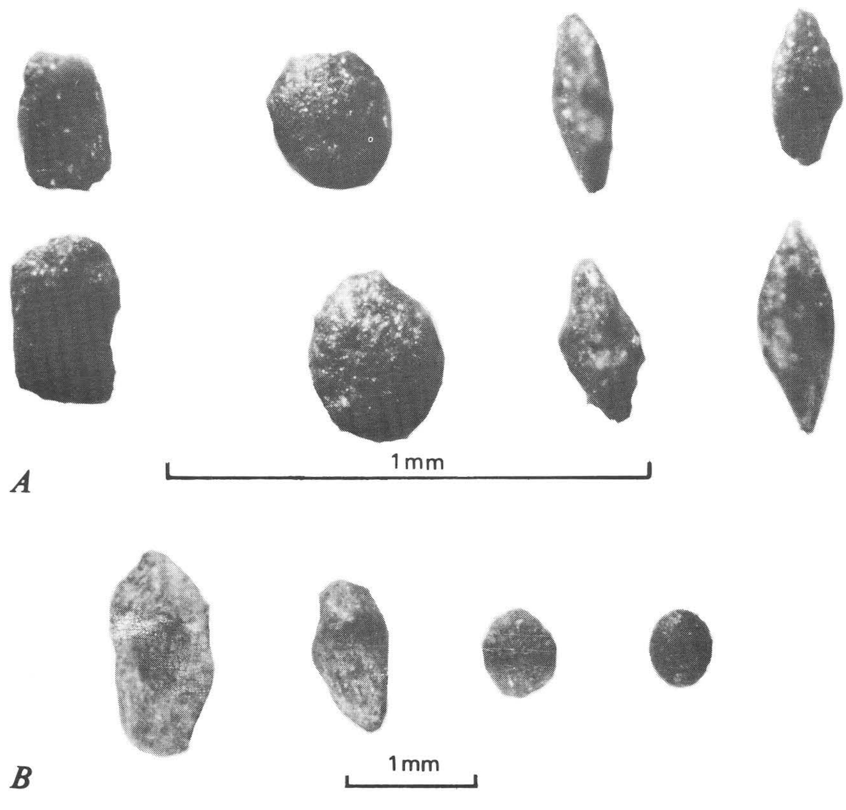

FIGURE 6.—Europium-rich monazite from Alaska. *A*. Typical grains are shown lying flat (column 2) and on end (columns 3 and 4). Rectangular grains (column 1) are rare. Grains at lower right show micaceous flanges. *B*. Series of dark monazite grains from a single placer concentrate from western Alaska, illustrating gradual release of grains from phyllitic matrix. *C*. Photomicrograph of Eu-rich dark monazite fragments showing turbidity due to numerous submicron-sized inclusions. Crushed grain is mounted in liquid with 1.73 refractive index.

directly related to the numerous inclusions found in all dark monazite grains, as shown in figure 6*C*.

The composition of dark monazite is generally similar to that of yellow monazite, but the differences in Eu and Th contents are significant. A comparison of the average of all known analyses of dark monazite with the average for as many yellow monazites as could be found in the literature shows that in the dark monazite, Eu is more abundant (Eu_2O_3 = 0.39 vs 0.05 percent) and Th is less abundant (ThO_2 = 0.73 vs 7.15 percent) than in the yellow monazite (Rosenblum and Mosier, 1982). Spectrochemical analyses of 11 dark monazites from Alaska average (in percent) La_2O_3 = 13.54, Ce_2O_3 = 29.00, Pr_2O_3 = 3.34, Nd_2O_3 = 14.0, Sm_2O_3 = 1.92, Eu_2O_3 = 0.31, Gd_2O_3 = 0.98, Y_2O_3 = 0.47, P_2O_5 = 22.36, ThO_2 = 0.88, SiO_2 = 8.35, TiO_2 = 0.66, Al_2O_3 = 1.98, Fe_2O_3 = 1.77, CaO = 0.28, MgO = 0.23; for a total of 100.07. Carbon in four samples averages 0.61 percent, and H_2O in three samples averages 0.94 percent.

On the basis of our studies and geochemical considerations, we conclude that dark monazite originates from low-grade regional metamorphism or low-grade contact metamorphism of shale in the temperature range of 150-300°C and that the rare-earth elements in dark monazite are probably derived from the disintegration of detrital igneous biotite in the shale.

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Alaskan geochemical field laboratories

By Richard M. O'Leary and Steven K. McDanal

The Anchorage geochemical field laboratory was in operation during the summer of 1980 at 5500 Oilwell Road, Elmendorf AFB, Alaska, as it has been since 1967. The laboratory is managed and operated by the Branch of Exploration Research in Golden, Colorado. It is equipped with a drying oven, rock crushers, pulverizers,

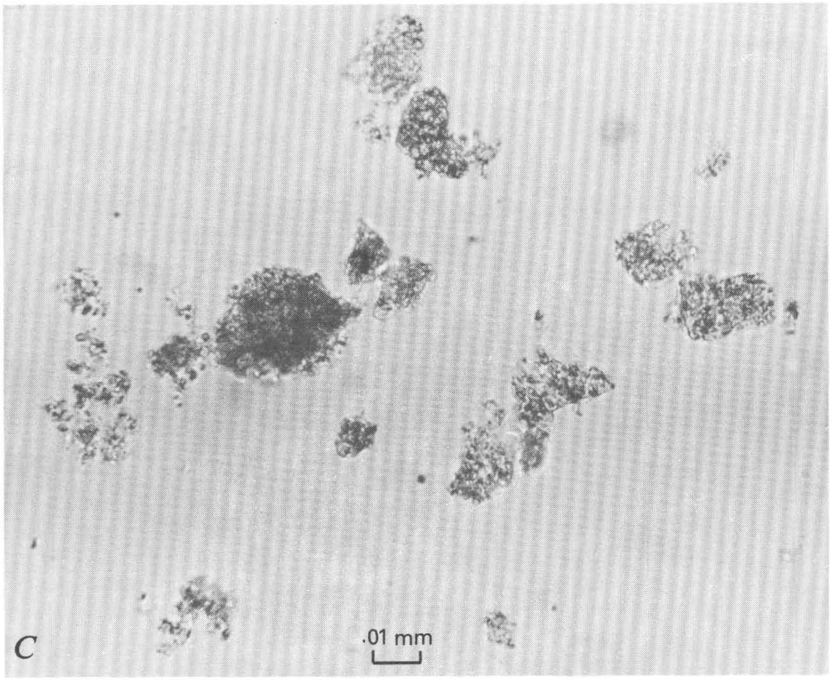

FIGURE 6.—Continued.

and sieve shakers to prepare rock, stream-sediment, glacial-debris, and soil samples. The analytical laboratory includes an emission spectrograph capable of generating semiquantitative analyses for 31 elements on each sample. Atomic absorption spectrophotometry and mercury-vapor detection instruments complement the analytical capabilities. A minicomputer system, added in recent years, is used to enter, edit, update, and retrieve analytical data from the Rock Analysis Storage System (RASS) while the laboratory is in normal operation. This system circumvents the keypunch operation and time lag involved in processing data through the Branch office in Colorado.

For the past several years the U.S. Geological Survey has also operated an analytical laboratory in Juneau, Alaska, where it supported many of the southeastern Alaska projects. In the spring of 1980 the laboratory was assigned to the Chugach Rare II project. In order to handle this work more efficiently, the laboratory, which is primarily a self-contained, wet-chemical, mobile unit housed in a 24-foot fifth-wheel trailer, was moved to the location of the Anchorage laboratory. The mobile unit's equipment includes an emission spectrograph and an atomic-absorption spectrophotometer. Branch personnel assigned to the Chugach Rare II project utilize the Anchorage laboratory's sample preparation facilities.

The field laboratories offer the advantage of being centrally located for fast mail service and are able to handle a large volume of samples in a short period of time. The quick turnaround of analytical results allows further sampling or checking of anomalous or critical areas while crews are still in their field areas.

In the two-month period in 1980 when both laboratories were in operation about 4,000 samples from the Mount Hayes, Bristol Bay-Ugashik-Karluk, Healy, Circle, Bradfield Canal, Medfra, and Petersburg quadrangles, and the Chugach Rare II project were processed. The results of the analyses are used mainly to generate geochemical maps that aid in mineral resource assessments and help locate possible mineral deposits.

Radiometric age file for Alaska

By Nora Shew and Frederic H. Wilson

The Alaska radiometric age file of the Branch of Alaskan Geology is a computer-based compilation of radiometric dates from the state of Alaska and the western parts of the Yukon Territory and British Columbia. More than 1800 age determinations from over 250 references have been entered in the file. References date back to 1958 and include both published and unpublished sources. The file is the outgrowth of an original radiometric age file compiled by Don Grybeck and students at the University of Alaska-Fairbanks (Turner and others, 1975).

In addition to age, each entry in the file is coded by quadrangle, latitude, longitude, sample number, rock type, dating method, and dated phase. References and additional comments pertinent to the entries are also given. Results from potassium-argon, rubidium-strontium, uranium-lead, uranium-thorium, lead-alpha, fission track, and samarium-neodymium dating methods are included. The compilation is maintained by regular updating as dates become available in the literature. Though probably not complete, the file contains most of the published dates. Radiometric age maps for various regions of the state are revised when warranted by availability of new age data.

Through use of computer programs, the data can be sorted by any combination of the coded factors; for example, a listing of granites ranging in age between 30-40 m.y. in east-central Alaska. This sorting capability makes the file amenable to a variety of uses, from rapid perusal of previous work in a particular area to examinations of large-scale distributions of rocks of a particular age in selected areas. The file has been used to produce radiometric age maps (Wilson and Turner, 1975a-e; Wilson and others 1979; Dadisman, 1980) and provides age data for various topical and regional studies. Currently the data bank is being utilized for a study of the distribution of ages throughout the state (Wilson and Shew, 1982). The file is maintained on the U.S. Geological Survey Multics computer in Menlo Park, California. Further information on the file, and its capabilities can be obtained from the authors.

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Apparent episodicity of magmatic activity deduced from radiometric age determinations

By Frederic H. Wilson and Nora Shew

Results of recent potassium-argon age studies in the Chignik region, Alaska, (Wilson, 1980; Wilson and others, 1982) have suggested a distinct episodicity in igneous activity during Tertiary time. To date work on the Aleutian magmatic arc indicates that plutonic activity took place along the present outer Pacific margin and in the northern Alaska-Aleutian Range batholith (Reed and Lanphere, 1973; Kienle and Turner, 1976; DeLong and others, 1978) in latest Cretaceous and earliest Tertiary time (70-58 m.y.) and was followed by a hiatus lasting until late Eocene (45 m.y.) time. Late Eocene to earliest Miocene (45-20 m.y.) magmatic activity was followed by a middle Miocene hiatus (10 m.y.). Since that time, magmatic activity in the Aleutian arc has been continuous.

The Alaska Branch radiometric age file (Shew and Wilson, 1982) was used statistically to examine the age distribution of dated rocks in Alaska. This statistical analysis assumes a random distribution of age determinations which, of course, is not true; with the large number of dates, however, we assume that any important episode is represented. We suggest that, in a crude sense, the general distribution of radiometric dates can be related to periods of magmatic activity. Although these periods may be overprinted by regional metamorphism, igneous activity is commonly related to episodes of metamorphism.

Histograms (fig. 7) of published dates and a few unpublished ones show that the gaps noted in magmatic activity in the Alaska Peninsula are widespread. Reed and Lanphere (1973) distinguished three events in the Alaska-Aleutian Range batholith; during Jurassic, latest Cretaceous and earliest Tertiary, and middle Tertiary time. Wilson (1980) defined the time extent of a late Miocene to Holocene age event. Here we extend the area to which this episodicity of activity applies.

Extending the Chignik region work to the Alaska Peninsula and Aleutian arc (defined as that part of Alaska south of 59° N. and west of 152° W.) as a whole (fig. 7*A*) shows the episodes noted at Chignik are true for the entire region. Wilson (1980) suggested a tectonic explanation for this; other authors (DeLong and others, 1978; Byrne, 1979) have also considered this episodicity.

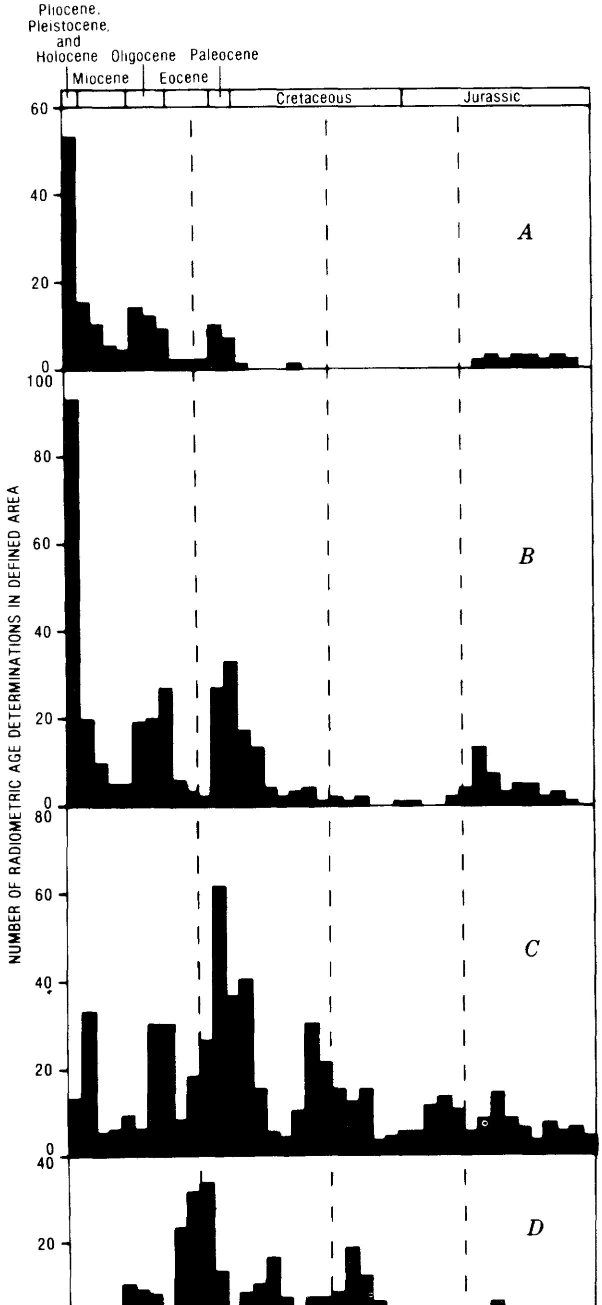

FIGURE 7.—Histograms showing distribution of radiometric ages for four regions of southern Alaska. *A*, Alaska Peninsula; *B*, southwestern Alaska; *C*, southern Alaska; *D*, southeastern Alaska.

In the Alaska Peninsula, the latest Cretaceous and earliest Tertiary event is related to anatectic(?) (Hudson and others, 1979) plutonism in the Kodiak, Semidi, Shumagin, and Sanak Islands and plutonism in the Aleutian Range (Reed and Lanphere 1973). The late Eocene to earliest Miocene event is related to magmatic activity in a magmatic arc existing during the time of deposition of the Tolstoi and Meshik Formations (Tolstoi-Meshik arc) (Wilson, 1980) and a plutonic event distinguished by Reed and Lanphere (1973) in the Alaska-Aleutian Range batholith. The last event, of late Miocene to Holocene age, is relatable to activity in the present Aleutian arc. If these events, particularly the activity in the Tolstoi-Meshik and Aleutian arcs, are related to subduction at the Aleutian or a paleo-Aleutian trench, then episodicity in subduction during Tertiary time is indicated. Wilson (1980) previously suggested that the Miocene hiatus could be related to subduction of the KulaPacific Range according to the model of DeLong and others (1978). An alternative interpretation would relate magmatic activity to coupling and uncoupling of this part of Alaska with the Pacific and Kula plates.

Expanding the area covered, we next looked at the distribution of ages from other parts of Alaska: (1) southwestern Alaska, that part of Alaska west of 153° W. and south of 64° N.; (2) southeastern Alaska, that part of Alaska, British Columbia, and the Yukon Territory between 129°30′ and 138° W. and between 51° and 62° N., and (3) southern Alaska, that part of Alaska and the Yukon between 59° and 64° N., and between 138° and 152° W.

The episodicity noted for the Alaska Peninsula holds true for not only southwestern Alaska (fig. 7*B*) but also southern Alaska (fig. 7*C*); however, southeastern Alaska (fig. 7*D*) has a distinctively different pattern. The peaks on the histogram, if these do represent geologic events, indicate that peaks of activity during Cretaceous and Tertiary time in southeastern Alaska occurred during minima in southwestern and southern Alaska. We suggest that a fundamental difference in tectonic history exists between southeastern

Alaska and southwestern and southern Alaska. Of particular importance is that dates in southeastern Alaska include a higher proportion of plutonic and metamorphic ages than those in southwestern Alaska which include many volcanic ages.

Two other events of some importance can be distinguished in the histograms shown in figure 7. The Jurassic event in the Alaska-Aleutian Range batholith of Reed and Lanphere (1973) is very apparent in the Alaska Peninsula, southwestern and southern Alaska histograms (fig. 7*A*, *B* and *C*). There is even a slight hint of an event of this age in southeastern Alaska. The other event is a middle Cretaceous (about 100 m.y.) event shown on the southern and southeastern Alaska histograms (fig. 7*C* and *D*). In southern Alaska, many dates can be related to small plutons and metamorphic rocks in the Alaska Range, whereas in southeastern Alaska many dates are on plutons suspected to be thermally reset.

We prepared histograms including ages to 500 m.y.; however, the number of age determinations older than 200 m.y. is low, and the peaks shown are very diffuse. These older peaks may not be significant because number of determinations is small. The peaks are in the following intervals in southern Alaska: 490-455 m.y. (Ordovician), 395-350 m.y. (Late Devonian and Early Mississippian), and 300-280 m.y. (Late Pennsylvanian and Early Permian). In southeastern Alaska the one older peak is in the interval 450-415 m.y. (Late Ordovician and Silurian).

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Correlation of northern Alaskan glacial deposits—a provisional stratigraphic framework

By Thomas D. Hamilton and David M. Hopkins

Mapping of surficial geology through the central Brooks Range has defined a consistent sequence of glacial deposits in most major drainage systems. These deposits are assignable to three principal intervals of late Cenozoic glaciation that were separated by long episodes of uplift and erosion (Hamilton, 1978, 1979a). Similar patterns of glaciation in other regions of Alaska allow provisional correlations with the Brooks Range record and with the eustatic sea-level chronology defined by Hopkins (1967, 1973) for northwestern Alaska (table 1).

The oldest interval of glaciation in the north-central Brooks Range, informally termed "Gunsight Mountain" (Hamilton, 1979a, b), is represented by erratic boulders of highly resistant lithologies that extend as much as 25 km beyond the limits of drift sheets of Pleistocene age. The deposits that initially contained the erratics were eroded by river systems that flowed 50-100 m above modern stream levels during a long interval that followed Gunsight Mountain time. Similar erratics occur north of the Alaska Range where they appear to be weathering out of the boulder-bearing upper part of the late Tertiary Nenana Gravel (Carter, 1981; R. M. Thorson, written commun., 1978). A comparable glacial phase on Seward Peninsula is recorded by the Skull Creek erratics of Sainsbury (1967; also see Hopkins and others, 1974). The extensive distribution of erratics north of the Brooks Range and in the Nenana Valley region north of the Alaska Range suggests a northern source for at least part of the precipitation that nourished glaciers during Gunsight Mountain time. Such a source is compatible with the marine sediment record, which indicates that permanent sea-ice cover did not develop over the Arctic Basin until about 0.9-0.7 m.y. ago (Herman and Hopkins, 1980).

The second major glacial interval, termed the Anaktuvuk River Glaciation in the central Brooks Range, was preceded and followed by major interglacial episodes of uplift and erosion (Hamilton, 1979a). Drift sheets are generally intact, with subdued end marines that commonly overlap the extensive erosion surfaces of post-Gunsight Mountain age. Drift assigned to the Anaktuvuk River interval has been greatly modified by stream erosion and by pedimentation and altiplanation processes and typically exhibits well-integrated drainage networks and slope angles less than 1°-2°. North of the Brooks Range, the drift has been broadly dissected by streams flowing 40-65 m above modern drainage levels. The Anaktuvuk River Glaciation probably was contemporaneous with the Browne Glaciation of the Nenana Valley, which was preceded by major valley incision and followed by 150-200 m of uplift along the north flank of the Alaska Range (Wahrhaftig, 1958, and oral commun.). A comparable drift sheet, informally named the "Sinuk" drift, and reported here for the first time, is recognized near the mouth of the Sinuk River on Seward Peninsula, where it is clearly of pre-Kotzebuan age (D. M. Hopkins, unpub. data, 1973). The same glacial event is represented by unnamed glacial till younger than the Anvilian transgression and older than the Nome River (Illinoian) Glaciation at California River on western Seward Peninsula

Table 1.—*Provisional correlation of Brooks Range glacial phases with other Alaskan glacial and sea-level records*

[----- indicates minor erosion interval; xxxxx indicates major erosion interval]

Brooks Range	Alaska Range	Seward Peninsula	Sea-level maxima and approximate dates (yr B.P.)[1]
Itkillik II	Wonder Lake; Riley Creek; Donnelly[2]	Mount Osborn	
-----	-----	-----	Middle Wisconsinan (40,000)
Itkillik I	Healy	Salmon Lake	
-----	-----	-----	Pelukian (125,000)
Sagavanirktok River	"Lignite Creek"[3]	Nome River	
xxxxx	xxxxx	xxxxx	Kotzebuan (>175,000; <250,000)
Anaktuvuk River	Browne erratics	"Sinuk"[4]	
xxxxx	xxxxx	xxxxx	Anvilian (>0.7; <$1.8X10^6$)
Gunsight Mountain erratics	Pre-Browne erratics	Skull Creek erratics	

[1]Hopkins, 1973; Hopkins and others, 1974; Hopkins, 1982.

[2]Names applied in McKinley Park, Nenana River, and Delta Valley regions, respectively.

[3]Informal name (R. M. Thorson, written commun., 1978). The post-Browne, pre-Healy "Dry Creek" glaciation of Wahrhaftig (1958) is no longer considered a firmly proven major glacial event (R. M. Thorson and N. D. Ten Brink, written commun., 1980).

[4]Informal name.

(Hopkins and others, 1974)[1].

The youngest sequence of glaciations in Alaska was not interrupted nor followed by any major erosional intervals; hence, drift sheets retain much of their original morphology and generally can be differentiated by deposits of two or more separate advances. In the central Brooks Range, this phase is represented by drift belonging to the Sagavanirktok River Glaciation and to drifts I and II of the Itkillik Glaciation (Hamilton, 1978, 1979a), which nest together in eastern parts of the region but diverge progressively to the west. West of about 156° W. long, the three drifts differ markedly in their occurrence. Itkillik II deposits are restricted to the highest cirques and valley heads, Itkillik I drift occurs widely through lower courses of mountain valleys, and Sagavanirktok River drift extends close to or beyond the limits of the most extensive older glacial deposits. The morphology of the Sagavanirktok River drift and its great extent in western parts of the Brooks Range suggest correlation with the widespread Nome River drift of the Seward Peninsula-Kotzebue Sound area (Hopkins, 1963, pl. 3), and Sagavanirktok River deposits are traceable into those of Nome River age on aerial photographs and Landsat imagery. On the basis of its relation to dated high-sea-level stands (table 1), the Nome River Glaciation probably took place sometime within the interval of about 200,000 to 125,000 yr B.P. Part of the Bering platform evidently remained submerged during this time (Hopkins, 1973), providing a nearby moisture source for extensive glaciation around the shores of the Bering Sea.

The subsequent Itkillik I glacial advance also is extensive within the western Brooks Range, where it was generated in cirques as low as 550 m altitude. This ice advance is correlated with the Salmon Lake Glaciation of the Seward Peninsula, which is younger than the Pelukian marine transgression of about 125,000 yr B.P., but older than 40,000 yr B.P. (D. M. Hopkins, unpub. radiocarbon date I-7710). The Itkillik ice advance in the Brooks Range also is older than the approximately 50,000-year range of carbon-14 dating, but is younger than strongly oxidized alluvium of probable Pelukian age (Hamilton, 1978). The last major ice advance, termed Itkillik II in the central Brooks Range, has been dated between 25,000 and 11,500 yr B.P. (Hamilton and others, 1980). Corresponding radiocarbon dates indicate that this event was synchronous with the last major Pleistocene glaciation of the Alaska Range (Ten Brink and Ritter, 1980; Weber and others, 1981), and it probably is equivalent to the Mount Osborn Glaciation of the Seward Peninsula.

The contrast in the extent of the Itkillik glacial advances in northwestern Alaska may be related to a secular decline in North Pacific water temperatures which, according to Sachs (1973), may have been as great as 4°C during Wisconsinan time. Such a temperature drop would have reduced the effectiveness of the North Pacific Ocean and its marginal seas as moisture sources and thus could have led to a much drier climate and less extensive glaciation in northwestern Alaska during the late Wisconsinan cold period.

[1]So-called Iron Creek drift was found in the subsurface beneath till of the Nome River Glaciation at Nome and at Iron Creek, 75 km to the northeast. It is uncertain whether these buried bodies of drift correspond to the early Pleistocene Skull Creek erratics or to the middle Pleistocene Sinuk drift. The name "Iron Creek Glaciation," used in several earlier publications on Seward Peninsula (for example, Hopkins and others, 1960; Hopkins, 1953, 1967) is therefore, dropped.

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NORTHERN ALASKA

(Figure 8 shows study areas discussed)

Precambrian metamorphic rocks from the Hub Mountain terrane, Baird Mountains quadrangle, Alaska

By Charles F. Mayfield, Miles L. Silberman, and Irvin L. Tailleur

In the northeast Baird Mountains quadrangle is a terrane of metamorphic rocks, herein called the Hub Mountain terrane, that may be unique

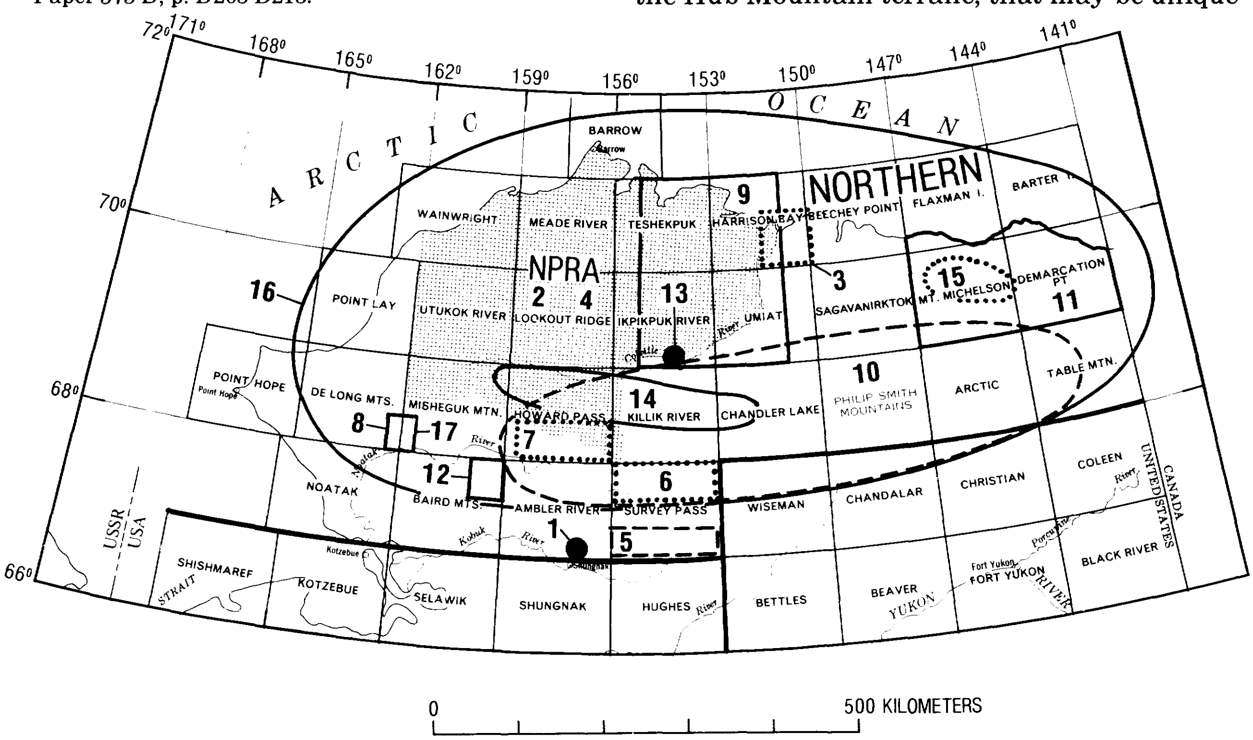

FIGURE 8.—Areas and localities in northern Alaska discussed in this volume. Authorship and applicable figures and tables relating to the numbered areas are listed below. 1, Armstrong, figure 17; 2, Blanchard and Tailleur, figure 21; 3, Carter, L. D., figure 22; 4, Carter, R.; 5, Cathrall, table 3; 6, Cathrall, table 3; 7, Ellersieck, Jansons, Mayfield, and Tailleur, figures 15 and 16; 8, Frank and Zimmerman, figure 10; 9, Galloway, figures 23 and 24; 10, Hillhouse and Grommé, figures 19 and 20; 11, Huffman, Kirk, and Molenaar, figure 18; 12, Mayfield, Siblerman, and Tailleur, figure 9, table 2; 13, McCoy; 14, Molenaar, Egbert, and Krystinik, figure 13; 15, Molenaar, Huffman, and Kirk, figure 14; 16, Nilsen, Brosgé, Moore, Dutro, and Balin, figure 12; 17, Zimmerman and Frank, figure 10.

in the southern Brooks Range because it appears to have escaped most of the thermal effects of the Cretaceous regional metamorphism. This study briefly describes the rocks that constitute this terrane and reports three new potassium-argon apparent ages that suggest it is composed of Precambrian metamorphic rocks.

Previous evidence for Precambrian metamorphic rocks in the southwest Brooks Range stems from interpretations of Turner and others (1978) and Turner and others (1979) based on their summaries of potassium-argon dates. They reported 12 Precambrian potassium-argon dates that range from 2,600 to 587 m.y. in addition to numerous Cretaceous and some Paleozoic dates. On the basis of an initial argon isochron analysis of the data, all five of the samples with dates greater than 1,300 m.y. were judged to contain excess argon resulting in anomalously old apparent ages. Seven samples with dates from 756 to 587 m.y. did not appear to contain excess argon; these suggest the existence of a late Precambrian regional metamorphic event (Turner and others, 1979). A major problem with the seven late Precambrian dates was that five of the samples had low potassium values, under 0.1 percent K_2O. Because a small amount of inherited argon can have a large effect on the apparent age of a low-potassium mineral and, conversely, a much smaller effect on a high-potassium mineral, some workers have been skeptical about the existence of a Precambrian metamorphic terrane. Therefore, the postulated existence of late Precambrian metamorphic rocks relied heavily on the two remaining samples which had much higher potassium concentrations. These two samples are from outcrops near locality 1 (fig. 9; table 2) in the northeastern part of the Baird Mountains quadrangle. One of the samples contained the mineral muscovite, which yielded the only Precambrian mica date in the southern Brooks Range metamorphic belt prior to this study. The date is significant because muscovite is less retentive of argon during post-crystallization thermal events than is amphibole. This occurrence needed to be tested, because if the two reported dates actually record an old metamorphic event that had not been too strongly affected by the Cretaceous metamorphic event, then this study of similar rocks in the surrounding Hub Mountain terrane should corroborate the previously published work and help to shed more light on the age of the metamorphic rocks in the southwest Brooks Range.

The Hub Mountain terrane is represented by the stippled areas on figure 9. It is a complex terrane composed of schists and gneisses that, before metamorphism, were sedimentary, volcanic(?), and plutonic rocks. Most metasedimentary rocks are fine- to coarse-grained quartz-mica schist interlayered with subordinate amounts of quartzite and marble. Common metamorphic minerals are quartz, mica, calcite, plagioclase, stilpnomelane, and garnet. Muscovite from coarse-grained quartz-mica schist at locality 2 (fig. 9, table 2) yields an age of 594 m.y., and a previously reported (Turner and others, 1979) date from a similar rock at locality 1 yields an age of 646 m.y. Before metamorphism, the sedimentary rocks were intruded by mafic igneous rocks, most of which were probably dikes and sills. In local areas in the northern and southern parts of the terrane, metabasite thinly interlayered with metasedimentary schist suggests that some of the mafic layers might have been tuffaceous. Common metamorphic minerals in the metabasite are green and blue-green amphibole, epidote, plagioclase, garnet, chlorite, and sphene. Green amphibole from a garnet-amphibolite at locality 3 (fig. 9; table 2) yields an age of 595 m.y., and a previously reported (Turner and others, 1979) amphibole date from a similar rock at locality 1 yields an age of 729 m.y. These ages should be considered minimum ages because some of their radiogenic argon may have been lost during the Cretaceous regional metamorphism. However, mineral assemblages and radiogenic dates consistently show that the Hub Mountain terrane was subject at least to high-grade greenschist-facies metamorphism in the Precambrian.

Prior to metamorphism, the sedimentary rocks in the Hub Mountain terrane were intruded by plutonic rocks with a variety of compositions including quartz monzonite, quartz diorite, and diorite. These igneous rocks are now metamorphosed, and their mineral components, which include stilpnomelane, garnet, and usually well-foliated green amphibole and (or) muscovite, suggest that they are partly, if not completely, recrystallized. Foliation, well developed in most places, is missing in some areas of the metadiorite. A few metabasite dikes cut the other rocks. Green amphibole from nonfoliated meta-

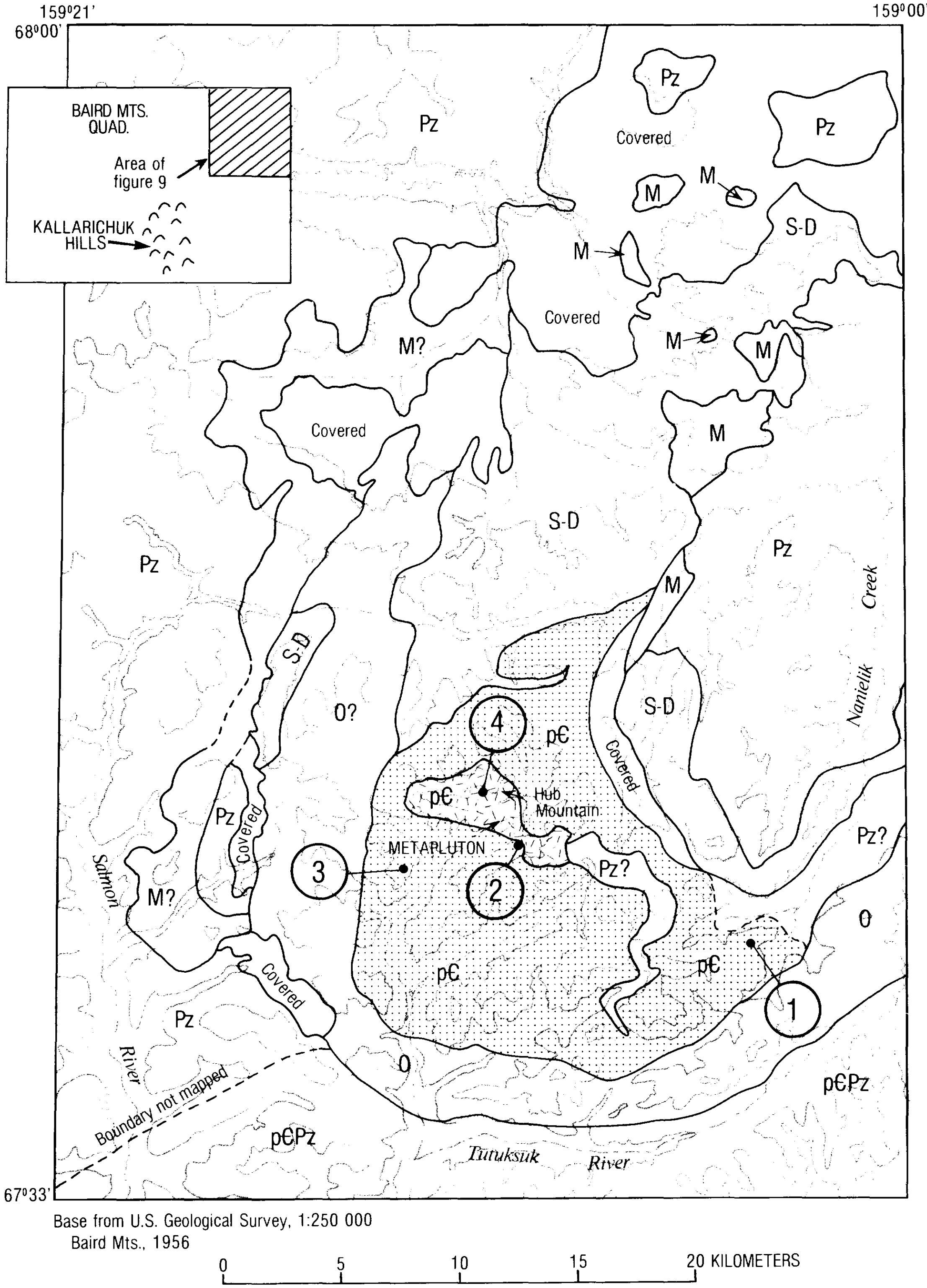

FIGURE 9.—Location of Hub Mountain terrane, stippled area, in the Baird Mountains quadrangle. Symbols refer to ages of deposition of sedimentary rocks or intrusion of igneous rocks: pЄ, Precambrian; pЄPz, Precambrian and (or) Paleozoic; Pz, Paleozoic; O, Ordovician; S-D Silurian and (or) Devonian; M, Mississippian. Numbers are locations of potassium-argon samples. Contour interval 1000 ft.

Table 2.--*Potassium-argon ages from metamorphic minerals in the Hub Mountain terrane, Brooks Range, Alaska*

[Constants used in calculation of ages: $\lambda e + \lambda e' = 0.581X10^{-10}$ yr^{-1}; $\lambda_\beta = 4.963X10^{-10}$ yr^{-1}; $^{40}K/K_{total} = 1.167X10^{-4}$ atom/atom.]

Map location number (fig. 9)	Sample number	Rock type	Mineral	K2O (weight percent)	$^{40}Ar_{(rad)}$ (moles/gX10^{-9})	$^{40}Ar_{(rad)}$ / $^{40}Ar_{(total)}$	Age (in m.y.)		Reference
1	74AF145-1[1]	Muscovite schist (metasediment)	Muscovite (I) Muscovite (II)	9.352 9.352	10.49 10.42	0.98 .97	648±39 644±26	646±19	Turner and others (1979)
1	74AF145-5[1]	Amphibole schist (metabasite)	Hornblende	.740	.9576	.96	729±22		Turner and others (1979)
2	78Md121[2]	Quartz-mica schist (metasediment)	Muscovite	9.27	9.381	.98	594±18		This report
3	76Md276EX[3]	Garnet amphibolite (metabasite)	Hornblende	.505	.5070 .5165	.85 .89	595±30		This report
4	76Md250HX[1]	Meta-diorite	Hornblende (I) Hornblende(II)	1.356 1.319	1.273 1.191	.96 .92	557±17 539±16	548±16	This report

[1]Analyses by D. L. Turner and Dianne Duvall, Geophysical Institute, University of Alaska.

[2]Analyses by M. L. Silberman, L. B. Gray, and M. Taylor, U.S. Geological Survey.

[3]Analyses by M. S. Robison, Teledyne Isotopes (argon) and Paul Klock, U.S. Geological Survey.

diorite at locality 4 (fig. 9; table 2) gives a Cambrian potassium-argon age of about 548 m.y. This date is too young by an undetermined amount because rocks of the plutonic complex intrude the surrounding metasedimentary and metabasite terrane which give Precambrian metamorphic ages and have been clearly metamorphosed to the same degree. It is unlikely the plutonic complex was metamorphosed in the Cambrian because such an event would have reset the potassium-argon apparent ages in the surrounding rocks.

The Hub Mountain terrane is thought to be an antiform ringed by early Paleozoic and Mississippian metasedimentary rocks. Regional metamorphic shear and recrystallization of the Paleozoic rocks probably occurred in the Cretaceous and thus show the extent of younger metamorphic overprint that also must have affected the Hub Mountain terrane. The Paleozoic sedimentary rocks are low-grade greenschist-facies rocks that were deposited as limestone, shale, conglomerate, and sandstone. The limestone is partly recrystallized to marble, but a few fossils are still preserved in it. The shale is now mostly slate, phyllite, or phyllitic schist. At one locality north of the Tutuksuk River, Tailleur and Carter report slate with well-preserved Ordovician graptolites (U.S. Geological Survey, 1975), indicating minimal shear and recrystallization at this locality. The conglomerate has been converted to quartzite with stretched pebbles. The Paleozoic metasedimentary rocks were locally intruded by mafic igneous rocks that have been metamorphosed to fine-grained greenstone. The metamorphic features of these rocks are believed to be the general effects of Cretaceous regional metamorphism on the area. The degree of Cretaceous recrystallization gradually decreases to the north such that Paleozoic metapelites in the south part of the map area (fig. 9) are mostly gray phyllite or phyllitic schist and in the north they are slate or gray phyllite. Although much of the Hub Mountain terrane has been more highly sheared and recrystallized, no attempt has yet been made to map a metamorphic boundary between older more highly recrystallized schist that underwent two periods of metamorphism and younger phyllite and marble that underwent only one period of metamorphism.

The strong radiogenic evidence for a late Precambrian greenschist-facies metamorphic event in the Hub Mountain area lends support to the Precambrian amphibole ages (Turner and others, 1978; Turner and others, 1979) from metabasites in the Kallarichuk Hills 50 km to the southwest (inset map, fig. 9) where it is much harder, if not impossible, to distinguish Precambrian from younger rocks because of the pervasive Cretaceous metamorphic overprint. Carden (1978) reported that a metabasite near locality 1 (fig. 9) contains the blueschist metamorphic mineral glaucophane. This mineral has not been seen in any of several dozen petrographic thin sections from elsewhere in the Hub

Mountain terrane, and therefore, the relative age relations of glaucophane-bearing assemblages to late Precambrian low-pressure metamorphic assemblages around Hub Mountain have not been determined with certainty. However, the occurrence of glaucophane in the Hub Mountain terrane coupled with late Precambrian dates from glaucophane, one in the Kallarichuk Hills and three from the Ambler River country (Turner and others 1978; Turner and others, 1979), lends some support to the contention of Carden (1978) and Turner and others, (1979) that the blueschist event in the southern Brooks Range is also late Precambrian.

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Petrography of nonultramafic rocks from the Avan Hills complex, De Long Mountains, Alaska

By Charles O. Frank[2] and Jay Zimmerman[2]

The general geology of the Avan Hills ultramafic complex has been discussed by Zimmerman and Soustek (1979). Nonultramafic rocks occur in two of the three major lithologic zones. In the transition zone (fig. 10), thin, widely spaced mafic rocks are interlayered parallel to foliation planes of volumetrically dominant ultramafic tectonites. The mafic zone consists almost entirely of gabbroic rocks with lesser amounts of anorthosite, troctolite, biotite granite, tonalite, and serpentinite (Zimmerman and others, 1981).

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In the field, the mafic zone is characterized by regular, laterally persistent compositional layering. Individual layers range in thickness from about 1.5 cm to several meters and are typically defined by changes in relative abundances of plagioclase and mafic minerals or by the appearance of rocks of entirely different bulk composition. The ubiquity of regular compositional layering suggests that most, if not all, of the rocks in the mafic zone were formed by cumulate processes.

Modal analyses of 51 nonultramafic rocks from both zones indicate that mafic material from the transition zone tends to be enriched in olivine relative to rocks from the mafic zone (fig. 11). No systematic differences have been observed in the textures of nonultramafic rocks from the transition and mafic zones.

Cumulate textures are not obvious in all rock types. The anorthosites examined have textures that indicate a cumulate origin, but in gabbroic and quartzofeldspathic rocks, secondary enlargement and, to a lesser extent, replacement reactions hinder the distinction between settled crystals and interprecipitate material (Jackson, 1961), producing textures that are suggestive but not absolutely indicative of cumulate origin. In the discussion that follows, rock types are classified by primary and accessory mineral content (Streckeisen, 1973) rather than by mechanics of crystallization based on textural evidence.

GABBROIC ROCKS

The gabbroic rocks include gabbro, olivine gabbro and gabbronorite.

Olivine and clinopyroxene typically occur either as single grains or as elongate or oval aggregates that comprise clinopyroxene alone; clinopyroxene and olivine; or clinopyroxene, olivine, and plagioclase. The clinopyroxene, which commonly replaces olivine, has mutually interfering boundaries, may be replaced by a second clinopyroxene, and is free of compositional zoning. Plagioclase, if present in the aggregates, occurs between clinopyroxene grains in amounts less than 5 percent. Brown or green amphibole may partly enclose the aggregates. Plagioclase outside the aggregates typically occurs as compositionally zoned anhedral grains with mutually interfering boundaries. Rarely, plagioclase remnants occur within younger plagioclase hosts.

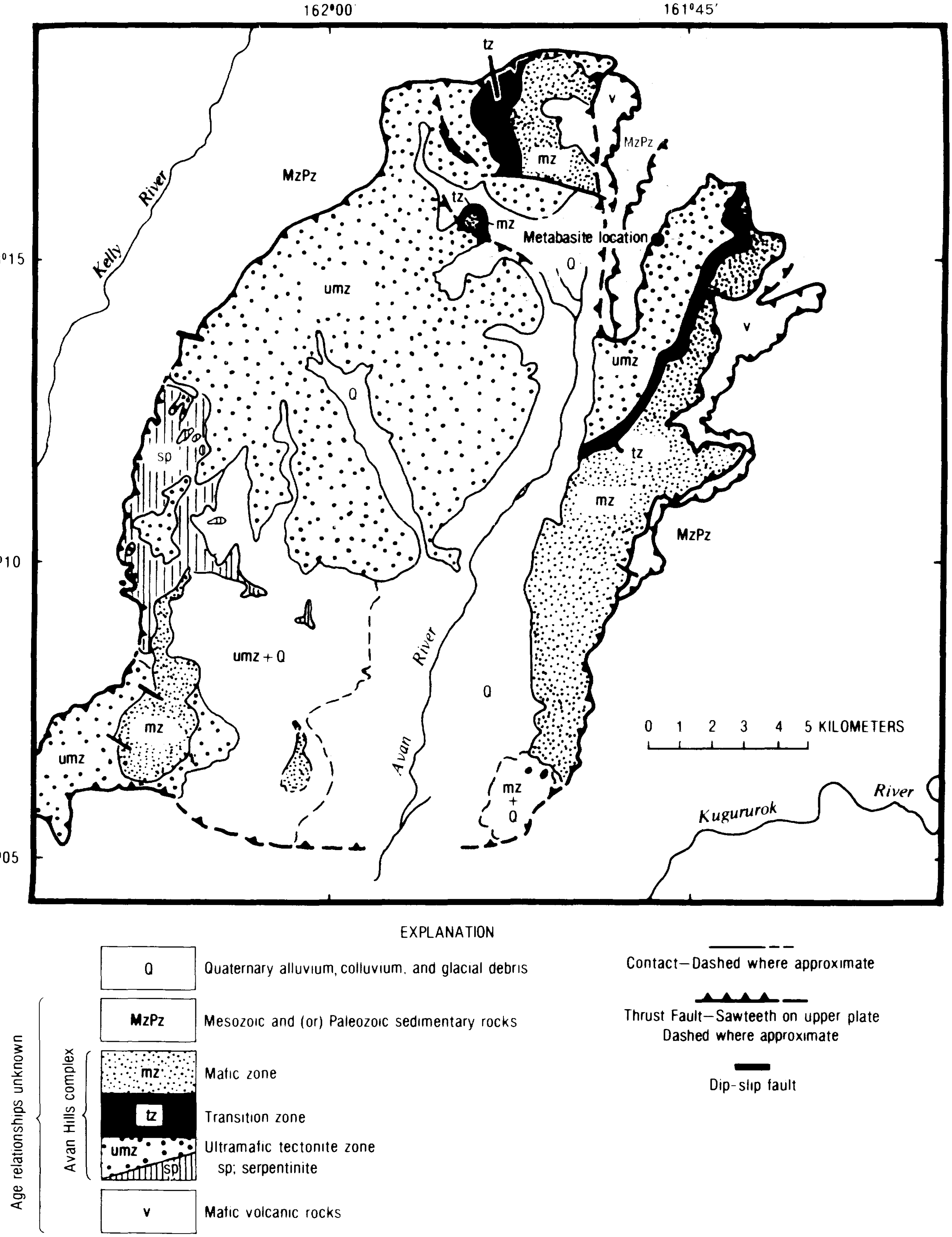

FIGURE 10.—Generalized geologic map of the Avan Hills ultramafic complex. Arrow indicates approximate location of metabasite discussed elsewhere in this Circular by Zimmerman and Frank.

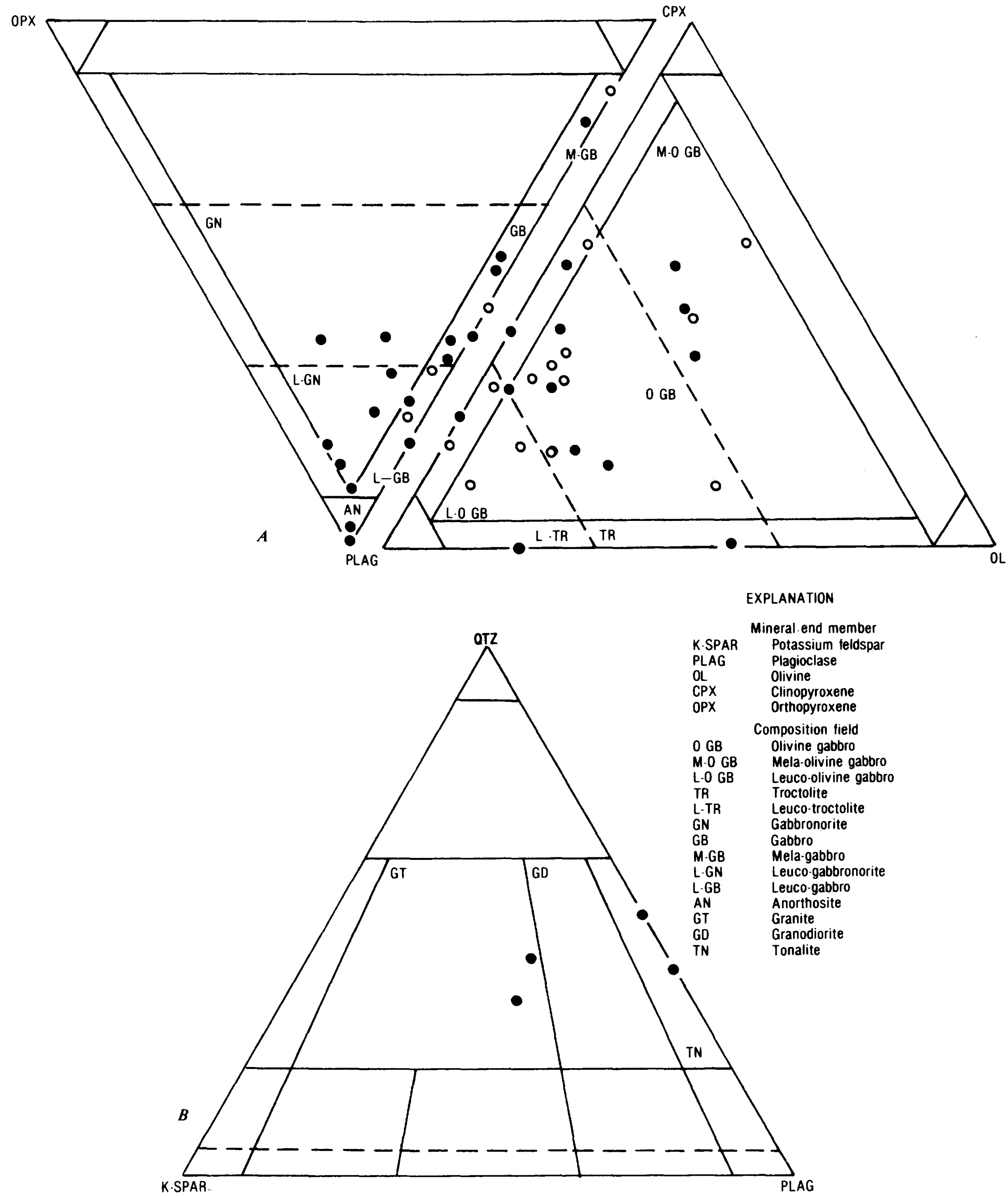

FIGURE 11.—Streckeisen (1973) diagrams showing modal composition of *A*, gabbroic and *B*, leucocratic coarse-grained igneous rocks from transition zone (circles) and mafic zone (dots) of the Avan Hills ultramafic complex. Note tendency toward decrease in modal olivine from transition to mafic zone.

Abundant brown amphibole is limited to rocks from the northernmost part of the western block of the complex (that part of the complex west of the Avan River; fig. 10). In this area, rocks are folded and locally mylonitized and typically contain strained minerals. Cataclastic textures occur in some samples.

Gabbro.—The gabbro contains from 13 to 79 percent plagioclase and from 15 to 57 percent clinopyroxene. Modal averages cluster at 45 and 70 percent plagioclase and at 40 and 25 percent clinopyroxene, defining gabbro and leucogabbro, respectively. Brown amphibole and opaque minerals can be present in combined amounts up to 19 percent in gabbro and up to 7 percent in leucogabbro. Combined amounts of olivine and orthopyroxene, if present, do not exceed 5 percent in either rock type. Other accessories include apatite and green amphibole.

A 3-cm thick leucogabbro (very close to anorthositic composition) contains small gabbro xenoliths and apparently intruded the host gabbro parallel to composition layering. Equant to tabular plagioclase grains are concentrically zoned with fresh material surrounding altered cores. Allanite and secondary(?) siderite are present in the rock. Feldspars show evidence of strain and incipient recrystallization at grain boundaries.

Olivine Gabbro.—The olivine gabbro contains plagioclase (11-68 percent), clinopyroxene (9-58 percent), olivine (5-29 percent), brown amphibole (0-18 percent), opaque minerals (0-14 percent), and orthopyroxene (0-4 percent).

Mela-olivine gabbros have more easily discernible interstitial material than the more leucocratic type. Olivine can occur as individual grains (with or without clinopyroxene rims) but is more commonly associated with aggregates of clinopyroxene and plagioclase.

A transition zone rock contains a plagioclase grain that was apparently broken, rotated, and surrounded by a forcefully injected clinopyroxene liquid that engulfed mineral aggregates elsewhere in the specimen.

Gabbronorite.—The major minerals of the gabbronorite are plagioclase (59-84 percent), clinopyroxene (9-31 percent) and orthopyroxene (9-25 percent). There are modes at 60 percent plagioclase (gabbronorite) and near 75 percent plagioclase (leucogabbronorite). Olivine, brown amphibole, opaque minerals, tourmaline, and apatite are accessories.

The orthopyroxene, commonly serpentinized, occurs as individual grains and in aggregate clusters associated with variable amounts of clinopyroxene, opaque minerals, tourmaline, and plagioclase. Clinopyroxene totally or partly rims many orthopyroxene grains, and brown amphibole partly surrounds individual clinopyroxene grains or the aggregate clusters. Pale-purple tourmaline is associated with the mineral aggregates or occurs as individual grains interstitial to plagioclase crystals.

TROCTOLITE

A troctolite collected from the deformed area of the western mafic zone contains 43.5 percent labradorite and 50.5 percent olivine together with clinopyroxene and pargasitic amphibole as reaction products and brown amphibole as an alteration product. Opaque minerals and green spinel are accessories. Oval, leucocratic mineral aggregates consist of mutually interfering, inclusion-free plagioclase and minor (<1 percent) clinopyroxene. Other, more mafic aggregates consist of olivine and lesser amounts of clinopyroxene, plagioclase, brown amphibole, green spinel, and an opaque mineral and impart a foliation to the rock.

Leucotroctolite from a thin layer oriented parallel to peridotite foliation near the top of the transition zone in the eastern block of the complex contains 76.7 percent plagioclase, 22.1 olivine (partly serpentinized), and about 1 percent chlorite and tremolite. The rock is characterized by widespread inter- and intragranular recrystallization of plagioclase feldspar. Relict crystals are surrounded by and (or) partly overprinted by both strained and apparently strain-free neoblasts. Common neoblast boundaries with 120° intersections suggest that annealing was the primary recrystallization process.

ANORTHOSITE

Two anorthosites, with distinct cumulate textures, consist of approximately 96 percent plagioclase and include green amphibole, biotite, clinopyroxene, zoisite, and an opaque mineral as accessories. Some clinopyroxene grains are partly rimmed by amphibole. Anhedral plagioclase grains are free of compositional zoning and have smooth, mutually interfering boundaries. Plagioclase remnants within later, differently oriented plagioclase crystals are rarely observed. A matrix of slightly finer grained feldspar commonly surrounds the larger grains.

QUARTZOFELDSPATHIC ROCKS

The quartzofeldspathic rocks were all collected from the western mafic zone. Tabular, subhedral to euhedral plagioclase grains have altered, more calcic cores that constitute 10-50 percent by volume of the individual grain and occur as individual crystals or in clusters with mutually interfering boundaries. Pleochroic, reddish-brown biotite with metamictic halos is associated with the plagioclase. Quartz forms anhedral, highly strained, composite grains with inclusions of euhedral to subhedral plagioclase and subhedral biotite. Quartz grain boundaries are lobate. Potassium feldspar occurs interstitially as anhedral, irregularly shaped grains between plagioclase and biotite and between quartz grains. Euhedral plagioclase, biotite, and quartz inclusions are common in potassium feldspars.

Biotite granite.—The two biotite granite samples contain plagioclase (34.8 and 39.1 percent), quartz (37.9 and 31.5 percent), potassium feldspar (19.7 and 22.8 percent), and biotite (7.3 and 6.6 percent) with zircon and pleochroic, blue tourmaline as accessories. The euhedral to subhedral, tabular plagioclase grains are normally zoned with calcic cores and oligoclase rims. In contrast to most rocks in the mafic zone, plagioclase in the biotite granite contains few polysynthetic twins. Strongly pleochroic, reddish-brown biotite occurs as subparallel, tabular grains, some partially altered to penninite. Potassium feldspar grains are anhedral and untwinned and typically contain euhedral plagioclase, biotite, and quartz inclusions. Anhedral quartz grains have lobate boundaries and commonly show undulose extinction.

Biotite tonalite.—Biotite tonalite was collected near one of the anorthosite layers described in a preceding section. Plagioclase (34.3 percent), quartz (33.6 percent), and biotite (29.7 percent) are the major phases, and opaque minerals, apatite, garnet, and tourmaline are accessories. Oligoclase/andesine plagioclase occurs in cyclically zoned euhedral to subhedral, strained crystals. Well-developed polysynthetic twinning is rare. Strained quartz grains with lobate boundaries occur in subparallel lensoidal aggregates. Subparallel alinement of biotite, quartz lenses, tabular plagioclase grains, and the opaque mineral imparts a strong foliation to the rock.

Tonalite.—This sample consists of plagioclase (55 percent) and quartz (36 percent) as major phases; potassium feldspar, biotite, and tourmaline are accessories, and muscovite and chlorite occur as alteration products. The plagioclase forms euhedral to subhedral, tabular grains with sericitized cores. Quartz occurs in subparallel lenses which, together with subparallel alinement of plagioclase crystals, produces a foliation.

This rock is not a plagiogranite (Coleman and Peterman, 1975) because it contains potassium feldspar and biotite. The crystallization of plagioclase earlier than, rather than synchronous with, potassium feldspar and quartz suggests that a classical granophyre model for the origin of the rock is not warranted. The conspicuous normal zoning of the euhedral and subhedral plagioclase grains and the presence of late differentiate phases (biotite, quartz, potassium feldspar, and tourmaline) suggest rapid crystal settling which trapped the surrounding liquid. Diffusion was retarded or ceased, and the interstitial liquid differentiated further in a semiclosed or closed system.

Detailed petrographic examination of these rocks revealed several areas for further study:

1. pressure, temperature, and strain conditions producing static recrystallization of plagioclase;
2. the origin and injection mechanism of late, mafic fluids near the transition zone-mafic zone boundary;
3. the significance of brown amphibole enrichment in highly deformed areas; and
4. genesis of the quartzofeldspathic rocks.

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Possible obduction-related metamorphic rocks at the base of the ultramafic zone, Avan Hills complex, De Long Mountains

By Jay Zimmerman[3] and Charles O. Frank[3]

The presence of metabasite (Miyashiro, 1973) at the base of the ultramafic tectonite zone of the Avan Hills complex suggests that low-pressure metamorphism accompanied obduction of the ophiolite. The rock was collected from the eastern part of the complex at the thrust contact between the ultramafic rocks and underlying mafic volcanic rocks of undetermined age and origin. Its location is shown in figure 10. Although the basal thrust contact of the complex has not been traced in its entirety, traverses have crossed into the underlying sedimentary or volcanic rocks at several locations around its circumference. High-temperature, low-pressure metamorphic rocks of the type described below have been found at no other location, and the sample may represent a unique occurrence in the Avan Hills complex.

The rock contains both nonfoliated and foliated layers. The nonfoliated part consists of plagioclase (38.5 percent), quartz (24.0 percent), high-iron cummingtonite (13.6 percent), and iron-rich orthopyroxene (11.7 percent), as major phases. Garnet, zircon, ilmenite, iron sulfide, and apatite are accessories. The plagioclase (andesine) exists as anhedral to subhedral, equant, untwinned grains, partly altered to epidote(?). Orthopyroxene grains are equant and partly to completely altered to serpentine. Quartz is present as inequant, strained crystals. Amphiboles are anhedral, embayed by quartz, and contain inclusions of plagioclase and opaque minerals. The amphibole has been identified as cummingtonite by optical methods, and a qualitative (X-ray) analysis suggests that the Fe:Mg ratio is high. The texture of the nonfoliated parts of the rock conforms to that of a fine-grained hornfels.

Foliated layers are enriched in orthopyroxene (29.7 percent), occurring as either altered or unaltered grains, and ilmenite (11.4 percent) but are depleted in quartz (5.0 percent). Zircon, tourmaline, and an iron sulfide are accessories. Unaltered cummingtonite and orthopyroxene crystals are highly embayed and contain numerous inclusions of quartz, plagioclase, and opaque minerals. Foliation is most obviously defined by planar concentrations of ilmenite although some optical alinement of orthopyroxene is apparent.

Occurrences of high-temperature metamorphic rocks at the structural bases of obducted ophiolites are not unusual (Woodcock and Robertson, 1977). Such rocks may exist as relatively continuous subjacent layers as in the Bay of Islands Complex of Newfoundland (Williams and Smythe, 1973) or as thin, discontinuous, structurally isolated fragments as in the Vourinos Complex of northern Greece (Zimmerman, 1972). Recent interpretations suggest that these rocks represent metamorphism of supracrustal rocks in the vicinity of subduction zones during early phases of obduction. Requisite temperature is attained from the residual heat of the ophiolite or from friction during emplacement (Woodcock and Robertson, 1977). The fragmentary distribution of typical metamorphic remnants is explained by attrition during subsequent thrust transport (estimated to be at least 120 km for the Avan Hills complex by Roeder and Mull, 1978).

The assemblage of cummingtonite, orthopyroxene, and plagioclase places the Avan Hills rock between the hornblende-hornfels and pyroxene-hornfels facies (Miyashiro, 1973). Garnet is typically rare in very low pressure metabasites produced by thermal (contact) metamorphism. Its occurrence in trace amounts in the Avan Hills rock may indicate a steeper thermal gradient and is compatible with elevated pressure caused by weight of the overlying ophiolite. The transition temperature indicated is somewhat higher than that of many ophiolite-related metamorphic rocks, which are typically

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low-grade amphibolites or greenschists. Rocks of similar composition have been reported from the Seward Peninsula (Throckmorton and Hummel, 1979), and the Avan Hills rock could be an allochthonous slice of the basement complex. Its restricted distribution, location at the base of the ophiolite, and the absence of similar rocks in the thrust sequences of the De Long Mountains, as well as evidence for a high-temperature, low-pressure origin, however, suggest the obduction-related alternative. If this is the case, the elevated temperature indicated by the assemblage implies that the ophiolite was a slice of thermally immature oceanic lithosphere and that obduction occurred within a few hundred kilometers of an active ridge system. Such evidence may ultimately place constraints on the size and (or) configuration of the ocean basin in which the Avan Hills complex and other ophiolites in the western Brooks Range originated.

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Significance of the Endicott Group for tectonic models of the Brooks Range

By Tor H. Nilsen, William P. Brosgé, Thomas E. Moore, J. Thomas Dutro, Jr., and Donna F. Balin

The Endicott Group, a thick sequence of chiefly Upper Devonian and Lower Mississippian clastic sedimentary rocks (Tailleur and others, 1967), crops out for a distance of about 1,000 km along the strike of the Brooks Range in northern Alaska (fig. 12). It contains both marine and nonmarine units and in the central and eastern Brooks Range includes, from oldest to youngest, the marine Hunt Fork Shale, marine and nonmarine Kanayut Conglomerate, nonmarine Kekiktuk Conglomerate, and marine Kayak Shale. The entire Endicott Group was thought to form a major Late Devonian and Early Mississippian offlap-onlap cycle in which the two conglomeratic units formed the middle nonmarine parts of the cycle. Tailleur (1973) concluded that the Endicott Group formed a major postorogenic clastic wedge shed southward from an uplifted area now to the north.

The Endicott Group has played a major role in the development of models to explain the origin and tectonic evolution of the Brooks Range. Several models have proposed detachment of the Endicott Group from a similar Devonian clastic wedge in the Innuitian foldbelt of the Canadian Arctic Islands by Mesozoic counterclockwise rotation of the Brooks Range away from the Arctic Islands about a pivot located near the Mackenzie Delta (Carey, 1958; Tailleur, 1969a, b, 1973; Rickwood, 1970; Freeland and Dietz, 1973; Richards, 1974; Boucher, 1978; and Newman and others, 1979). Some models have proposed detachment of the Brooks Range from the Innuitian foldbelt by left-lateral translation along a fault extending along the present shelf-slope break north of the Arctic Islands to the Porcupine megashear (Herron and others, 1974; Yorath and Norris, 1975; Dutro, 1981). Other models propose that the Brooks Range and Innuitian foldbelt have always been separated and in their present relative positions (Churkin, 1973a, b, 1975; Norris, 1973, 1974). The Endicott Group and the Canadian Arctic Islands wedge would thus have been deposited in separate basins and derived from separate northern source areas. The intervening Canada Basin, according to these models, could be as old as late Precambrian or Paleozoic.

Churkin and Trexler (1980) suggested that the Brooks Range, with Chukotka in eastern Siberia, originally formed a northwest-extending prong of North America that was offset to the northeast by right-lateral translation along the Porcupine megashear. Jones (1980) derived the Brooks Range from an original southerly position by right-slip on the Tintina and Kaltag faults, inferring that the Canada Basin existed as early as the Permian.

All of these models for the Brooks Range require clear understanding of the geology of the Endicott Group and the Canadian Arctic Islands clastic wedge. The Canadian Arctic Islands wedge is as thick as 5,000 m, extends southwestward from Ellesmere Island to Banks Island, and consists mostly of quartz and chert detritus transported southwestward. Source areas located to the north and east included the North Greenland foldbelt, Caledonian orogen, and Greenland Shield (Embry and Klovan, 1976). The shallow-marine to fluvial part of the wedge ranges in age from early Givetian to middle Famennian. Broad uplift took place in late Frasnian time, and extensive orogenic folding and uplift occurred in the latest Devonian or Early Mississippian.

The Endicott Group consists of two distinct sequences that must be distinguished in the regional tectonic framework. The first sequence may be entirely allochthonous and consists of the Upper Devonian Hunt Fork Shale, the Upper Devonian Noatak Sandstone, the Upper Devonian Kanayut Conglomerate, and the Mississippian Kayak Shale. The allochthonous sequence is as thick as 4,000 m, rests conformably on older Paleozoic marine rocks, and crops out in the central part of the Brooks Range in a series of north-transported thrust plates. The fluvial Kanayut Conglomerate, entirely Upper Devonian (Famennian), is as thick as 3,000 m and forms the middle part of a major delta that prograded southwestward and then

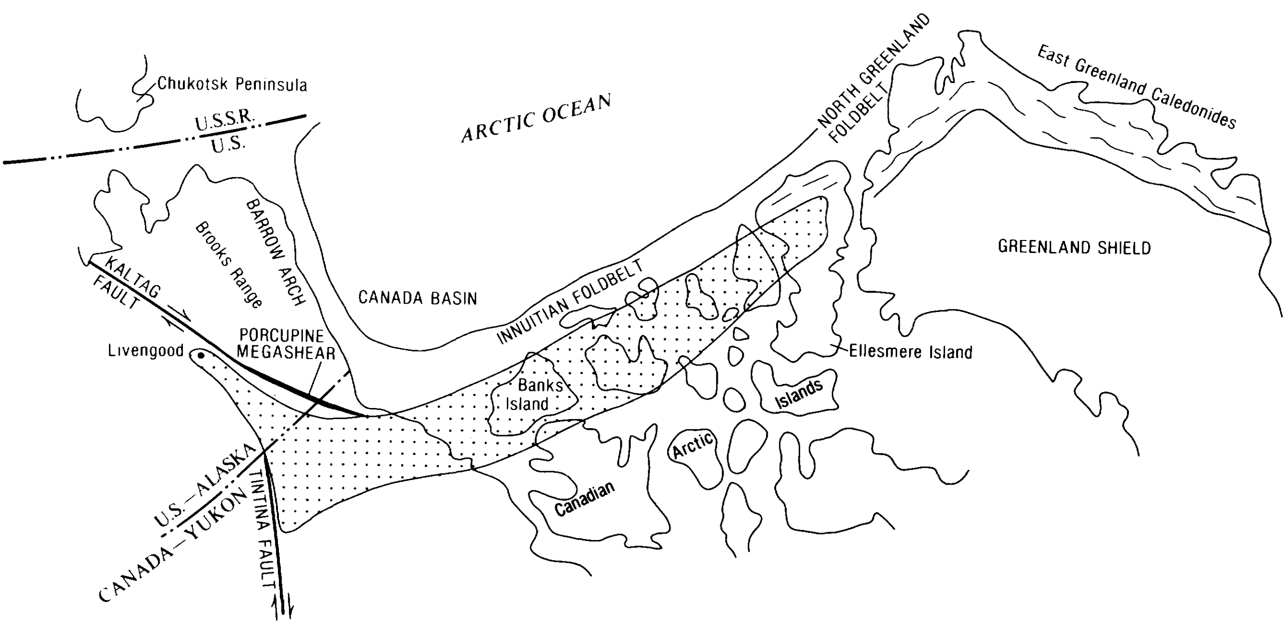

FIGURE 12.—Sketch map showing relation of Canadian Arctic Islands clastic wedge and submarine fan deposits of Nation River Formation (stippled pattern) to Brooks Range and Tintina and Kaltag faults.

retreated (Nilsen, Moore, and Brosgé, 1980; Nilsen, Moore, Dutro, and others, 1980; Nilsen, Brosgé, and others, 1980).

The autochthonous sequence consists of the fluvial Kekiktuk Conglomerate and marine Kayak Shale. It is generally several hundred meters thick, rests unconformably on deformed pre-Upper Devonian sedimentary, volcanic, and metamorphic rocks or Devonian granite, and crops out north, east, and south of the allochthonous sequence, separated from it by thrust faults. The Kekiktuk Conglomerate appears to be entirely Early Mississippian in age and forms a thin cover of southwest-transported fluvial deposits. Both the Kanayut and Kekiktuk are composed chiefly of chert, quartz, and quartzite clasts, but the proportions of each component differ in each unit.

The present distribution of the allochthonous and autochthonous sequences of the Endicott Group can be accounted for by two reconstructions. The allochthonous sequence may have been thrust northward, as suggested by Tailleur and Snelson (1969) and Roeder and Mull (1978). It originally could have been deposited on a separate plate to the south or on the south margin of the Brooks Range basement, then thrust over the autochthonous sequence during Mesozoic orogeny. This reconstruction suggests that the two sequences of the Endicott Group may be completely unrelated.

A second reconstruction suggests that the allochthonous sequence was deposited in a central trough located in the area of the present-day Brooks Range, that was flanked by uplifted areas which were later eroded down to permit deposition of the Kekiktuk Conglomerate around the margins of the Kanayut. Later compression during the Mesozoic produced folding and thrusting more or less in place (Dutro and others, 1976). This reconstruction does not require long-distance tectonic transport, and the Kekiktuk should underlie the Kanayut only on the faulted northern and southern flanks of the Range.

Restoration of the allochthonous and autochthonous sequences of the Endicott Group with either the Arctic Islands clastic wedge or the marine Upper Devonian rocks of the Nation River Formation of east-central Alaska presents formidable problems. The Arctic Islands wedge is mostly older than the Endicott Group, has no equivalent to either the Mississippian Kekiktuk Conglomerate or Kayak Shale, does not generally lie on or near granitic plutons, and is generally finer grained than both the allochthonous and autochthonous sequences. The Nation River Formation is a major coarse-grained submarine fan deposited by west-flowing turbidity currents (Nilsen and others, 1976). It is overlain and underlain by deep-marine shale units and appears to form a part of the continental margin of North America, unaffected by large-scale movements along the Tintina fault and Porcupine megashear. There is no regional geologic or sedimentologic evidence for a landmass that would have simultaneously shed detrital material northward into the basin where the Endicott Group sediments accumulated and southward into the basin in which the sediments of the Nation River were deposited.

Left-lateral translation of the Brooks Range from the Arctic Islands restores the two Devonian clastic wedges to a subparallel position with sediment transport to the southwest. Counterclockwise rotation of the Brooks Range away from the Arctic Islands yields subparallel wedges transported in opposite directions. Neither translation nor rotation resolves differences in age and clast size between the two Devonian wedges. Keeping the wedges in their present positions produces an unlikely juxtaposition of the fluvial deposits of the Endicott Group and submarine fan deposits of the Nation River Formation. Restoration of the Endicott Group southeastward along the Tintina fault juxtaposes the thick fluvial deposits of the Endicott Group with deep-marine facies of the Antler orogenic belt, an unlikely scenario. Restoration of the Endicott Group southwestward along the Porcupine megashear to a position adjacent to marine Devonian rocks of the Livengood area is not supported by facies changes, sediment transport data, or clast composition data.

In conclusion, examination of the geology of the Endicott Group and related Upper Devonian and Lower Mississippian rocks of Alaska and Arctic North America at this time fails to provide a suitable reconstruction for the Brooks Range on the basis of the important Devonian clastic wedges. The most reasonable scenario seems to be left-lateral slip along the northwest edge of the Canadian Arctic Islands, for it at least provides juxtaposition of subparallel belts characterized by reasonably similar geologic histories. However, even this reconstruction requires better understanding of the differences between the depositional framework and tectonic history of the allochthonous and autochthonous sequences of the Endicott Group before it can be favored.

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Depositional facies and reservoir potential of the Fortress Mountain Formation, central North Slope

By Cornelius M. Molenaar, Robert M. Egbert, and Lee F. Krystinik

The Fortress Mountain Formation of early Albian (Early Cretaceous) age is a thick clastic wedge that was deposited in a foredeep in front of the Brooks Range orogen. A stratigraphic field investigation of this unit was made during a one month period in 1980 in the central part of the southern foothills between the Chandler and lower Nuka Rivers (Rs. 2-31 W.), a distance of 280 km. The purpose of this study was to assess the hydrocarbon reservoir potential of the Fortress Mountain Formation in the subsurface to the north. Because of structural complexities and the discontinuous nature of the outcrops, we emphasized interpretation of depositional environments and measurement of paleocurrent direction to project facies trends. This report is a summary of a more complete report by Molenaar and others (1981).

The Fortress Mountain Formation is as thick as 3000 m and consists of shale, sandstone, and conglomerate (Patton and Tailleur, 1964; and Chapman and others, 1964). The sandstone is greenish-gray, moderately to poorly sorted, very fine- to coarse-grained graywacke in the southern part of the outcrop area and becomes finer grained to the north. The sandstone consists mostly of lithic grains in a clayey, chloritic matrix. The conglomerate is mostly poorly sorted and disorganized, and consists of pebbles and cobbles with minor boulders. The clasts are composed of variable proportions of chert, mafic igneous rocks, and sandstone or quartzite with lesser amounts of limestone and shale. Porosity in both the sandstone and conglomerate appears to be low.

The Fortress Mountain Formation along the outcrop belt consists largely of deep-water deposits that were rapidly deposited into a relatively steep-sided asymmetric basin (fig. 13). Deposits interpreted to be nonmarine were noted in only two areas: the upper 150 m of Castle Mountain (T. 10 S., R. 3 W.) and at least 180 m of section on the southwest side of Ekakevik Mountain (T. 10 S., R. 22 W.). The abrupt transition from nonmarine to shallow marine to deep marine deposits indicates a very narrow shelf and a steep basin slope. Many of the deep-water deposits generally fit into the turbidite fan-facies model of Mutti and Ricci Lucchi (1972). Most of the thick conglomerate units are interpreted to be inner fan channel or submarine canyon facies (fig. 13).

The formation appears to have been deposited from many point sources along the ancestral Brooks Range as indicated by changes in conglomerate clast composition in different areas and abrupt lateral facies changes along the depositional strike. An example of this is the inferred relation between the section at Fortress Mountain (T. 10 S., R. 5 W.) and the section at Castle Mountain, two conglomerate-capped mountains about 16 km apart along an east-west line. The 300-m-thick conglomerate unit at Fortress Mountain, which is interpreted to be a submarine canyon facies, grades to a predominantly sandstone unit 2.5 km north across a syncline. To the east it thins and apparently grades into a predominantly shale section in the lower part of the 3000-m-thick type section of the Fortress Mountain Formation near Castle Mountain. These relations and the many apparently discontinuous thick lenses of coarse clastic material support the interpretation that

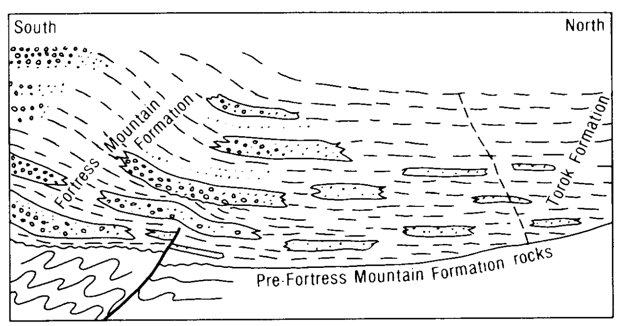

FIGURE 13.—Diagrammatic cross section across southern foothills showing inferred facies relations of Fortress Mountain and Torok Formations prior to intense folding. Horizontal beds in upper left are alluvial and shallow-marine facies, tilted beds are slope shales and submarine canyon facies, and lower horizontal beds are basinal shale and turbidite facies. No scale intended.

the Fortress Mountain Formation is composed of a number of, in part, overlapping deep-sea fan complexes, each one of limited lateral extent.

Rapid downdip changes into the basin are also apparent, even though direct correlations are usually not possible because the area has been telescoped into complex folds and faults. However, tubidites of the Fortress Mountain Formation are finer grained and more thinly bedded to the north. All the thick conglomerate sections are located on the south side of the outcrop belt. The rapid change from conglomerate to sandstone noted at Fortress Mountain is also exemplified by the long, well-exposed, but highly folded section along the Kiligwa River (Tps. 5, 6, 7, and 8 S., R. 28 W.). In the southern part of the section, there are many conglomerate units that are interpreted to represent inner-fan channel deposits. About 16-20 km north, in the vicinity of the Brady anticline, the Fortress Mountain Formation consists of thinner bedded, finer grained turbidites of an outer-fan or basin-plain facies. These finer grained rocks of the Fortress Mountain Formation grade into the lower part of the Torok Formation to the north (fig. 13).

Paleocurrent directions from the outer-fan or basin-plain facies associations indicate a current flow to the east-northeast. Although these sands were derived from the south, the turbid flow directions were influenced by an easterly slope of the depositional basin. In the northern foothills, seismic data indicate a gentle south-dipping basin slope during the time of deposition of the lower part of the Torok Formation and (or) Fortress Mountain Formation (Molenaar, 1981a, b). Thus the northern part of the southern foothills area may coincide with the depositional axis of the Colville basin during at least part of Fortress Mountain time (fig. 13). If that were true, very little sandstone or a more distal facies would be expected in the subsurface farther north. However, because of the many point sources contributing the coarse clastic material of the Fortress Mountain Formation, individual fans could extend farther north.

The best opportunity for finding thicker sandstone packages of the Fortress Mountain Formation in the subsurface would be in the southern part of the northern foothills, where large anticlinal structures are present. However, on the basis of the lithic character of the sandstone studied, reservoir properties would probably be poor.

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Cretaceous-lower Tertiary depositional relations, northeastern Alaska

By Cornelius M. Molenaar, A. Curtis Huffman, and Allan R. Kirk

Recent stratigraphic fieldwork in the Sadlerochit Mountains area of northeastern Alaska indicates that the total Cretaceous section is relatively thin and consists largely of deep-water deposits. This report briefly summarizes the evidence and presents a revised interpretation of the depositional history of these strata.

In the Canning River-Sadlerochit Mountains area, the Kemik Sandstone Member of the Kongakut Formation—a shallow-marine sandstone at the base of the pebble shale transgression—rests unconformably from northeast to southwest on truncated Lower Triassic to Jurassic strata (Mull and Kososki, 1977, p. 21). The Kemik is overlain by the pebble shale member of the Kongakut Formation, a transgressive shale about 75 to 150 m thick that is widespread throughout northern Alaska.

Foraminiferal data from outcrops in the area indicate the pebble shale member is Neocomian (Early Cretaceous) in age (Palmer and others, 1979, locs. 16 and 19). The Sadlerochit Mountains area is suggested here to be a continuation of the Barrow arch structural trend, and the major unconformity and transgression cited is the same as that recognized in the Barrow and Prudhoe Bay areas.

A stratigraphic section measured on the south side of Ignek Valley in the NE1/4 sec. 6, T. 2 N., R. 26 E. (fig. 14) is thought to be representative of Cretaceous strata in the area surrounding the Sadlerochit Mountains. On the basis of foraminifers and radiolarians, and on the bentonite content and organic nature of the shale, the lower 125 m of the Colville Group in Ignek Valley is correlated by Detterman and others (1975, p. 34) with the lowest part of the Colville—the Shale Wall Member of the Seabee Formation of Turonian (Late Cretaceous) age—in the Umiat area to the west. The reported major unconformity between the pebble shale member

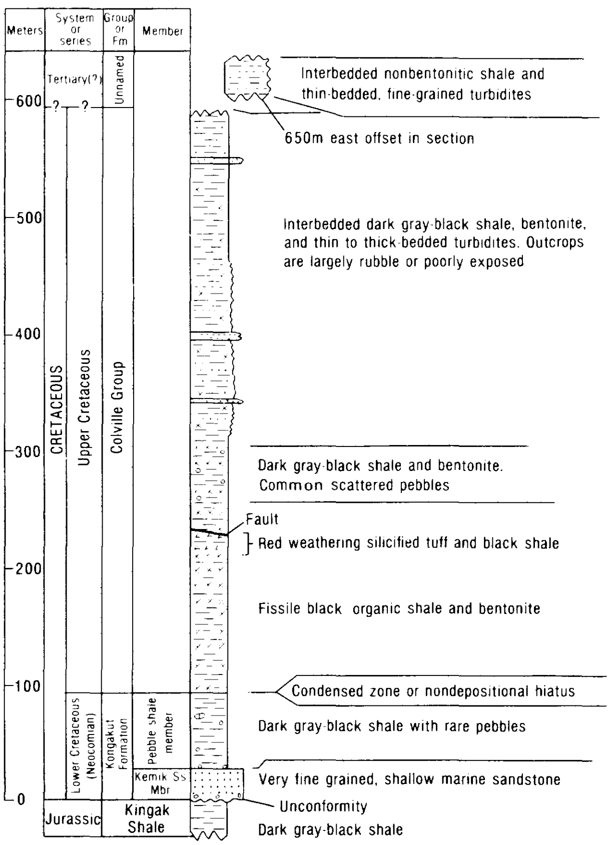

FIGURE 14.—Cretaceous section in Ignek Valley, northeastern Alaska.

and overlying Upper Cretaceous shales (Detterman and others, 1975, p. 32) is interpreted here to be a condensed or nondepositional zone that was never subjected to subaerial or wave-base erosion.

One or two red-weathering zones, which brightly color the subdued outcrops around the Sadlerochit Mountains, occur in the upper part of or just above the organic shale of the Colville Group. The red coloration is due to surface oxidation of silicified or devitrified tuff in the section. A fault of unknown, but probably not large, displacement cuts out section just above the red zone. Micropaleontologic data indicate that the upper half of the continuous section is Late Cretaceous in age. It consists of bentonitic shale containing several zones of fine-grained turbidites. Most of these turbidite beds are thin, but a few are as thick as 3 m. Sole markings are common. Outcrops of these sandstone beds are generally fractured and reduced to rubble.

A 30-m-thick section of thin-bedded turbidites and nonbentonitic shale is infolded in the Ignek Valley syncline a few hundred meters east of the section just described. Two similar-appearing turbidite sequences, one exposed 19 km to the west on the west bank of the Canning River and the other exposed 23 km to the northeast along the Katakturuk River, have been dated as probable Paleocene in age on the basis of palynomorphs (Palmer and others, 1979, locs. 18 and 8). If the correlation of these sections with the top of the Ignek Valley section is correct, the total Cretaceous section is only 600 to 700 m thick.

In the Arctic Creek-lower Kekiktuk River area (T. 2 N., Rs. 30 and 31 E.) east of Ignek Valley, most, if not all, of the sandstones that have been mapped as the Cretaceous Nanushuk Group (Reiser and others, 1971) are here interpreted to be turbidites. Most are interbedded with bentonitic shales and therefore are of probable Late Cretaceous age inasmuch as bentonites are not common in the Early Cretaceous of the North Slope. The occurrence of the Albian ammonite *Paragastroplites spiekeri*, reported by Detterman and others (1975, p. 29) in the Arctic Creek area, indicates the presence of Nanushuk-age strata in this area, but our observations suggest that no more than a limited part of the section is latest Early Cretaceous; the remainder is Late Cretaceous in age.

Paleocurrent directions determined from the few good exposures of Cretaceous and lower Tertiary turbidites throughout the area around the Sadlerochit Mountains indicate an average current flow to the east-northeast. The turbidites were probably derived from deltas to the west or southwest rather than from canyon-fed systems.

Available paleontologic data and physical relations based on interpretations of depositional environments suggest a depositional pattern of easterly or northeasterly prograding deltaic-to-slope-to-basinal deposits in which the downlapping post-Neocomian deposits become younger from southwest to northeast. This pattern is well documented by seismic and well data in the thick interval of the Torok Formation and Nanushuk Group (Albian, Early Cretaceous) in National Petroleum Reserve in Alaska (Molenaar, 1981a, b). This same depositional pattern is thought to continue into northeastern Alaska in younger strata. The condensed section or probable absence of Aptian and Albian rocks in the Ignek Valley-Sadlerochit Mountains area could be due to the higher elevation of that area (though still in deep water) relative to the Colville basin axis to the south at that time, so that it did not receive the south and southwesterly derived sediments until Late Cretaceous time.

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The Story Creek and Whoopee Creek lead-zinc-silver occurrences, western Brooks Range, Alaska

By Inyo F. Ellersieck, Uldis Jansons[4], Charles F. Mayfield, and Irvin L. Tailleur

Geologic investigations in the northwestern Brooks Range during the past five years have identified a province of zinc-lead-silver occurrences hosted by sedimentary rocks of late Paleozoic age. These sulfide mineral occurrences are all in rocks of the Brooks Range allochthon (Ellersieck and others, 1979), which comprises a distinctive succession of Devonian through Lower Cretaceous formations. Two sulfide occurrences, at Story Creek and Whoopee Creek in the Howard Pass quadrangle, differ from occurrences further west at Red Dog (Plahuta and others, 1978), Drenchwater (Nokleberg and Winkler, 1978), and Ginny Creek (Mayfield and others, 1979).

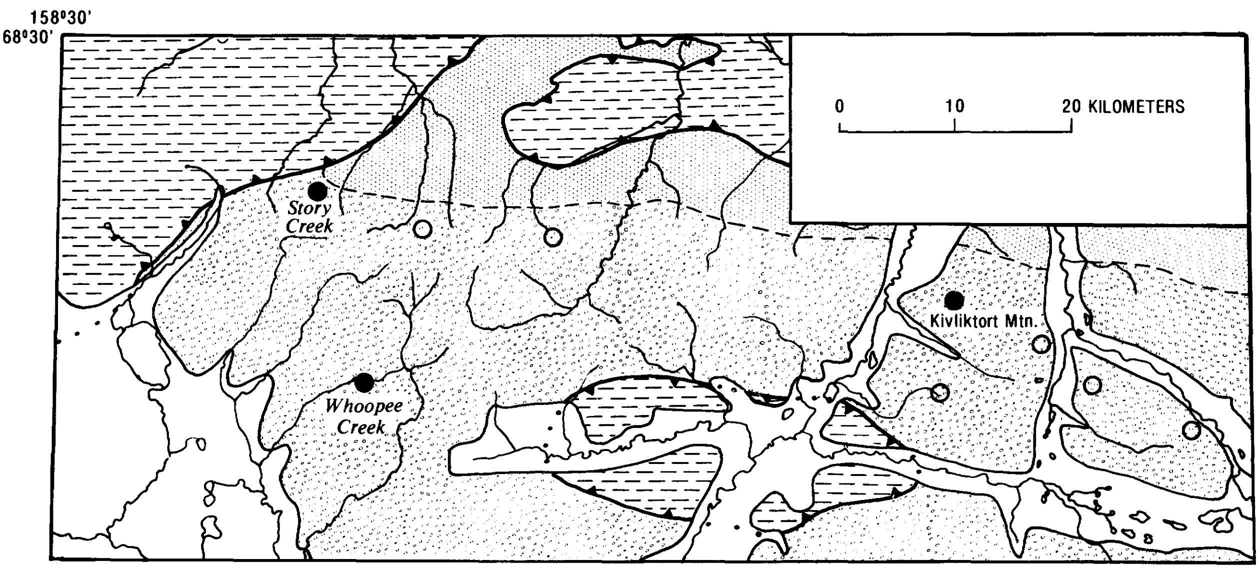
The Story Creek occurrence is located on a topographic bench south of upper Story Creek, a tributary of the Kuna River. The best exposure of sulfides is in a small stream cut at lat 68°23′25″ N., long 157°55′50″ W. (figs. 15 and 16). Three types of sulfide mineralization are present in this outcrop: banded massive sphalerite and galena, brecciated sphalerite with galena matrix, and shale-chip breccia with overgrowths of quartz on the chips and galena and sphalerite in the interstices. A grab sample of high-grade banded sulfide contained 0.43 percent Cu, 34 percent Pb, 28.8 percent Zn, 1.2 ppm Au, and 940 ppm (27 tr oz/ton) Ag. Other composite or grab samples from within the same mineralized zone contained from 1.5 to 15.5 percent Pb, 1.5 to 50 percent Zn, and 35 to 500 ppm Ag. The apparent width of the mineralized zone at this location is 10 to 15 m. Float and outcrops of the quartz-galena-sphalerite breccia have been traced for a distance of 3 km along a N. 65° E. trend. The nearly straight trend, in an area where the bedding is tightly folded and faulted, makes it likely that the breccia is in a crosscut-

[4]U.S. Bureau of Mines, Anchorage, Alaska.

ting fracture zone, rather than conformable with bedding. On the southwest end of this fracture system, a northwest-flowing creek exposes several parallel breccia zones that also contain galena and sphalerite. Some of these breccias contain open vugs, or vugs filled with an unidentified bituminous material. A few milky quartz veins parallel axial-plane cleavage in the host rock, but it is not known whether these are related to the mineralized fractures.

The host rock at Story Creek is an interval at the top of the Stuver Member of the Upper Devonian Kanayut Conglomerate and the bottom of the Lower Mississippian Kayak Shale. It is predominantly dark-reddish-brown siltstone and silty shale, with interbedded sandstone and minor amounts of dolomite and coal. The bedding is tightly folded and distrupted by numerous thrust faults. Axial planes of the folds strike northeast and dip to the south. The host rock is in thrust contact above Carboniferous through Lower Cretaceous sedimentary rocks typical of the Brooks Range allochthon. It is probably in thrust contact beneath sandstone of the Kanayut Conglomerate to the south, although this contact becomes an overturned stratigraphic contact within 10 km eastward.

The Whoopee Creek sulfide occurrence is located 17 km south of the Story Creek occurrence, in the Noatak River drainage at lat 68°13′45″ N., long 157°51′20″ W. (figs. 15 and 16). A S. 65° E.-trending, steeply dipping, 6-m-wide fracture zone contains siltstone breccia with a matrix of quartz, galena, sphalerite, and minor amounts of carbonate. Mineralized breccia contained as much as 0.24 percent Cu, 44 percent Zn, 4.4 ppm Au, 460 ppm (13.4 tr oz/ton) Ag, and 3,700 ppm Cd; however, these high values are not found together in the same sample. Mineralized breccia is exposed or found in float over a distance of at least 1 km, but the limits of the mineralization have not been determined.



EXPLANATION

Quaternary deposits

Misheguk Mountain, Copter Peak, Ipnavik(?), and Picnic Creek allochthons, undivided

BROOKS RANGE ALLOCHTHON

Cretaceous to Lower Mississippian rocks, undivided

Lower Mississippian and Upper Devonian Endicott Group: includes Kayak Shale, Kanayut Conglomerate, and Hunt Fork Shale

Contact—Dashed where approximate

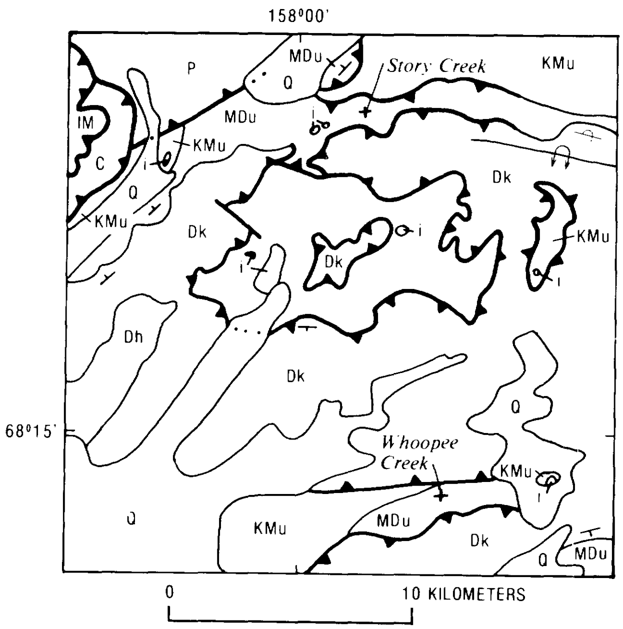
Thrust fault—Sawteeth on upper plate. Dotted where inferred

Sulfide occurrence

Geochemical anomaly

FIGURE 15.—Generalized geologic map of part of the south-central Howard Pass quadrangle, showing the locations of lead-zinc-silver sulfide occurrences (solid circles) and stream-sediment geochemical anomalies (open circles).

The Story Creek and Whoopee Creek occurrences are in the same stratigraphic interval and similar structural settings. In the hillside immediately above Whoopee Creek is a thrust sheet of a sandy interval of the Kanayut Conglomerate, which is continuous with the sheet immediately south of the Story Creek occurrence.



EXPLANATION

- Q — Quaternary deposits
- IM — Misheguk Mountain allochthon
- C — Copter Peak allochthon
- P — Picnic Creek allochthon

BROOKS RANGE ALLOCHTHON

- KMu — Cretaceous to Mississippian rocks, undivided
- i — Jurassic(?) and Mississippian igneous rocks, undivided
- MDu — Lower Mississippian Kayak Shale and Upper Devonian Stuver Member of Kanayut Conglomerate, undivided
- Dk — Upper Devonian Kanayut Conglomerate
- Dh — Upper Devonian Hunt Fork Shale

Contact

Thrust fault—Sawteeth on upper plate. Dotted where inferred

Anticline

Strike and dip of beds

Inclined

Overturned

+ Lead-zinc-silver mineral occurrence

FIGURE 16.—Sketch map of area near Story Creek and Whoopee Creek sulfide occurrences.

These sulfide occurrences differ in several ways from previously described occurrences in the Brooks Range allochthon. The exposed sulfides occur primarily in veins or breccias, instead of being predominantly stratiform, as at Red Dog or Drenchwater, or disseminated, as at Ginny Creek (Mayfield and others, 1979). The lead to zinc ratios indicate that lead is as abundant as, if not more abundant than, zinc. Silver contents of the samples are much higher than any reported from other mineral occurrences in the Brooks Range allochthon, possibly reflecting the larger proportion of galena.

Rocks similar to the host rocks at Story Creek and Whoopee Creek underlie large areas in the Endicott Mountains, Mulgrave Hills, and western Baird Mountains. Although the circumstances leading to the mineralization are currently unknown, there is a good chance that they were duplicated elsewhere in this extensive terrane.

Regional stream-sediment geochemical sampling gave the initial indications that led to the discovery of Story Creek and Whoopee Creek sulfide occurrences. A belt of geochemical anomalies extends from Story Creek along the northern front of the Brooks Range east to the Nigu River (Theobald and Barton, 1978). A brief reconnaissance revealed cobbles of high-grade zinc mineralization in float at Kivliktort Mountain, and zinc and lead mineralization at several other localities (Jansons and Parke, 1981). These results indicate that there is a high potential for lead-zinc-silver occurrences of the Story Creek type in rocks of Middle Devonian to Mississippian age over at least 2000 km^2 of the Howard Pass quadrangle.

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Petrography and cathode luminescence of carbonate rocks at Bornite, Alaska

By Augustus K. Armstrong

The Bornite copper deposit is on Ruby Creek in the Cosmos Hills on the southwest side of the Brooks Range, Alaska. Runnells (1963) gives a summary of early investigations of the deposit as described by R. H. W. Chadwick (written commun., 1960) and subsequently Runnells (1965, 1969) published detailed accounts of the mineralogy and sulfur istopes. Additional description of the deposit is given by Mayfield and Grybeck (1978). The author's recent study of cores and of thin sections made from core samples from four exploration drill holes at Bornite developed some significant new information about the mineralized carbonate rocks and their history.

The carbonate rock section at Bornite is formed by breccia fragments from micrometers to meters in size. Dolomite breccia clasts preserve many relic sedimentary features and fossil fragments. Relic textures in some dolomites suggest that they were derived from lime mudstones (micrites). Some dolomite clasts have a well-preserved micropellet to peloid fabric; others contain dolomite pseudomorphs of stromatoporoids, tabulate and rugose corals, ostracodes, brachiopods, gastropods, and pelecypods. These features indicate a shallow marine, subtidal environment of deposition. Subtidal to supratidal environments of deposition are indicated by clasts that contain birdseye structure, stromatolites, and edgewise pebble conglomerates.

At Pardners Hill, about 500 m west-southwest of the Bornite drill sites, there are outcrops of Devonian carbonate rocks that do not show the brecciation that is characteristic of the section penetrated by the Bornite drill holes. Sedimentary features observed in outcrop include interformational conglomerates, algal mats, cut-and-fill structures, laminated beds, and birdseye structures. Cavities within the laminated structures are filled by zoned spar dolomite. The outcrops on Pardners Hill support the deduction that a significant part of the Devonian section was deposited within an intertidal to supratidal environment.

The ubiquitous breccias at Bornite seem to represent brecciation of the entire Devonian carbonate section. The clasts are angular to subrounded (fig. 17*A*). Flowage of the marble fabric and syntaxial overgrowth on the calcite crystals indicate that the breccias were formed previous to the dynamic metamorphism which formed the marbles in the cores. Direct evidence for the origin of the breccia is not available. R. H. W. Chadwick (written commun., 1960) proposed that the breccias were patch-reef talus deposits, but the low percentage of corallum and stromatoporoid fragments argues against this type of environment. The subangular to subrounded clasts, their tight packing, abundant stylolites, and matrix indicate that the breccia could have formed through extensive solution of the carbonate section and collapse of the residual rock mass.

The carbonate rocks and, in particular, the dolomites studied by cathode luminescence reveal a complex history of metasomatic replacement not evident under a transmitted light microscope. The dolomites adjacent to and associated with the ore bodies all glow in red and deep-red zones. The vein fillings, in particular, show large red-zoned dolomite crystals. The red color is believed to indicate the presence of Mg and Fe introduced in the lattice of the dolomite by iron-rich hydrothermal fluids. Metasomatic replacement from fluids in the veins into adjacent dolomite clasts increased the rhomb size within the clasts and produced the characteristic red and dark-red zonation (fig. 17*B*) seen under the luminoscope.

Cathode luminoscope examination of samples from the Pardners Hill outcrop of Devonian dolomites, which are unbrecciated, reveals only limited areas of red-glowing dolomite, which are confined to small veins leading to vugs also

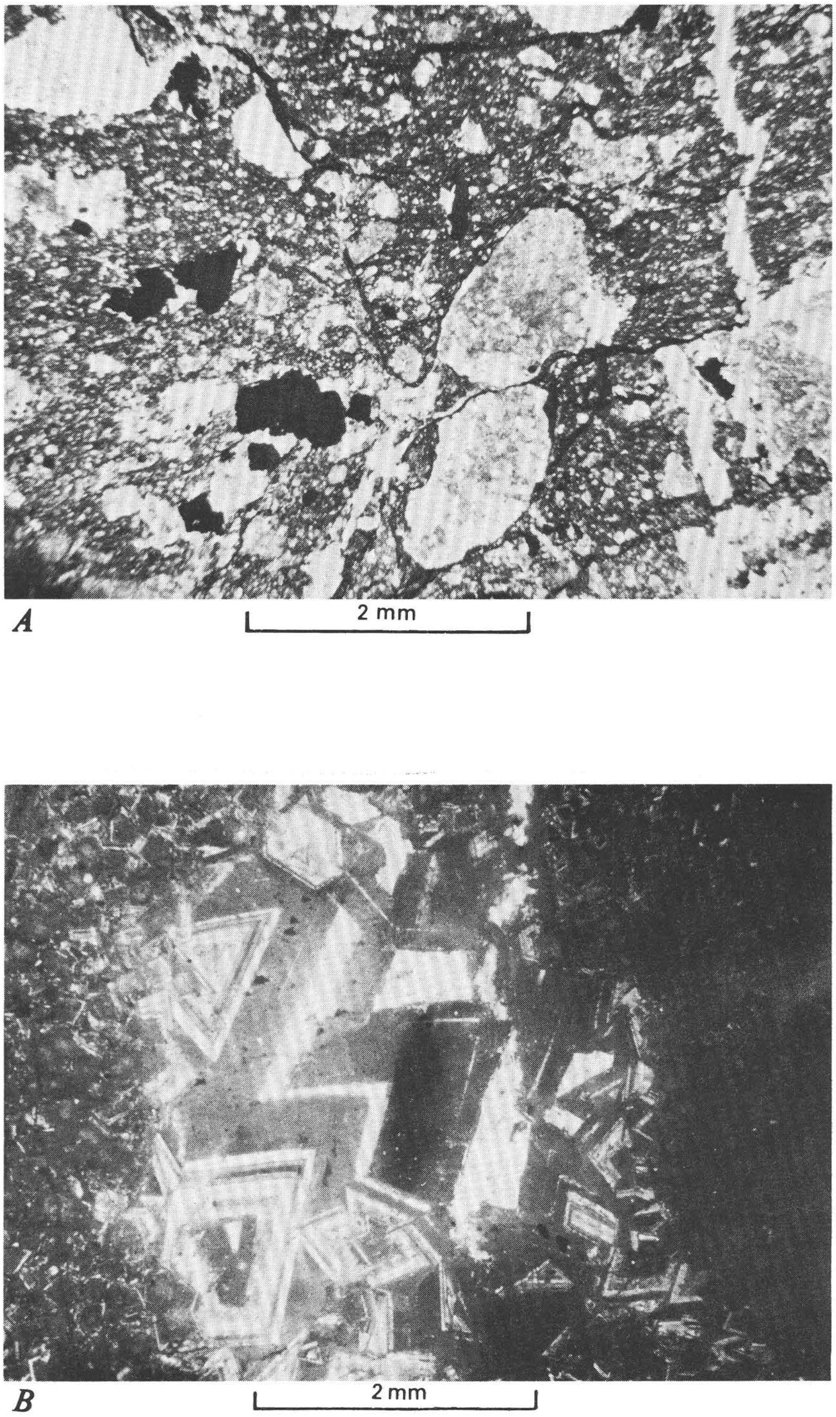

FIGURE 17.—Photomicrographs of Bornite dolomite: *A*, Dolomite breccia clasts in a matrix of pyrite-rich dolomite. Specimen is from a drill hole at a depth of 670.9 m. *B*, Vein filling of cathode luminescence-zoned dolomite from a drill hole at a depth of 597.4 m adjacent to a massive sulfide ore body.

lined with red-glowing dolomite. The bulk of the Pardners Hill rock is not reactive or light emitting under cathode luminescence.

At Bornite, spar dolomite adjacent to the ore bodies glows dark and bright red under cathode luminescence, whereas about 500 m away from the ore bodies only small veinlets and vugs of spar dolomite glow. The characteristic red and dark-red banding under cathode luminescence suggests hydrothermal zoning that might indicate proximity to adjacent ore bodies, and cathode luminescence of this type could be an exploration tool applicable over a broad region of the south flank of the Brooks Range.

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Potential mineralized target areas in the Brooks Range schist belt are characterized by anomalous stream-sediment geochemistry, magnetic and lithologic signature

By John B. Cathrall

Geochemical anomalies from minus-80-mesh stream-sediment samples from the Survey Pass quadrangle, Brooks Range, Alaska, suggest that the known zone of copper- and zinc-bearing stratiform volcanogenic sulfide occurrences found in the Ambler River quadrangle to the west (Grybeck and Nokleberg, 1978, Marsh and Cathrall, 1982) extends eastward into the Survey Pass quadrangle. This volcanogenic sulfide mineralization has been found in the schist belt located along the southern flank of the Brooks Range.

The schist belt, often referred to as the "copper belt," is composed primarily of low-grade metamorphic Devonian (possibly Precambrian) schist including Devonian metafelsitic schist and mafic metavolcanic rocks of uncertain age. West of Reed River, both the metafelsitic schist and mafic metavolcanic rocks are prevalent. However, geologic mapping in the schist belt east of Reed River did not reveal outcrops of metafelsitic schist.

West of Reed River to the west edge of the Survey Pass quadrangle, the prospective copper belt is reflected by samples containing anomalous copper, zinc, lead, silver, barium, antimony, molybdenum, tungsten, bismuth, and boron (table 3). The drainage areas with anomalous geochemical values are associated primarily with metafelsitic schist and mafic metavolcanic rocks. The aeromagnetic data (J. W. Cady and S. W. Hackett, oral commun., 1980) within this section of the schist belt show that the anomalous drainage areas have a characteristic magnetic signature. The axes of aeromagnetic highs wrap around these areas, which in turn lie within aeromagnetic lows where magnetic anomalies greater than approximately -20 gammas are absent.

East of Reed River, stream-sediment samples from drainage areas just west and east of Walker Lake contain anomalous concentrations of copper, zinc, lead, silver, barium, antimony, and boron. Donald Grybeck, (oral commun., 1980) reports finding rock containing anomalous concentrations of copper, lead, silver, barium, and molybdenum. Although metafelsitic schist, the suggested volcanogenic massive sulfide host rock, does not crop out in this area, the mafic metavolcanic rocks, commonly associated

Table 3.--*Threshold values, number of values, range of values, and 95th percentile for selected elements from stream-sediment samples, Survey Pass quadrangle, Alaska*

[Values are reported in parts per million. Leaders (--) indicate no data or insufficient data. Analysts: E. F. Cooley and R. M. O'Leary]

Element	Threshold value	Number of values	Range of values	95th percentile
Ag	0.5	79	0.5 - 10	--
B	300	1,387	<10 - 2,000	247
Ba	1,500	1,477	20 - >5,000	1,180
Be	5	1,358	<1 - 200	3.9
Bi	20	49	20 - 500	--
Cu	150	1,492	<5 - 1,500	109
La	1,000	623	<50 - >1,000	1,125
Mo	15	715	<5 - 70	12
Pb	70	1,470	<10 - 500	72
Sb	100	103	100 - 200	--
Sn	20	222	<10 - 200	41
Th	200	17	<200 - 1,000	--
W	50	25	50 - 1,000	--
Zn	300	1,497	5 - 1,500	196

with the schist, are present in many places. The axes of aeromagnetic highs appear to wrap around the anomalous drainage areas, anomalous rock sample sites, and the areas underlain by mafic metavolcanic rocks. These areas lie within aeromagnetic lows similar to the potential target areas to the west.

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Evidence from stream-sediment geochemical and biogeochemical data, mineral occurrences, and Landsat images for potential mineralized target areas in the Brookss Range, Alaska

By John B. Cathrall

Stream-sediment geochemical data, mineral occurrences, chemical analyses of periphyton (assemblages of algae, fungi, and bacteria that are attached to or live on submerged objects in streams or lakes), and Landsat images delineate two areas as targets for the occurrence of mineral deposits in the northern third of the Survey Pass quadrangle where unmetamorphosed to low-grade metamorphosed Devonian and Mississippian sedimentary rocks crop out. The potential target areas, which contain anomalous amounts of tin, tungsten, molybdenum, bismuth, silver, copper, lead, zinc, barium, lanthanum, thorium, and beryllium in stream-sediment samples (table 3), are: (1) in and adjacent to the drainage encompassing the westernmost loop of the continental divide in the northwestern part of the quadrangle; and (2) in and adjacent to the Coalet Creek drainage south of Plateau Mountain.

In the first target area, follow-up stream traverses located mineralized quartz breccia containing sphalerite, galena, and chalcopyrite.

In the Coalet Creek target area, some parts of the streams appeared distinctly red colored in aerial reconnaissance. Close examination of some of these areas revealed that the color is due to red pigmentation in the profuse growth of periphyton. Periphyton from one of the streams contained high concentrations of silver (10 ppm), copper (1,500 ppm), cadmium (30 ppm), arsenic (500 ppm), nickel (1,000 ppm), and zinc (5,000 ppm). It appears that the periphyton has utilized the elements supplied to the streams as micronutrients, which have caused profuse growth and pigmentation of the periphyton within a section of the stream. Downstream where there is no additional source of these micronutrients, the profuse growth and red color diminish.

Landsat images (J. R. Le Compte, oral commun., 1980) delineated a prominent circular feature and iron-oxide color anomalies in each of the target areas. The circular features suggest the presence of concealed intrusive bodies. The association of anomalous values of tin, tungsten, molybdenum, thorium, and lanthanum found in the stream-sediment samples suggests that mineralization in these areas is from a differentiated felsic source. However, geologic mapping and geophysical data collected to date can neither verify nor negate the presence of intrusive bodies of any kind. The iron-oxide color anomalies visible on computer-enhanced Landsat images occur within and adjacent to the circular features. These iron-oxide-colored areas appear much like those found at other places in Alaska, many of which have proved to be sites of hydrothermal alteration and mineralization associated with plutons (J. R. Le Compte, oral commun., 1980).

Uranium investigations—northeastern Alaska

By A. Curtis Huffman, Allan R. Kirk, and Cornelius M. Molenaar

A reconnaissance of northeastern Alaska has confirmed previously reported uranium anomalies and revealed a heretofore unreported occurrence. Mesozoic and Cenozoic rock units were examined in detail and samples collected for delayed-neutron analyses. Paleozoic sedimentary rocks exposed in the Sadlerochit and Shublik Mountains and the Franklin and Romanzof Mountains (fig. 18) were examined and collected in much less detail. The Okpilak batholith and Jago stock of the Romanzof Mountains (fig. 18) were sampled on a 5- to 6-km grid pattern where possible. From this and previous studies of North Slope and Brooks Range sedimentary rocks we conclude that with the possible exception of the Lower Mississippian Kekiktuk Conglomerate and the Middle and Upper Triassic Shublik Formation in the vicinity of the Okpilak batholith, and the batholith itself, there is little potential for uranium mineralization on the North Slope.

Anomalous uranium values have previously been reported from the phosphatic deposits of the Shublik Formation in the vicinity of the Shublik Mountains (Brosgé and Reiser, 1976). We examined and resampled these known occurrences (fig. 18). Analytic data on these samples are not yet available, but total-count scintillometer readings in the field indicate values as high as 5-6 times background values, primarily in the interbedded phosphatic limestone and shale in the upper third of the formation. Anomalous readings were also obtained from the overlying Upper Triassic Karen Creek Sandstone where it contained reworked and incorporated phospate nodules from the Shublik.

Brosgé and Reiser (1976) also reported anomalous uranium values from the Okpilak batholith in the Romanzof Mountains, noting that the highest values were obtained in the core of the granitic body. Sable (1977) mapped several phases of granite as well as several types of dikes contained within the batholith. Our preliminary results show little mineralization in the dikes and about 2-3 times background values in the porphyritic phase of the granite. Readings from the inner or coarse-grained phases of the granite, however, are commonly 4-5 times background values whether within areas that Sable mapped as the coarse-grained facies or as the variable fine- to coarse-grained facies (fig. 18).

Although the exact age of emplacement of the granite is questionable, Sable (1977, p. 57) prefers a Devonian date on the basis of several lines of evidence. One of his principal reasons is

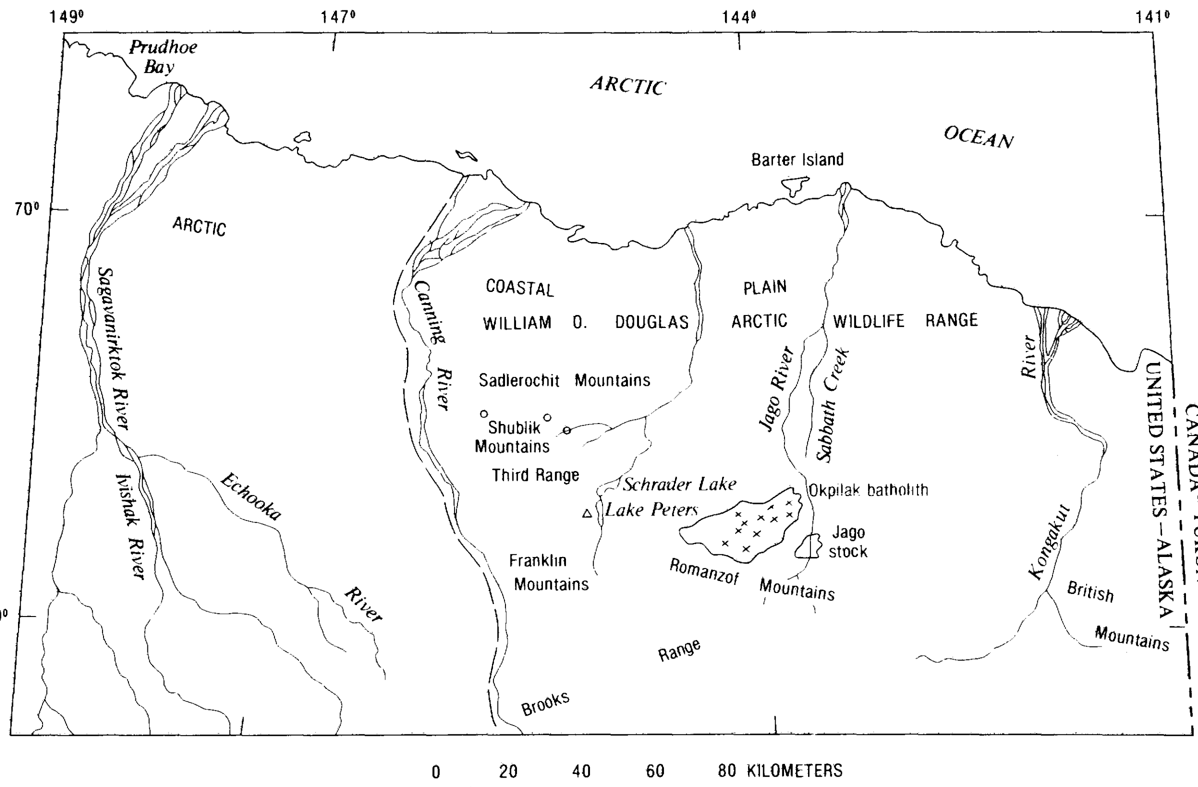

FIGURE 18.—Area of investigation showing locations of anomalous total-count scintillometer readings in rocks other than the pebble shale: O, Shublik Formation; Δ, Kekiktuk Conglomerate; X, granite.

that nowhere does the granite intrude the Kekiktuk Conglomerate, and the presence of quartz-tourmaline clasts in the Kekiktuk Conglomerate in at least one exposure suggests to him that they are reworked granitic detritus. In the vicinity of Lake Peters (fig. 18), the Kekiktuk appears to be a fining upward fluvial sequence in which carbonaceous siltstone and shale overlie conglomeratic channel deposits that contain coalified plant fragments. Scintillometer readings on the conglomerate are 2-3 times background values, whereas those on the fine-grained deposits are as much as 7 times background values.

A. C. Huffman (unpub. data) has documented the presence in the central and western North Slope of Cretaceous and younger rock units that appear favorable as potential uranium host rocks and has speculated as to the reasons for the absence of mineralization; as one of the primary reasons, he cites the apparent absence of nearby uranium source rocks. This criterion is satisfied in the eastern North Slope by the proximity of the Okpilak batholith. However, preliminary results of our examination of all exposed Cretaceous and Tertiary rock units on the eastern North Slope have revealed no evidence that the granite was exposed at any time between the Mississippian and Pleistocene. Several formations, such as the Prince Creek Formation (Upper Cretaceous), Sagavanirktok Formation (Tertiary), and Gubik Formation (Quaternary) as well as the nonmarine beds of questionable age exposed along Sabbath Creek (fig. 18) appear to satisfy most of the criteria for good host rocks, but no evidence of either granitic debris or uranium mineralization was found in the outcrops of any of these units.

Lower Cretaceous pebble shale was sampled in a number of localities in the central and eastern North Slope. Although it exhibits a distinctive anomaly on gamma ray logs in the subsurface and scintillometer readings may be as high as 4-5 times background values on the outcrop, delayed-neutron analyses give no indication of anomalous uranium content so the high total count readings are probably due to potassium.

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Cretaceous overprint revealed by paleomagnetic study in the northern Brooks Range

By John W. Hillhouse and Sherman Grommé

Oriented samples were collected from Devonian and Mississippian sedimentary rocks in the northern Brooks Range in an effort to use paleomagnetic poles to investigate the paleoposition of arctic Alaska. The purpose of the study was to test whether arctic Alaska has undergone large scale rotation relative to the North American craton, as speculated by Carey (1958) and Tailleur (1973). Realizing that the thrust sheets of the Brooks Range may have undergone local rotations, we located our paleomagnetic sites hundreds of kilometers apart and in several structural levels of the range, the rationale being that consistent paleomagnetic poles from such a broad region would accurately reflect possible large-scale movements of the autochthonous basement of the North Slope. The sites are distributed in a narrow zone extending 450 km along the continental divide from Arctic Village to the Ivotuk well site (fig. 19).

The paleomagnetism of the Upper Devonian rocks is represented by 384 oriented cores from 11 stratigraphic sections within siltstone and sandstone members of the Kanayut Conglomerate of the Endicott Group. The sedimentary rocks have intensities of natural remanent magnetization (NRM) that range from 10^{-5} to 10^{-1} A/m and average about 10^{-3} A/m. Within stratigraphic sections, the magnetic directions of the weakly magnetized specimens generally show large angular dispersion about a steep downward mean direction. For most sites, treatment of cores in alternating fields (A.F.) of 10 to 30 mT decreases the directional dispersion, although angular standard deviations remain rather large (25°) for many of the sites. Fine-grained hematite and magnetite contribute to the NRM, as indicated by thermal demagnetization experiments and other rock magnetic tests.

As a first step toward determining the age of the magnetization, we tested whether the remanence originated before the Devonian rocks were deformed. The fold test compares the angular dispersion of virtual geomagnetic poles (VGP) before and after the results are corrected for tilt of the bedding. If the magnetization predates deformation, then application of the tilt corrections will decrease the angular dispersion of VGP's, provided that all corrections are not similar. Our 11 stratigraphic sections have sufficiently different structural attitudes to provide a significant fold test. The results conclusively fail the test despite A.F. treatment. The tilt corrections move the site VGP's from a high-latitude cluster to a streaked distribution along a longitudinal great circle, making a significant increase in angular dispersion. We conclude that the characteristic magnetization postdates Cretaceous deformation of the Brooks Range; no trace of a primary depositional remanence remains. Combining the 11 site VGP's without correcting them for tilt of the bedding gives an overall mean paleomagnetic pole at 59° N., 197° E. ($\alpha95 = 12°$) in the southeastern Bering Sea.

Mississippian limestone of the Lisburne Group was sampled at four sites (totaling 103 cores) near Kurupa Lake and the Killik River. In this region, the Lisburne Group crops out in a thin thrust sheet that structurally underlies klippe to the south consisting of Upper Devonian sedimentary rocks. The pale-gray limestone is coarsely crystalline and free of silty or sandy detritus. Dissolution surfaces and dark-gray chert beds (50 cm - 3 m thick) give the formation a banded appearance. During the coring, we avoided the chert beds and nodules, which were too hard for the drill bit to penetrate.

The limestone gives intensities of NRM averaging 5×10^{-4} A/m, about 10 times weaker than the mean NRM of the Devonian rocks. NRM directions for two sites (13, 14) were tightly clustered about steeply inclined means in the southwest quadrant; the remaining sites gave widely dispersed directions. Vector diagrams obtained from pilot thermal demagnetization runs were analyzed to determine the directional components of the NRM. In general, thermal treatment of the limestones to 450°C unblocks 95 percent of the initial remanence. Heating the specimens beyond 500°C produces erratic directional changes as the specimens acquire spurious thermoremanence in the furnace. Samples from sites 13 and 14 attain stable characteristic magnetizations at temperatures above 200°C. Sites 12 and 15 yielded erratic directional changes during thermal demagnetization, which failed to isolate stable, characteristic magnetizations. Therefore, the internal-

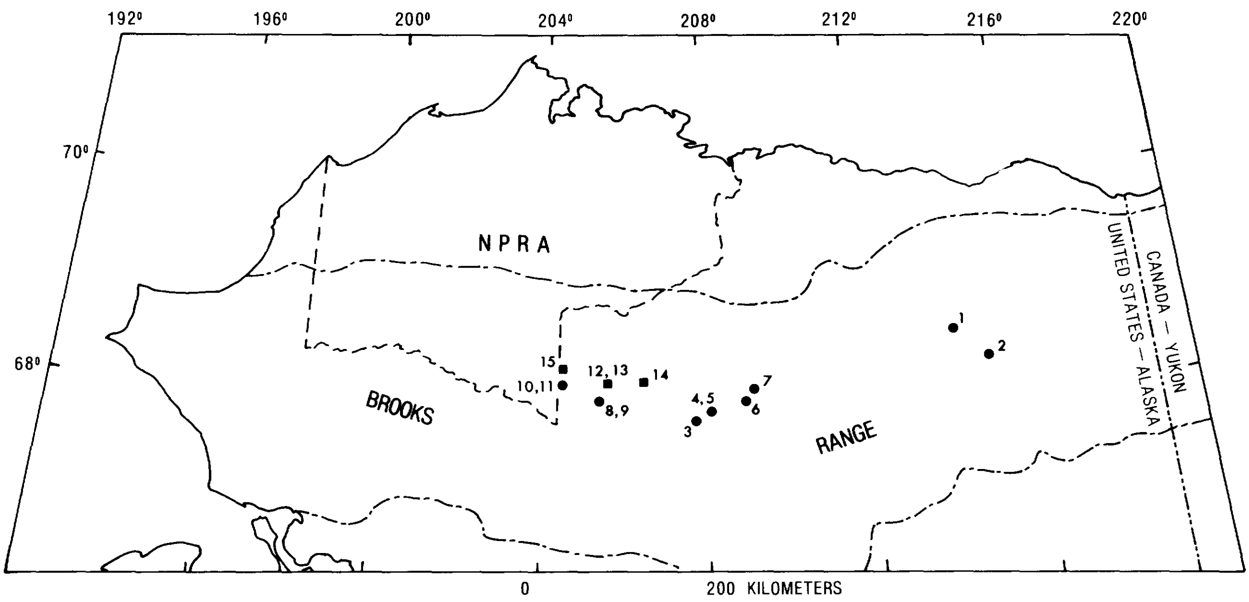

FIGURE 19.—Locations of paleomagnetic sample sites in the Kanayut Conglomerate (dots) and Lisburne Limestone (squares), northern Alaska.

ly inconsistent results from these two sites are omitted from further discussion.

After thermal cleaning to 400°C, the two Lisburne sites give significantly different mean directions, which become more divergent when the bedding corrections are made. Two sites do not constitute a statistically significant fold test; however, the large divergence of mean directions suggests that not all of the magnetization was acquired during formation of the limestone. The characteristic magnetization of site 13 is overprinted by a steep, southwestern component similar to the *in situ* magnetic direction of site 14. Applying bedding corrections to the overprint and the mean direction of site 14 increases the angular difference between the means. Therefore, the steep component appears to be a post-deformational overprint, as was observed in the Upper Devonian sedimentary rocks. Whether the characteristic magnetization of site 13 originated during formation of the limestone remains open to debate, until supporting evidence is obtained from a positive fold test.

In the central Brooks Range (Killik River, Chandler Lake, and Arctic quadrangles), sedimentary rocks have apparently lost their Paleozoic magnetic signatures. The original remanence has been replaced by a normally polarized magnetic overprint that postdates the Cretaceous orogeny of the Brooks Range (Mull, 1979). The final stages of the orogeny occurred in Albian time, when the uplifted stack of thrust sheets began shedding debris into the basin to the north. When compared to the reference curve for North American apparent polar wander (Irving, 1979), the paleomagnetic pole corresponding to the overprint implies a Cretaceous age of magnetization (fig. 20). Of course, assigning ages on the basis of paleomagnetic pole positions is a dubious practice in tectonically active regions such as Alaska. However, isotopic evidence from the southern Brooks Range suggests that a thermal event occurred in the Cretaceous, lending support to a Cretaceous age of remagnetization. Most plutonic rocks in the southern Brooks Range give potassium-argon ages of 130-100 m.y. (Turner and others, 1979), but lead isotopic ages indicate that the plutons originated much earlier, approximately 375 m.y. ago (Dillon and others, 1980). The resetting of the potassium-argon system is attributed to a Cretaceous thermal event. In addition, the commencement of argon retention partly occurred during the Cretaceous normal epoch (110-81 m.y.), providing a likely explanation for the normal polarity of the overprint.

The maximum temperatures attained by the sedimentary rocks in our sampling localities can be deduced petrographically. Well-preserved textures and the absence of recrystallized minerals limit the maximum temperatures to approximately 300°C. In the Survey Pass quadrangle, 50 km south of sites 8-15, conodonts in limestone of the Lisburne have colors indicative of heating in the range 200-300°C (Nelson and Grybeck, 1980). Although conodont temperatures are not available directly from our sample localities, regional trends based on the Survey Pass study suggest that heating did not exceed 200°C in the sampled area. Such mild heating was capable of remagnetizing the Paleozoic sandstones and limestones, because the thermal stability of the original remanence was inherently low. Theoretically, magnetite domains having laboratory-determined blocking temperatures of 425°C (as in limestone of the Lisburne) will be demagnetized in nature if

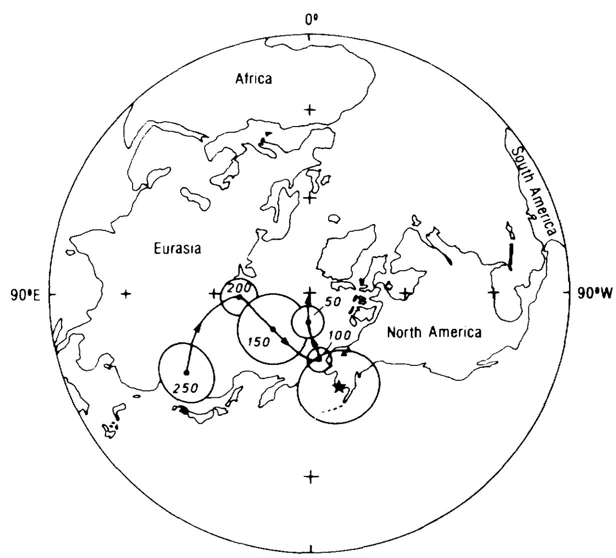

FIGURE 20.—Comparison of overprinted Kanayut Conglomerate VGP (star) with curve of North American apparent polar wander, given at 50 m.y. intervals (dots). Ninety-five percent confidence limits for paleomagnetic poles are shown as circles. Sample sites in northern Alaska indicated by triangles.

heated to 200°C for 10 m.y. (Dunlop and Buchan, 1977). Hematite domains with blocking temperatures below 525°C will be demagnetized under similar conditions in the natural environment. Once unblocked, the magnetic domains will become alined with the ambient field. In addition, new carriers of magnetic remanence with high thermal stability will develop if chemical conditions favor the growth of authigenic hematite. These processes can account for the magnetic overprint that is shared by magnetite and hematite domains in the Kanayut Conglomerate.

The core of the southern Brooks Range consists of schistose metaclastic rocks primarily of the greenschist facies. Our sample localities, which are far north of the schist belt, received a mild but prolonged heating on the outer edge of the regional metamorphic zone. The immense remagnetized region, which comprises most of the northern Brooks Range, is probably the product of rapid uplift and cooling after Cretaceous overthrusting of the southern allochthons onto the North Slope terrane. The paleomagnetic pole position and the uniform polarity of the overprinted directions suggest that the magnetization was locked in during the middle Cretaceous normal epoch.

Our results fail to prove or disprove the hypothetical rotation of arctic Alaska and the rift-origin of the Canada Basin. The thrust tectonics of the Brooks Range may indeed be a product of such a rotation. However, in the area sampled, the thermal disturbance created by the thrusting has destroyed the Paleozoic magnetic signature that is required to prove the rotation hypothesis.

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Preliminary geothermal isograd map, NPRA

By David C. Blanchard and Irvin L. Tailleur

Estimates of subsurface rock temperatures have been obtained by downhole temperature measurements in 22 of the 34 wells drilled during current exploration of the National Petroleum Reserve in Alaska (NPRA). These estimates of static formation temperatures (derived at a few specific depths from temperatures recording progressive interaction between rocks in the borehole and the cooler drilling fluid) yield geothermal gradients between the points where temperatures were determined. Temperature gradients bear on source-rock maturation, diagenesis, hydrocarbon generation, and migration, so obtaining usable estimates of formation temperatures is an important element of petroleum exploration. Contouring of local gradients between permafrost and total depth extends a preliminary characterization of the geothermal regime of the Prudhoe Bay region to the east (Tailleur and Engwicht, 1978). The trends, variations, and anomalies in the contours of the small-scale map, figure 21, afford another basis for deducing gross geologic conditions in the North Slope region.

Two sets of temperature data are being used to derive undisturbed rock temperatures. The first is part of routine well-logging practice and is recorded by maximum-reading thermometers attached to the wire line a variable distance above the bottom of the logging tool. The loggers record the time drilling fluid stops circulating and the times at which successively run tools reach the bottom of the hole or bottom of the logged interval. The temperatures measured reflect in general the progressive warming of the static fluid column in contact with recently drilled rock. The method used to obtain the static formation temperature is based on the Horner Plot (Dowdle and Cobb, 1975) principle but modified to plot bottom-hole static temperature as a function of time in hours since circulation ceased versus temperature recorded.

The second set of data comprise dual temperature logs, one run at the beginning and the other at the end of the logging operations when drilling to total depth is completed. These extraordinary logs provide second data sets for 14 of the 22 wells and the only data set for 2 Barrow wells, where temperatures at the shallow depths are too low to be recorded with present logging techniques. The two continuous curves measure the fluid to wall adjustment at any depth interval while the drilling mud is undisturbed for as long as 60 hours. The temperatures and time differentials at any given depth permit extrapolation to the static formation temperature at that depth. Agreement between results derived from the independent data sets indicates that the estimated undisturbed temperatures in most wells are a satisfactory base from which to calculate geothermal gradients across given intervals.

Despite the low density of data points, the preliminary geothermal isograd map (fig. 21) permits some regional geologic interpretation. This interpretation is based on the premise that higher gradients are likely to reflect: (1) less conductive materials between the heat source and the Earth's surface, (2) a greater supply of heat at the source, or (3) a significantly shorter distance from the source to the surface. Several anomalies on the gradient map correlate with seismically determined basement highs; most notable are the Meade arch and the Fish Creek platform.

A gradient of 47°C/km in the South Meade Test Well is the steepest measured on the North

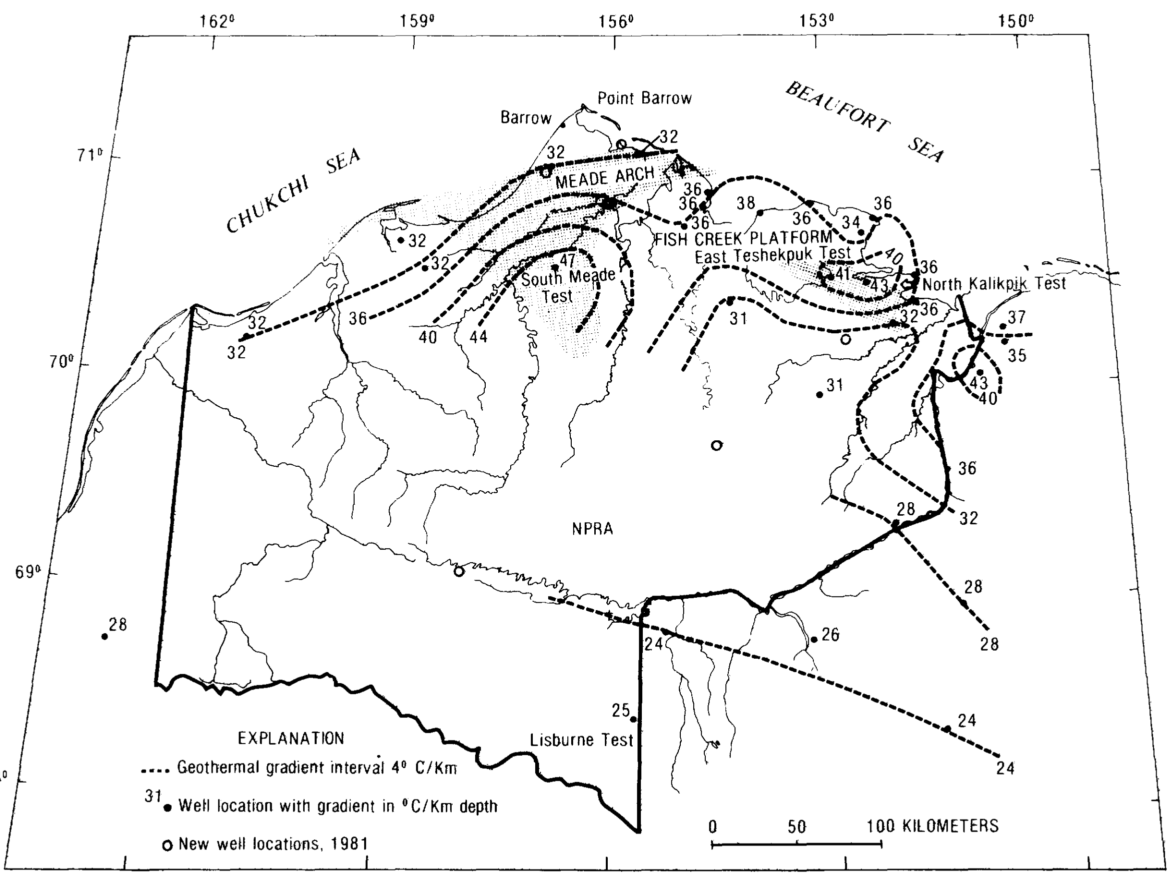

FIGURE 21.—Preliminary geothermal isograd map, central North Slope. Shaded areas represent basement highs.

Slope and exceeds most gradients reported for other epicontinental basins. Preliminary judgement indicates neither anomalous proximity to basement nor a sediment-conductive anomaly and presumes, therefore, an anomalous heat supply. A source of substantial heat might be relatively young felsic rocks at depth. The Devonian granite at the bottom of East Teshekpuk Test Well (Bird and others, 1978), for example, may account for the lower anomaly enclosing Teshekpuk and the shallower North Kalikpik Well. Furthermore, at least four conjecturally similar plutons in the Brooks Range induce thermal springs. A Bouguer gravity map (S. I. Gutman, written commun., 1980) shows a low about Teshekpuk that could reflect the granite; South Meade, however, is on the margin of a stronger and larger gravity low.

Our scattered data show geothermal gradients lessening to the south. This trend correlates with evidence for substantially increased depths to basement: the isostatic depression compensating the Brooks Range, inferred from reconnaissance gravity data (Barnes, 1977); the sixfold thrust fault imbrications drilled by the Lisburne Test Well; regional evidence that the ubiquitous subsurface sequence underlies the drilled and other allochthonous sequence in the disturbed belt; and, tenuously, the absence of coherent seismic reflections above 4 and 5 seconds two-way traveltime.

Six new wells scheduled for completion in FY 81 will add significantly to the data base overall and will nearly double it for the southern part of the reserve. Also, a joint effort with the Geothermal Studies Projects staff is being initiated to correlate their geothermal gradient measurements in the permafrost and the projections made by the NPRA group. With this work, it will be possible to construct a more complete picture of regional geothermal gradients in the reserve.

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NPRA data release, 1980

By Robert D. Carter

Since the inception of the Geological Survey's petroleum assessment of the National Petroleum Reserve in Alaska (NPRA) in June 1977, every effort has been made to release the data generated to the public as soon as possible. Following the operational needs of the ongoing exploration program, current information is gathered and delivered in reproducible form to the National Oceanic and Atmospheric Administration (NOAA) in Boulder, Colorado for processing and release. Seismic data acquired in 1979 are typical of the yearly seismic programs and include 93 sections or profiles, 9 shotpoint location maps, a shotpoint location tape, and 727 velocity analyses. This information was gathered along 3012 line-kilometers located in the north-central, southwestern, and far south areas of the 96,000-km^2 reserve. These data, along with geophysical and geologic interpretive reports prepared by contract personnel, have been forwarded to NOAA. In addition, a magnetic tape containing gravity data for the 1974 through 1980 surveys and logs and histories on five of the wells completed in 1980 have also been sent to NOAA. These data should be available to the public early in 1981. Inquiries should be addressed to:

National Geophysical and Solar-Terrestrial Data Center,
NOAA/EDIS
(D-621) 325 Broadway
Boulder, Colorado 80303
Telephone: (303) 497-6541

Releases to date include nearly 16,000 line-kilometers of seismic data acquired from 1972 through 1978 along with interpretive reports; well logs and histories on each of 28 wells drilled from 1964 through 1979; micropaleontology and palynology reports on 38 wells; velocity surveys on 16 wells; and 8 miscellaneous geologic and geophysical reports. Copies of seismic field tapes in SEG-Y format obtained from 1972 through 1978 were made available in November 1980.

The practice of providing, by telephone, weekly reports on the progress of the drilling wells from the Anchorage and Menlo Park offices of ONPRA was continued in 1980. Copies of field prints of well log runs were also made available locally.

Terraces of the Colville River Delta region, Alaska

By L. David Carter and John P. Galloway

Terraces flanking the lower Colville River and the Itkillik River in the Harrison Bay quadrangle (fig. 22) can be classified into three groups (I, II, and III) on the basis of relative elevations. The terraces are underlain by unconsolidated deposits of the Gubik Formation (Black, 1964) that record a complex marine and alluvial history spanning perhaps 3.5 m.y. Data from these unconsolidated deposits, when considered in terms of the sequence of marine transgressions established for northwestern Alaska by Hopkins (1967), permit speculation regarding the age range of each terrace group.

Terrace I, the oldest and most extensive, occurs on both sides of the Colville River (fig. 22). West of the Colville River, it is bounded by degraded scarps that separate it from terrace III and by bluffs along the Colville River. An east-trending scarp at the northeast corner of this part of terrace I is inferred to be an ancient sea-cliff. To the west, the seacliff becomes indistinct, and beyond this point terrace I can not be distinguished from terrace III. Surface elevations range from about 36 m near the ancient sea-cliff to about 44 m near the Colville River bluffs.

East of the Colville River, the western boundary of terrace I is a degraded scarp that is traceable from the south edge of the map area to within about 5 km of Harrison Bay. Beyond this point, terraces I and II cannot be differentiated. Elevations at the top of the scarp range from 23 m in the north to 60 m at the south margin of the map area. Surface elevations increase southeastward away from the scarp to a maximum of about 75 m where terrace I terminates against low uplands underlain by Tertiary deposits (Beikman, 1978).

Unconsolidated deposits underlying terrace I west of the Colville River range from 14 to 20 m in thickness and include marine and nonmarine deposits. The oldest beds are fossiliferous marine gravelly sand and sandy silt that unconformably overlie pre-Quaternary rocks (Carter and others, 1977; O'Sullivan, 1961; Schrader, 1904). Amino acid analyses of *Hiatella arctica* shells from several locations show that the marine beds were deposited during a single transgression (J. K. Brigham, written commun., 1980). The remainder of the unconsolidated deposits consist of sand to gravelly sand alluvium of at least four different ages. The oldest of the alluvial beds overlies the marine deposits, is generally free of detrital wood, and

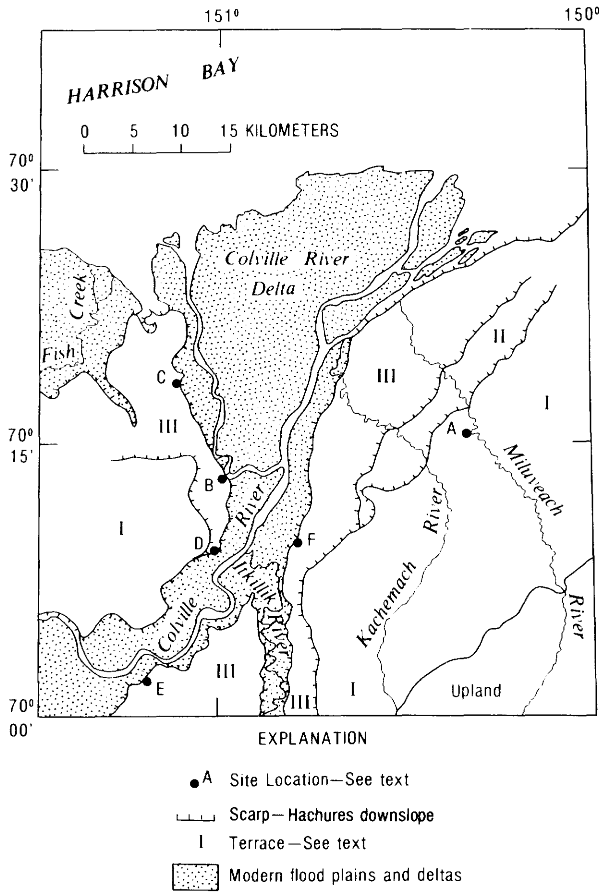

FIGURE 22.—Map showing terraces of Colville River delta region and locations of sites and features mentioned in text.

contains a pollen assemblage that indicates deposition during a glacial episode (Nelson, 1981). Younger alluvium contains abundant detrital wood, some of which has been identified as *Picea* (spruce) *sp.* (R. C. Koeppen, written commun., 1976). Because spruce does not now grow north of the Brooks Range, its presence suggests that these beds were deposited during an interglacial or warm interstadial period. Still younger deposits are devoid of detrital wood and may represent other glacial episodes. The alluvial stratigraphy suggests that terrace I is an alluvial plain that consists of a number of alluvial surfaces formed over a long period of time but differing from one another only slightly in elevation.

Although deposits underlying terrace I east of the Colville River have not been extensively studied, one particular exposure (site A, fig. 22) has provided information that bears on the age range of this geomorphic feature. At this locality 7 m of marine deposits is overlain by 2.5 m of pebbly sand that is presumed to be nonmarine. The marine section consists of 2.5 m of mud overlain by 1.5 m of fossiliferous, gravelly beach sand, which is in turn overlain by 3 m of mud and sand. Amino acid ratios determined for shells of *Hiatella arctica* collected from the gravelly beach sand show that this bed is correlative with the marine beds that underlie terrace I west of the Colville River (J. K. Brigham, written commun., 1980). The underlying mud, however, contains an ostracode fauna that includes *Rabilimis paramirabilis* (Swain, 1963) (E. M. Brouwers, written commun., 1980). This species is not known to occur with *Rabilimis septentrionalis* (Brady, 1866) and *Heterocyprideis sorbyana* (Jones, 1856), both of which are common in the marine deposits of terrace I west of the Colville River. Therefore, the *R. paramirabilis*-bearing mud at site A is interpreted to represent an older marine transgression than the gravelly beach sand at site A. Inasmuch as *R. paramirabilis* occurs in deposits of the Beringian transgression at Nome but not in younger beds (D. M. Hopkins, oral commun., 1980), the older marine unit at site A is correlated with the Pliocene Beringian transgression. The beach deposits and overlying marine mud and sand at site A and the marine beds of terrace I west of the Colville River may correlate with deposits of Hopkins' (1967) next younger Anvilian transgression, which is early Pleistocene. If this is true, terrace I is early Pleistocene or younger.

Terrace II occurs only on the east side of the Colville River delta (fig. 22) and is bordered by low scarps that trend northeast-southwest. It appears to consist of a single surface to which broad alluvial terraces of the Miluveach and Kachemach Rivers are graded. Maximum surface elevations are about 17 m northeast of the Miluveach River, but elevations increase southwest of the Miluveach River to a maximum of about 24 m at the southern terminus of the terrace. At the few sites examined, materials underlying the terrace are pebbly sand. Northeast of the Miluveach River, the nearly constant elevation of the terrace surface suggests that the underlying deposits are marine, whereas increasing elevations southwest of the Miluveach River suggest that the deposits there are alluvium. If these inferences are correct, the terrace and its deposits may represent a single marine transgression, most probably the middle Pleistocene Kotzebuan transgression, which is the marine transgression next younger than Anvilian in Hopkins' (1967) Alaskan sequence.[5]

Terrace III occurs on both sides of the Colville River delta, between the Colville and Itkillik Rivers, and east of the Itkillik River. West of the Colville River delta, terrace III is a single surface that is traceable from the coast inland 16 km to the sea-cliff mentioned earlier, and up the Colville River an additional 10 km. Elevations on the terrace surface range from about 10 m near the coast to approximately 18 m at its southern terminus. Between the Colville and Itkillik Rivers, terrace III also appears to be a single surface and may be equivalent to the terrace III surface west of the Colville River. Surface elevations range from about 18 m in the north to about 24 m at the south edge of the map area. East of the Itkillik River, terrace III also may be a single surface that is possibly correlative with the surface west of the river. However, to the north, bordering the east side of the Colville River delta, terrace III may consist of two or more terraces. Maximum elevations

[5]Hopkins' 1967 paper lists a marine transgression between the Anvilian and Kotzebuan transgressions that he called the Einahnuhtan transgression. However, recent work has shown that the beds for which the Einahnuhtan transgression was named were deposited during the Kotzebuan transgression (D. M. Hopkins, oral commun., 1980).

range from about 8 m near Harrison Bay to about 24 m at the south edge of the map area.

Materials underlying this terrace group are poorly exposed; exposures examined consist of sand to gravelly sand. Marine fossils have not been collected from these deposits, but proboscidian remains were found at site B (H. J. Walker, written commun. to R. E. Nelson, 1975). Accordingly, the deposits are interpreted as alluvium.

Fossil wood collected from terrace III deposits at site C was identified as *Alnus* (alder) *sp.* (R. C. Koeppen, written commun., 1978) and radiocarbon dated as older than 50,600 yr (USGS-675). *Populus* (poplar) *sp.* wood was collected at site B (H. J. Walker, written commun. to R. E. Nelson, 1975), and wood identified as spruce (R. B. Miller, written commun., 1979) was collected at sites D, E, and F. The spruce at site D yielded a radiocarbon age of greater than 48,000 yr (USGS-631), and the spruce from site F proved to be more than 48,500 years old (USGS-630). The presence of spruce, poplar, and alder suggests deposition during an interglacial or relatively warm interstadial period, and the radiocarbon dates show that deposition took place more than 48,000 years ago. We believe that terrace III and its underlying deposits most likely were formed during the last interglacial episode (Sangamonian), which corresponds to the Pelukian transgression of Hopkins (1967). However, formation during one or more Illinoian or early Wisconsinan interstadials is a possible alternative.

In summary, we believe terrace I to be a complex alluvial plain that is younger than the Anvilian transgression and older than the Kotzebuan transgression. Terrace II may consist of a single surface underlain by alluvial and marine deposits formed during the Kotzebuan transgression. Terrace III consists of alluvial terraces that are most likely Pelukian in age but may have formed during one or more Illinoian or early Wisconsinan interstadials.

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Grain-size analyses of 20 eolian sand samples from northern Alaska

By John P. Galloway

Grain-size parameters were determined for 20 samples collected from a Pleistocene dune field in northern Alaska. The dune field covers more than 7000 km^2 of the National Petroleum Reserve in Alaska (fig. 23) and has been described by Carter (1981). Standard preparation techniques as described by Folk (1974) were used. The sand fraction was analyzed by sieving and the fine fraction by hydrophotometer. A standard computer program (Pierce and Good, 1966) for grain-size analysis was used to generate a cumulative curve and graphical statistics for each sample.

Eighteen of the samples contained 90 percent or more sand. Mud content (silt and clay) ranged from 12.5 percent to less than 0.5 percent. For those samples with less than 6 percent mud, most suspended matter settled before a hydrophotometer reading could be made. For those samples with 6 percent or more mud, nearly all the fines had settled out after the first six hydrophotometer readings (7.0 Ø, fine silt), and further readings were erratic. Accordingly, a limit of 7.0 Ø was used in the program for generation of the cumulative curves from which the graphical statistics were derived (Folk and Ward, 1957). Figure 24 is a composite of all the cumulative curves.

Table 4 lists the graphical statistics. For convenience, samples are arranged by increasing mean. Mean grain size ranged from 3.3377 Ø to 2.0630 Ø with an overall mean of 2.6838 Ø (fine sand). According to sorting limits set by Folk (1974), one sample was poorly sorted, and the other 19 samples ranged from moderately to well sorted. Ninety percent of the samples were either nearly symmetrical or finely skewed, ranging from -0.0957 Ø to 0.3715 Ø. Although there are problems associated with the use of graphical statistics as the sole criteria to identify eolian deposits, data presented in table 4 are not inconsistent with graphical statistics used to describe other inland dune fields (Ahlbrandt, 1979). Overall, the eolian sand samples reported here lack granules, consist of fine sand, show a tendency toward being moderately to moderately well sorted, and are generally symmetrically distributed about the mean.

Table 4.—*Graphical statistics resulting from grain size analyses of 20 eolian sand samples*

Sample number	Mean Ø(phi)	Percent sand	Percent silt	Sorting Ø (phi)	Skewness Ø (phi)
79ACr 152e2	3.3377	87.502	12.498	0.7097	-0.0592
79ACr 036c2	3.1519	95.213	4.787	.4234	-.0004
79ACr 036b2	3.1261	88.032	11.968	.7857	-.0110
79ACr 006a	3.0338	94.780	5.220	.6018	-.0957
80ACrII 017	3.0251	93.210	6.790	.7669	.2235
79ACr 027	2.9718	92.166	7.834	.6878	-.0518
79ACr 156k2	2.9692	93.630	6.370	.8844	-.2140
79ACr 152f2	2.8184	91.709	8.219	.8276	-.0773
79ACr 074q	2.7524	98.866	1.134	.4564	-.2922
78ACrII 055a	2.6949	92.249	7.751	1.0853	.4719
79ACr 152g3	2.6103	95.111	4.889	.8571	.0669
78ACrII 46a	2.6076	98.779	1.221	.5223	.0453
79ACr 106e	2.5352	99.475	.525	.5331	.0023
79ACr 104c	2.4965	99.836	.164	.4598	-.0202
78ACrII 55c-2	2.4307	97.652	2.348	.6819	.0447
79ACr 156j2	2.3285	99.054	.946	.7657	.0143
79ACr 088b	2.2885	99.213	.787	.5477	.0957
79ACr 075d	2.2787	99.302	.698	.4530	.0732
78ACrII 55b	2.1573	96.141	3.859	.8519	.3715
79ACr 156i1	2.0630	99.819	.181	.3858	.2196
Mean---------	2.6838				

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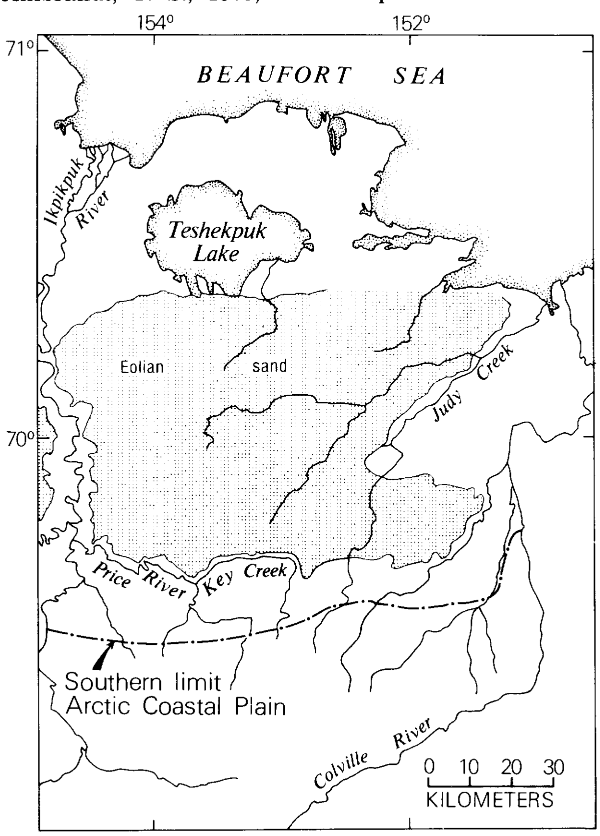

FIGURE 23.—Map of Pleistocene dune field in northern Alaska.

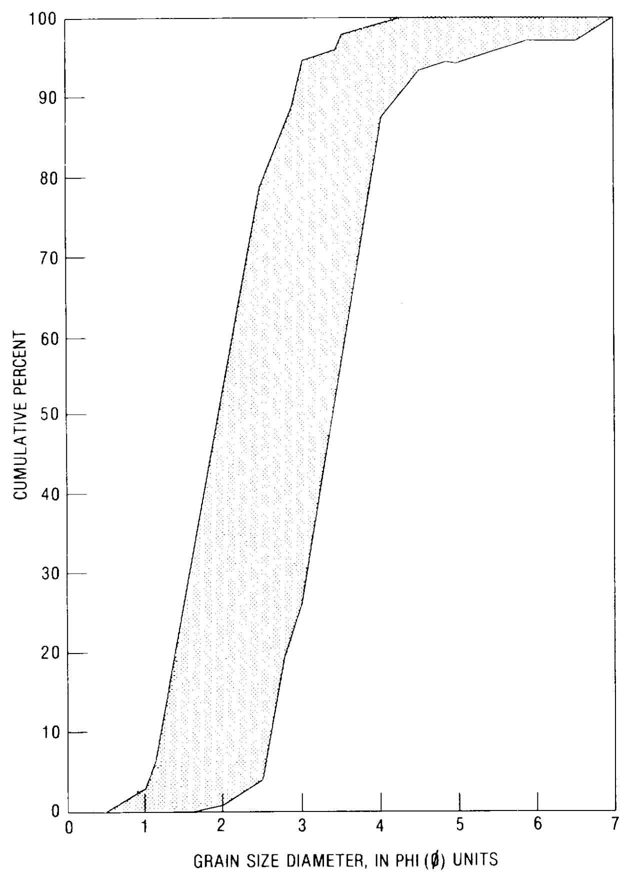

FIGURE 24.—Composite of cumulative percent vs. grain size of eolian sand samples, northern Alaska.

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Nutrient limitation in two arctic lakes

By George A. McCoy

Some lakes in the National Petroleum Reserve in Alaska may be subject to cultural eutrophication because of the influx of domestic or industrial wastes or disturbances of surficial material or vegetation. In order to assess and define the factors that limit primary productivity in arctic lakes, two lakes on the flood plain of the Colville River near Umiat were selected for study, one lake as a control, the other for experiments with fertilization. The lakes' areas are about 10,500 m^2, and their mean depths are about 1.4 m. In 1979, the first year of the study, laboratory cultures of *Selenastrum capricornutum* Printz in water from the lakes were spiked with various combinations of nutrients. The hypothesis developed from this procedure was tested in the second year of the study by adding fertilizers directly to the experimental lake.

The bioassays performed on the two lakes in the arctic foothills indicated that phosphorus was limiting early in the growing season and that both lakes tended toward a state of nitrogen and phosphorus co-limitation later in the growing season. Additions of trace elements and organic growth factors did not stimulate growth in the laboratory cultures.

In 1980 one lake was fertilized with phosphorus pentoxide (P_2O_5) in June and early July and with ammonium nitrate (NH_4NO_3) in late July and early August. Total phosphorus concentration in the lakes was 0.006 to 0.05 mg/L and total nitrogen, 0.1 to 0.5 mg/L before fertilization. Phosphorus fertilizer was added in sufficient quantity to add 0.10 mg/L phosphorus to the experimental lake and nitrogen fertilizer to 2.0 mg/L. Phosphorus additions caused an approximately tenfold increase in primary production as indicated by the number of organisms, chlorophyll concentration, and biomass of periphyton and phytoplankton. The nitrogen additions caused another order-of-magnitude increase in primary production over that promoted by the phosphorus additions alone. By early September the fertilized lake was bright green and had a Secchi transparency of 10 cm or less. The pH in this lake had risen from about 7.0 to 9.3. The pH in the control lake and the phytoplankton and periphyton density were unchanged from the previous year, and this lake was transparent to its maximum depth 2.2 m. Nitrogen fertilizer added to the control lake in August had no effect on primary productivity or pH.

These results have been interpreted to indicate that phosphorus is limiting early in the growing season and that the amount of phosphorus added was sufficient to cause the experimental lake to be nitrogen-limited. The results obtained by adding fertilizer to these lakes are exactly those predicted from the laboratory bioassays.

EAST-CENTRAL ALASKA

(Figure 25 shows study areas discussed)

Graptolites discovered in the Woodchopper Volcanics

By Michael Churkin, Jr., James H. Trexler, Jr., and Claire Carter

Early Devonian graptolites within a volcanic rock section along the Yukon River were found by Churkin and Trexler during fieldwork of summer 1980. The graptolites are from a relatively thin (0.5 m-thick) shale bed within the Woodchopper Volcanics exposed on the south bank of the Yukon River 5.5 km downstream from the mouth of Woodchopper Creek (fig. 26). The shale is in an interval of thinly interbedded shale, volcanic sandstone, and tuff that forms a recessive unit within the cliff-forming basalt pillow lavas and breccias, all in the Woodchopper Volcanics near its type locality (Mertie, 1930).

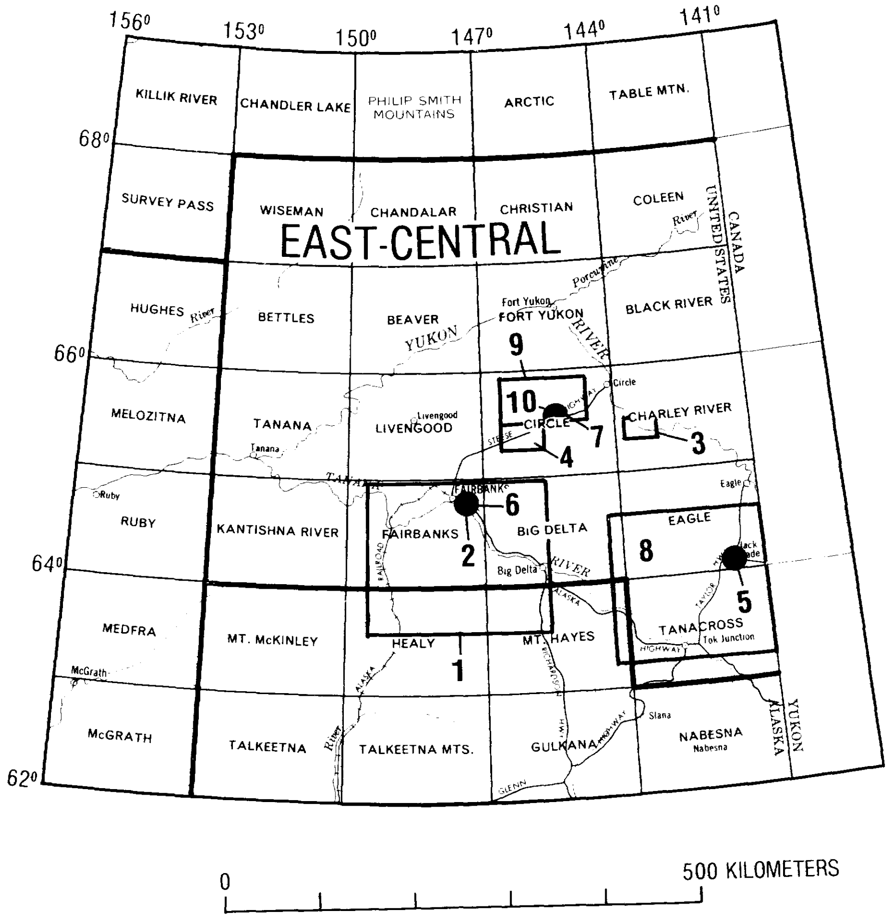

FIGURE 25.—Areas and localities in east-central Alaska discussed in this volume. Authorship and applicable figures relating to the numbered areas are listed below. 1, Ager, figure 34; 2, Burrows, Emmett, and Parks, figures 35, 36, and 37; 3, Churkin, Trexler, and Carter, C., figures 26 and 27; 4, Cushing, Foster, Laird, and Burack, figures 28 and 29; 5, Foster and O'Leary, figure 32; 6, Krumhardt, figure 38; 7, Tripp, Detra, and Nishi; 8, Weber, figure 31; 9, Weber and Foster, figure 30; 10, Yeend, figure 33.

The age of the Woodchopper Volcanics and correlation with non-volcanic sequences about 70 km further east in the Charley River quadrangle has been a problem (Churkin and Brabb, 1968). Shelly fossils from limestone closely associated with the volcanic rocks of the Woodchopper have been dated as Devonian (Mertie, 1930; Brabb and Churkin, 1969; Oliver and others, 1975). However, Churkin and Brabb (1968), along with earlier workers, emphasized the complex faulting in the Woodchopper area and questioned the stratigraphic relation of the limestone to the volcanic rocks. More recently, Lane and Ormiston (1976) assigned an Early Devonian age to a 50-m section of limestone that has volcanic rocks exposed both above and below. The basis of their age determination was a study of the conodonts and corals from the limestone (fig. 27). The contacts between the limestone and the pillow lavas and volcaniclastic rocks of the Woodchopper were not described, nor are the relations clear in the outcrop.

Subsequently, corals were found in limestone clasts from basalt breccias within the Woodchopper Volcanics. The corals were identified by W. A. Oliver, Jr., (written commun., 1980) as *Haplothecia?* sp. A., *Phillipsastraea* sp., and *Favosites* sp. According to Oliver, these corals are probably Late Devonian (Frasnian). We lavas. Shelly-fossiliferous limestone clasts in basalt breccia and tuff commonly occur together

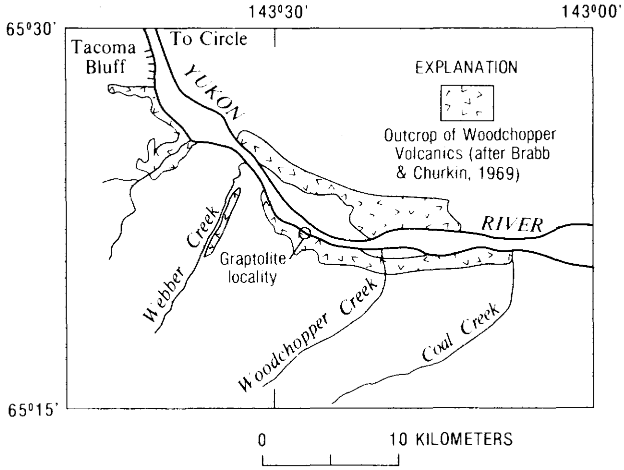

FIGURE 26.—Graptolite locality in the Woodchopper Volcanics.

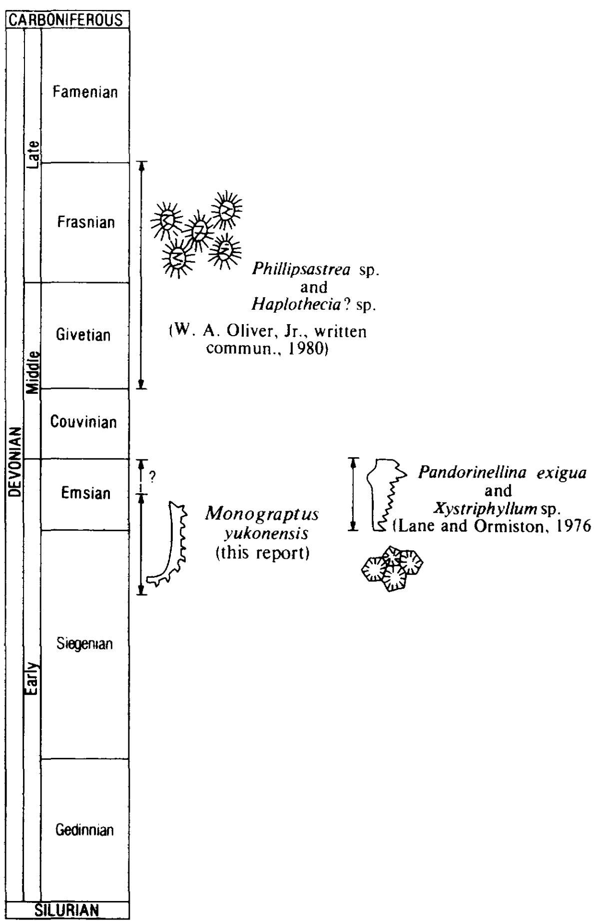

FIGURE 27.—Correlation of faunas from the Woodchopper Volcanics.

basalt breccia and tuff commonly occur together in other parts of Alaska where Devonian flows are found (Churkin, 1975).

Last summer's discovery consists of two fragmentary specimens of *Monograptus yukonensis* Jackson and Lenz. This cosmopolitan species is very distinctive and is restricted to the Early Devonian. One specimen is a straight, distal fragment of a rhabdosome, and the other is a curved, proximal fragment. According to Lenz and Jackson (1971), the single most distinctive character of *M. yukonensis* is the J-shaped rhabdosome (curved proximally, straight distally). The specimen from the Woodchopper Volcanics exhibits this characteristic shape and closely matches the dimensions cited by Lenz and Jackson (1971). *M. yukonensis*, one of the youngest known graptoloid graptolites, occurs in the Lower Devonian (Pragian) *M. yukonensis* Zone of the Road River Formation in the Richardson Mountains of Yukon Territory and in the southern District of Mackenzie (Lenz and Jackson, 1971). In Alaska, it has been found in the Road River Formation in the western Ogilvie Mountains near the Yukon boundary (Churkin and Brabb, 1965) and in southeastern Alaska on Prince of Wales Island. At the latter locality, it occurs in association with *Monograptus craigensis* Jaeger and marine shelly faunas and is considered to be Pragian (upper Siegenian through lower Emsian) and possibly younger (Churkin and others, 1970).

The graptolites discovered here are nearly the same age as the shelly fossils reported by Lane and Ormiston (1976) from outcrops of limestone associated with pillow basalts of the Woodchopper on the north side of the Yukon River. The Late(?) Devonian age of the corals reported by Oliver suggests that a large part of the Devonian may be represented by the Woodchopper (fig. 27). The Woodchopper Volcanics, previously considered to be of Early(?) and Middle Devonian age, is here considered to be of Early, Middle(?), and Late(?) Devonian age. Further study is necessary to determine if the many faults in the area may have juxtaposed basaltic flows of different ages.

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Description and preliminary interpretation of folds and faults in a small area in the Circle B-4 and B-5 quadrangles, Alaska

By Grant W. Cushing, Helen L. Foster, Jo Laird, and Anna C. Burack

A 15-km-long, northwesterly trending zone of torlike outcrops, as much as 10 m high, of complexly folded metamorphic rocks was discovered during the course of reconnaissance geologic mapping in the Circle B-4 and B-5 quadrangles (fig. 28). The tors seem to delineate the margin of a greenschist-facies metamorphic terrane, that differs somewhat in lithology and that appears more complexly folded than the terrane to the northwest of the tors. The metamorphic rocks of both of these terranes are of probable Paleozoic and (or) Precambrian age.

The rocks composing the tors and the terrane immediately to the southeast are impure quartzite, quartzitic schist, thin pelitic schists, and a mappable unit of calcareous chloritic schists with distinctive feldspar porphyroblasts. At least three generations of folds deform these rocks, and the folding seems most intense and the fold patterns are most disrupted in the zone of tors.

A foliation or schistosity S_1 is well defined throughout the terrane and appears to be most commonly parallel or subparallel to lithologic layering (except at fold hinges). The lithologic layering is believed generally to represent original bedding. This schistosity is parallel to axial planes of tight folds which are rarely observed in the field.

Second-generation folds, the most prominent set of folds in this terrane, deform the S_1 schistosity. Folds of this generation range from relatively open recumbent to tight recumbent and also include chevron and isoclinal folds (fig. 29). A second schistosity, S_2, has developed locally as a result of the mechanical rotation of the preexisting S_1 schistosity. In the pelitic rocks, it is observed as an axial planar fabric, but only rarely. The amplitudes and wavelengths of these folds range from a microscopic scale to approximately 10 m. Their axial plane is subhorizontal and strikes about east-west.

The third generation of folds, deforming all previous structural elements, is relatively open folds which are widely observed throughout the B-4 and B-5 quadrangles, varying only slightly in style and observable wavelengths and amplitude. They range from open symmetric to slightly overturned asymmetric with wavelengths and amplitudes less than approximately 50 cm. Generally associated with the folds is a slip cleavage that is subparallel to the axial plane.

The rocks to the northwest of the zone of tors are primarily quartzite and quartzitic schist with minor pelitic, mafic, and calcareous layers. The quartzite commonly contains muscovite and plagioclase and locally minor biotite, and chlorite is not abundant. In many places, the quartzite contains scattered grains of blue-gray, clear, or cloudy-white quartz that are considerably larger than the average grain size of the rock.

The metamorphic rocks to the northwest, compared to those to the southeast, are more quartzitic and less chloritic and locally contain the scattered large quartz grains. The calcareous greenschist with feldspar porphyroblasts, which is a mappable lithology in the southeastern terrane, is not found in the northwestern terrane. A group of magnesium and aluminum calc-silicate rocks is found in the northwestern but not the southeastern terrane. The pelite and quartzite in the southeastern terrane are more calcareous than those in the northwestern terrane.

The rocks of the northwestern terrane have a well-developed foliation, like the S_1 schistosity of the southeastern terrane, but the second-generation folds, which are so prominent in the terrane south of the tors, are not evident. However, relatively gentle open folds of the third generation are clearly present. This less deformed terrane to the northwest is intruded by a small granitic body and by numerous small felsic hypabyssal bodies of probable Tertiary age (fig. 28). Felsic intrusions have not been found in the southeastern terrane.

A northeasterly trending lineament visible on aerial photographs and Landsat imagery and evident in the topography is believed to represent the trace of a fault, probably with left-lateral strike-slip movement (fig. 28). The offset cannot yet be conclusively demonstrated, but the northwestern and southeastern terranes may be offset about 2 km.

Although the above presented lithologic and structural data suggest differences in the two metamorphic terranes described, the cause of these differences and their significance is not yet known. These two metamorphic terranes may have been widely separated and juxtaposed by thrusting, the more highly deformed southeastern terrane being thrust over the less deformed northwestern terrane. The zone of tors may mark the leading edge of the upper plate. Alternatively, they may be closely related terranes that were displaced only a few hundred or a few thousand meters by large nappelike folds; the more deformed terrane may represent the nose of the fold and the less deformed

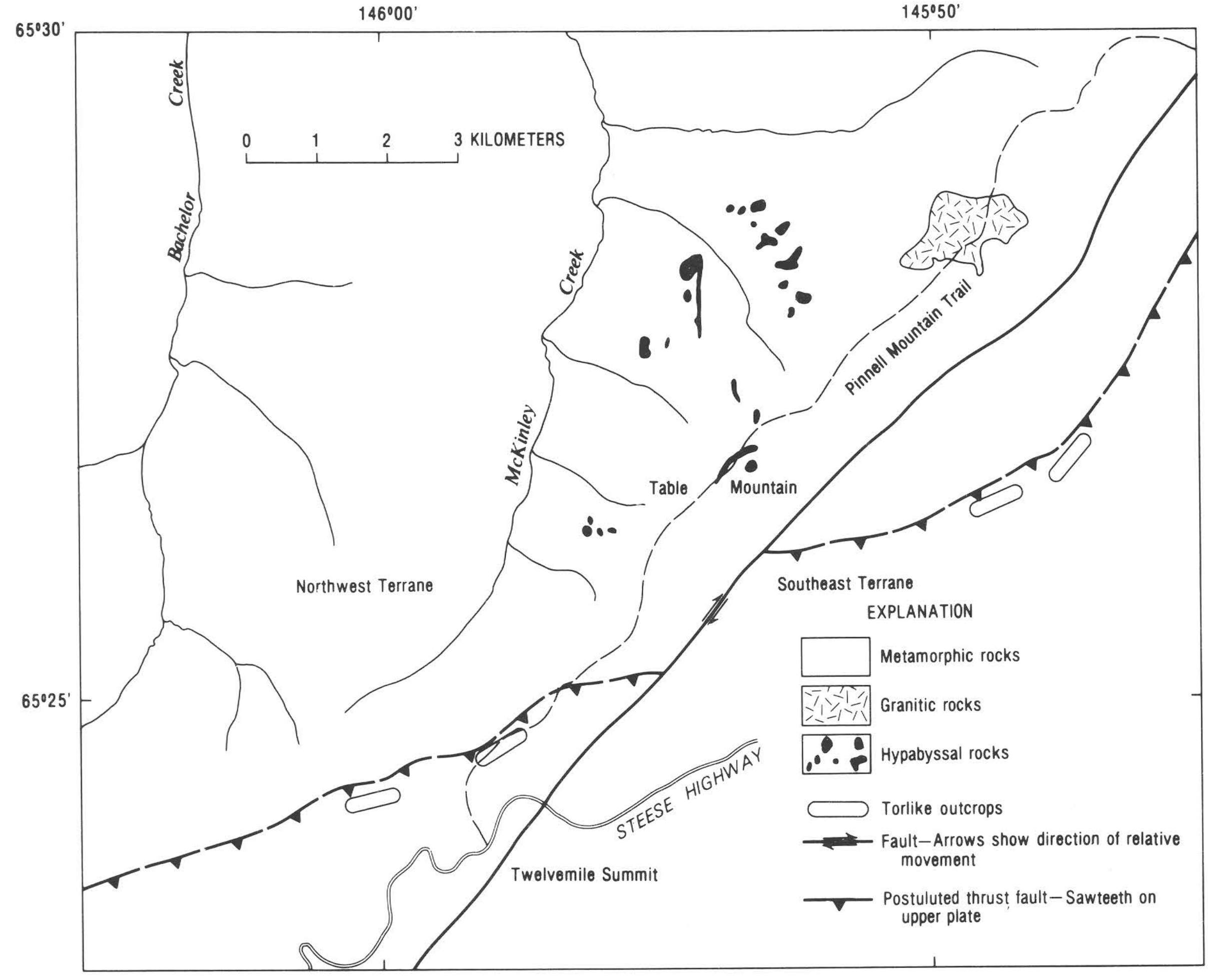

FIGURE 28.—Generalized geologic map of small portion of Circle B-4 and B-5 quadrangles.

terrane a fairly flat-lying and less disturbed limb. Detailed petrologic work is now in progress on the metamorphic rocks of the two terranes and may yield data useful in deciphering the structural history of the area.

Tertiary(?) conglomerate and Quaternary faulting in the Circle quadrangle, Alaska

By Florence R. Weber and Helen L. Foster

Sedimentary rocks of probable Tertiary age in the northwestern part of the Circle quadrangle, which were originally described by Mertie (1937, p. 175), are probably more extensive in occurrence then formerly considered. Several previously undescribed exposures were discovered during the course of reconnaissance geologic mapping in the summer of 1980, including especially well exposed sections in the bluffs north of the mouth of Loper Creek (fig. 30).

At the Loper Creek locality, deposits 60 to 90 m thick of fairly well to poorly consolidated Tertiary(?) conglomerate are overlain by 60 m of unconsolidated Quaternary alluvial gravel, all of which have been dissected into serrate ridges.

The conglomerate forms a brownish-yellow outcrop at the base of the bluff. Eighty percent of the clasts are well-rounded gray and black chert, 10 percent are white quartz, and the remainder consists of several kinds of lithic fragments, mostly gray and light-green phyllite. Clasts composed of igneous rock were not found. The matrix is a coarse sand of the same general composition as the clasts but containing much more lithic material with many angular granules. The clasts are as large as 20 cm in diameter, but about a 3-cm diameter is most common.

Beds of medium-grayish-brown sandstone are interlayered in the conglomerate and compose about 30 percent of this Loper Creek section. The sand grains are very coarse and consist of 30 percent clear quartz, 50 percent gray, green, and

FIGURE 29.—Typical recumbent second generation folds in the southeast terrane. S_1 schistosity is essentially parallel to lithologic layering at this locality.

maroon phyllite, and 20 percent lignite and clayey rocks. The grains are all angular, and the cement is argillaceous and iron-stained. The finer grained sandstone layers are a little better consolidated than the conglomerate, and a few of them bear brown plant impressions, mainly of stems or stemlike fragments.

Within the coarse layers, the conglomerate is poorly sorted. Unoriented fragments of lignite, as large as 20 cm in diameter, are present locally. Chunks of coal in float, probably derived from this formation, have been collected near the junction of the North Fork and Preacher Creek proper.

The overlying alluvial gravels are distinguished from the conglomerate at Loper Creek primarily by color and composition. The alluvial gravels are light gray and less resistant than the conglomerate and are made up of well-rounded clasts of which 60 percent are schistose quartzite and metamorphosed grit, 30 percent white quartz, 10 percent maroon and green phyllite or slate, and very little chert. They were probably derived mostly from the metamorphic terrane to the south and not from the Crazy Mountains to the north where chert is a major lithologic component.

The conglomerate section on the North Fork of Preacher Creek, originally described by Mertie (1937, p. 175), shows the same characteristics as the conglomerate at Loper Creek. It is coarse conglomerate composed primarily of chert or quartzite clasts with a muddy lithic wacke matrix. The deposits are also poorly sorted and contain a few beds of "salt and pepper" sandstone.

Conglomerate exposed on a tributary of Albert Creek forms a badlands-type topography and is light gray or pink. Very light colored layers consist of small, well-rounded white quartz pebbles with very little gray or black chert. The pink conglomerate consists mostly of chert pebbles in a sandy matrix. The pink color is derived from minute maroon or pink clay particles, possibly the weathering products of maroon slates that crop out in the Crazy Mountains. A few limonite-cemented beds are present. Hard, rounded boulders, as large as 45 cm in diameter, of chert pebble conglomerate are also a rock type found in outcrop immediately to the north. The chert pebbles in the conglomerate are probably several-generation stream gravels; lithic fragments could be derived from the Crazy Mountains, and the

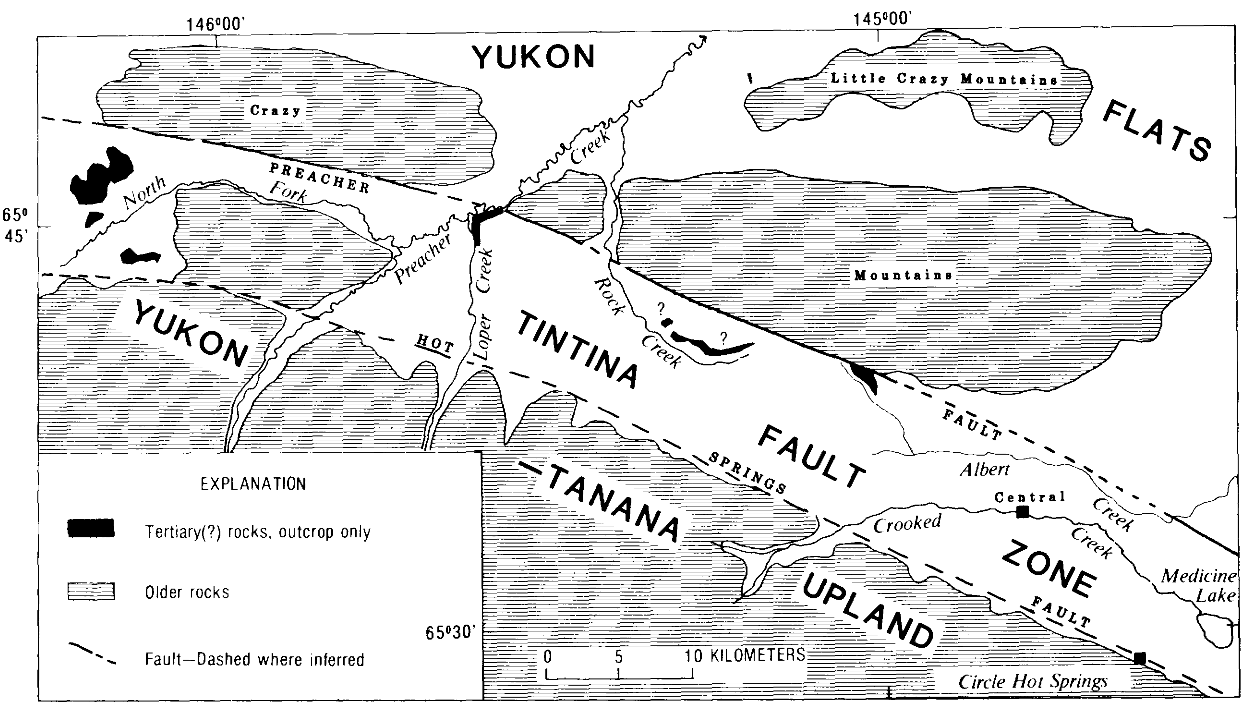

FIGURE 30.—Location of outcrops of Tertiary(?) rocks and Quaternary faults in northwestern part of Circle quadrangle.

lignite from nearby previously deposited sediments.

The age of the conglomerate is unknown but is most probably Tertiary, as Mertie (1937, p. 175) suggested. A Tertiary age is indicated because the unit is much less consolidated and the clasts differ in composition from the slightly metamorphosed polymictic conglomerate of Mesozoic age in the nearby Livengood quadrangle (Chapman and others, 1971), and yet it is better consolidated than the Quaternary terrace deposits bordering the Yukon River (Williams, 1962).

The lithology of the conglomerate, the poor sorting, absence of orientation of clasts, and presence of the argillaceous material in the matrix of the conglomerates suggest rapid local deposition, possibly as debris flows into a relatively shallow body of water. Bedded fines, good sorting, orientation of pebbles, and discontinuous channel deposits, rather than the fairly continuous sandstone layers that occur in these deposits, might be expected in fluvial environments.

The conglomerate may have been deposited in a depression, possibly a graben or trench along the Tintina fault zone or in a broad river valley following a fault zone. The Tertiary(?) deposits appear to be cut off on the north by a fault that borders the south side of the Crazy Mountains. The fault crosses Preacher Creek below the mouth of Loper Creek where the contact of the conglomerates with older slate, quartzite, and greenstone is exposed. This is called the Preacher fault after this exposure on Preacher Creek.

To the west this fault may be continuous with a major fault in the Victoria Creek valley (Chapman and others, 1971). In the hills east of Preacher Creek, a 6-m-high scarp marks the trace of the fault. Farther to the southeast, the fault trace disappears beneath modern alluvial deposits. The conglomerate is absent on the north side of the fault, and some alluvial gravels are displaced and may be thinner on the north side.

The north side of the fault appears to be up relative to the south side, and some right-lateral movement is suggested by possible slight offsets of Preacher Creek where the fault crosses it and by other possible slight offsets of smaller streams. However, at present there is insufficient evidence to prove right-lateral offset.

A distinct scarp 6 to 13 m high is evident in the surficial deposits near Medicine Lake (Davies, 1972, p. 215). This faulting may be an extension of the Preacher fault or a separate fault. Faulting in this area occurred in Quaternary time, possibly into the late Pleistocene, because the fault trace cuts some alluvium but is obscured by the Yukon River terrace gravels and by the most recent alluvium and colluvium. This faulting may represent the most recent visible effects of movement on the Tintina fault system in the Circle quadrangle. If this is so, much later movement appears to have occurred on this fault system on the northwest end of its trend than on the southeast end in Canada where the latest recognized movement is early Tertiary (Tempelman-Kluit, 1979, p. 26).

The south side of the depression in which the Tertiary(?) gravels may have been deposited is marked by the very straight mountain front on the Yukon-Tanana Upland which may also be a fault trace (Davies, 1972, p. 214), here called the Hot Springs fault (fig. 30). The inferred fault is named for Circle Hot Springs, which is located on the fault trend (fig. 30). Scarps along the fault, other than the mountain front itself, have been obscured by mass wasting processes.

Minor, recent seismic activity has been detected along the Tintina fault zone (Gedney and others, 1972), and within the past 10 years at least four moderate earthquakes have been recorded in the Crazy Mountain area (L. D. Gedney, oral commun., 1980).

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Two new tephra localities in the Yukon-Tanana Upland

By Florence R. Weber

A discontinuous layer of white volcanic ash as much as 1 cm thick was found during a study of soil profiles on a terminal moraine at the mouth of an unnamed tributary of Joseph Creek, about 11 km northwest of the old village of Joseph (fig. 31). This ash was found below 6 cm of turf and soil. It contains glass, plagioclase feldspar, hornblende, and hypersthene and is similar in mineralogy to that described for the rhyodacitic White River Ash (Lerbekmo and Campbell, 1969).

Prindle Volcano (fig. 31) is only 100 km from Joseph, but no trace is known of an ash fall related to its eruption and the volcano is of basaltic composition. Furthermore, the Prindle Volcano lava flow underlies the White River Ash and is older (H. L. Foster, oral commun., 1980).

The White River Ash has a wide distribution in extreme central-eastern Alaska and the adjoining parts of Canada—in the Yukon and Northwest Territories. It was derived from a vent in the eastern Wrangell Mountains and has been radiocarbon dated at approximately 1,500 years B. P. (Lerbekmo and Campbell, 1969).

Excavations on moraines 9.5 km to the southwest in the same tributary valley revealed no ash layer, so the Joseph Creek locality places the presently known westward extent of the ash northwest, but only slightly beyond the limit originally estimated by Bostock (1952), on the basis of widely spaced stations (fig. 31).

A second ash was located during the summer of 1980 on the upper part of the Middle Fork of the Fortymile River, 17 km southeast of Mount Harper (see "0" on fig. 31). The tephra at this site is exposed in a 16.5-m-high cutbank on the west side of the river. The white ash is of variable thickness, as much as 13 cm, and is made up either of a single thick layer or of several closely spaced thin layers. It is significantly disrupted by cryoturbation. This ash is found about 1 m from the surface below 60 cm of loess and near the top of an 11-m section of alternating thin layers of fine, gravelly sand and silt with scattered thin bands of organic material. The finely and continuously layered beds may represent sediments deposited in a lake or ponds developed when the Middle Fork drainage was temporarily blocked by a glacier extending across its course at the mouth of Molly Creek, 5.6 km downstream. The glacier, of unknown age, is identified as the most extensive (and perhaps one of the oldest) emanating from the Mount Harper area.

A cursory petrographic examination indicates no similarity to the White River Ash, and no ash with White River characteristics has been found in this area. Plagioclase is present, but hornblende is rare or absent, and hypersthene is

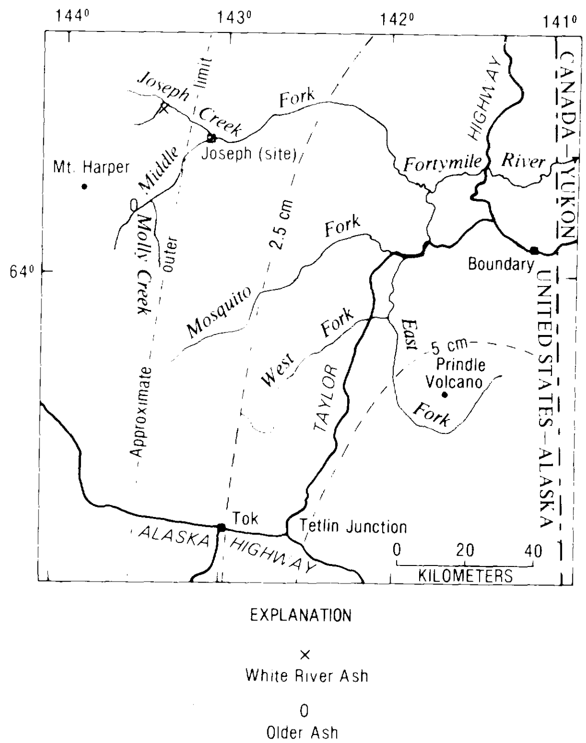

FIGURE 31.—Location of two new tephra localities in Yukon-Tanana Upland. Dashed lines indicate approximate thicknesses of the White River Ash (Bostock, 1952; Lerbekmo and Campbell, 1969).

absent. Several volcanic ashes have been described from the interior of Alaska (Péwé 1975, p. 79-81). Further work is planned to determine the affinity and age of this tephra.

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Mineralized zones in bedrock near Miller Creek, Circle quadrangle

By Richard B. Tripp, David E. Detra, and James M. Nishi

Several unreported mineralized zones in bedrock were located in T. 8 N., R. 3 W., near Miller Creek in the Yukon-Tanana Upland metamorphic terrane, during a reconnaissance geochemical survey of the Circle quadrangle that was done as part of the Alaska Mineral Resource Assessment Program (AMRAP).

The most notable zone found is 0.6 m thick and traceable for about 6 m. Spectrographic analysis of the heavy-mineral-fraction of bulk samples of mineralized quartzite and schist taken from the zones indicated anomalous amounts of copper (2,000 ppm) and zinc (3,000 ppm). In places the zone contains crystalline pods and stringers of galena associated with cerussite, mimetite, and iron oxides. Some of the galena that was isolated and analyzed for trace elements using a semiquantitative spectrographic method (Grimes and Marranzino, 1968) showed anomalous amounts of tin (>2,000 ppm), bismuth (2,000 ppm), antimony (1,000 ppm), silver (5,000 ppm), cadmium (1,000 ppm), and copper (1,000 ppm). Preliminary scanning of the galena with an electron microscope did not reveal the nature of the included tin, bismuth, and antimony, but their presence suggests a moderate to high temperature origin.

Other sites in the Miller Creek area have irregularly shaped sulfide zones as much as 6 m in diameter with finely disseminated grains and crystals of arsenopyrite and some pyrite. Spectrographic analysis of the heavy-mineral-fraction of a bulk rock sample from one of these sites indicates anomalous amounts of gold (20 ppm), copper (2,000 ppm), antimony (700 ppm), lead (200 ppm), tungsten (100 ppm), and bismuth (100 ppm).

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Gold found in bedrock of Lost Chicken Creek gold placer mine, Fortymile area, Alaska

By Helen L. Foster and Richard M. O'Leary

Gold was detected in sheared and altered diorite in a bedrock cut in the Lost Chicken placer gold mine, Fortymile area, Eagle quadrangle, Alaska. Gold in bedrock has only rarely been found directly associated with placer deposits in the Fortymile area and other placer districts of the Yukon-Tanana Upland.

Lost Chicken Creek is a small tributary, only about 2 km long, of the South Fork of the Fortymile River (fig. 32). The upper part of the creek has been modified by a ditch tapping water from tributaries to Chicken Creek and by small reservoirs. The creek in most places flows over alluvial fill and peat-covered terrace gravels, mostly unrelated to the present Lost Chicken Creek.

Bedrock in the Lost Chicken drainage area is believed to be mostly altered diorite and quartz diorite near the margin of a pluton. Some metamorphic contact rocks may also exist, but exposure of bedrock in the area is very limited. The bedrock differs from that of neighboring Chicken Creek on the west, which is underlain mostly by Tertiary sedimentary rocks, and that of the next unnamed creek, to the east, which is underlain by greenschist-facies metamorphic rocks (Foster, 1976).

Lost Chicken Creek has been placer mined discontinuously since about 1895. In the past few years, the lower part of the valley has been hydraulically mined.

When the mine was visited in early September 1980, good cuts in greenish-gray highly sheared diorite were exposed about 2,000 m above the mouth of the creek. The rock is cut through by many calcite veinlets, which appear to be mostly fracture fillings. It contains abundant sulfides, and much of it is stained yellow and orange-brown. Six grab samples of bedrock were collected from one cut and analyzed for gold using the atomic absorption method. Two of the six samples contained 0.10 ppm gold. Gold was not detected in the other four samples (limit of detection 0.05 ppm). The samples that contained gold were both stained orange-brown, and one contained large amounts of visible sulfides.

Gold has been found in quartz-calcite veins and calcite veinlets cutting greenschist-facies metamorphic rocks about 6 km northwest of the mouth of Lost Chicken Creek (Purdy Quartz mine). Also, traces of gold occur in calcite veinlets cutting greenschist-facies metamorphic rocks just north of Stonehouse Creek about 7 km north of the mouth of Lost Chicken Creek. A third occurrence of gold occurs in sheared granitic rock exposed along the Mosquito Fork of the South Fork of the Fortymile River about 6 km above (west) of the mouth of Lost Chicken Creek and nearby at the head of Lilliwig Creek (Mertie, 1937, p. 244). These limited bedrock occurrences of gold are the only ones presently known in the Fortymile area that are near or associated with significant gold placer deposits.

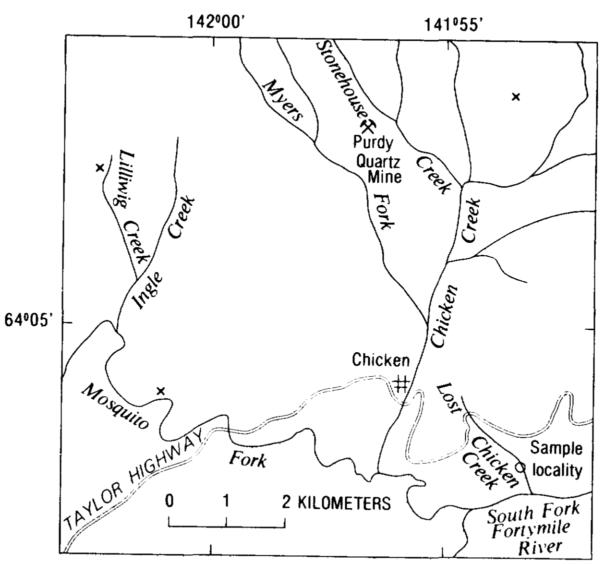

FIGURE 32.—Sketch map showing location of grab samples from Lost Chicken gold placer mine. Symbols "X," and "⚒" indicate localities where gold has been found in bedrock.

Placer gold from Lost Chicken Creek differs in appearance and composition from that of neighboring creeks but, in itself, is not uniform, and its appearance and composition vary considerably, as shown by the following results from semiquantitative six-step spectrographic analyses.

[Looked for but not detected: Fe, Be, B, Co, Cr, La, Pt, Sb, Sc, Sn, V, Y, Zn, Zr, Ga, In, Yb, Ir, Rh, Ru. N, not detected at the given limit of detection; G, greater than 10 percent. Analyst: A. L. Sutton]

	Ag ppm	Cu ppm	Pd ppm	
Reddish yellow placer gold	20,000	1,000	N2	
Small rounded nuggets	G	70	N1	possible inclusions
Small flattened nuggets	30,000	599	7	
Rods	70,000	200	N1.5	inclusions (Robinsonite?)
Leaves	G	150	N1.5	

The bedrock analyses suggest that some Lost Chicken placer gold is probably very locally derived from sheared mineralized diorite and quartz diorite, but the differences in appearance and composition of the gold and the physiographic aspects of Lost Chicken Creek suggest that much of it has been derived elsewhere and has a long and complex history.

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Placers and placer mining, Circle District

By Warren E. Yeend

Approximately 500 miners were working in the Circle district during the summer of 1980. This extensive activity has been referred to by the Alaskan newspapers as a modern-day Alaskan gold rush. Placer gold has been mined in the Circle district since 1893, and roughly 730,000 troy ounces of gold has been produced during the past 87 years. The gold is concentrated in alluvial and colluvial deposits in the stream valleys draining into Birch and Crooked Creeks in the east-central part of the Circle quadrangle.

Mining operations typically include a sluice box with 1 or 2 bulldozers or a front-end loader. The old hydraulic method is still employed in the remote area on the North Fork of Harrison Creek where abundant water is available (fig. 33). Small suction dredges are also used. Generally two to five miners are employed at any one operation. During a typical shift of 8 to 12 hours, 200 to 1200 yd^3 of gravel is washed. Gold values ($500/ounce) range from $0.50 to $17.00 per yd^3 and are more commonly $3.00 to $8.00 per yd^3.

The gravels being mined are generally composed of locally derived quartzite, and quartz mica schist and, rarely, porphyritic granite. A 1- to 2-m-thick layer of muck commonly must be thawed and stripped from the gravel before mining begins. The gold-bearing gravel beneath the muck ranges from 2 to 5 m in thickness and rests on fractured and locally weathered bedrock. Generally, 0.5 to 1 m of bedrock is ripped and washed to ensure complete recovery of gold that may have worked into the bedrock joints and fractures. The gold particles, 0.5 to 3 mm in diameter, are commonly flattened and have smooth rounded edges. Nuggets of more than 1 troy ounce are rarely found. Because gold particle size does not systematically decrease in a down-valley direction in all drainages the gold probably comes from more than one specific source area.

Heavy minerals observed by microscopic examination of concentrates include ilmenite, garnet, zircon, magnetite, hematite, tourmaline, rutile, pyrite, sphene, and limonite. Cassiterite, scheelite, and wolframite are also sporadically present in the heavy fraction, occasionally in sizable amounts. Because almost all the heavy-mineral fragments are subangular to angular, their source is probably local. A few very well rounded pale-purple to clear zircon grains may be second-cycle minerals derived from a fossil placer.

The gold source is not immediately apparent. There are no recognized fossil placers in the area, and a local lode source has not been discovered. The gold occurrence, seemingly unrelated to outcropping metamorphic and plutonic bedrock types, and a lack of systematic down valley decrease in the size of gold fragments would seem to suggest a fossil placer source, the evidence of which has all but been destroyed; all, except for the gold and possibly the few well-rounded zircon grains. This meager evidence for a fossil placer interpretation is not conclusive, but bears consideration.

Whatever the source, a sizeable resource of gold still remains in the Circle district. Many of the gold-bearing stream channels previously left unmined have become attractive as a result of the recent increases in gold prices. A moderately large, low-grade, but as yet largely unevaluated, gold resource may be contained in the extensive valley-fill deposits in the lower reaches of Crooked and Birch Creeks, as well as in the broad topographic trough on the south side of the Crazy Mountains.

Pollen studies of Quaternary-age lake sediments in the Tanana Valley, Alaska

By Thomas A. Ager

Pollen analysis of sediment cores from Blair Lakes (Fairbanks B-1 quadrangle), Harding Lake (Big Delta B-6 quadrangle), an unnamed pond on the Delta-age moraine (Péwé, 1975) near Delta Junction (Mount Hayes D-4 quadrangle), and Quartz Lake (Big Delta A-4 quadrangle) in the Tanana Valley (fig. 34) permit refinement of the regional Quaternary history of vegetation and climate. The pollen records from these Tanana Valley lakes, supplemented by previous studies in the region by Ager (1975) and Matthews (1974), are the basis for interpreting the vegetational history of the valley during the past approximately 35,000 years. The regional pollen record is most detailed for the past 16,000 years.

FIGURE 33.—Hydraulic mining on North Fork of Harrison Creek, Circle quadrangle.

During the middle Wisconsinan interstadial interval, the lowlands of the Tanana Valley had a muskeg vegetation cover, probably patchy in distribution. Spruce trees were important elements of this vegetation, along with willows, dwarf birch, ericales (for example, blueberry, labrador tea), herbs, and *Sphagnum* moss. Alders were present in the Tanana Valley at that time, but these shrubs were much less common than they have been in interior Alaska since middle Holocene time. At the end of the interstadial (about 27,000 years ago) the climate shifted to a more continental regime, characterized by aridity, severe winter temperatures, and high winds. It is likely that late Wisconsinan summer temperatures were lower than present-day summer temperatures in the region. Regional lowland vegetation shifted from partially forested muskeg to herbaceous tundra characterized by a predominance (in the pollen assemblages, at least) of grass, sedge, sage (*Artemisia*), willows, and numerous herbs. Trees and shrubs were exceedingly rare; spruce and tree birches may have been eliminated entirely from Alaska. Larch, balsam poplar, and quaking aspen may have survived in a few scattered sites, but there is no direct evidence of this yet. Evidence from western Alaska (among others, Ager, 1981; Colinvaux, 1964; Schweger, 1976) and elsewhere in Alaska and northwestern Canada (for example, Colinvaux, 1967; Cwynar and Ritchie, 1980) indicates that herbaceous tundra vegetation covered most of the unglaciated parts of these regions and the Bering land bridge during the late Wisconsinan glacial interval.

By about 14,000 years ago, a vegetation change began in the Tanana Valley and other areas of the state such as the Yukon Delta (Ager, 1981). This change is reflected in pollen records as a dramatic increase in percentages of pollen of dwarf birch (*Betula nana* and (or) *B. glandulosa*). Pollen percentages of other shrubs also increased. The pollen evidence suggests that the predominant vegetation of Alaska was shifting from a predominantly sparse, herbaceous tundra type to a more mesic shrub tundra type. This change appears to have been a vegetational response to a probable climatic shift from relatively cool and dry summers to warmer and moister summers.

By about 11,000 years ago, poplar trees (probably *Populus balsamifera*) began to spread rapidly throughout interior Alaska. Poplar pollen is poorly preserved in most sediments, but a core from the deepest part of Harding Lake in the Tanana Valley (Nakao and others, 1980) yielded a well-preserved record for poplar between about 11,300 and 9,500 years ago (Ager, unpub. data). By about 9,500 years ago, spruce trees began to spread throughout the interior, perhaps after migrating from northwest Canada. The population of alders expanded rapidly in interior Alaska beginning about 8,400 years ago. A regional decline in spruce occurred about 7,500 to 6,000 years ago. By about 6,000 years ago, spruce populations returned to stable conditions. No detectable vegetation changes have occurred in the Tanana Valley during the past 6,000 years.

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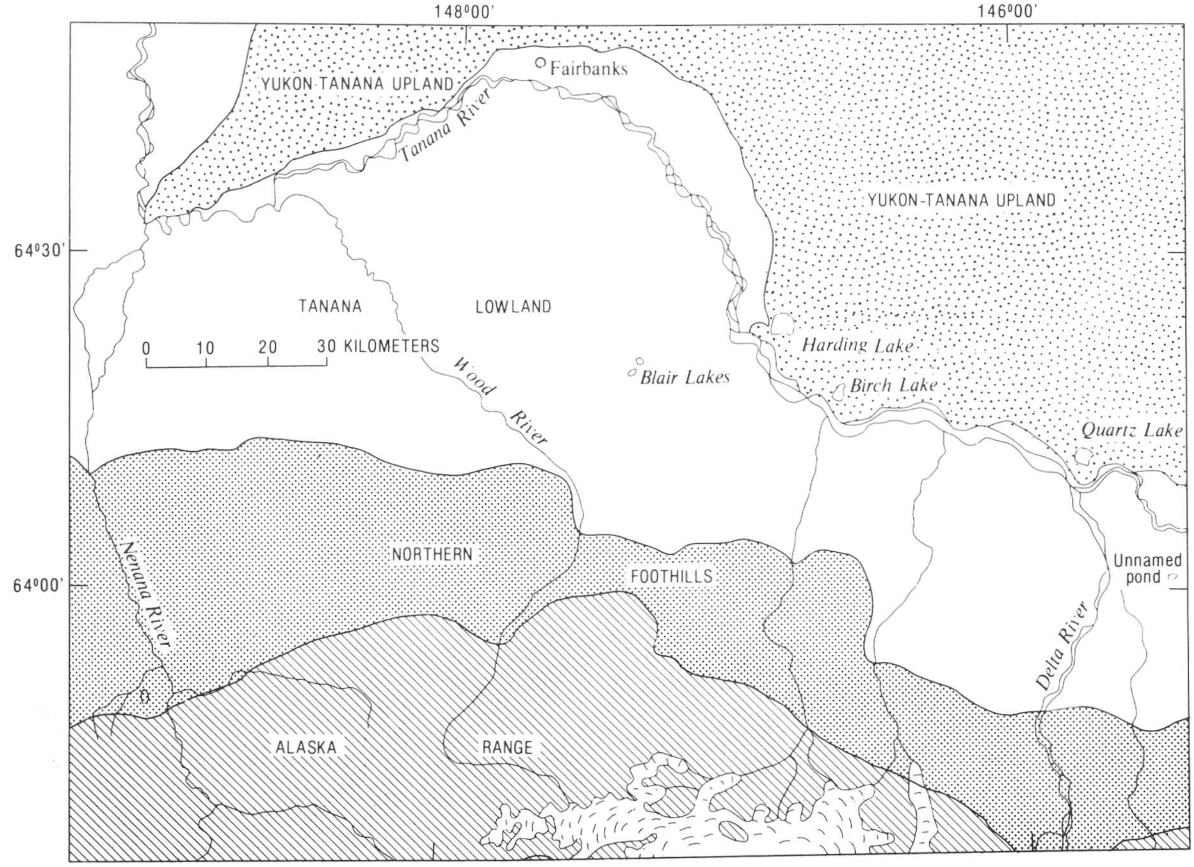

FIGURE 34.—Index map of middle Tanana Valley and adjacent areas. Sediment cores were obtained from Blair Lakes, Harding Lake, Birch Lake, Quartz Lake, and unnamed pond east of Delta River.

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Sediment transport in the Tanana River in the vicinity of Fairbanks, Alaska

By Robert L. Burrows, William W. Emmett, and Bruce Parks

To facilitate design and operation of engineering structures on the Tanana River and regulate gravel extraction from the river in the vicinity of Fairbanks, the U.S. Army Corps of Engineers, Alaska District, requested that the U.S. Geological Survey collect and evaluate sediment-transport and river-hydraulic data during periods of principal runoff, beginning in 1977.

Measurements of the sediment load of the Tanana River in the vicinity of Fairbanks, Alaska, for 1977-79 show that suspended-sediment transport rate, G_S, in tons per day, relates to water discharge, Q, in cubic feet per second, where r is the correlation coefficient, as

$$G_S = 1.210 \times 10^{-7}\, Q^{2629}\ (r^2 = 0.950)$$

The bedload-transport rate is usually 1 to 3 percent of the suspended-sediment transport rate (fig. 35).

Data collected at Fairbanks and upstream from Fairbanks near North Pole (fig. 36) show little difference in size distribution of suspended sediment between the two locations. The median particle size of suspended sediment is generally in the silt range, but at some low-water discharges, the median particle size is in the very fine sand range.

The median particle size of bedload near North Pole is generally in the gravel range, but at some low transport rates, the median particle size is in the medium sand range. At Fairbanks, data collected in 1977 indicate median particle sizes of bedload comparable to those of the upstream location, whereas data collected in 1978 indicate a marked decrease in median particle size of bedload between the two locations. Bed-material data collected in 1979 show that generally coarser material was transported at both sites but indicate an even greater decrease in particle size of bedload between North Pole (median particle size in the

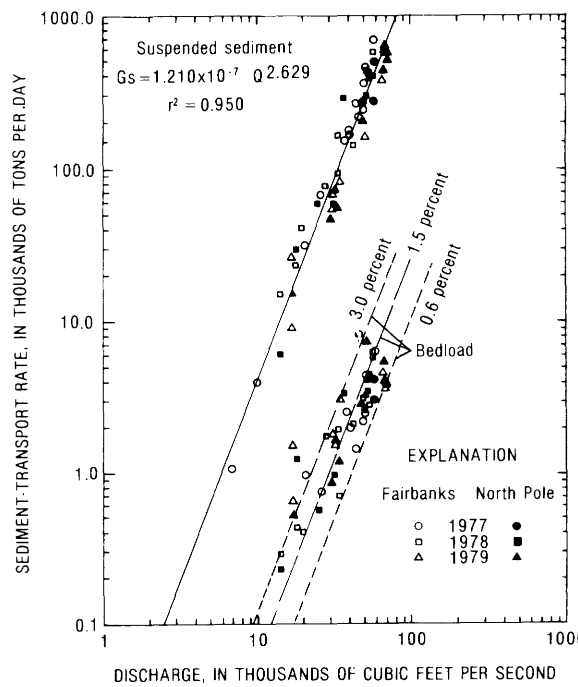


FIGURE 35.—Sediment-transport rate as a function of discharge, Tanana River at Fairbanks and near North Pole.

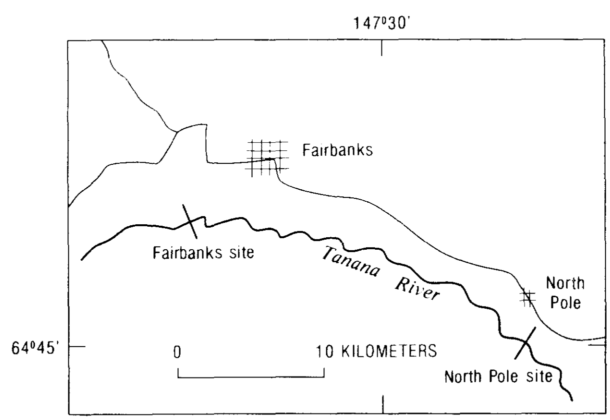


FIGURE 36.—Data collection sites for Tanana River sediment-transport study.

coarse gravel range) and Fairbanks (median particle size in the medium gravel range in the main channel but in the fine sand range in the overflow part of the channel) (fig. 37). For both locations and at all water discharges and sediment-transport rates, particles constituting the suspended load are significantly smaller than particles constituting the bedload.

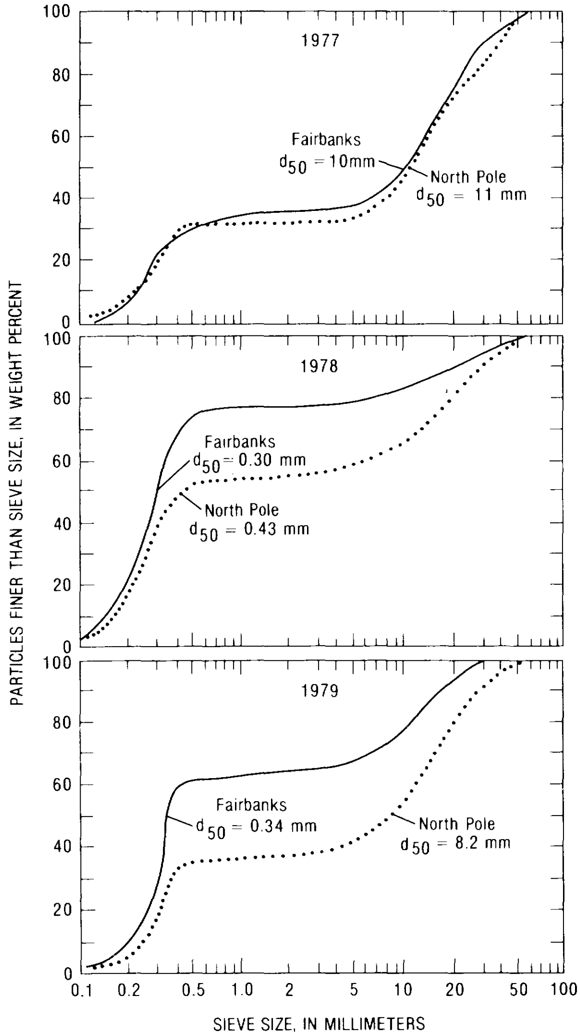


FIGURE 37.—Particle-size distribution of bedload; comparison of Fairbanks and North Pole by year.

Geohydrology of the Fairbanks North Star Borough, Alaska

By Andrea Krumhardt

The geohydrology of the Fairbanks North Star Borough, Alaska, is a continuing program designed to provide basic hydrologic information for land-use planning. From 1975 to 1980, emphasis has been largely directed toward uplands hydrology. Increasing pressure to subdivide lots into smaller parcels in the lowlands has shifted the emphasis to these areas. The major concern is that closely spaced private septic systems may contaminate the aquifer, particularly where the water table is less than 5 m below the land surface.

Observation wells located within the lowland study area (fig. 38) are being tested on a quarterly basis for one year to determine seasonal variation in water quality. Water levels are being measured monthly for one year to determine seasonal flow directions. In addition, water from private wells located in subdivisions with varying lot sizes is being tested for water quality characteristics.

The water is being tested for hardness and the following constituents: arsenic, iron, nitrate, chloride, phosphate, sulfate, dissolved solids, and fecal coliform bacteria. Of 53 analyses returned from the U.S. Geological Survey central laboratory, the range of reported values for hardness and the above constituents (except fecal coliform bacteria) are as follows:

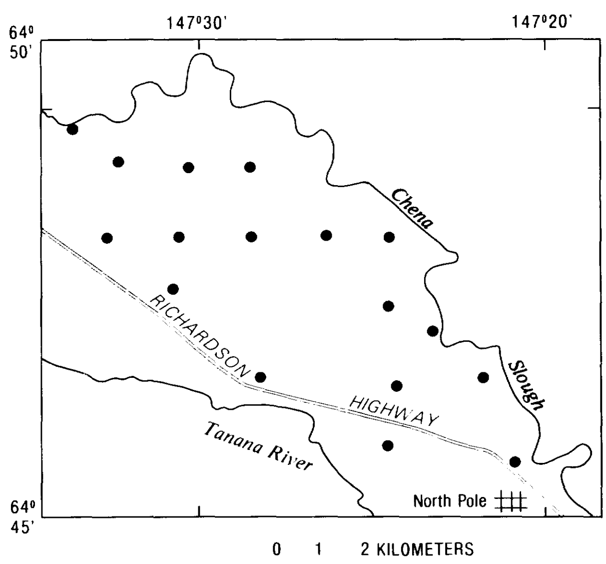


FIGURE 38.—Location of observation wells (dots) in lowland study area, Fairbanks North Star Borough Project.

Hardness	130 -342.1	mg/L
Arsenic	0 - 27	ug/L
Iron	.09 - 53	mg/L
Nitrate	.0 2.1	mg/L
Chloride	.8 - 15	mg/L
Phosphate	.0 - 2.0	mg/L
Sulfate	.9 - 31	mg/L
Dissolved solids	162 -342	mg/L

To date, iron is the only constitutent that has exceeded any limits set by the U.S. Environmental Protection Agency. Fecal coliform bacteria tests were performed in the Fairbanks Subdistirct Office. Except for one sample which showed 1 colony per 100 mL, all results were negative.

WEST-CENTRAL ALASKA

(Figure 39 shows study areas discussed)

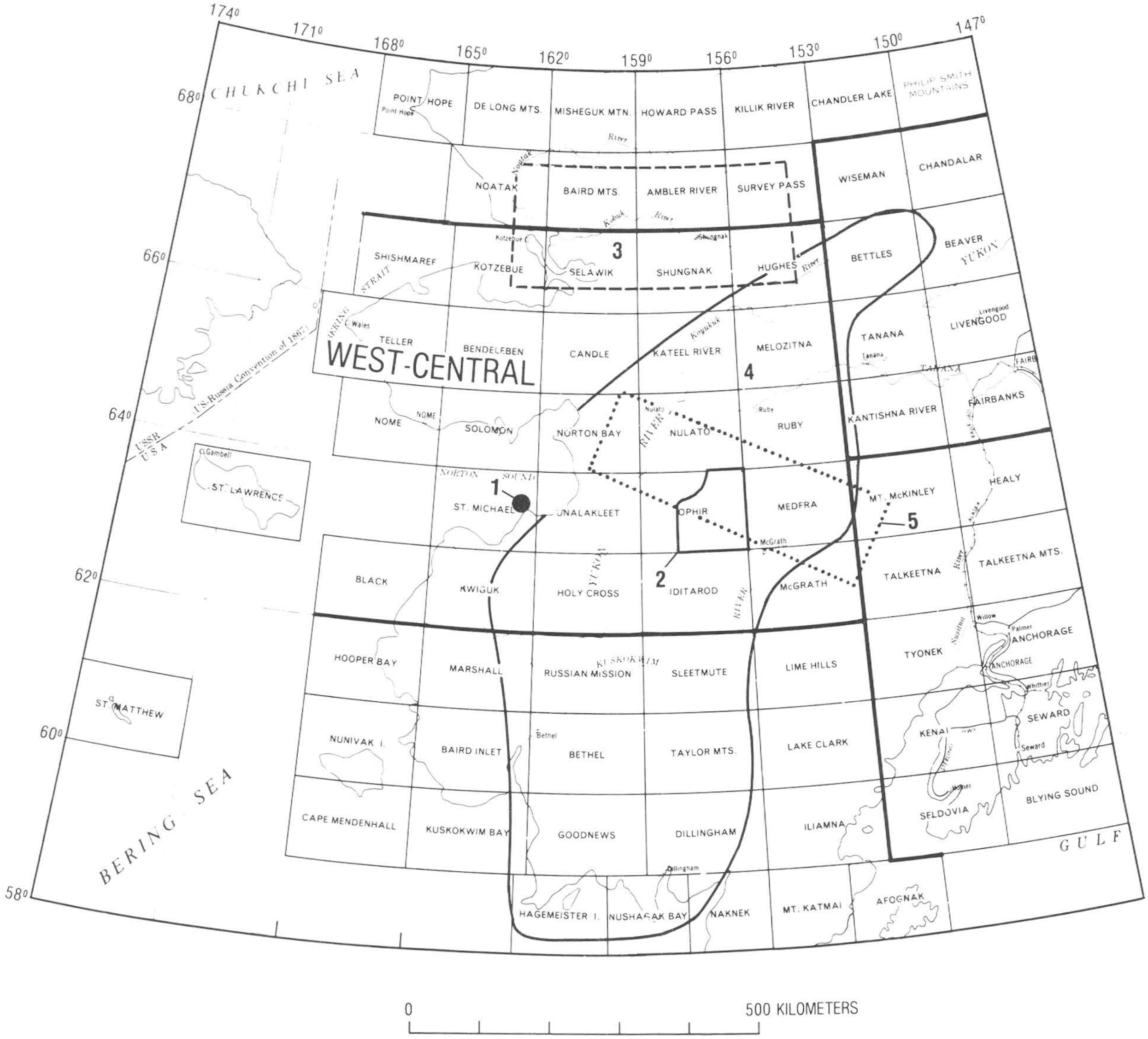


FIGURE 39.—Areas and localities in west-central Alaska discussed in this volume. Authorship and applicable figures relating to the numbered areas are listed below. 1, Ager and Bradbury, figures 45 and 46; 2, Chapman, Patton, and Moll, figure 40; 3, Childers and Kernodle, figure 47; 4, Moll and Patton, figure 41; 5, Patton and Moll, figures 42, 43, and 44.

Preliminary summary of the geology in eastern part of Ophir quadrangle

By Robert M. Chapman, William W. Patton, Jr., and Elizabeth J. Moll

Eleven bedrock units that range in age from Precambrian(?) and early Paleozoic through Cretaceous and (or) Tertiary and a unit of Quaternary unconsolidated deposits have been identified in the initial geologic reconnaissance of the eastern part of the Ophir quadrangle. The interpretations and data in this report are tentative, pending completion of petrologic studies, paleontologic and geochemical analyses, and additional field research.

Low-grade metamorphic rocks comprising phyllite, greenschist, pelitic schist, quartzite, calcareous schist, marble, and dolomitic marble form northeast-trending outcrop belts in the central and northern parts of the map area (fig. 40). Marble and dolomitic marble were found only at the north and south ends of the belt in the central part of the map area. In the absence of evidence for a definitive age, a Precambrian(?) and early Paleozoic age is assigned to the metamorphic rocks on the basis of their

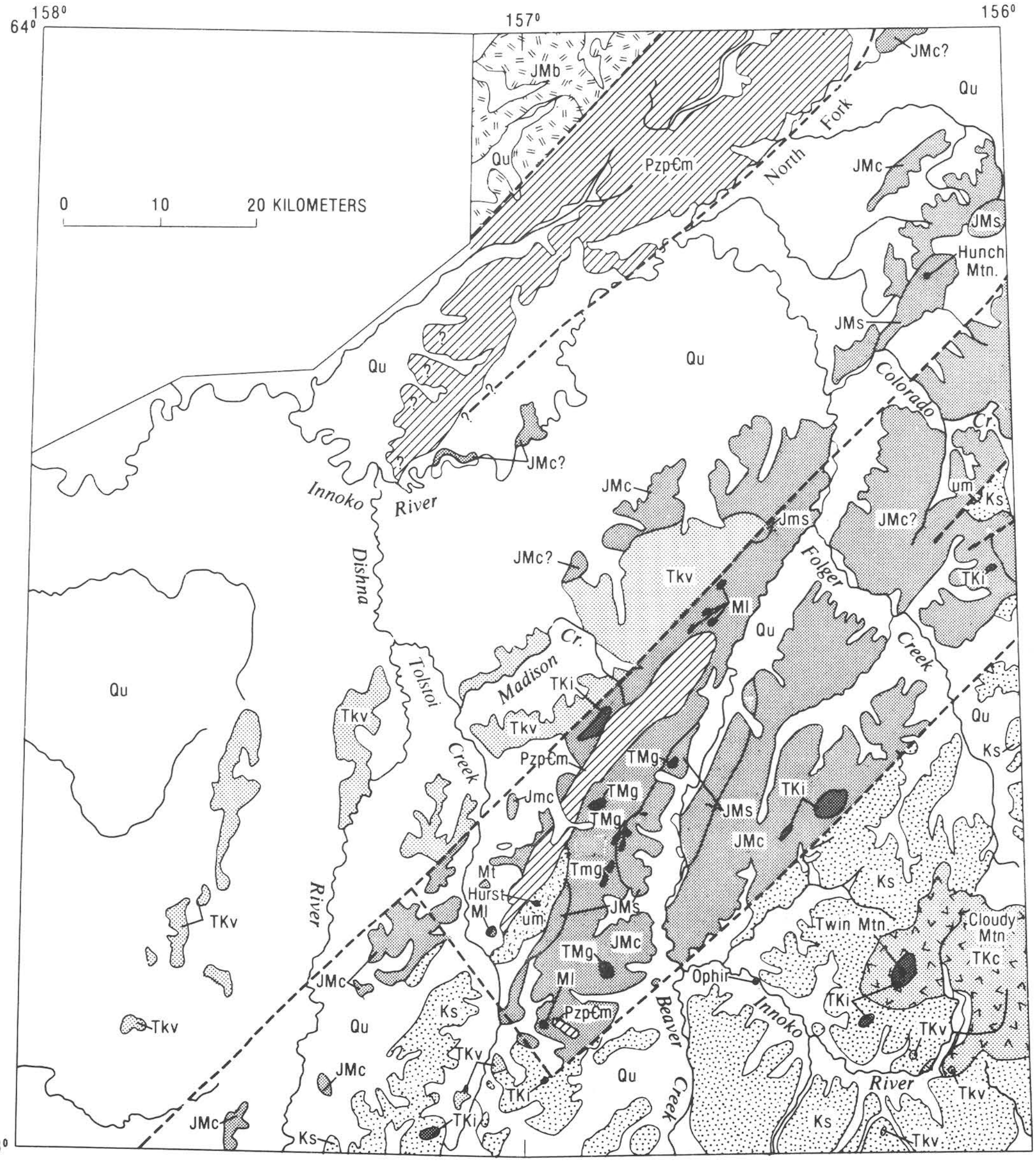
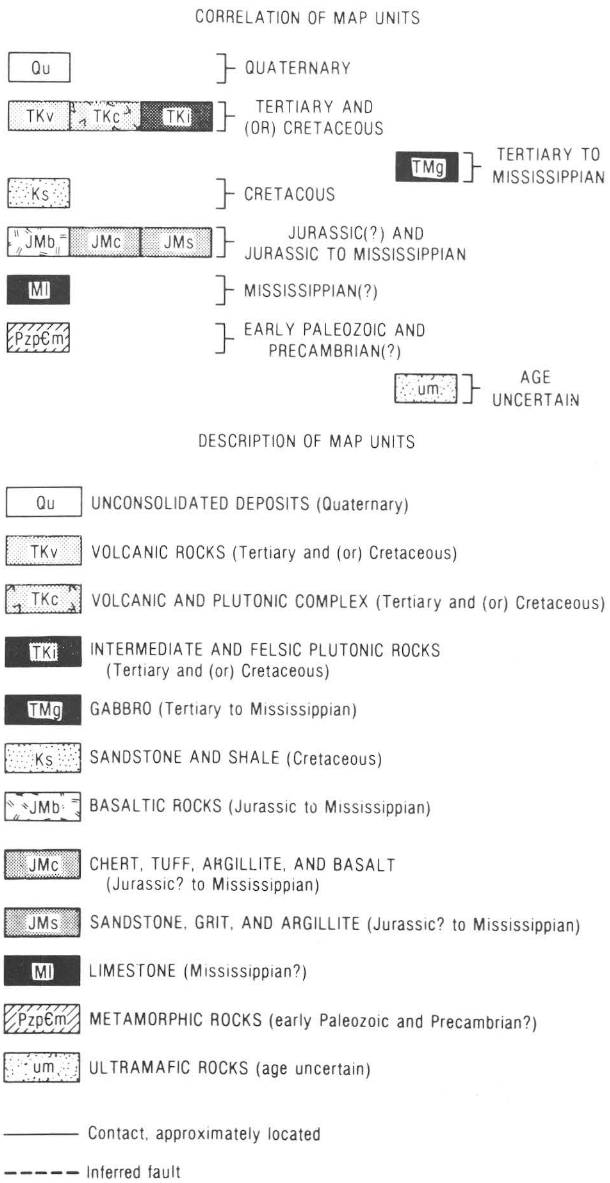


FIGURE 40.—Preliminary geologic map of the eastern part of Ophir quadrangle.

similarity in setting, lithology, and metamorphic grade to rocks of this age in the Nulato, Ruby, and Medfra quadrangles.

A chert, tuff, argillite, and basalt unit, which includes radiolarian-rich chert, crystal-lithic tuff, and some sandy beds, trends northeast through the central part of the map area. The stratigraphic sequence and ages represented in this structurally complex unit are not yet adequately known. Radiolarians from three localities near lower Folger Creek have been studied, and Mississippian and Late Triassic ages, respectively, were determined for two of the collections (B. K. Holdsworth, written commun., 1978); the third is assigned an Early Pennsylvanian age (D. L. Jones, oral commun., 1980). This unit is coextensive along strike to the northeast with lithologically similar cherty units in the Medfra quadrangle that contain radiolarians and other fossils of Mississippian, Pennsylvanian, and Triassic and Early Jurassic(?) ages (Patton and others, 1980), and with a chert and argillite unit in the Ruby quadrangle that contains radiolarians, originally assigned latest Devonian and Mississippian ages (Chapman and Patton, 1979) but now identified as Mississippian (B. K. Holdsworth, oral commun., 1980).



Six isolated small bodies of impure, slightly recrystallized, light-gray limestone were found within, and in apparently conformable sequence with, the chert, tuff, argillite, and basalt unit; four of these are 10 km northeast of the head of Madison Creek, and two are east of upper Tolstoi Creek (fig. 40). Crinoid and other fossil fragments were noted at two of the localities near Madison Creek. The limestone 6 km southwest of Mount Hurst yielded a reworked assemblage of conodonts. Anita G. Harris (written commun., 1980) reports that these conodonts "are incredibly broken up and the fragments are ragged and crumb-like and have sharp angular breaks," which she attributes "to possible boring by fungi and algae coupled with hydraulic activity." This mixed fauna contains "Late Devonian elements (*Palmatolepsis*), earliest Mississippian (Kinderhookian) elements (*Siphonodella*), indigenous and (or) redeposited late Early Mississippian (Osagean) elements (*Gnathodus texanus pseudosemiglaber*), and various other longer ranging, though strictly Mississippian, species. Thus the age of the rock is no older than latest Osagean; it could be even younger if the Osagean conodonts are also redeposited." Similar small lenticular beds of limestone in the northwest corner of the Medfra quadrangle contain Mississippian conodonts and foraminifers (Patton and others, 1980). Therefore, a probable Mississippian age, no older than late Early, is suggested for the limestone unit in the map area.

A sandstone, grit, and argillite unit, mapped near Mount Hurst, along the Innoko River, and at Hunch Mountain, is characterized by partly calcareous, medium-dark-gray graywacke sandstone, dark-gray argillite-shale, and grits with clasts of chert, argillite, tuff, and other rocks. No fossils were found. The age of this unit and the nature of its contact with the chert, tuff, argillite, and basalt unit, from which the clasts may have been derived, are uncertain. A conformable contact between these units is suggested, but unproven, by field data, and a Mississippian to Early Jurassic(?) age is assigned to both units. These units may be correlative with the Permian graywacke and Mississippian chert and argillite units of the Ruby quadrangle (Chapman and Patton, 1979), and with parts of the Mississippian and Pennsylvanian chert and limestone and the Triassic and possible Lower Jurassic cherty tuff, crystal and lithic tuffs, and volcanic breccia units in the Medfra quadrangle, 10-20 km northeast of Hunch Mountain (Patton and others, 1980).

The unit of basaltic rocks at the north edge of the map area is extrapolated from mapping in adjacent parts of the Nulato and Ruby quadrangles and probably includes mafic volcanic, tuffaceous, and sedimentary rocks typical of the Rampart Group (Chapman, 1976). The unit is inferred to be in unconformable, possibly thrust-fault, contact with the metamorphic rocks and is presumably Mississippian to Jurassic in age.

The sandstone and shale unit includes both Lower and Upper Cretaceous sandstone, siltstone, shale, conglomerate, and volcanic graywacke that are moderately deformed by folds and faults. The structural trend is northeast, and beds generally dip steeply. Fossils are present, but not abundant, in parts of the unit. Cretaceous rocks have been subdivided in the Medfra quadrangle (Patton and others, 1980), and in part of the Ophir and adjacent Iditarod quadrangles (Bundtzen and Laird, 1980), but further field study is needed to relate these units to the Cretaceous rocks throughout the Ophir map area.

Ultramafic rocks, chiefly partly serpentinized peridotite, dunite, and pyroxenite that typically are weathered dark yellowish orange to yellowish brown, form a prominent, barren, northeast-striking ridge topped by Mount Hurst. The age and contact relation of these rocks are uncertain. A small body of serpentinite, probably unrelated to the unit at Mount Hurst, lies near upper Colorado Creek on the fault(?) contact between the chert, tuff, argillite, and basalt unit and the Cretaceous sandstone and shale.

Small bodies of medium- to coarse-grained gabbro intrude, or are tectonically emplaced in, rocks of Mississippian to Jurassic age in the area between Tolstoi Creek and Innoko River. Contact relations are uncertain, and the gabbros are assigned a provisional age in the range of Mississippian to Tertiary.

The other igneous rocks are of Cretaceous and (or) Tertiary age on the basis of field relations and their affinities with similar rocks in adjacent areas. They are divided into three units and described in field petrologic terms. The volcanic and plutonic complex at Cloudy Mountain is largely basaltic to trachyandesitic flows, tuffs, and some equivalent coarser grained hypabyssal rocks. Latest Cretaceous and earliest Tertiary (60-70 m.y.) radiometric dates were obtained from part of this complex in the Medfra quadrangle (Patton and others, 1980). The intermediate and felsic plutonic rocks unit includes fine-grained monzonite, quartz monzonite, and granite that intrude and hornsfelsically alter both Cretaceous and older rocks in the map area. Relative ages of the monzonitic body of Twin Mountain and the surrounding volcanic and plutonic complex are uncertain, but complexes of similar rocks at Cloudy Mountain, Page Mountain, and Alone benchmark in the Medfra quadrangle yield latest Cretaceous and earliest Tertiary potassium-argon ages (Patton and others, 1980). Two small plutons of medium-grained granite, surrounded by hornfels aureoles, intrude the Cretaceous sandstone and shale unit near the head of Tolstoi Creek. A small gabbro body associated with the pluton east of Tolstoi Creek has been thermally altered by the granite. The gabbro probably is a part of this plutonic event and is not necessarily correlative with the gabbros of the Mount Hurst area. The unit of volcanic rocks includes rhyolitic, dacitic, trachyandesitic, and basaltic flows, rhyolite tuffs and domes, and numerous rhyolite-dacite dikes, too small to show in figure 40, that intrude the Cretaceous sandstone and shale. This unit is assigned a latest Cretaceous and (or) earliest Tertiary age, on the basis of radiometric dates from similar volcanic rocks in the Ruby and Medfra quadrangles (Silberman and others, 1979).

The Quaternary unconsolidated deposits include undifferentiated silt, sand, and gravel mainly of alluvial and colluvial origin. In contrast to large areas to the northeast in the Ruby and Kantishna River quadrangles, eolian dunes and sheetlike deposits in the lowlands and thick loess on the hills are inconspicuous or absent in this map area. Placer-mine tailings, dating from 1906 and later, occupy many stream valleys in the vicinity of upper Innoko River, Madison and Tolstoi Creeks, and Colorado and Folger Creeks. In 1980 there were six small-scale placer gold mining operations within the map area and at least three others nearby in the Iditarod and McGrath quadrangles.

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Preliminary report on the Late Cretaceous and early Tertiary volcanic and related plutonic rocks in western Alaska

By Elizabeth J. Moll and William W. Patton, Jr.

Reconnaissance mapping and K-Ar dating have documented a major episode of subaerial calc-alkalic volcanism and related plutonism of Late Cretaceous and early Tertiary age in a vast region of western Alaska stretching from St. Matthew and St. Lawrence Islands on the Bering Sea shelf eastward to the front of the Alaska Range and northward to the Arctic Circle (fig. 41).

The volcanic and plutonic rocks are poorly studied over most of this region, but recent studies in the Medfra quadrangle (Patton and others, 1980; Silberman and others, 1979) and in the Goodnews-Hagemeister Island quadrangles (Hoare and Coonrad, 1978; Wilson, 1977) have provided a total of 41 potassium-argon dates on volcanic and plutonic rocks ranging in age from 59 to 73 m.y. In addition, 23 potassium-argon dates between 59 and 79 m.y. are available from elsewhere in western Alaska (fig. 41) and from St. Lawrence and St. Matthew Islands.

This paper describes the occurrence of these Late Cretaceous and early Tertiary volcanic and plutonic assemblages in the Unalakleet, Ophir, and Medfra quadrangles where they have been studied in greatest detail (fig. 41). For descriptive purposes, the rocks are divided into three groups: volcanic piles, volcano-plutonic complexes, and intrusive rocks.

Volcanic piles

The volcanic piles crop out in gently folded, faulted-bounded, northeast-trending synclines and range in composition from basalt to rhyolite. In the eastern part of the Unalakleet quadrangle more than 1,000 m of volcanic rocks is preserved in a broad, fault-bounded syncline. The lower part of this sequence is composed of dark columnar-jointed mafic and intermediate flows and the upper part of interbedded mafic and intermediate flows and felsic flows and domes. To the east in the Nowitna River drainage, a gently folded syncline contains at least 1,500 m chiefly of dark trachyandesite flows overlain locally by small rhyolite domes. Farther to the east in the Sischu Mountains, more than 500 m of dacite and rhyolite flows and domes occurs in a broad, down-faulted syncline.

Volcano-plutonic complexes

The volcano-plutonic complexes are circular, down-faulted stacks of mafic and intermediate flows and shallow hypabyssal rocks intruded by small granitic cores. Although the volcanic rocks have been altered by the granitic cores, both rock types give similar potassium-argon ages and have similar major- and trace-element contents. The composition of an average flow is trachyandesitic, and that of the average granitic core is the plutonic equivalent of a trachyandesite—that is, monzonite or quartz mon-

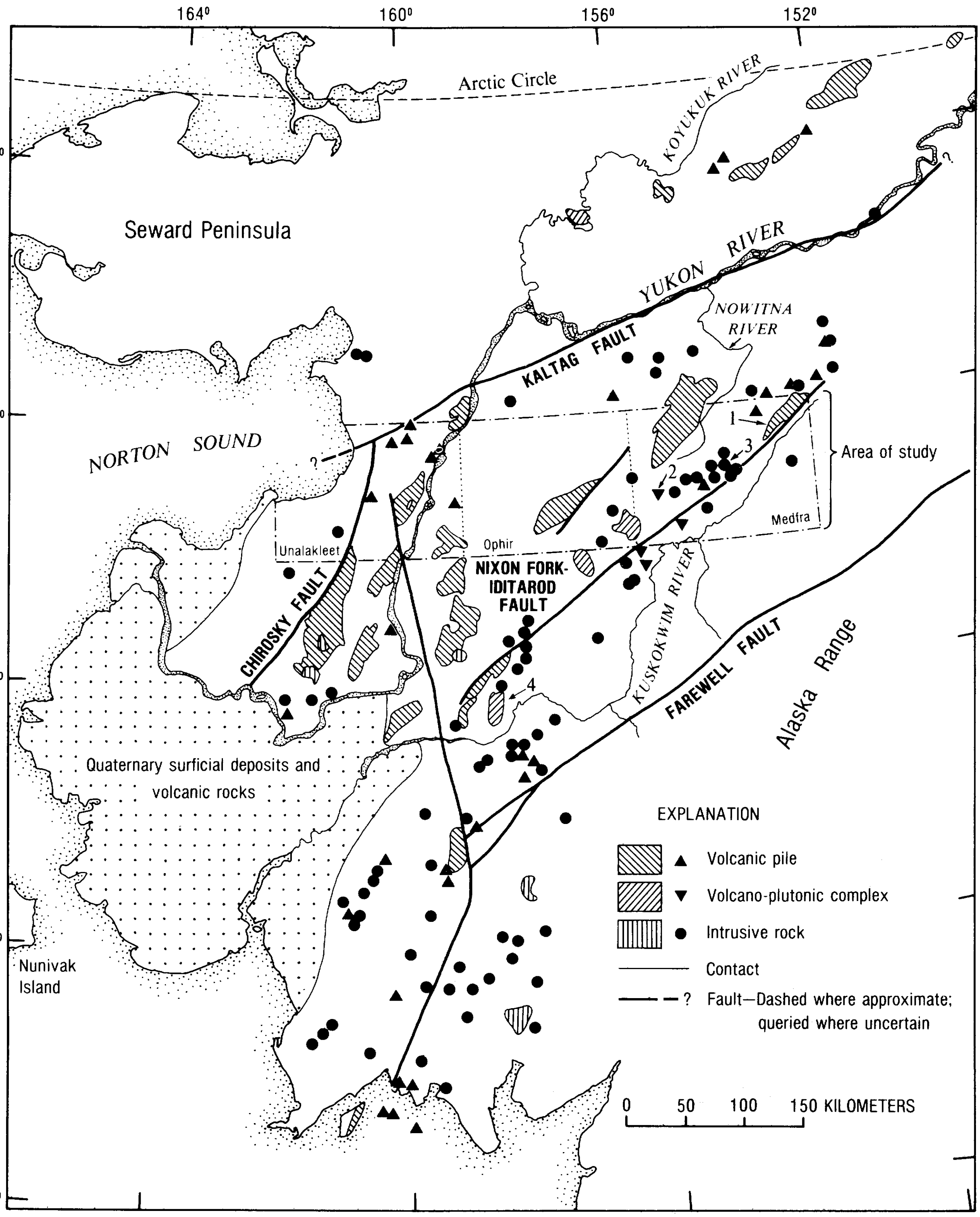

FIGURE 41.—Map showing the distribution of Late Cretaceous and early Tertiary (75-55 m.y.) volcanic and plutonic rocks in western Alaska, excluding the Seward Peninsula and the Alaska Range. Key to localities mentioned in text: 1, Sischu Mountains; 2, Page Mountain; 3, Von Frank Mountain; and 4, Horn Mountains.

zonite. Seven of these complexes have been described by Bundtzen and Laird (1980), Patton and others (1980), and Cady and others (1955) in a northeast-trending belt that extends over 250 km southwest from Page Mountain in the Medfra quadrangle to the Horn Mountains in the Sleetmute quadrangle. Additional complexes may occur along the poorly known segment of this belt that extends through the southern and central Iditarod quadrangle.

Intrusive rocks

Intrusive rocks in the area generally occur as small (1- to 9-km-diameter) stocks or dikes and sills. Their compositions range from gabbro and monzogabbro to granite, but most of the rocks are monzonite, quartz monzonite, or granite (Streikeisen, 1973). Most of the stocks appear to be compositionally homogeneous, but some are strongly zoned. One such body, at Von Frank Mountain, is composed chiefly of monzonite but has patches of monzogabbro along the rim and a core of quartz monzonite. The dikes and sills are generally composed of trachyandesite, dacite, or rhyolite. Large numbers of these dikes and sills are found throughout the map area, but most are too small to be shown on the map (fig. 41).

Chemistry

Our chemical data, which come entirely from the Medfra quadrangle, show that the volcanic and plutonic rocks there (fig. 41) are calc-alkalic and characterized by high K_2O contents. K_2O is about 2-4 percent at 58 percent SiO_2 and about 4-6 percent at 72-75 percent SiO_2. Most analyzed samples have between 53 and 75 percent SiO_2 and therefore are classified as basaltic andesite, trachyandesite, dacite, and rhyolite.

Mineralogy

Basalt and trachyandesite in both the volcanic piles and complexes generally have 1-20 percent phenocrysts of chiefly plagioclase and clinopyroxene, subordinate opaque minerals and orthopyroxene, and rare olivine. The groundmass texture is either trachytic or intersertal/intergranular. Trachytic groundmasses have strongly alined plagioclase microlites, granular pyroxene grains, scattered magnetite, and interstitial K-feldspar. Intersertal and intergranular rocks are similar to the trachytic rocks but generally have more phenocrysts, less strongly alined plagioclase microlites, and dark glass instead of K-feldspar in the intersertal rocks.

The dacite and rhyolite domes, flows, and hypabyssal rocks are generally more altered than the more mafic flows. Unaltered samples have less than 15 percent phenocrysts of embayed quartz, plagioclase, sanidine, biotite, opaque minerals, and rare hornblende in a groundmass of fine-grained interlocking quartz, K-feldspar, and opaque minerals.

Monzogabbro contains olivine, orthopyroxene, clinopyroxene, plagioclase, and more than 10 percent K-feldspar. Gabbro is similar to monzogabbro but has less K-feldspar. Monzonite, quartz monzonite, quartz monzodiorite, and monzodiorite have 25-40 percent mafic minerals (biotite, hornblende, orthopyroxene, and clinopyroxene) and 60-75 percent plagioclase, K-feldspar, and quartz. Granite has 85-95 percent K-feldspar, quartz, and plagioclase, and 5-15 percent mafic minerals of chiefly biotite and subordinate opaque minerals and hornblende.

Tectonic significance

The volcanic rocks are found in five separate northeast-trending fault-bounded geologic terranes that lie between the north front of the Alaska Range and the Koyukuk basin on Norton Sound (see Patton and Moll, 1982). The presence of the volcanic rocks in all of these diverse geologic terranes and the fact that locally they overlap the faulted boundaries establishes conclusively that no major structural dislocation has occurred between the terranes since Late Cretaceous time.

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Structural and stratigraphic sections along a transect between the Alaska Range and Norton Sound

By William W. Patton, Jr., and Elizabeth J. Moll

Regional geologic mapping and stratigraphic studies have been completed along a transect that extends for 400 km from the north front of the Alaska Range to the Koyukuk basin bordering Norton Sound (fig. 42). This report briefly summarizes the broad structural and stratigraphic features of this heretofore little known region and provides some tectonic interpretations for its diverse geologic terranes. Within this transect, the bedrock is subdivided into six northwest-trending fault-bounded belts, each of which differs in some important respects from the adjoining belts (fig. 43). Although major structural dislocations have occurred along their boundaries, we do not regard these belts as separate tectonostratigraphic terranes (Beck and others, 1980) or "microplates" because each belt shares some stratigraphic or structural features in common with the belts that adjoin it. A pre-Late Cretaceous tectonostratigraphic boundary may occur, however, along the Susulatna fault (fig. 44) where continental Paleozoic and Mesozoic carbonate and terrigenous rocks of the northern Kuskokwim Mountains are juxtaposed with oceanic-island arc upper Paleozoic and Mesozoic radiolarian chert and mafic volcanic rocks of the Innoko River area (Patton, 1978).

The major stratigraphic and structural features of each of these belts are shown in figures 43 and 44 and briefly described below:

Northern Kuskokwim Mountains belt—The stratigraphic succession in this belt comprises more than 5,500 m of highly folded and faulted Ordovician to Devonian platform carbonate rocks which rest unconformably upon metasedimentary and metavolcanic rocks of Precambrian age and are overlain by about 4,000 m of moderately deformed late Paleozoic and Mesozoic terrigenous sedimentary rocks (Patton and others, 1980). The terrigenous deposits were derived in large part from the underlying carbonate and metamorphic rocks and, in contrast to the terrigenous deposits of the Innoko River belt, contain little volcanic debris.

Minchumina lowlands belt—Sparse bedrock exposures in this belt consist chiefly of slightly schistose Ordovician to Devonian shaly limestone and chert which are interpreted to be a deep-water offshore facies of the lower Paleozoic platform carbonate rocks in the northern Kuskokwim Mountains belt. At the top of the section, the upward gradation of the deep-water deposits into shallow-water limestone and dolomite suggests a seaward progradation of the carbonate platform in Devonian time. In the Telida Mountains, the deep-water beds are underlain by an assemblage of quartzite, quartz grit, and quartz-feldspar grit which tentatively is assigned a Precambrian or earliest Paleozoic age on the basis of correlation with similar assemblages to the northeast in the Kantishna River and Livengood quadrangles (Patton and others, 1980). The source area and depositional

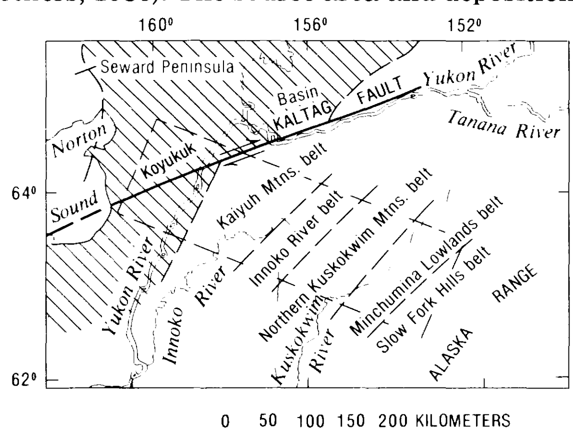

FIGURE 42.—Index map of west-central Alaska showing location of transect and geologic belts. The lined pattern represents rocks of Cretaceous age in the Koyukuk Basin.

environment of these coarse clastic rocks are uncertain.

Slow Fork Hills belt—The bedrock in this belt is composed of sheared grit, quartzite, and quartz-mica schist which are interpreted to be a metamorphic facies of the quartzite and grit beds of the Telida Mountains in the adjoining Minchumina lowlands belt (Patton and others, 1980).

Innoko River belt—In this belt an oceanic-island arc assemblage of Mississippian and Pennsylvanian radiolarian chert and Triassic and Jurassic tuff, basalt, and diabase forms a tectonic package that appears to have been thrust southeastward onto Precambrian and early Paleozoic continental metasedimentary rocks. The oceanic-island arc strata are presumed to have originated in the "Yukon-Koyukuk sea"—a marginal ocean basin that occupied the present site of the Koyukuk basin during late Paleozoic and early Mesozoic time (Patton, 1979).

Kaiyuh Mountains belt—In this belt continental metasedimentary rocks of Precambrian and early Paleozoic age are overlain by two separate thrust sheets of mafic-ultramafic rocks (ophiolites) that appear to have their roots in the Koyukuk basin. The lower thrust sheet is

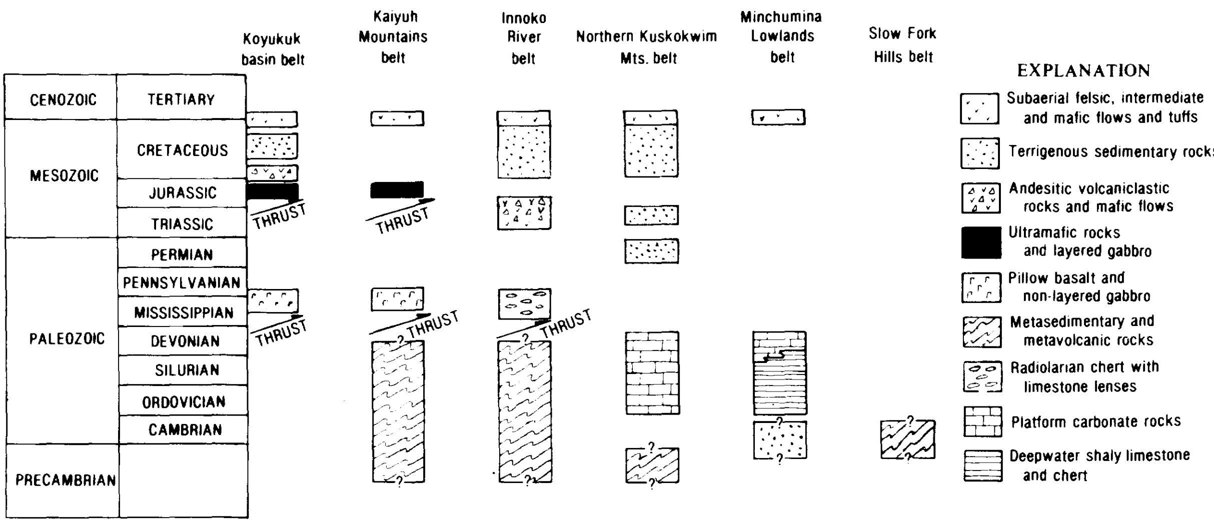

FIGURE 43.—Generalized stratigraphic sections of geologic belts along transect shown in figure 42.

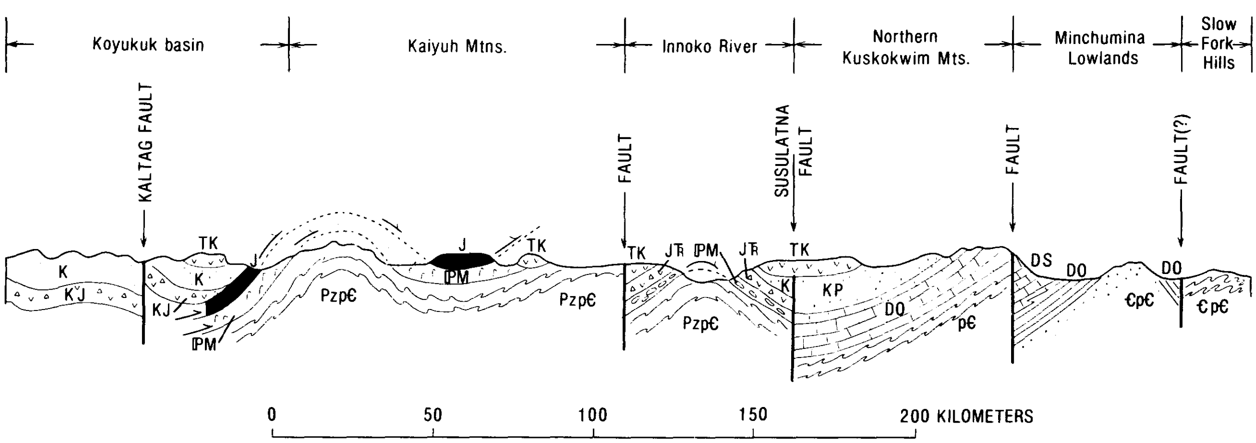

FIGURE 44.—Diagrammatic cross sections along transect shown in figure 42. See figure 43 for explanation of lithology. Symbols for ages of rocks: €p€, Precambrian or Cambrian; DO, Ordovician to Devonian; DS, Silurian and Devonian; J, Jurassic; JTR, Triassic and Jurassic; K, Cretaceous; KJ, Jurassic and Cretaceous; KP, Permian to Cretaceous; p€, Precambrian; PM, Mississippian and Pennsylvanian; Pzp€, Precambrian and Paleozoic; TK, Cretaceous and Tertiary.

composed of pillow basalt, diabase, nonlayered gabbro, and chert, and the upper sheet of ultramafic rocks and layered gabbro. "Blueschist" mineral assemblages are found in the base of the lower thrust sheet as well as in the underlying metasedimentary rocks. The Mississippian and Pennsylvanian age assignment of the lower thrust sheet is based upon fossils; the Jurassic age assignment of the upper sheet is based upon potassium and argon determinations from similar ultramafic-mafic assemblages on strike to the northeast of the Kaiyuh Mountains. Field evidence suggests that the Jurassic age represents an emplacement age rather than the age that these rocks were formed in the upper mantle and lower crust (Patton and others, 1977).

Koyukuk basin belt—This belt comprises a broad depression filled with as much as 8,000 m of highly folded middle and Upper Cretaceous terrigenous sedimentary rocks, chiefly marine volcanic graywacke in the lower part and shallow-water marine and nonmarine paralic deposits in the upper part (Patton, 1973). In the central part of the basin, the terrigenous strata are underlain by Lower Cretaceous marine andesitic volcanic rocks and at the margin of the basin, they overlap onto the late Paleozoic and Mesozoic ophiolite assemblages of the Kaiyuh Mountains. Geologic and geophysical mapping elsewhere around the margin of the Koyukuk basin suggests that the ophiolites dip inward beneath the basin (Patton and others, 1977).

Summary

The three southeastern belts are composed of early Paleozoic continental margin deposits that rest unconformably on a Precambrian metamorphic basement and are overlain by late Paleozoic and Mesozoic successor basin deposits. This nonvolcanic assemblage is juxtaposed on the northwest by late Paleozoic and Mesozoic island-arc and ophiolite assemblages that appear to have been thrust southeastward across a Precambrian(?) and early Paleozoic metamorphic complex from a root zone at the margin of the Koyukuk basin. A tectonostratigraphic or "microplate" boundary may separate these two contrasting assemblages along the Susulatna fault, but the presence of similar metamorphic basement rocks on both sides of the fault casts some doubt on this interpretation. Subaerial volcanic rocks of Late Cretaceous and early Tertiary age (Moll and Patton, 1982) that locally overlap the boundaries of the belts indicate that major structural dislocations between the belts were completed by Late Cretaceous time.

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Quaternary history of vegetation and climate of the Yukon Delta-Norton Sound area

By Thomas A. Ager and Platt Bradbury

Pollen analysis of lake-sediment cores and peat cores from northern Yukon Delta and St. Michael Island in southern Norton Sound is being used to reconstruct a detailed history of vegetation and climate of this region during the past approximately 42,000 years. Lacustrine-sediment cores as much as 15 m long were obtained in 1978 and 1979 from ice-covered lakes within maar craters on the volcanic island of St. Michael (figs. 45 and 46). A radiocarbon date from the base of the longest core from

Zagoskin Lake is greater than 39,000 years. Pollen analysis of this 15-m core is in progress, but preliminary interpretations suggest that the lower 10 m of the core represents the time interval from about 42,000 to 14,000 years B.P. (Ager, 1982). Vegetation in this unglaciated region during that part of the Wisconsinan glacial interval was herbaceous tundra; the climate during most of that long interval was arid and windy, with cold severe winters and summers that were probably dry and cooler than today (about 10°C July mean temperature). An interval of somewhat milder climate may have occurred about 30,000 years ago (estimated age), probably a middle Wisconsinan interstadial event. The vegetation changes are subtle during this "interstadial," however, with only minor increases in shrub pollen types and greater diversity of herbaceous taxa.

About 14,000 years ago, the vegetation began to change to a mesic shrub tundra (Ager, 1980). This change is interpreted to reflect a vegetational response to a climatic change to warmer, moister conditions in summer, and perhaps greater snowfall in winter. The transition from herbaceous tundra to mesic shrub tundra probably occurred over a period of at least 2,000 years in this part of Alaska. During this time sea level began to rise, partly flooding the Bering land bridge (Hopkins, 1967, 1973).

The next major vegetation change in this region occurred about 7,000 years ago when alder shrubs quickly spread through the landscape, occupying gullies, hollows, and other habitats that afford protection from winter winds. Some pollen evidence from peat cores and lake-sediment cores (for example, Puyuk Lake, fig. 45) suggests that alder shrubs were present in the Yukon Delta-Norton Sound area perhaps as early as 12,000 years ago. The sudden rise in percentages of alder pollen about 7,000 years ago suggests, therefore, that an important climatic threshold was crossed at that time that permitted the alder population to expand rapidly. Increased summer warmth is a likely factor in this event, as this vegetation change seems to coincide with the Hypsithermal climatic interval.

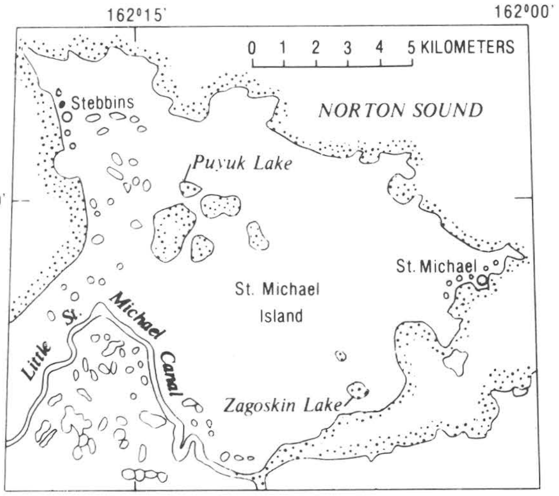

FIGURE 45.—Map showing Puyuk and Zagoskin Lakes on volcanic St. Michael Island.

Spruce trees began to invade the lower Yukon River east of Norton Sound about 5,500 years ago. Wind-transported spruce pollen appears in lake sediment cores at about that time on St. Michael Island and other sites in the Yukon Delta; all the sites studied in the region thus far are located in tundra settings at least 50 km beyond the present limit of spruce trees.

Diatom stratigraphy of the 15-m core from St. Michael Island is being studied by J. P. Bradbury. Preliminary zonation of diatom assemblages closely parallels the pollen zonation. Full-glacial diatom floras are characterized by low species diversity; many of the "glacial" sample assemblages include a new species of *Cyclotella*. Post-14,000-year-old diatom assemblages are characterized by much greater species diversity and a relative abundance of temperate diatom types.

Three volcanic ash layers have been recognized in the 15-m-core from St. Michael Island. Ash I is about 16,000 years old and is probably of local origin (Stuart Island?). Ash II is about 14,500 years old and also appears to be of local origin. Ash III is about 4,000 to 5,000 years old but is probably from a distant source (Alaska Peninsula?); it is an important marker horizon in western Alaska.

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FIGURE 46.—Winter coring operation on ice-covered lake on St. Michael Island.

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Hydrologic reconnaissance of the Kobuk River basin

By Joseph M. Childers and Donald R. Kernodle

Hydrologic surveys were conducted during August 1979 and during March and April 1980 to furnish data for the Kobuk River basin. This important drainage, containing about 31,000 km², is a corridor for increasing subsistence and non-subsistence uses such as fishing, hunting, and use of cabin sites and other forms of recreation, by both residents and visitors.

Water-resources data collected during these trips at various sites on the Kobuk River and selected tributaries included flood surveys, discharge measurements, and water-quality assessments. Benthic invertebrates were also collected.

River travel during August was accomplished using rafts, and sites were visited by raft or helicopter. The rafts and all supplies and equipment were flown into Walker Lake, near the headwaters of Kobuk River. From this departure point the field party traveled 460 km downstream to Kiana and visited 30 sites. River discharge was 31 m^3/s at Kobuk River just above Walker Lake outlet and increased to 1,400 m^3/s below Jade Creek (fig. 47). Measured unit runoff (discharge divided by drainage area) in August 1979 increased from about 0.03 $(m^3/s)/km^2$ in the upper basin to more than 0.08 $(m^3/s)/km^2$ in the middle basin. Some of that increase in discharge and unit runoff was caused by abnormally high rainfall.

Winter transportation during the March-April 1980 trip was provided by snow machines with sleds. Starting from the village of Kotzebue, the survey party traveled upriver as far as Kobuk village where unseasonably high temperatures brought about poor snow and ice conditions. These conditions prohibited any further work, and only 10 sites were visited. Measured discharge of the Kobuk River increased from 35 m^3/s above Kobuk to 53 m^3/s at a site below Kiana. Late winter low-flow unit runoff of about 0.002-.003 $(m^3/s)/km^2$ is probably available from most of the Kobuk River basin, but values approaching no flow may be expected in some basins such as the Squirrel River near Kiana where runoff was only 0.0002 $(m^3/s)/km^2$ in March 1980.

Evidence of ice jams found at the tops of

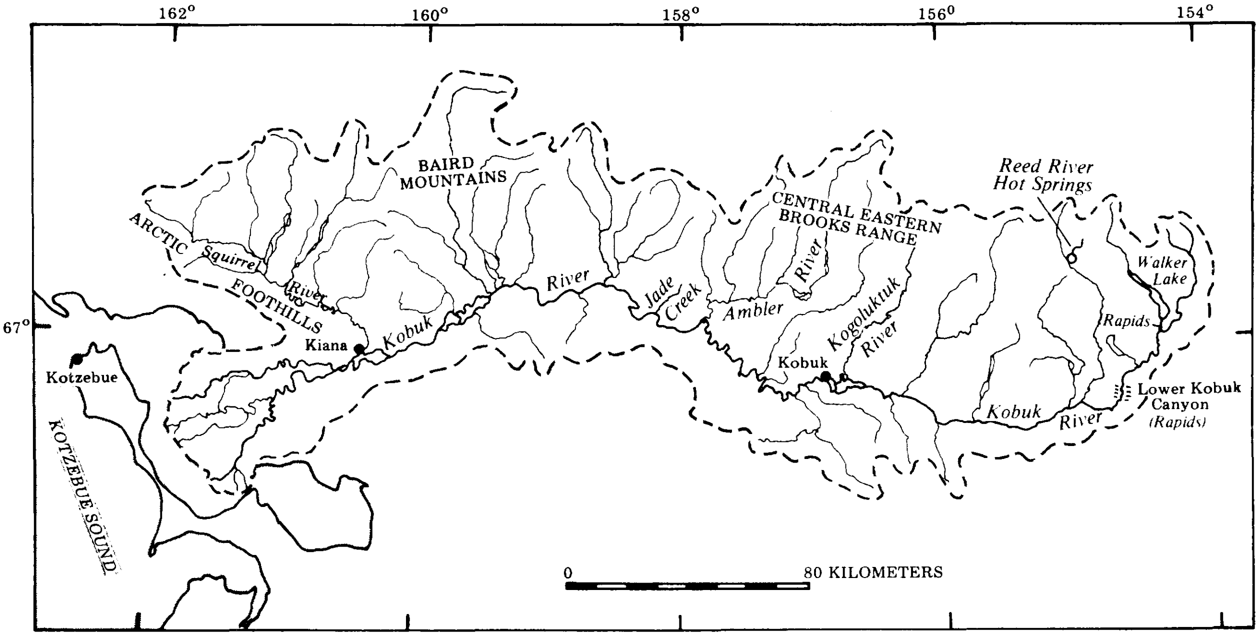

FIGURE 47.—Kobuk River basin, Alaska (dashed envelope).

banks along the Kobuk River at most sites surveyed indicates that ice-jam flooding is a hazard along the river. Maximum evident flood (MEF) high-water marks were also found along the tops of the banks at most sites and were used to compute peak discharges. MEF unit runoff decreased from 0.86 $(m^3/s)/km^2$ above Walker Lake to 0.19 $(m^3/s)/km^2$ at Kobuk.

The Kobuk River from the Walker Lake outlet to its mouth is generally a wide, smooth-flowing stream, and it offers good boating conditions. However, rapids below the Walker Lake outlet and in Lower Kobuk Canyon are navigation hazards.

Water-quality information gathered at the sites along the Kobuk and its tributaries during both trips included specific conductance values of less than 350 μmhos/cm^2 at 25°C for both summer and winter trips. One thermal spring, Reed River Hot Springs, had a conductance of 750 μmhos/cm^2 and a water temperature of 47.5°C. Elsewhere, river water temperatures were less than 18°C during the summer trip and only slightly above freezing during the winter.

SOUTHWESTERN ALASKA

(Figure 48 shows study areas discussed)

Sedimentology of the Cretaceous Kuskokwim Group, southwest Alaska

By John E. Decker and Joseph M. Hoare

Sedimentary features within the type locality of the Cretaceous Kuskokwim Group were studied during a paleomagnetic sample collecting traverse across the central Kuskokwim Mountains. The outcrops visited occur along cut-banks of the Kuskokwim River between the villages of Sleetmute and Aniak (fig. 49). The Kuskokwim Group in this region is composed of moderately well indurated, thin- to thick-bedded graywacke and shale that has been deformed into broad open folds. Our preliminary data indicate that these rocks were deposited as fine-grained sand and silt turbidites in nested depositional lobes within a regionally subsiding continental trough.

The graywacke beds range in thickness from less than 1 cm to about 2 m and are composed dominantly of medium- to very fine-grained sandstone. Beds are commonly graded with 2- to

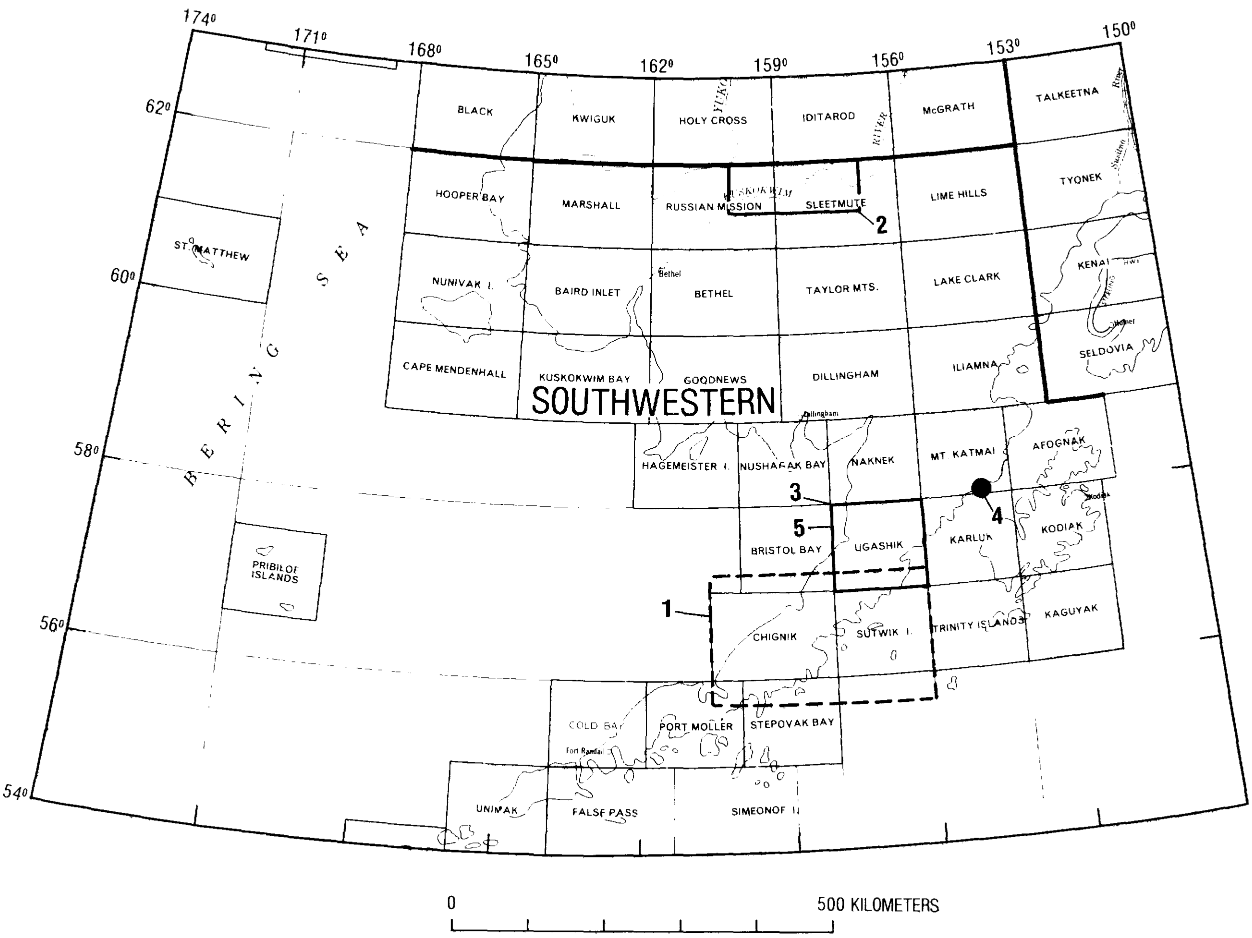

FIGURE 48.—Areas and localities in southwestern Alaska discussed in this volume. Authorship and applicable figures relating to the numbered areas are listed below. 1, Case, Cox, Detra, Detterman, and Wilson, figure 54; 2, Decker and Hoare; figure 49; 3, Detra, Detterman, Cox, Wilson, and Theodore, figure 53; 4, Miller, Detterman, and Case, figures 50, 51, and 52;5, Wilson and Shew, figure 53.

10-cm-thick beds displaying complete (Ta-e) or nearly complete (Tb-e) Bouma sequences (facies C of Mutti and Ricci Lucchi, 1972). Thinner beds more typically lack the lower (Ta or Ta,b) Bouma intervals (facies D of Mutti and Ricci

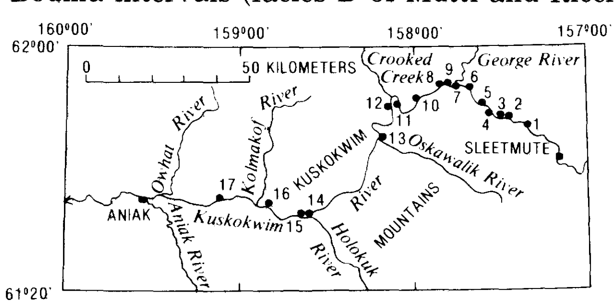

FIGURE 49.—Map showing outcrop localities where sedimentary features of Kuskokwim Group were studied on river traverse across the Kuskokwim Mountains between Sleetmute and Aniak. Numbered outcrop localities discussed in text.

Lucchi, 1972), and thicker beds generally lack the upper Bouma intervals and consist of amalgamated sequences of successively repeated Ta or Ta,b intervals (facies B of Mutti and Ricci Lucchi, 1972). Although the sequence generally is even bedded, thick beds locally have channeled bases with rare coarse-grained sand or granule lag deposits and shale rip-ups. Groove casts are common on the bases of graywacke beds, but flute casts are rare. Carbonized plant material is locally abundant along bedding surfaces. Stratal disruptions and soft-sediment folds related to submarine slumping and landsliding (facies F of Mutti and Ricci Lucchi, 1972) are common at outcrop 17 but were observed only rarely at other localities.

The shale beds generally are from 1 to 10 cm thick but locally are as thick as 30 cm. The shale typically is a medium- to dark-gray siltstone or silty claystone; black claystone was observed at only one locality.

Sandstone:shale ratios are quite variable, averaging from 10:1 to 20:1 in sandy areas and about 1:1 or 2:1 in shaly areas. We estimate an overall average of about 5:1, but this ratio may be too high because large covered areas and valleys are most likely underlain by the sections richer in shale. Beds commonly are arranged in systematic megacycles. Four outcrops studied (4, 6, 13, and 14) have graywacke beds thickening and coarsening upward, each cycle ranging from 3 to 10 m in thickness and generally repeated several times. Outcrop 12 has a symmetrical distribution of beds, and outcrop 15 shows a nonrepeating thinning- and fining-upward megacycle. The other outcrops show no discernible megacycles; however, rocks at several of these localities are poorly exposed and critical areas are covered.

The repeated association of facies B, C, and D, dominantly thickening- and coarsening-upward megacycles, even bedding, absence of channel deposits, and scarcity of redeposited strata indicate that these rocks were deposited as overlapping depositional lobes, either in an outer-fan environment (Mutti and Ricci Lucchi, 1972) or as prodelta aprons at the base of a regionally subsiding basin. Although we cannot distinguish between outer-fan and prodelta environments on the basis of our data, we favor the prodelta model for two reasons:

1) Our traverse of more than 100 km across the Kuskokwin Group represents more than 12 km of strata (Cady and others, 1955) that we believe to have been deposited in the same depositional environment. Age-equivalent interbedded graywacke and shale sections that have been described to the southwest by Hoare (1961) and to the northeast by Patton and others (1980) and Bundtzen and Laird (1980) suggest a regionally similar depositional setting and an absence of other deep-sea fan components (for example, inner-fan and middle-fan facies associations). Such a thick regionally similar depositional system implies that ongoing basin subsidence kept the basin active and allowed the same facies to be deposited repeatedly throughout the depositional cycle. Filling of a preexisting basin, on the other hand, would produce a whole suite of facies associations progressing from deep marine through nonmarine.

2) Conglomerate occurs throughout the section, particularly along the margins of the basin in the vicinity of older bedrock where clast compositions reflect the local source (Cady and others, 1955; Hoare, 1961). The conglomerate beds interfinger laterally and are sharply overlain by interbedded graywacke and shale (Hoare, 1961). Assuming that the conglomerates are the most proximal deposits and that the intertonguing graywacke and shale are similar to our depositional lobes, an outer-fan model would be difficult to accept because middle-fan and inner-fan facies are absent.

We speculate that major rivers carried dominantly fine-grained suspended material into a regionally subsiding continental trough whose uplifted margins shed conglomeratic detritus locally into the basin. Subsidence rates were high enough to allow the fine-grained material to be deposited as depositional lobes that covered the basin floor throughout its depositional cycle.

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Lower Cretacoeous (Albian) rocks on the Alaska Peninsula

By John W. Miller, Robert L. Detterman, and James E. Case

Lower Cretaceous (Albian) rocks are here well documented for the first time on the Alaska

Peninsula. The strata are exposed in sea cliffs on the east side of Katmai Bay (fig. 50), in the Mount Katmai A-3 quadrangle. Lower Cretaceous (Neocomian) rocks have been known from the Mount Katmai quadrangle for several years (Parkinson, 1960; Jones and Detterman, 1966; Magoon and others, 1976). Albian rocks were reported in the Chitina Valley many years ago (Moffit and Capps, 1911; Moffit, 1918, 1938) and in the Matanuska Valley more recently (Jones, 1967; Grantz *in* Jones, 1967).

Albian fossils were first collected at Katmai Bay in 1979 by R. M. Egbert, T. N. Smith, and G. W. Petering, who were measuring a stratigraphic section. The full significance of the collection was not apparent until the fossils were identified. In 1980 we visited the site as part of the Alaska Mineral Resource Assessment Program (AMRAP) in the Ugashik and Karluk quadrangles, to collect additional material and to determine the thickness and stratigraphic relations of the Albian rocks. Only the sea-cliff exposure was investigated at this time; additional field studies will probably reveal the presence of Albian strata in other parts of the quadrangle.

The unnamed Albian sequence is about 82 m thick and consists mainly of thick-bedded fine- to medium-grained gray to green sandstone with abundant carbonaceous debris and a few calcareous concretions (see schematic stratigraphic section—fig. 51). The middle part contains a few pebbles as large as 4 cm and is locally crossbedded. Two partially covered units of siltstone and shale are present in the upper half of the section. At the faulted lower contact, the beds rest on the Naknek Formation (Upper

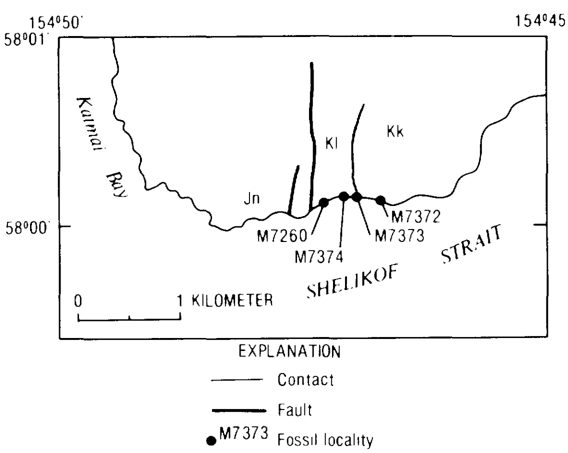

FIGURE 50.—Sketch map of east side of Katmai Bay showing three localities where fossils were collected from unnamed rocks of Early Cretaceous (Albian) age (Kl) and one locality (M7372) in the Kaguyak Formation (Kk), of Late Cretaceous age. A fault separates the Cretaceous rocks from the Naknek Formation (Jn), of Late Jurassic age.

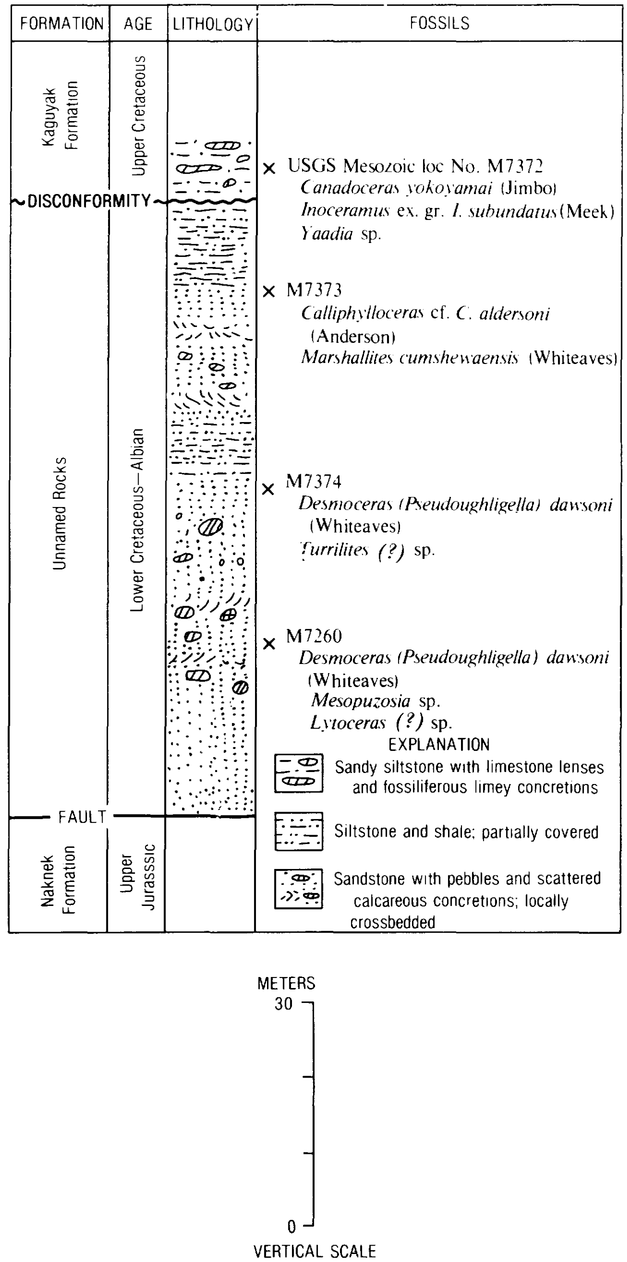

FIGURE 51.—Schematic generalized stratigraphic section of Lower Cretaceous (Albian) rocks east side of Katmai Bay.

Jurassic). The upper contact appears to be conformable with the overlying Kaguyak Formation (Upper Cretaceous) but is more likely a disconformity, because a hiatus of considerable magnitude is represented by the contact between the two units.

Three collections of fossils have been made from the Albian strata. The lowest (M7260), about 21 m above the lower contact (figs. 50 and 51), contained *Desmoceras* (*Pseudoughligella*) *dawsoni* (Whiteaves) (fig. 52a, b), *Mesopuzosia* sp., and *Lytoceras*(?) sp. The middle collection (M7374), about 46 m above the lower contact (figs. 50 and 51), contained *Desmoceras* (*Pseudoughligella*) *dawsoni* (Whiteaves) and *Turrilites*(?) sp. (fig. 52d). The highest collection (M7373) contained *Calliphylloceras* cf. *C. aldersoni* (Anderson) (fig. 52e, f) and *Marshallites cumshewaensis* (Whiteaves) (fig. 52c), collected about 70 m above the lower contact (figs. 50 and 51).

Many of these species of ammonites are characteristic of strata of the *Desmoceras* (*Pseudoughligella*) *dawsoni* zone of probable late Albian age in the upper Chitina Valley (Jones, 1967, p. 4). A similar fauna with the upper Albian ammonite *Mortoniceras* from the sandstone member of the Haida Formation in the Queen Charlotte Islands, British Columbia, is described by McLearn (1972, p. 5). *Desmoceras* (*Pseudoughligella*) *dawsoni* (Whiteaves) also occurs in upper Albian beds containing *Mortoniceras* in California.

The recognition of this previously unknown stratigraphic unit in the Katmai Bay area is an important contribution regarding the distribu-

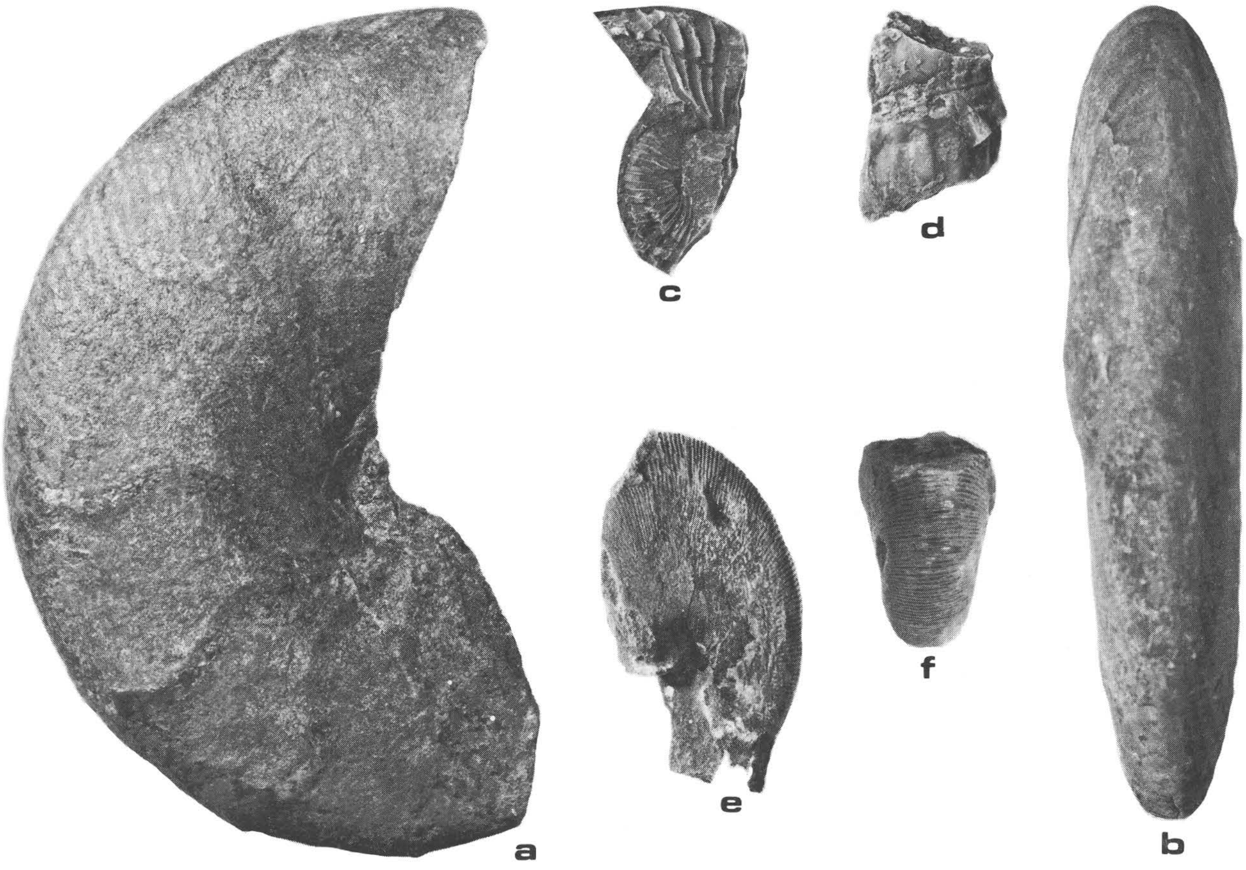

FIGURE 52.—Specimens of *Desmoceras* (*Pseudoughligella*), *Calliphylloceras*, *Marshallites*, and *Turrilites*?. All specimens natural size. Collected by J. E. Case, R. L. Detterman, and J. W. Miller except as noted. a, b, *Desmoceras* (*Pseudoughligella*) *dawsoni* (Whiteaves). Lateral and ventral views. USGS Mesozoic loc. M7260. Collected by T. N. Smith.; c, *Marshallites cumshewaensis* (Whiteaves). Lateral view, rubber cast. USGS Mesozoic loc. M7373.; d, *Turrilites*? sp. Ventral view. USGS Mesozoic loc. M7374.; e, f, *Calliphylloceras* cf. *C. aldersoni* (Anderson). Lateral and ventral views. USGS Mesozoic loc. M7373.

tion and stratigraphy of Albian rocks on the Alaska Peninsula.

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Preliminary results of potassium-argon age determinations from the Ugashik quadrangle, Alaska Peninsula

By Frederic H. Wilson and Nora Shew

Early and preliminary results of potassium-argon dating work on samples from 12 sites in the Ugashik quadrangle indicate a continuation of the geologic trends seen in the Chignik and Sutwik Island quadrangles to the south (Wilson, 1980). Tertiary volcanic and hypabyssal rocks apparently fall into two age groups: early Tertiary—late Eocene to earliest Miocene and late Tertiary and Quaternary—late Miocene to Holocene (fig. 53).

The older group of rocks is found primarily in the southern part of the quadrangle on the northwest side of the Aleutian Range. The rocks are, at least in part, equivalent to the Meshik Formation (Eocene and Oligocene) and include hornblende andesite and dacite, pyroxene andesite and possible basalt. Preliminary ages range from 42 to 21 m.y. on the basis of determinations on whole-rock and hornblende samples. As in the Chignik and Sutwik Island quadrangles, biotite is a rare primary mineral phase.

The late Tertiary rocks are found in general along the Pacific coast in the southern part of the quadrangle and on the northwest side of the Aleutian Range in the northern part of the quadrangle. A major break in the volcanic arc trend in the vicinity of Wide Bay may indicate a transition between two segments of the Aleutian arc.

The volcanic rocks in the northern part of the quadrangle, including Blue Mountain and The Gas Rocks, yield Pleistocene ages, whereas the rocks in the southern part of the quadrangle yield late Miocene to Pleistocene ages. A single 9.25-m.y. age determination on the south edge of the Agripina Bay batholith (Wilson, 1981; sample 77AWs9) may indicate plutonism the

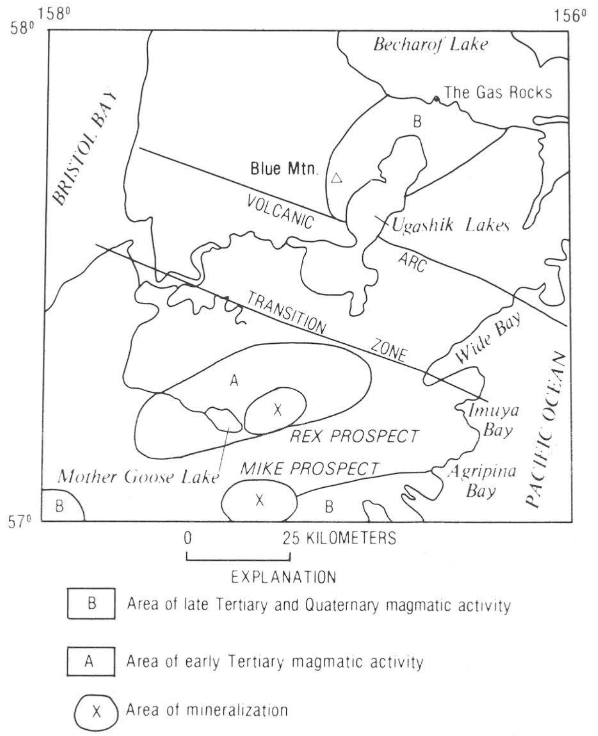

FIGURE 53.—Sketch map of Ugashik quadrangle showing areas of magmatic activity and mineralization.

same age as the Devils batholith (Wilson, 1980; Detterman and others, 1981).

A number of porphyry-type copper prospects are being dated in the Ugashik quadrangle. Age determinations by M. L. Silberman (written commun., 1980) and by us have shown that the Rex prospect of Bear Creek Mining Company in the mountains west of Wandering Creek is of probable earliest Oligocene age. Preliminary age determinations at Rex suggest a multiple intrusion history. A few very preliminary age determinations indicate that the Mike prospect of Bear Creek Mining Company, a molybdenum prospect south of Needle Lake, is apparently of Pliocene age.

In the coastal area south of Imuya Bay (the Agripina Bay batholith), a number of potential porphyry-type prospects have been located. These are expected to yield primarily Pliocene ages.

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Reconnaissance geochemical studies in the Bristol Bay, Ugashik, and Karluk quadrangles, Alaska

By David E. Detra, Robert L. Detterman, Dennis P. Cox, Frederic H. Wilson, and Ted G. Theodore

Geochemical reconnaissance studies in the Bristol Bay, Ugashik, and Karluk quadrangles were completed with collection and analyses of more than 1,200 stream-sediment and heavy-mineral-concentrate samples during 1979 and 1980. At least two areas of possible metallic mineral occurrence have been delineated by results of the analyses, geologic fieldwork, and data provided by Bear Creek Mining Company.

One area of interest (fig. 53, area REX) is located on an unnamed mountain to the northeast of Mother Goose Lake, between Volcano and Wondering Creeks and surrounds the Rex porphyry copper prospect identified by Bear Creek Mining Company in 1977. The streams draining this area, particularly to the south, west, and north, show gold (20 to 1000 ppm) and silver (2 to 200 ppm) anomalies in the nonmagnetic heavy-mineral concentrates with associated less significant Pb (50 to 1500 ppm), Cu (10 to 3000 ppm), and W (100 to 150 ppm) anomalies. Most of the exposed rock in the area is a hypabyssal hornblende dacite to andesite that has been intruded into the Tolstoi Formation (Tertiary) as used by Burk (1965). Country rocks are hornfelsic, silicified, and propyllitically altered sandstone and shale. Both the intrusive and sedimentary rocks are cut by dikes of andesitic composition, some of which show considerable pyritization and others of which are quite fresh. The bulk of the alteration and mineralization occurs on the upper, western side of the mountain where the more significant anomalies seem to be associated with the dike swarms.

The second area of interest (fig. 53, area MIKE) is in the region surrounding the Mike prospect at Painter Creek Pass near the south-central margin of the Ugashik quadrangle (shallow drilling at the Mike prospect in 1978 by Bear Creek Mining Company revealed molybdenite mineralization, but the grade indicated by drilling results was considered too low to justify further work). This area shows many zonally distributed clusters of element values significantly above the mean concentrations obtained from the entire sample population of heavy-mineral concentrates from the three quadrangles. Anomalous values for molybdenum (180 ppm; 20-1500 ppm)[6], copper (269 ppm; 150-2000 ppm), zinc (3050 ppm; 1500-20,000 ppm), and tungsten (360 ppm; 100 ppm) occur in samples from drainage basins in the center of the area; this central area is surrounded by drainage basins from which samples anomalous in copper (269 ppm; 150-10,000 ppm), lead (390 ppm; 50-5000 ppm), zinc (3050 ppm; 500-20,000 ppm), arsenic (1610 ppm; 500-3000 ppm), silver (15 ppm; 3-500 ppm), or bismuth (20-50 ppm) were collected. This area is a dike and sill complex of diorite or andesite and hydrothermally altered rhyolite, all of which have

[6]The first number is the mean concentration of element in three-quadrangle area calculated from those heavy-mineral-concentrate samples in which a measurable concentration was detected. The following numbers indicate range of values of the element found in the heavy-mineral concentrates collected from the discussion area.

intruded sandstone and siltstone of the Naknek Formation (Upper Jurassic) as used by Burk (1965). The sedimentary rocks have been strongly hornfelsed, sericitized, and, in places, mineralized by quartz-molybdenite-pyrite stockworks. Individual sills, which are as much as 60 m thick, have been dated at 3-4 m.y. This area of mineralization appears to be along a trend of porphyry-type mineralization defined in the Chignik and Sutwik Island quadrangles in which anomaly clusters show a similar zonal distribution of metals (Cox and others, 1982).

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Aeromagnetic survey of the Chignik and Sutwik Island quadrangles

By James E. Case, Dennis P. Cox, David E. Detra, Robert L. Detterman, and Frederic H. Wilson

The recently completed aeromagnetic survey of the Chignik and Sutwik Island quadrangles, on the Alaska Peninsula, is one of the most detailed systematic surveys ever made across a modern "volcanic arc" setting where both Holocene volcanic edifices and older volcanic centers occur (fig. 54).

Four main groups of magnetic anomalies have been identified in the area by Case and others (1982):

(1) Ovoid clusters of high-amplitude positive and negative anomalies over Holocene volcanic edifices at Mount Veniaminof, Black Peak, and Aniakchak Crater.

(2) Ovoid clusters of positive and negative anomalies over volcanic and plutonic complexes near Kujulik Bay, Chiginagak Bay, Sutwik Island, and Nakchamik Island. Isotopic ages of igneous rocks from these older groups tend to fall into three ranges: 40-30 m.y. for igneous rocks in the Kujulik Bay and Sutwik Island areas, 10-5 m.y. for those in the Chiginagak Bay and Nakchamik Island areas (Wilson and others, 1981) and a few isotopic ages from other igneous complexes are as old as 50-40 m.y. If the igneous complexes are related to subduction in the Aleutian trench, the site of igneous activity has remained nearly unchanged (within ±75 km) with respect to the present trench axis since about 50 m.y. ago.

(3) Large positive ovoid magnetic anomalies occur over Cenozoic intrusive rocks that range in age from 38 m.y. to about 1.0 m.y. These rocks are dioritic to granodioritic in composition and presumably are related to subduction in the Aleutian trench. Several of the plutons have mineralization of porphyry copper-molybdenum type, so that their distinctive aeromagnetic expression is a regional ore guide. Cox and others (1982) have identified 14 zoned clusters, or "centers of mineralization," and most of these centers occur directly on or near the flanks of positive aeromagnetic anomalies. Local areas of extreme hydrothermal alteration are expressed by flat magnetic gradients or by negative aeromagnetic anomalies, and such anomalies may represent a more specific ore guide.

(4) Broad, gentle-gradient anomalies occur over sedimentary basins on both the Bristol Bay side and Pacific side of the Alaska Peninsula. Local high-frequency anomalies related to volcanic intercalations or small intrusions are superimposed on the deep basin anomalies. The magnetic basement is 2 km or deeper, on both the Bristol Bay and Pacific margins of the peninsula, but the basins are interrupted on both sides of the peninsula by volcanic-plutonic complexes that range in age from at least 38 m.y. to 1.0 m.y.

Chignik anticline, which might, at first glance, appear to be a favorable target for exploratory drilling for oil and gas, has a core of Mesozoic sedimentary rocks, but many small, high-amplitude magnetic anomalies occur in the core region. These anomalies may indicate magnetic materials in the Naknek Formation, the presence of many small intrusive bodies, or the presence of Tertiary volcanic rocks forming part of the Meshik Formation below thrust faults. Thus, the Chignik anticline should not be considered a major site for oil and gas exploration until much additional detailed geophysical work has been done.

Prominent, northwest-trending aeromagnetic lineaments and local mapped faults, nearly perpendicular to the regional geologic grain, indicate that the volcanic arc is segmented on a fine scale (segments 5-20 km apart) as well as on the coarse scale (segments 50-150 km apart), defined by offset trends of lines of volcanoes.

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SOUTHERN ALASKA

(Figure 55 shows study areas discussed.)

Seismic studies in southern and southeastern Alaska
By Christopher D. Stephens and John C. Lahr

The seismotectonic framework along the southern margin of Alaska is primarily controlled by north-northwestward-directed convergence of the Pacific plate with the North American plate. Strain that accumulates between the two plates is periodically released in large, shallow earthquakes that may involve as much as 20 m or more of slip on rupture surfaces that can extend for hundreds of kilometers. Since 1958, nearly all of this margin has broken in three large (M_S 7 and larger) earthquakes that occurred in 1958 along the Fairweather fault in southeastern Alaska (M_S 7.9), in 1964 beneath Prince William Sound and extending southwest seaward of Kodiak Island (M_S 8.4), and in 1979 beneath the Chugach and St. Elias Mountains near Icy Bay (M_S 7.1; fig. 56). The area between the rupture zones of the 1964 Prince William Sound and 1979 St. Elias earthquakes, termed the Yakataga seismic gap, has not broken in a large earthquake since at least 1899-1900 and is expected to be the location of

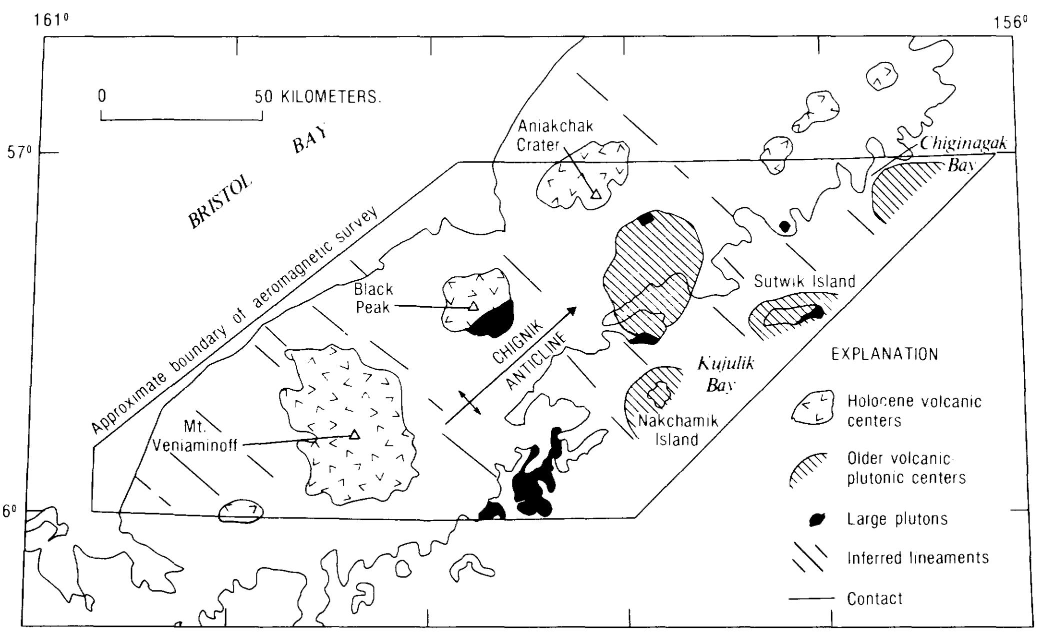


FIGURE 54.—Simplified sketch map of the Alaska Peninsula showing large Holocene volcanic centers, older volcanic-plutonic centers inferred from geologic and aeromagnetic data, large plutons, inferred transverse lineaments, and approximate boundary of aeromagnetic survey of parts of Chignik and Sutwik Island quadrangles. Geologic data modified from Beikman (1980) and Detterman and others (1979).

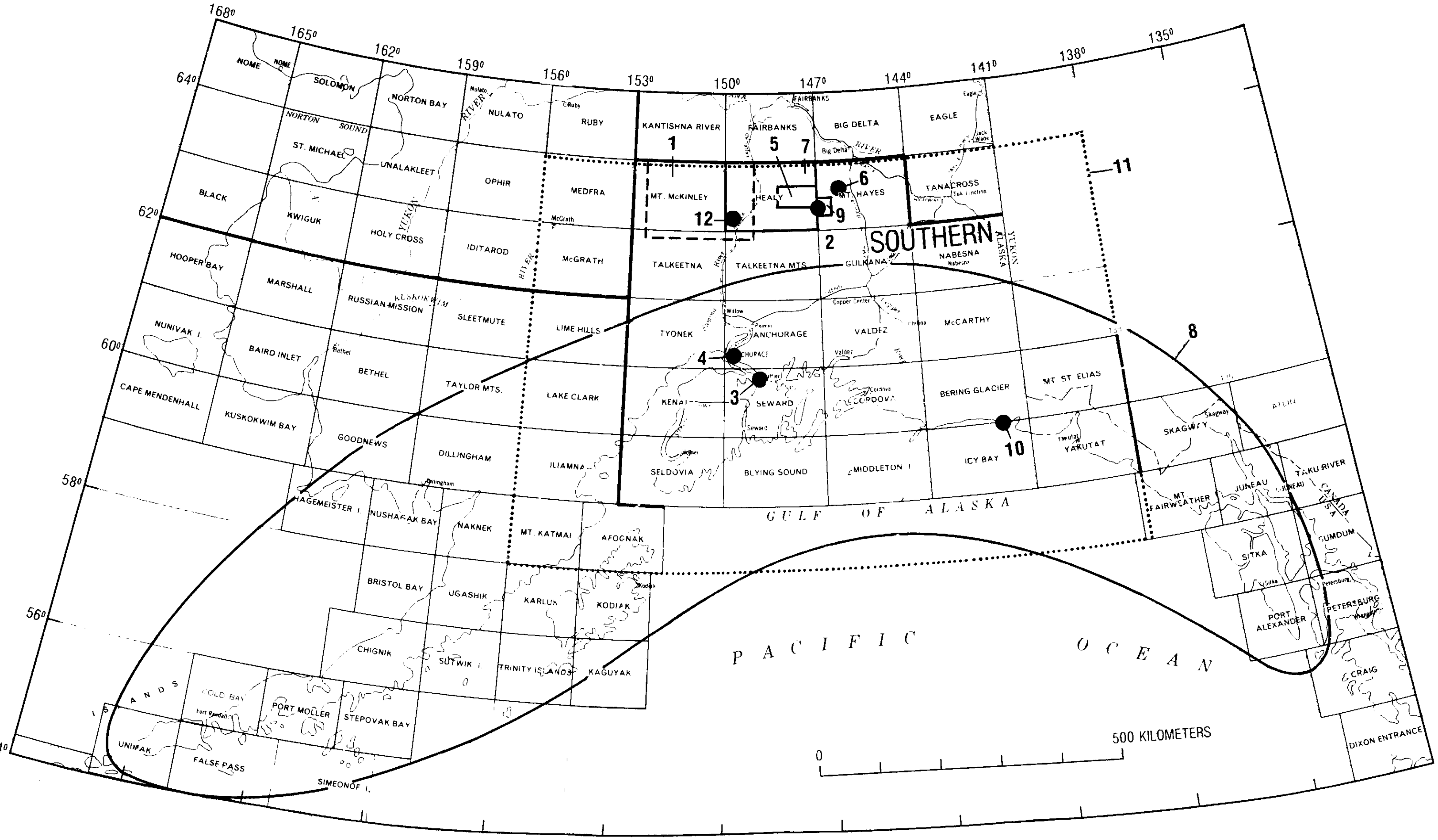
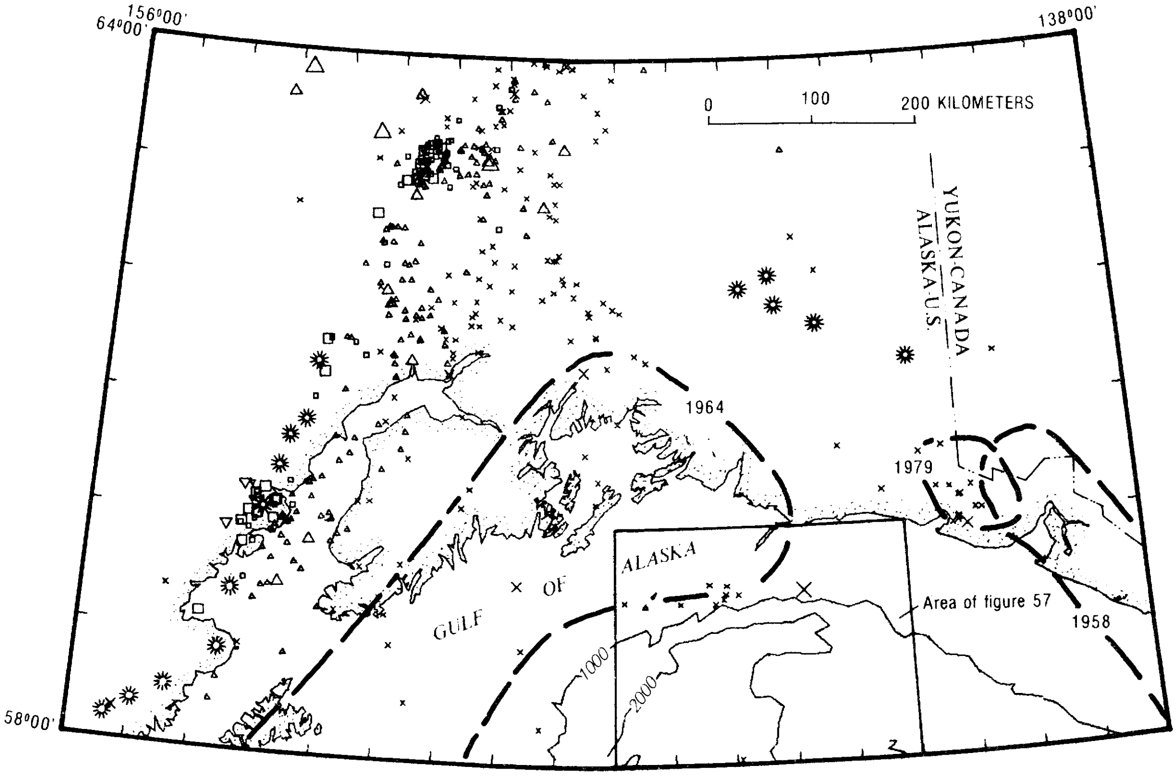


FIGURE 55.—Areas and localities in southern Alaska discussed in this volume. Authorship and applicable figures and tables relating to the numbered areas are listed below. 1, Ager, figure 67; 2, Aleinikoff, Nokleberg, and Herzon, figures 59 and 61 and table 5; 3, Bartsch-Winkler and Garrow, figures 70 and 71; 4, Cowing and Emanuel; 5, Csejtey, Cox, Evarts, Stricker, and Mullen, figure 62 and table 6; 6, Evenson, Stephens, G., Curtin, and Tripp, figures 65 and 66; 7, King, figure 64; 8, Nilsen; figure 58; 9, Nokleberg, Zehner, and Miyaoka, figures 59 and 60; 10, Plafker, Hudson, Rubin, and Dixon, figures 68 and 69, table 7; 11, Stephens, C., and Lahr, figures 56 and 57; 12, Wardlaw, figure 63.

one or more large earthquakes within the next few decades (Lahr and others, 1980; McCann and others, 1980). Within this region, the plate boundary is both complex and poorly defined, so that it is difficult to estimate with certainty the probable extent of future ruptures (Lahr and Plafker, 1980). Both the 1964 and 1979 earthquakes involved thrust motion on shallow planes dipping at low angles to the northwest, and it is likely that future large earthquakes occurring within the Yakataga gap will have similar mechanisms. With an estimated potential slip of almost 4 m (Lahr and Plafker, 1980), a single gap-filling earthquake could be as large as M_W 8.3.

Since the 1979 St. Elias earthquake, an increased emphasis has been placed on monitoring the seismicity within and surrounding



EXPLANATION

DEPTH, IN KILOMETERS	MAGNITUDE		
	3 ≤	4 ≤	5 ≤
0 ≤	×	×	×
50 ≤	△	△	△
100 ≤	□	□	□
150 ≤	▽	▽	▽

FIGURE 56.—Map showing distribution of earthquake epicenters for magnitude 3 and larger events that occurred between October 1979 and September 1980. Sources are Stephens, Fogleman, and others (1980), Stephens, Lahr, Fogleman, Helton, and others (1980), and unpublished data. Symbol size is proportional to magnitude, and symbol type varies with depth as indicated. Heavy dashed lines indicate extent of rupture zones of large earthquakes based on distribution of aftershocks (after Sykes, 1971; Stephens, Lahr, Fogleman, and Horner, 1980). Quaternary volcanoes are indicated by stars (after King, 1969).

the Yakataga gap for possible short-term premonitory changes in spatial or temporal patterns. Since at least September 1974, when the regional coverage of the USGS high-gain seismic network east of Cordova was expanded to near its present configuration, the level of seismic activity east of about 146° W. has been low compared to that in the Prince William Sound -Kenai Peninsula - Cook Inlet region (for example, Stephens and Lahr, 1979). Prior to February 1979, most of the seismic activity in the eastern part of the network was concentrated in two broad zones, one centered on the area of the aftershock zone of the St. Elias earthquake, and the other centered about 100 km northeast of Kayak Island. A review of the available record of seismicity that preceded the 1979 St. Elias earthquake showed that within the immediate vicinity of the aftershock zone during a six-month period before the mainshock the number of events above magnitude 1.8 (the estimated level of completeness) increased 45 percent, and the regional b-value (a measure of the frequency distribution of magnitudes) apparently decreased relative to a comparable six-month period about one year earlier (Stephens, Lahr, Fogleman, and Horner, 1980). In addition, a prominent clustering of events occurred near the southeast corner of the aftershock zone in the six months preceding the mainshock. The record of seismic activity since 1974 is not yet sufficiently complete to determine how unusual these observed short-term variations are.

It is more difficult to recognize and determine the significance of variations in seismicity patterns that might occur in the Yakataga gap region because the magnitude threshold for completeness is higher than in the vicinity of the St. Elias aftershock zone. Within the available record of seismicity for the gap region, there appears to be some evidence that the level of activity northeast of Kayak Island fluctuates with time; the rate of activity for events of magnitude 2 and above was lower between October 1979 and September 1980 than in the six-month period preceding the 1979 St. Elias earthquake. In the offshore area southwest of Kayak Island (fig. 57), the level of seismic activity during the period between October 1979 and September 1980 appears to have increased and is higher than in the six months preceding the St. Elias earthquake. On September 4, 1980, a magnitude 5.2 m_b (PDE) earthquake occurred about 60 km southeast of Kayak Island. This is the largest earthquake to occur either offshore or onshore within the gap region during the past year. Our current program of monitoring the seismic

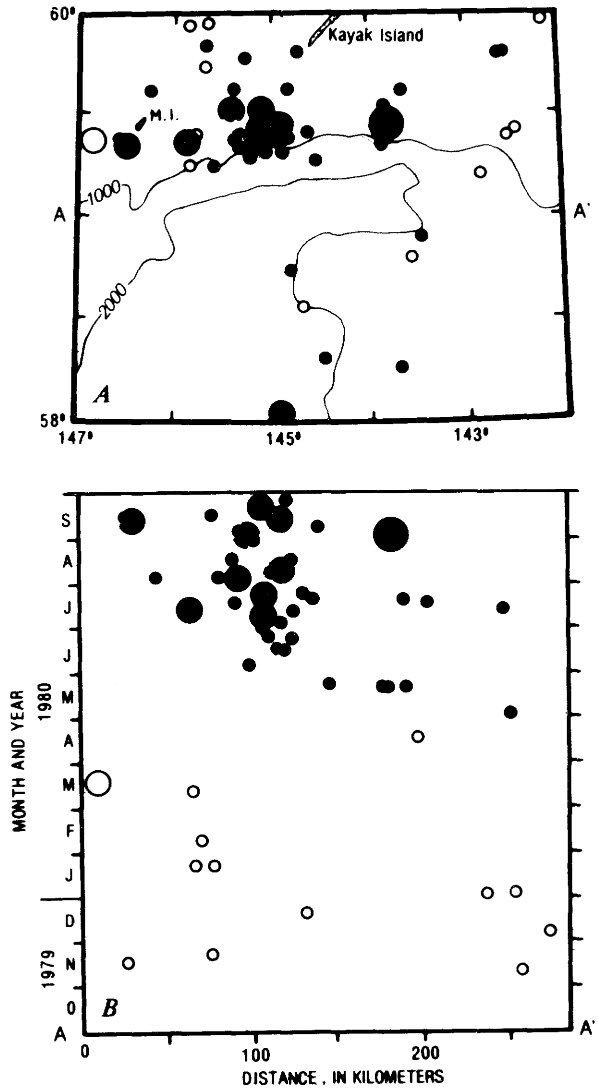


FIGURE 57.—Seismicity plot (magnitude 2 and larger events) south of Kayak Island from October 1979 through September 1980. *A*, Map view. *B*, Space-time plot, events projected onto line A-A′. In both upper and lower parts, solid symbols correspond to events that occurred after April 1980. Three symbol sizes correspond to magnitude as follows: small - less than 3.0; intermediate - 3.0 to 4.9; large - 5.0 and larger. MI is Middleton Island.

activity within and around the Yakataga gap and of completing the earlier seismic record since 1974 should provide a stronger basis for interpreting variations in seismicity patterns as possible premonitory indicators.

To date, no strong-motion records have been obtained in the United States closer than 25 km from the causative fault of a magnitude 7 earthquake (D. Boore, oral commun, 1980), or within 100 km of a magnitude 8 earthquake. Since the 1979 St. Elias earthquake, several new strong-motion stations have been installed within or near the Yakataga seismic gap. Currently, a total of 10 USGS strong-motion instruments, including three new installations made in the summer of 1980, are sited within the gap region between Kayak Island and Icy Bay. These and other nearby strong-motion instruments will assure that a high-quality suite of accelerograms will be obtained for the major earthquake(s) expected in the next few decades.

Several new high-gain seismic stations have been added to the USGS network in southern Alaska. An array of four vertical component and one three-component high-gain seismic stations was installed around Bradley Lake on the Kenai Peninsula near Homer. The array will be used to monitor the regional seismicity around the lake, which is a site being studied by the Army Corps of Engineers for a proposed hydroelectric facility. Recording for these stations began in November 1980. Another high-gain station was installed near Juneau to improve the control for locating earthquakes in southeastern Alaska.

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Accretion model for the Cretaceous Chugach terrane, southern Alaska

By Tor H. Nilsen

The Cordilleran orogen of western North America is now recognized by many workers to consist of a large number of diverse terranes, many of which are allochthonous to the North American continent and appear to have accreted to it in Mesozoic and Cenozoic time (see summary in Coney and others, 1980). Most of Alaska has now been subdivided (Berg and others, 1978; Jones and Silberling, 1979) into various terranes that contain different types of rocks of different ages. Although paleomagnetic and faunal evidence indicates that at least some terranes in central and southern Alaska originated far to the south of their present latitude, the migration history and time of accretion of most are poorly known or unknown. It is also generally unclear whether most terranes arrived individually in Alaska or aggregated elsewhere to be later transported northward to final collision and accretion in Alaska.

This report summarizes recent sedimentologic, petrographic, structural, and paleomagnetic data from the Upper Cretaceous Chugach terrane of southern Alaska and proposes a model for its formation, aggregation as part of a larger block, northward movement, and accretion to Alaska. The Chugach terrane has previously been described as an accreted Mesozoic terrane (Plafker and others, 1977), and its structure, stratigraphy, sedimentology, and petrography have been studied by a number of workers in various areas (see summary in Nilsen and Zuffa, 1982).

The Chugach terrane crops out extensively across southern Alaska for almost 2000 km (fig.

58). Berg and others (1972) named it for outcrops of graywacke, argillite, slate, conglomerate, volcanic rocks, chaotic melanges, and granitic plutons in the Chugach Mountains. They indicated that it extended southeastward to Baranof Island and southwestward to Kodiak Island and the Shumagin and Sanak Islands. They divided the Chugach terrane into an older landward part, consisting mostly of melange containing blocks of various lithology, and a younger seaward part consisting mostly of slate and graywacke. The melange, herein referred to as the melange subterrane, contains fossils as old as Triassic in blocks and as young as Maestrichtian in the matrix (Plafker and others, 1977). The slate and graywacke, herein referred to as the flysch subterrane, contain fossils chiefly of Maestrichtian age (Jones and Clark, 1973). Faults that generally dip landward separate the two parts of the Chugach terrane from each other and from older rocks to the north and younger rocks to the south; however, in southeast Alaska northwest-trending strike-slip faults have complicated this pattern considerably. The older rocks to the north include parts of the Peninsular, Wrangellia, and Alexander terranes (fig. 58) of Coney and others (1980).

The flysch subterrane has been inferred to be a trench deposit by numerous workers (Berg and others, 1972; Moore, 1973; Budnik, 1974a, b; Nilsen and Bouma, 1977; Nilsen and Moore, 1979) on the basis of its tectonic setting, deformation history, sedimentologic characteristics, paleocurrent data, and petrology. It may represent the longest and best preserved ancient trench-fill sequence in the world. Petrographic analysis of rock fragments indicates that it was derived from older accretionary wedge deposits and a magmatic arc to the north that became increasingly dissected eastward (Zuffa and others, 1980). Turbidite facies associations and paleocurrent determinations indicate major westward longitudinal transport of sediment down the axis of the trench and accumulation as narrow and elongate fans (Nilsen and Zuffa, 1982). Slope deposits are preserved along the north edge of the trench-fill facies, and submarine canyons cutting across this area probably fed some sediment laterally to the trench. The highly deformed and regionally metamorphosed melange subterrane to the north may partly represent landward trench-slope deposits that included thick and abundant olistostromal accumulations.

The flysch subterrane contains gabbro, diabasic sheeted dikes, and basalts of oceanic affinity at various locations (Tysdal and others, 1977). Granitic batholiths, interpreted to be products of anatectic melting of the subduction complex, intrude the subterrane and decrease in age eastward from about 60 m.y. on Sanak Island to 43 m.y. on Baranof Island (Hudson and others, 1979).

North of the Chugach terrane are coeval Upper Cretaceous sedimentary rocks in south-central and southwestern Alaska that form a belt of forearc basin or arc-trench gap deposits. These units crop out on the Alaska Peninsula (Mancini and others, 1978; Detterman, 1978), in the Cook Inlet area (Magoon and others, 1980), and north of Cook Inlet (Grantz, 1964). They consist of nonmarine, shallow-marine, and deep-marine clastic sedimentary deposits transported southward and derived chiefly from a magmatic arc. The belt generally grades southward into offshore and deeper marine deposits.

North of the forearc basin belt are Upper Cretaceous and lower Tertiary quartz diorite to granite batholiths referred to as the Iliamna-McKinley phase of the Alaska Range-Talkeetna Mountains batholithic complex (Hudson, 1979a, b). Moore and Connelly (1979) concluded on the basis of K_2O trends that the batholiths were generated in response to northwest-directed subduction.

I believe that the belts of Upper Cretaceous trench-fill, forearc basin, and magmatic arc rocks form a related group that developed in response to Late Cretaceous subduction. In addition to accretion of the Chugach terrane (Plafker and others, 1977), the associated forearc basin and magmatic arc belts must be accounted for in a reasonable tectonic model. The larger block, a megaterrane, most likey accreted as a unit, because otherwise the trench-fill, forearc basin, and magmatic arc belts would probably have been dismembered and reoriented with respect to one another.

Paleomagnetic studies of southern Alaska rocks, as well as reinterpretations of the record of northeast Pacific sea-floor spreading, are compatible with this model. Although paleomagnetic data from the Chugach terrane are sparse, those reported by Stone and Packer (1976, 1979) and Hillhouse and Gromme (1977)

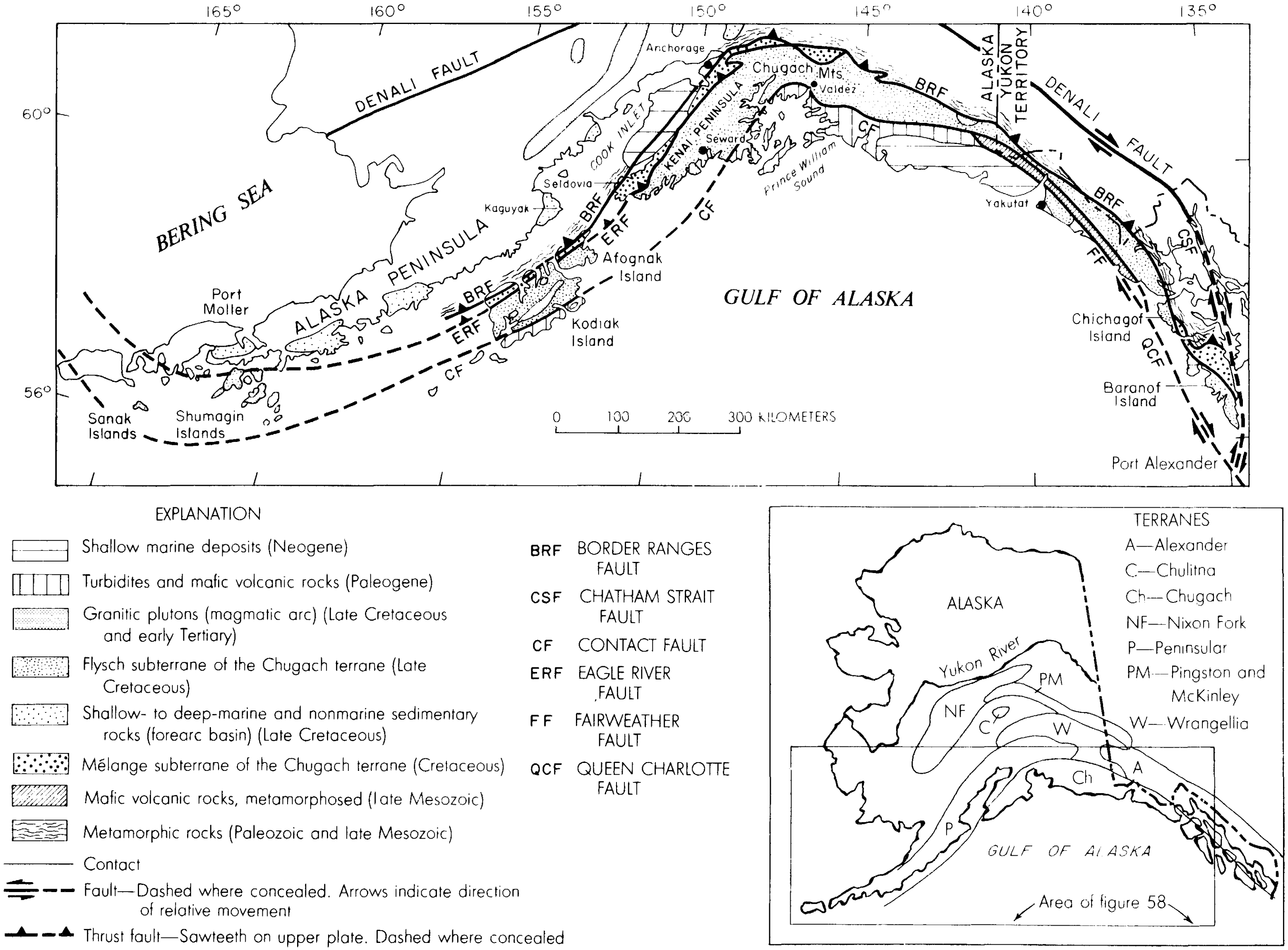

FIGURE 58.—Distribution and geologic framework of Chugach terrane. Inset map shows terrane distribution from Coney and others (1980).

from Jurassic rocks that underlie the forearc basin deposits, parts of the flysch subterrane, and Eocene rocks that unconformably overlie the forearc basin deposits indicate that the Jurassic to Eocene strata were originally deposited and magnetized far to the south of their present latitude. Stone (1980; oral commun., 1980) concluded on the basis of paleomagnetic data that the Chugach, Peninsular, and Wrangellia terranes, as well as Cretaceous turbidites of the Kuskokwim Group in west-central Alaska, appear to have been connected since Late Jurassic time and have undergone similar northward movement since the Jurassic. He suggests a Late Cretaceous paleolatitude of about 20° N. for these terranes. Byrne's (1979) reconstructions of the plate history of the northeast Pacific indicate that southern Alaska was situated south of its present latitude and moved independently of North America during the Early Tertiary.

The time of accretion of the megaterrane to Alaska is not well constrained by paleomagnetic, structural, or sedimentologic data. It probably accreted at least before middle Miocene time because the early Miocene to Holocene Yakataga Formation of southeast Alaska, which rests unconformably on Paleogene turbidites on the southwest flank of the Chugach terrane, contains glacial deposits in its upper part that suggest a high-latitude setting (Armentrout, 1980).

If one can accept that the megaterrane, which includes the Peninsular and Wrangellia terranes and possibly others such as the Nixon Fork, Chulitna, and Alexander, accreted to Alaska as a large block during the early and middle Cenozoic, a number of major problems become apparent, as well as solutions to other problems. For example, rather than requiring that a great variety of Mesozoic terranes of varying ages accrete to southern Alaska at various times in the Mesozoic, the megaterrane permits accretion of smaller blocks elsewhere in the late Mesozoic, prior to development of the Chugach arc system. After the Chugach trench-forearc basin-magmatic arc complex was superimposed on the previously accreted smaller blocks, the entire system of accreted blocks was then transported northward to Alaska in the Cenozoic.

The Aleutian Abyssal Plain contains thick and voluminous turbidities as old as Eocene that are petrographically similar to south-transported Eocene turbidities on the south flank of the flysch subterrane (Stewart, 1976). A suitable and sufficiently large provenance for the abyssal plain turbidites has not been clearly recognized to date. The large northward-migrating megaterrane could have been the source for both the presently onland and offshore turbidites.

The ultimate place of origin of the flysch subterrane and the larger megaterrane remains wholly speculative. Some paleomagnetic data suggest that Jurassic and older components of the megaterrane may have formed subequatorially. Possible places of origin for the Upper Cretaceous trench-fill deposits of the Chugach terrane and associated forearc basin and magmatic arc belts are the areas of western Mexico and central America and the Pacific Northwest-southern British Columbia. Although the present orientation of the megaterrane suggests northwest-directed subduction, it may originally have been east-directed. More paleomagnetic studies of both limbs of the Chugach terrane are needed to clarify its history of translation and rotation.

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Geologic setting of the Maclaren metamorphic belt, Mount Hayes A-6 and B-6 quadrangles, eastern Alaska Range, Alaska

By Warren J. Nokleberg, Richard E. Zehner, and Ronny T. Miyaoka

The Maclaren metamorphic belt, first named by Smith and Turner (1973, 1974), occurs south of the McKinley strand of the Denali fault in the south-central and southwestern part of the Mount Hayes quadrangle in the eastern Alaska Range (figs. 59, 60). The Maclaren metamorphic belt is bounded to the north by the East Susitna batholith which consists of a suite of regionally metamorphosed and deformed granitic plutons of Cretaceous and Tertiary age (figs. 59, 60) (Turner and Smith, 1974). The Maclaren metamorphic belt and the East Susitna batholith constitute the Maclaren River terrane, a unique tectonostratigraphic terrane south of the Denali fault (Nokleberg and others, 1981). The Maclaren River terrane is bounded to the south by the Wrangellia terrane, and the Broxson Gulch thrust system of Stout (1976) separates the two terranes.

From south to north, the three major map units forming the belt are argillite and metagraywacke, phyllite, and schist (fig. 60). These three units dip moderately to steeply north, and each lower grade unit dips northward underneath the next higher grade unit, thereby forming an inverted metamorphic sequence (fig. 60).

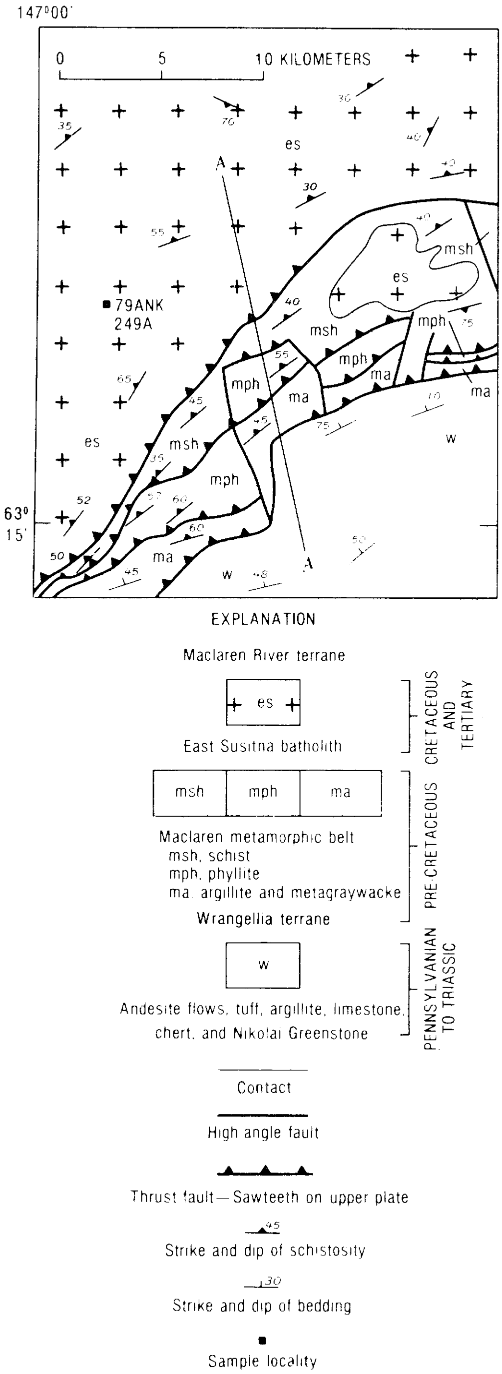

FIGURE 59.—Simplified geologic map of part of Mount Hayes A-6 and B-6 quadrangles showing location of East Susitna batholith, Maclaren metamorphic belt, and Wrangellia terrane. Surficial units omitted. Sample location and number in East Susitna batholith denotes locality for rock selected for uranium-thorium-lead and potassium-argon isotopic studies (Aleinikoff and others, this volume). See figure 60 for cross section *A-A'*.

Similarly, the Maclaren metamorphic belt is thrust under the East Susitna batholith to the north along the steeply northdipping Meteor Peak fault (fig. 60). Contacts between the three units are generally faults with intense shearing and abrupt change of metamorphic facies at each contact. Local faulting along contacts has substantially reduced thicknesses of units in the Maclaren metamorphic belt. The approximate maximum structural thickness of each unit is 2000 m for the argillite and metagraywacke unit, 2400 m for the phyllite unit, and 2000 m for the schist unit.

Each unit of the Maclaren metamorphic belt is intensely deformed. Bedding is mostly transformed into slip-cleavage in the argillite and metagraywacke unit and almost entirely transformed to schistosity in the phyllite and schist units. Isoclinal folds of the first generation are developed in bedding and are common in the argillite and phyllite. Where observed in the same outcrop, slip-cleavage or schistosity occurs parallel to the axial planes of the first-generation isoclinal folds, and this parallelism indicates that these planar structures are also first-generation structures. In the phyllite and schist units, schistosity is locally refolded into isoclinal folds of the second generation. A second-generation axial plane schistosity commonly occurs parallel to the axial planes of the second-generation folds. The axial planes of both first-and second-generation folds and parallel slip-cleavage and schistosity generally dip moderately to steeply north, parallel to the major faults between units. Locally, second-generation folds are asymmetrical and verge southward, whereas first-generation folds show opposite sense of asymmetry. These relations indicate at least two periods of movement during regional-grade metamorphism and penetrative deformation, the second of which consisted of south-directed folding and thrusting.

The argillite and metagraywacke unit is composed predominantly of volcanic graywacke and siltstone, andesite, and lesser calcareous and quartzose mudstone. The volcanic parentage is indicated by relict plagioclase microlites and relict plagioclase phenocrysts with complex twinning and normal and oscillatory zoning. The argillite and metagraywacke unit is moderately to strongly recrystallized and strongly schistose. About 20 to 40 percent of rocks in the unit consist of relict minerals, mainly quartz, graphite, plagioclase, and opaque minerals. No fossils have been found in the argillite and metagraywacke unit.

In general, metamorphic grade increases from the argillite and metagraywacke unit in the south to the schist unit in the north. In the argillite and metagraywacke unit, the common metamorphic minerals are quartz, plagioclase, white mica, chlorite, actinolite, clinozoisite, calcite, and graphite. Average grain size is about 0.01 to 0.05 mm. In the phyllite unit, the common metamorphic minerals are quartz, plagioclase, biotite, white mica, chlorite, actinolite, clinozoisite-epidote, calcite, and sparse garnet and graphite. Average grain size is about 0.05 to 0.1 mm. In the schist unit, the common metamorphic minerals are hornblende, garnet, biotite, white mica, plagioclase, and sparse epidote and calcite. Average grain size is about 0.1 to 0.3 mm. Commonly, poikiloblastic plagioclase grains in the schist unit contain cores with relict phyllite and (or) argillite and metagraywacke defined by schistose aggregates of graphite, white mica, and clinozoisite. This textural relation indicates that the schist unit is derived from prograde metamorphism of the argillite and metagraywacke and phyllite units.

In thin section, the dominant metamorphic texture in the Maclaren metamorphic belt is a strongly developed schistosity. Relict grains in the argillite and metagraywacke unit show abundant fractures, shears, and undulose extinction. In all three units, porphyroblasts commonly exhibit granulated margins and undulose extinction and are locally displaced along schistosity. Rolled garnet occurs locally in the schist unit. These textures indicate mainly syntectonic crystallization of most metamorphic minerals in the Maclaren metamorphic belt.

From south to north, the principal prograde metamorphic changes in the Maclaren metamorphic belt are: a gradual disappearance of graphite, chlorite, and actinolite in the phyllite unit; a reduction in the amount of clinozoisite-epidote, and a change from clinozoisite to epidote in the phyllite unit; the appearance of biotite and garnet in the phyllite unit; the appearance of hornblende in the schist unit; a gradual decrease of relict minerals in the argillite and metagraywacke and phyllite units; and a coarsening of metamorphic mineral grain size. These prograde changes indicate a gradual transition from lower-greenschist-facies metamorphism in the argillite and metagraywacke unit in the south to amphibolite-facies metamorphism in the schist unit to the north. Locally in areas of intense shearing, particularly near contacts between units, the higher grade rocks in the schist and phyllite units show indications of retrograde metamorphism. These indications include reaction of hornblende, biotite, and (or) garnet to actinolite, chlorite, and (or) clinozoisite. This retrograde metamorphism is commonly developed near or along the axial plane schistosity of the second generation.

In summary, the Maclaren metamorphic belt exhibits the classical features of prograde Barrovian-type metamorphism developed in wallrocks adjacent to a copmosite batholith that was intruding into a tectonically active

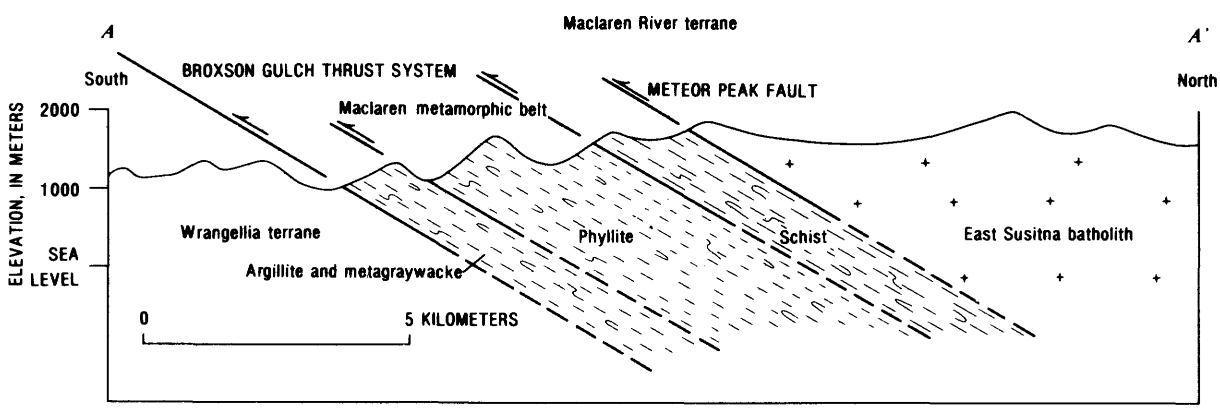

FIGURE 60.—North-south cross section showing East Susitna batholith, various units of Maclaren metamorphic belt, and Wrangellia terrane in Mount Hayes A-6 and B-6 quadrangles, eastern Alaska Range, Alaska. See figure 59 for location of cross section.

area. Inversion and tectonic shortening of the Maclaren metamorphic belt, retrograde metamorphism, and thrusting of the East Susitna batholith over the Maclaren metamorphic belt probably occurred during southward-verging thrusting along the Broxson Gulch thrust system in Tertiary to recent time (Nokleberg and others, 1981).

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Age of intrusion and metamorphism of the East Susitna Batholith, Mount Hayes B-6 quadrangle, eastern Alaska Range, Alaska

By John N. Aleinikoff, Warren J. Nokleberg, and Paige L. Herzon

The East Susitna batholith occurs immediately south of the Denali fault in the south-central and southwestern part of the Mount Hayes B-6 quadrangle in the eastern Alaska Range. The batholith consists of a series of regionally metamorphosed and deformed plutonic rocks ranging in composition mainly from quartz diorite to quartz monzonite. The batholith is bounded to the south by the Maclaren metamorphic belt which is a tectonically shortened, prograde Barrovian-type metamorphic belt (Smith and Turner, 1974; Nokleberg and others, 1981). The East Susitna batholith and the Maclaren metamorphic belt are separated by the Meteor Peak fault which is interpreted as a tectonically shortened intrusive contact (Nokleberg and others, 1981). The East Susitna batholith and the Maclaren metamorphic belt to the south constitute the Maclaren River terrane, a unique tectonostratigraphic terrane south of the Denali fault (Nokleberg and others, 1981). Previous geochronologic studies of the East Susitna batholith in the Mount Hayes B-6 quadrangle by Turner and Smith (1974) show potassium-argon ages ranging from 67.9 to 48.4 m.y. for biotite and hornblende from granitic gneiss, migmatite, and amphibolite forming part of the batholith.

Because of moderate discordance in the potassium-argon ages of biotite and hornblende from single rocks and large discordance in ages between rocks from the East Susitna batholith, a highly deformed and metamorphosed sample of schistose quartz diorite (sample 79ANK249A, fig. 59 and table 5) was chosen for uranium-thorium-lead isotopic study of zircon and sphene and potassium-argon isotopic study of biotite. Zircon, sphene, and biotite were extracted using a Wilfley table and heavy liquids. The zircons are remarkably clear and occur in two distinct morphologies. More than 95 percent of the zircons are doubly terminated with a typical length to width ratio of about 3. Less than 5 percent of the zircons are needlelike with a length to width ratio of at least 15; these zircons could not be separated for isotopic analysis. Many zircons contain inclusions but show no visible evidence of inherited material. Four size fractions of zircon were analyzed for uranium, thorium and lead isotopes. Standard extraction chemistry was used, modified slightly from Krogh (1973). Isotope ratios were measured on a 12-inch NBS mass spectrometer with on-line digital processing.

Isotopic data from the analyses of zircon form a linear array with a lower concordia intercept at 70±7 m.y. (table 5, fig. 61). The upper concordia intercept is poorly known because the locations of the data points are very near the lower intercept and because of the small amount of scatter in relation to the best-fit line. However, the data provide excellent evidence for the occurrence of inherited radiogenic lead. As shown by the upper concordia intercept, the age of the source rock(s) forming the magma for the schistose quartz diorite may have been middle Precambrian (Proterozoic Y). Uranium is relatively enriched in the finer grains, a typical characteristic of zircons first noted by Silver and Deutsch (1963). Inversely, the decrease in

$^{207}Pb/^{206}Pb$ ages with decreasing grain size indicates that coarser grains contain more inherited radiogenic lead than the finer grains. Surprisingly, there is no visible evidence, such as cores or overgrowths, for inheritance of old lead. We interpret the zircon data as indicating that the schistose quartz diorite was intruded about 70 m.y. ago. The source rock for this quartz diorite was probably 800 to 1900 m.y. old (Proterozoic Y). Additional isotopic analyses are needed to refine the age of the source rock.

A sample of sphene from the schistose quartz diorite was also analyzed for uranium, thorium, and lead isotopes (table 5, fig. 61). Biotite was analyzed for potassium and argon isotopes. The isotopic analysis of sphene indicates a 56-m.y. age (table 5, fig. 61), and the isotopic analysis of biotite indicates a 56.9±3.9-m.y. age. Sphene forms ragged inequant crystals. The long dimensions of the sphene crystals generally parallel the prominent schistosity in the sample, which is most visibly defined by parallel plates of biotite. These textural relations indicate that both sphene and biotite are metamorphic minerals. Minor poikiloblastic hornblende crystals occurring along schistosity in the schistose quartz diorite indicate amphibolite-facies metamorphism. These textural and paragenetic relations suggest that the isotopic ages of 56 m.y. for sphene and 57 m.y. for biotite represent the age of regional-grade amphibolite-facies metamorphism.

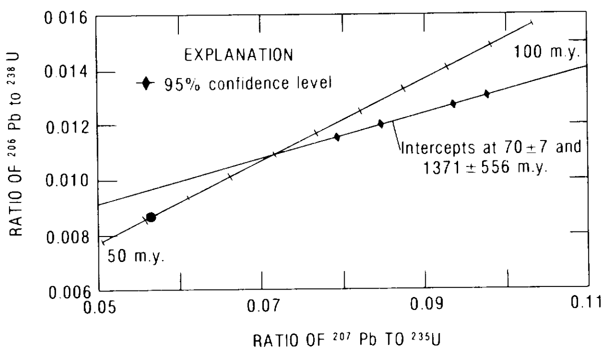

FIGURE 61.—Concordia diagram of zircon (diamonds) and sphene (circle) from the schistose quartz diorite of the East Susitna batholith, Mount Hayes B-6 quadrangle, eastern Alaska Range. Least-squares best fit line through the four zircon points indicates an igneous intrusive age of 70 ± 7 m.y.

These studies show that: (1) the source rock for the magma that formed the schistose quartz diorite was probably 800 to 1900 m.y. old; (2) the schistose quartz diorite was intruded about 70 m.y. ago; (3) the schistose quartz diorite was subsequently regionally metamorphosed and deformed under conditions of the amphibolite facies about 57 m.y. ago; and (4) in regionally metamorphosed and deformed granitic rocks in this area, the age of intrusion, as indicated by uranium-thorium-lead zircon isotopic studies, may be substantially older that the age of metamorphism, as indicated by uranium-

Table 5.—*Uranium-thorium-lead isotopic data for zircon and sphene from quartz diorite of the East Susitna batholith, Alaska Range*

[Sample location: 63°20′43″N., 146°55′26″W. (fig. 59). Age constants: $\lambda^{235}U = 0.98485X10^{-9}/yr$, $\lambda^{238}U = 0.155125X10^{-9}/yr$, $\lambda^{232}Th = 0.0475X10^{-9}/yr$, and $^{235}U/^{238}U = 1/137.88$. n.d., not determined.]

Sample 79ANK249A	Concentration (ppm)			Atomic percent*				Age (m.y.)			
Size fraction	U	Th	Pb	^{204}Pb	^{206}Pb	^{207}Pb	^{208}Pb	$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$	$^{208}Pb/^{232}Th$
(-150+200)	379.3	80.0	4.9	0.023	87.35	5.059	7.57	84	94	374	93
(-200+250)	403.8	78.7	5.1	.032	86.86	5.130	7.97	81	91	356	99
(-325+400)	513.7	n.d.	6.1	.035	87.19	4.953	7.82	77	82	234	--
(-400)	545.2	n.d.	6.3	.056	86.45	5.129	8.37	74	77	184	--
(+150)S**	78.1	14.5	2.5	1.019	42.59	17.016	39.37	56	56	56	3

*Blank corrected.

**Sphene; all other samples are zircons.

thorium-lead studies of metamorphic sphene and by potassium-argon isotopic studies of metamorphic biotite and hornblende. Previously reported potassium-argon isotopic ages, particularly those ranging from 60.2 to 48.4 m.y., for granitic rocks in this area (Turner and Smith, 1974), probably represent mainly ages of regional grade metamorphism.

The geochronology of the East Susitna batholith is of considerable importance to the regional tectonics of southern and southeastern Alaska. Forbes and Smith (1973) first suggested that the only known rocks correlative with the Maclaren River terrane on the north side of the Denali fault occur several hundred kilometers to the southeast in the Coast Plutonic Complex (Tracy Arm terrane of Berg and others, 1978). This study supports this interpretation and also indicates a Precambrian source for part of the East Susitna batholith. The nearest known Precambrian source would be somewhere along the western margin of the North America Cordillera, possibly in the Tracy Arm terrane of Berg and others (1978).

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The Denali fault system of Alaska: The case for minor rather than major Cenozoic lateral displacement

By Béla Csejtey, Jr., Dennis P. Cox, Russell C. Evarts, Gary D. Stricker, and Michael W. Mullen

The Denali fault system of southern Alaska, first described as a major strike-slip system by St. Amand (1957), forms a northward-convex surface trace across the entire width of the State for a distance of more than 1200 km. Along its central part in Alaska, near its apex, the system is split into two strands (Grantz, 1966): the Hines Creek fault on the north, and the McKinley fault on the south (fig. 62). Although the Hines Creek fault has been essentially inactive for about the past 95 m.y. (Wahrhaftig and others, 1975), the rest of the system, including the McKinley fault, shows evidence of Cenozoic movement (Grantz, 1966). The Hines Creek fault and the eastern part of the Denali fault system do form boundaries between dissimilar tectonostratigraphic terranes (Beikman, 1980; Csejtey, Foster, and Nokleberg, 1980), and elsewhere the Denali fault appears to offset a number of geologic features (Reed and Lanphere, 1974; Jones and others, 1980). Consequently, the Denali fault system, with the exception of the Cretaceous Hines Creek fault, has been postulated in the literature to have several tens or even hundreds of kilometers of right-lateral offset in Cenozoic time (Smith, 1974; Forbes and others, 1974; Reed and Lanphere, 1974; Turner and others, 1974; Hickman and others, 1977; Lanphere, 1978; Jones and others, 1980).

The authors' recent reconnaissance investigations along the McKinley fault segment of the Denali fault system, in the eastern part of the Healy quadrangle, yielded metamorphic-facies and isograd data which suggest that such large Cenozoic lateral offsets do not exist. Mapping in the Healy quadrangle further indicates that the McKinley fault in central Alaska is not a boundary fault between dissimilar tectonostratigraphic terranes. This evidence implies that the eastern part of the Denali fault system must have developed in Cenozoic time along an old Cretaceous suture (Csejtey, Foster, and Nokleberg, 1980; Csejtey, Stricker, and others, 1980).

A simplified geologic map of the area that yielded the information on the McKinley fault in the Healy quadrangle is shown in figure 62.

In this area, several superimposed thrust sheets were emplaced during the accretionary development of southern Alaska in middle Cretaceous time (Csejtey and others, 1978; Csejtey, 1979; Csejtey, Foster, and Nokleberg, 1980). The structurally lowest position is occupied by a Lower Cretaceous flysch sequence. Thrust on top of the flysch is an Upper Triassic carbonaceous sequence of limestone, calcareous shale, and sandstone containing diabase sills of probable Jurassic age. Thrust on top of the Upper Triassic rocks is a Paleozoic, probably upper Paleozoic, sequence of mafic volcanic rocks, varicolored radiolarian chert, and shale. After thrusting, the area was intruded by granitic plutons of early Late Cretaceous and Tertiary ages (Turner and Smith, 1974).

Probably after emplacement of the Cretaceous plutons, the eastern part of the study area was regionally metamorphosed, both north and south of the McKinley fault. This metamorphism appears to be part of a large regional feature named the Maclaren metamorphic belt by Smith (1974). In the study area, the metamorphic grade increases eastward toward the core of the metamorphic belt, from lower greenschist facies to lower amphibolite facies assemblages (definition of metamorphic facies after Turner, 1968). The metamorphic gradient is about the same on both sides of McKinley fault, and rocks with similar metamorphic mineral assemblages, that is, of similar degree of metamorphism, occur opposite each other across the fault (table 6). The similarity in grade across the fault has also been noted by Wayne Brewer of the University of Wisconsin (written com-

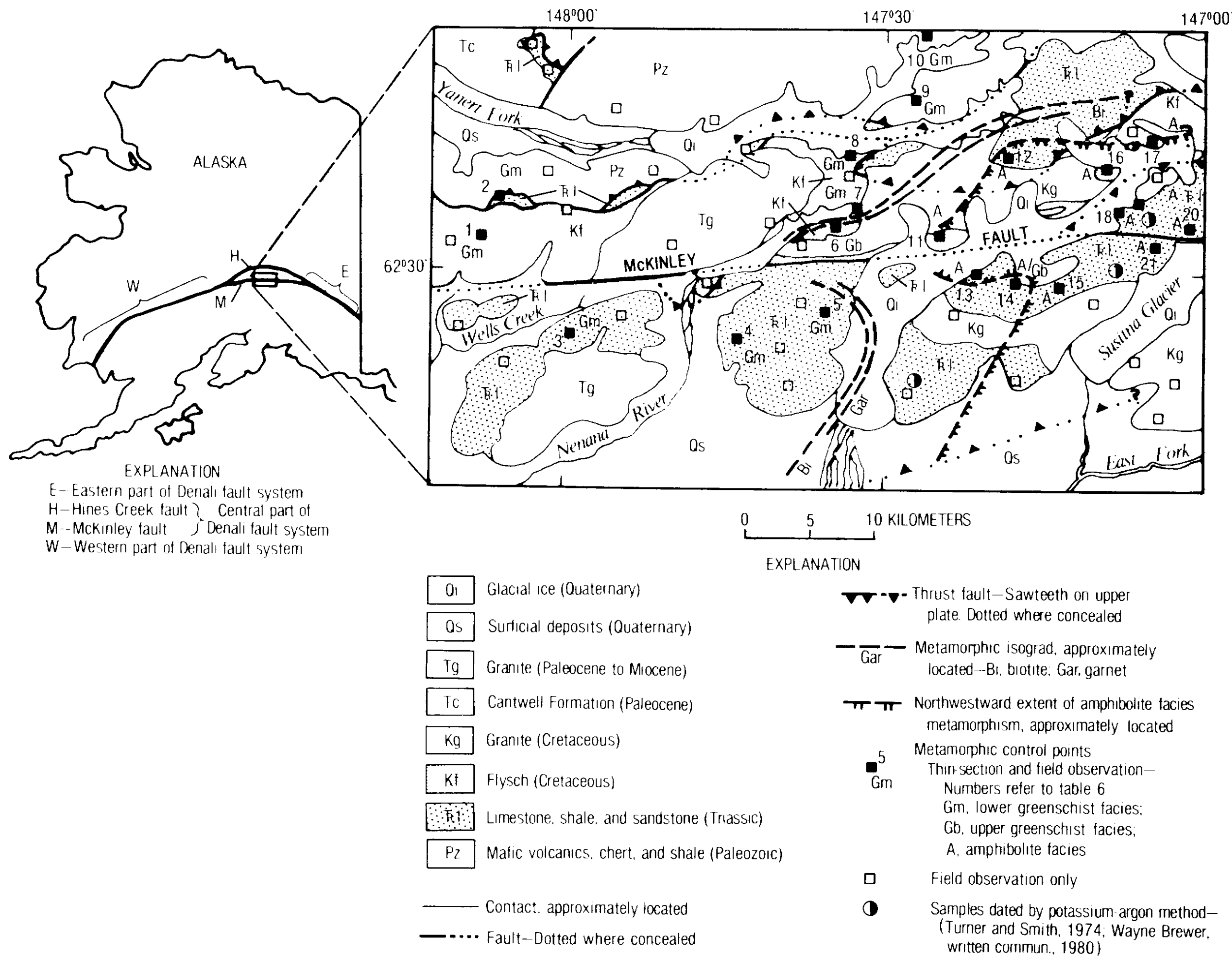

FIGURE 62.—Location of Denali fault system in Alaska, and generalized geologic map of area in eastern part of Healy quadrangle.

Table 6.--*Metamorphic mineral assemblages along the McKinley fault in the eastern part of the Healy quadrangle*

[Map numbers indicate sample locations in figure 62. Abbreviations as follows: Act-actinolite; Ab-albite; And-andesine; Bi-biotite; Cc-calcite; Clpx-clinopyroxene; Clzo-clinozoisite; Di-diopside; Ep-epidote; Gar-garnet of undetermined composition; Gross-grossular garnet; Hb-hornblende; Kfp-potassium feldspar; Ogcl-oligoclase; Pc-plagioclase; Qtz-quartz; Sph-sphene; Stilp-stilpnomelane; Tour-tourmaline; Wm-white mica, including muscovite; Wo-wollastonite. Gm-lower greenschist facies metamorphism, with white mica only. Gb-upper greenschist facies metamorphism, with biotite. A-amphibolite facies metamorphism.]

NORTH OF McKINLEY FAULT

Map number	1	2	6	7	8	9	10	11	12	16	17	18	19	20
Field number(s)	80ACy-18	80ACx-101	80ACy-72	80ACy-71	80ACy-61	78ACy-176	78ACy-177	80ACx-109 (A, C, C-1)	80ACy-88	80ACy-87	80ACy-86	80ACx-116 (A1, A2, A3, B)	80ACx-117 (A, B)	80ACy-78, 82F, 82F
Lithologic assemblages: Calcic		Qtz, Cc; minor Wm							Di, Trem, And/Labr, Qtz; minor Gar, Cc			B: Di, Cc, Qtz, And, Clzo; minor Sph, Wm	B: Qtz, Cc, Ep, Clzo, Act; minor Sph, Wm	82E: Cc, Qtz, Act or Trem, Hb(?); minor, Gar, Wm, Ep, Sph, Clzo, Di(?)
Lithologic assemblages: Peletic and calc-pelitic	Qtz, Wm, Ab		Qtz, Bi; minor Ab, Gar	Qtz, Wm; possibly minor Ab	Wm, Qtz, Stilp	Qtz, Wm; minor Stilp	Qtz, Wm			Hb, Bi, Qtz; minor And, Gar	Qtz, Bi, Wm; minor Gar, Tour, Pc (prob. And)		A: Qtz, Bi, Kfp; minor Pc (And?), Wm, Gar, Ep	78: Qtz, And, Bi; minor Gar, Kfp
Lithologic assemblages: Quartzo-feldspathic								C: Qtz, Bi, Wm, Clzo, Minor Gar C-1: Qtz, Bi; minor And, Kfp				A-1: Qtz, Bi, Pc, Kfp; minor Gar, Wm A-3: Qtz, And, Bi, Kfp; minor Gar, Wm		
Lithologic assemblages: Basic								A: Act, Clpx, Pc (prob. And), Sph, minor Qtz				A-2: Hb, And, Bi, Chlor; minor Qtz, Wm		82F: Hb
Assigned metamorphic facies	Gm	Gm	Gb	Gm	Gm	Gm	Gm	A?	A	A	A?	A	A	A

SOUTH OF McKINLEY FAULT

Map number	3	4	5	13	14	15	21
Field number(s)	80ACx-100	80ACx-115	80ACy-85 (A, B, C)	78ACy-174	80ACx-105 (A, B)	80ACx-106 (A, B, C)	80ACx-107 (A, B, C)
Lithologic assemblages: Calcic	Qtz, Cc; minor Wm	Qtz, Cc; minor Wm	Wm, Qtz, Cc	Qtz, Di, Wm, Cc, And; minor Sph, Gross, Wo	B: Qtz, Cc, Wm, Bi; minor Gross, Sph, Act or Trem, Tour, Ogcl		C: Cc, Act or Trem, minor And, Bi, Qtz
Lithologic assemblages: Pelitic and calc-pelitic					A: Qtz, Bi, Wm	A: Qtz, Bi; minor Wm, Gar, Tour	A: Otz, Bi; minor Wm, Gar, Tour
Lithologic assemblages: Basic						B: Bi, Act; minor Cc, Qtz C: Hb, And; minor Qtz	B: Act, And, Bi; minor Qtz
Assigned metamorphic facies	Gm	Gm	Gm	A	A or Gb	A	A

mun., 1980). Rocks of the lower greenschist facies are fine grained and only partially recrystallized, whereas amphibolite-facies rocks to the east are fine to medium grained and fully recrystallized and generally have a distinct metamorphic foliation. Two approximately located isograds, marked by the first appearance of biotite and garnet, respectively, appear to approach the McKinley fault, both from the north and the south, at about the same general location. More significantly, an extensive area of the lower amphibolite facies metamorphism, defined in this paper (after Turner, 1968) by the presence of andesine (identified optically), appear to trend across the McKinley fault without any obvious horizontal offset (fig. 62 and table 6).

Potassium-argon age determinations on metamorphic rocks from both sides of the McKinley fault in and near the study area (Turner and Smith, 1974; Wayne Brewer, written commun., 1980) yielded ages of about 80 to 55 m.y., indicating that the metamorphism is probably Late Cretaceous, possibly early Tertiary in age.

Because of the relative paucity of control points and the lack of more definitive isotopic age determinations, tectonic interpretations based on the metamorphic data from the Healy quadrangle should be considered tentative. Nevertheless, we believe the data permit us to conclude, with a fair degree of confidence, that the possible horizontal offset along the McKinley fault since the metamorphism is not more than about 10 km.

The metamorphic data from the Healy quadrangle also suggest that the correlation of the Maclaren and Ruby Range post-Early Cretaceous metamorphic belts proposed by Forbes and others (1974), Smith (1974), and by Turner and others (1974), as well as the 400 km of dextral offset along the Denali fault system implied by this correlation, is not likely because rocks of the Maclaren metamorphic belt are not significantly offset by the McKinley fault and because major displacement on the Hines Creek fault is older than 95 m.y. (Wahrhaftig and others, 1975).

Jones and others (1980) have pointed out that large apparent horizontal offsets of allochthonous terranes do occur along the western part of the Denali fault system. By way of reconciling our data with theirs, we suggest that several hundreds, perhaps even a few thousands of meters of vertical separation, in addition to the possible horizontal offset of a few kilometers, did occur along the Denali fault system in Cenozoic time. Such movements would produce large apparent horizontal offsets in a region of generally flat-lying superimposed thrust sheets but little displacement of apparently steeply-dipping metamorphic isograds.

The 40 km of dextral offset along the McKinley fault in the last 40 m.y. proposed by Reed and Lanphere (1974) on the basis of correlation between the McGonagall and Foraker plutons in the Mount McKinley and Talkeetna quadrangles seemed convincing prior to our study. We propose now that most of the offset in plutonic contacts is apparent and results fortuitously from relatively minor fault displacement of an irregularly shaped plutonic body.

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Smithian and Spathian (Early Triassic) conodont faunas from the Chulitna terrane, south-central Alaska

By Bruce R. Wardlaw

The Chulitna terrane is a key tectonostratigraphic unit in the Upper Chulitna district, Alaska. It contains a distinctive suite of rocks found nowhere else in Alaska or North America (Jones and others, 1979). The Chulitna terrane (fig. 63*A*) is composed of ophiolitic rocks of Late Devonian age; late Paleozoic flysch, conglomerate, sandstone, chert, and limestone; Early Triassic limestone; Late Triassic basalt, limestone, and redbeds; and Jurassic and Cretaceous clastic rocks. This report documents the occurrence of conodonts in the Early Triassic limestone that forms the uppermost part of the unit mapped by Jones and others (1979, pl. 1) as sedimentary rocks of Devonian to Early Triassic age (fig. 63*A*, ₮ Ds).

Nichols and Silberling (1979) reported on the ammonite fauna recovered from the Early Triassic limestone. It contains *Meekoceras gracilitatis* and an associated fauna that is in common with that of the early and middle Smithian rocks of the northern miogeosyncline in Idaho, Nevada, and Utah; the common faunas suggest a strong relation for the displaced Chulitna terrane to the conterminous United States. The conodont faunas have been analyzed to see if any supporting data can be obtained.

Conodonts were recovered from two samples. One sample (M5027), from locality 20 (Silberling and others, 1978; Jones and others, 1979), was made up of the limestone matrix of the ammon-

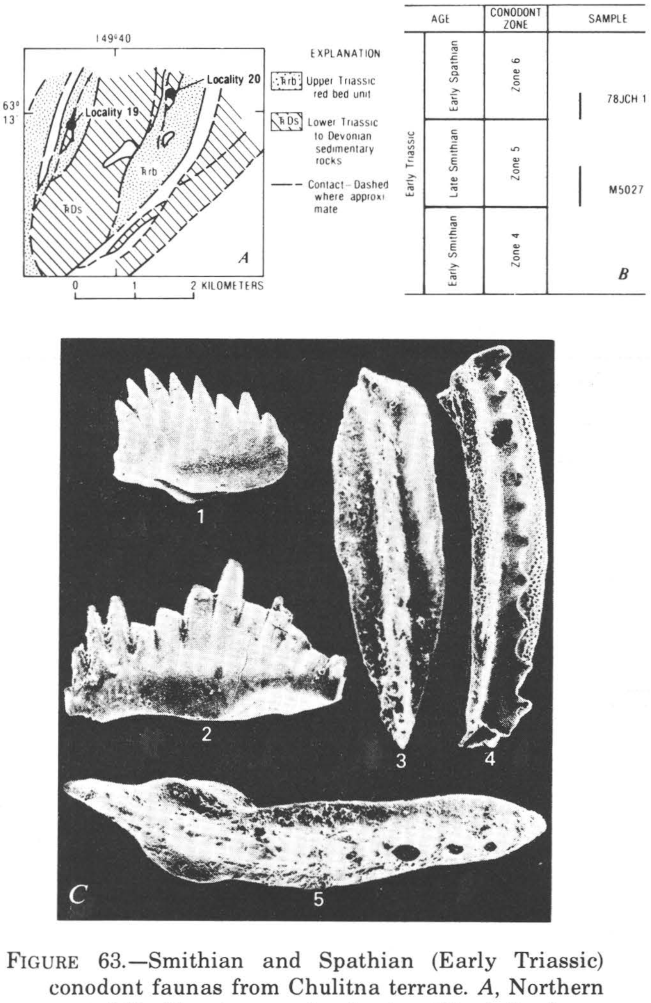

FIGURE 63.—Smithian and Spathian (Early Triassic) conodont faunas from Chulitna terrane. *A*, Northern part of Chulitna terrane showing localities (after Jones and others, 1979), *B*, age of sample; *C*, Conodonts, 1, *Neospathodus waageni* Sweet, X60; 2, "*Neospathodus*" *conservativus* (Muller), X60; 3, *Neogondolella silberlingi* Wardlaw and Collinson, X60; 4, *Neogondolella jubata* Sweet, X100; 5, *Neogondolella tozeri* Wardlaw and Collinson, X60.

ite collection, a kind of rock that commonly yields abundant conodonts. The sample was processed in acetic acid, residues were separated with heavy liquids, and abundant conodonts were obtained. The other sample (78JCH-1), from locality 19 (Silberling and others, 1978; Jones and others, 1979), represented a radiolarian, phosphatic chert about 10 cm thick that caps a meter-thick exposure of ammonite-bearing limestone. The sample was processed in dilute hydrofluoric acid, and radiolarians and a few conodonts were recovered.

The Early Triassic ammonite bed (M5027) yielded:

Ellisonia triassica Muller
Neogondolella silberlingi Wardlaw and Collinson
Neogondolella tozeri Wardlaw and Collinson
Neospathodus waageni Sweet

This conodont fauna (fig. 63*C*) indicates a late Smithian age, representing the Early Triassic Zone 5 of Collinson and Hasenmueller (1978). This fauna overlaps the range of the ammonite *Meekoceras gracilitatis* in the early late Smithian. *Neospathodus waageni* commonly dominates offshore faunas in the Smithian. *Neogondolella silberlingi* is a rare form that is found sporadically in the northern miogeosyncline of Idaho, in offshore basinal faunas, in western Nevada outboard of miogeosynclinal deposits, and in possibly deeper water faunas than those of the miogeosyncline. This fauna probably represents offshore deep-water deposition. The conodont color alteration index (CAI) is 4.0, indicating maximum host rock temperatures of 190-300° C. The color of the conodont is retained during treatment with hydrofluoric acid even though the skeletal material is changed from apatite to fluorite. The color of a conodont (below a CAI of 5.0) is due to the organic material in the interlayers of the mineral skeleton.

The Early Triassic phosphatic chert (78JCH-1) yielded:

Neogondolella jubata Sweet
"Neospathodus" conservativus (Muller)

This conodont fauna (fig. 63*C*) indicates an earliest Spathian age, representing the Early Triassic Zone 6 of Collinson and Hasenmueller (1978).

"Neospathodus" conservativus is intermediate in its distribution, and *Neogondolella jubata* tends to represent offshore faunas in the northern miogeosyncline of Idaho, Nevada, and Utah. The 4.0 CAI of this sample indicates that the rocks at the two localities had similar thermal histories.

The faunas recovered represent offshore deep-water deposition. They do not indicate a specific paleolatitude or indicate affinity with any particular faunas elsewhere in North America. They do fit well into the suite of nearshore-offshore faunas that are more typical of faunas of western North America and the Canadian Arctic than of Europe, Asia, and elsewhere.

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Reconnaissance geochemical survey of the Healy quadrangle

By Harley D. King

Geochemical samples including stream sediments, glacial debris, and heavy-mineral concentrates were collected from 600 sites in the Healy quadrangle during the 1980 field season, the first of a two-year study. Analyses of the minus-80-mesh fraction of the stream-sediment and glacial-debris samples have been completed, and some preliminary interpretation has been done. Values considered anomalous for various elements including copper, gold, lead, and zinc were found in samples from several areas. Histograms for these elements are shown in fig. 64. An area with anomalous gold values from 0.7 to 3.0 parts per million (ppm) was located in the Lignite Creek drainage south of Marguerite Creek in the north-central part of the quadrangle. The highest gold value, 3.5 ppm, is from a tributary to the East Fork of the Chulitna River in the southwest part of the quadrangle. Southeast of the Lignite Creek area with anomalous gold values is an area with anomalous lead values (from 70 to 300 ppm) in the Healy Creek and Coal Creek area. An area with anomalous copper values from 300 to 700 ppm was located in the center of the quadrangle in the upper Revine and Moose Creeks drainages. Anomalous zinc values, from 400 to 650 ppm, were found in the Last Chance Creek and Moose Creek drainages in the north-central part of the quadrangle. High zinc values were also found on tributaries to Yanert Fork of the Nenana River at the lower end of Yanert Glacier.

Geochemical exploration using englacial debris

By Edward B. Evenson[7], George C. Stephens[8], Gary C. Curtin, and Richard B. Tripp

The U.S. Geological Survey, in conjunction with Lehigh and George Washington Universities, conducted detailed sampling of the englacial debris load of Trident Glacier, Mount Hayes quadrangle, during August of 1980. The objective of the study is to determine if the sampling of debris entrained as medial moraines can be used as an effective geochemical exploration technique for the evaluation of mineral resource potential in mountainous areas drained by compound valley glaciers.

Trident Glacier, which drains a catchment area of 322 km^2, receives contributions from 31 separate, small catchment basins and has 30 well-developed medial moraines (fig. 65). Each medial moraine was sampled and studied in detail along a 3.2-km-long traverse across the compound trunk glacier (figs. 65 and 66).

From each moraine, fine-grained englacial meltout debris and water-washed sediment were collected for geochemical analysis. These samples are being processed at Lehigh University, and heavy-mineral concentrates from the fine-grained samples are being analyzed by semi-

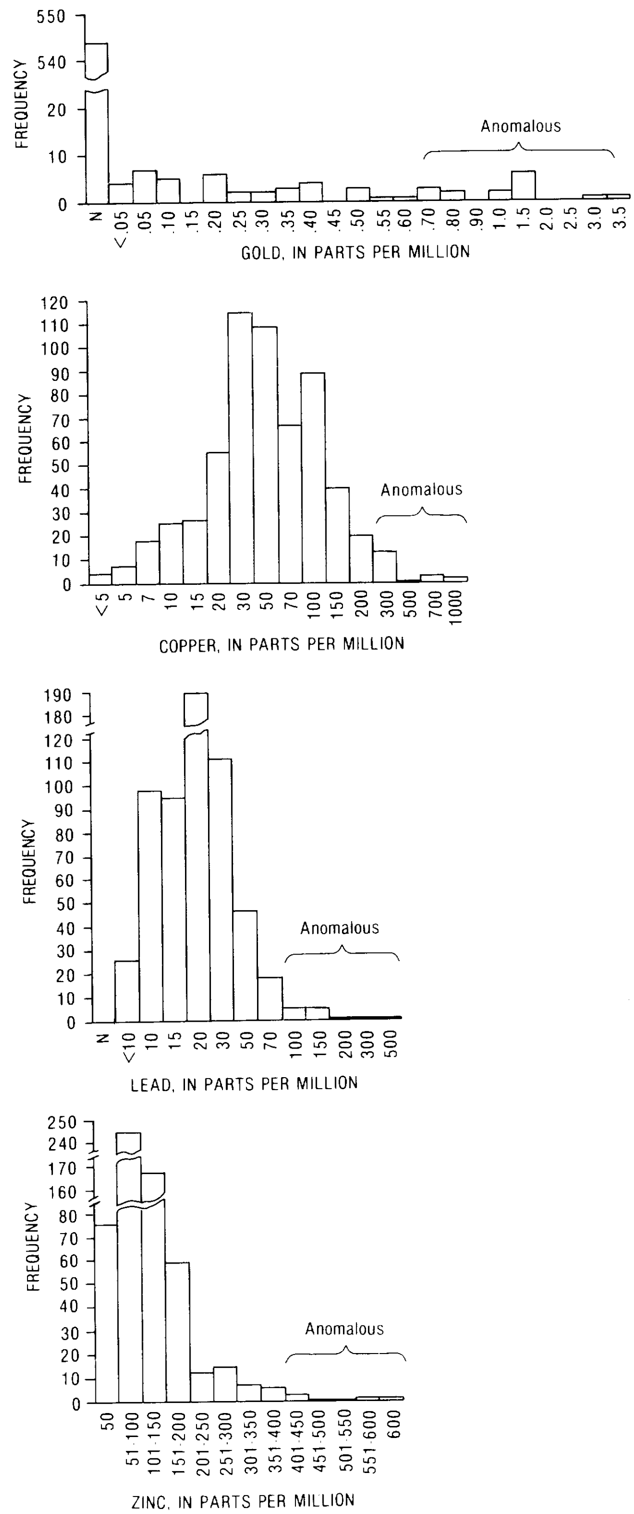

FIGURE 64.—Histograms for gold, copper, lead, and zinc in minus-80-mesh stream-sediment samples, Healy quadrangle, Alaska.

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quantitative emission spectroscopy for 32 elements. In addition to the sampling of fine-grained sediment, 100 pebbles were collected at each site, and the lithology and mineralization of 50 cobbles and 50 boulders were determined. Numerous mineralized samples were collected, and the results to date demonstrate that much information on bedrock composition and mineralization can be obtained from this type of investigation. Preliminary results indicate the occurrence of sulfide mineralization in at least two separate areas, the locations of which can be determined by moraine tracing. When geochemical and lithologic analyses are completed, the information gained from this study will be compared to that generated by the Alaskan Mineral Resource Assessment Program geochemical study on the same area to determine the relative sensitivity and applicability of each mineral exploration sampling program and to determine the applicability of detailed moraine sampling as an aid in geologic mapping.

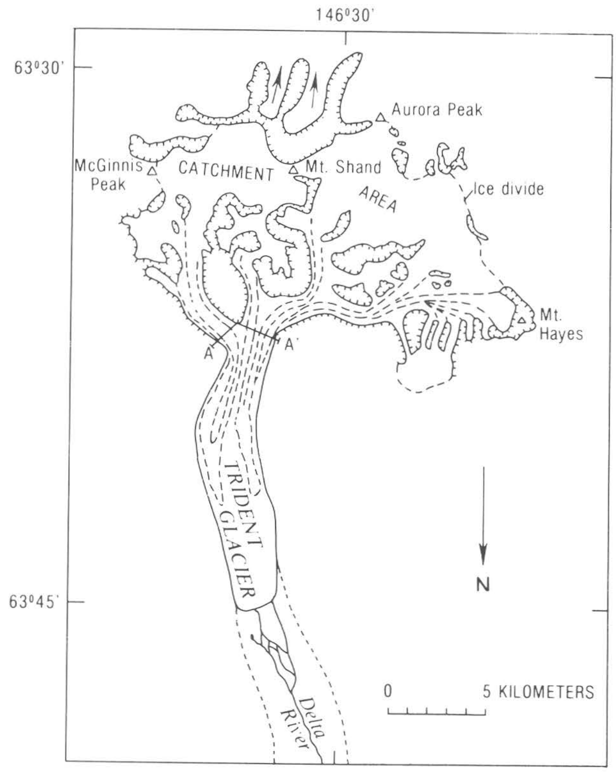

FIGURE 65.—Map of Trident Glacier showing some of the prominent medial moraine systems. A total of 30 medial moraines were identified and sampled along traverse A-A.' Arrows show direction of glacial movement.

Quaternary history of vegetation in the north Alaska Range

By Thomas A. Ager

Pollen samples collected in 1977 and 1979 from lacustrine-sediment cores from six localities (fig. 67) and numerous outcrops in the north Alaska Range (Mount McKinley National Park and adjacent areas) have been analyzed to permit the first reconstruction of vegetation history of the region during the Wisconsinan glacial interval and the Holocene. It is now possible to construct an outline of the vegetation history of the north Alaska Range during the past 40,000 years or more. A continuous pollen record from Eightmile Lake (fig. 67) provides a history of vegetation for the northern foothills back to about 15,000 years B.P. Cores from five other localities have basal ages ranging from 13,000 to 5,700 years B.P.

Only a few samples from the Alaska Range appear to date to middle Wisconsinan interstadial time. Some of those samples are associated with infinite radiocarbon ages, so they may represent vegetational conditions that existed prior to the middle Wisconsinan interstadial. The available fragmentary evidence suggests, however, that the north Alaska Range region was probably treeless during middle Wisconsinan time, whereas today spruce trees occur up to 1000-m altitude. The "mid-Wisconsinan" vegetation appears to have been an herbaceous tundra with scattered low shrubs of willow and dwarf birch; spruce and other trees grew in at least some sites in the Tanana Valley at that time (Matthews, 1974; Ager, 1982).

During the late Wisconsinan glacial interval, the vegetation of the Alaska Range was sparse, discontinuous herbaceous tundra growing in those areas not covered by glaciers, perennial snowfields, or barren boulder fields and outwash plains. The climate during this glacial interval was generally arid with very cold, windy winters; summers were probably significantly cooler than they are at present in this region. The most extensively vegetated areas of the region were the northern foothills, most of which lay beyond the limit of late Wisconsinan glaciers. The herbaceous tundra in the Alaska Range appears to have been rather similar to the lowland vegetation in the Tanana Valley during the late Wisconsinan glacial interval (Ager, 1975). This close similarity of vegetation

FIGURE 66.—Vertical aerial photograph showing location of sample traverses (see fig. 65) across medial moraines of Trident Glacier.

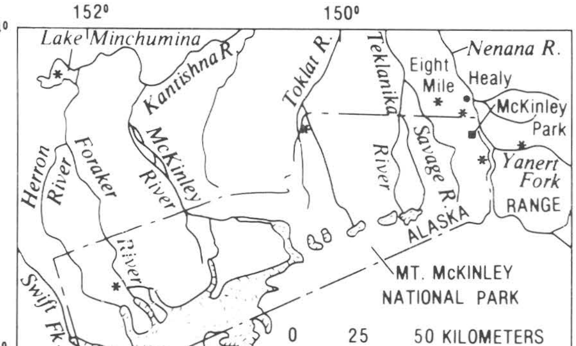

FIGURE 67.—Mount McKinley National Park and adjacent areas of Alaska Range showing sample sites (stars).

from both mountains and lowlands suggests that altitudinal zonation of vegetation was much less pronounced than today or any other time during the Holocene. Variability within this regional herb tundra was probably controlled primarily by relative availability of moisture.

By roughly 14,000 years ago, a vegetation change began in the Tanana Lowland to the north, probably in response to a climatic shift to warmer summers and probably increasing availability of moisture (Ager, 1975). This shrub tundra vegetation consisted of low shrubs such as dwarf birch, willows, ericales and herbaceous plants such as sedges, grasses, and assorted forbs. Radiocarbon dates from Eightmile Lake indicate that shrub tundra began to invade the northern foothills of the north Alaska Range (fig. 67) up to an altitude of at least 700 m as early as 13,500 years ago.

By about 11,000 years ago, summers became sufficiently warm and probably dry enough to permit the rapid spread of poplar trees (*Populus*) in the Tanana Lowland and many other areas of Alaska and Yukon Territory. One of the earliest known appearances of *Populus* in the Alaska Range occurred roughly 10,000 years ago in the Delta River headwaters area (Tangle Lakes area), according to pollen and macrofossil data reported by C. E. Schweger (unpub. data). *Populus* spread from the Tanana Valley to at least the northern foothills of the Alaska Range in the Nenana Valley at about the same time.

The next significant regional change in vegetation began about 9,500 years ago in the Tanana Valley, when spruce rapidly spread throughout the lowlands (Ager, 1975). The area around Delta River was an early invasion route into the Alaska Range, where spruce are known to have spread to the headwaters of the Delta River at Tangle Lakes by about 9,100 years ago (C. E. Schweger, unpub. data). In the Nenana Valley and areas to the west of that drainage, however, the spruce invasion appears to have been delayed until perhaps 7,500 years ago or later. Spruce and alder invaded almost simultaneously in the Nenana Valley area, whereas spruce preceded alder in the Tanana Valley by about 1,000 years.

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Holocene marine terraces and uplift history in the Yakataga seismic gap near Icy Cape, Alaska

By George Plafker, Travis Hudson, Meyer Rubin, and Kirk L. Dixon

A sequence of as many as three marine terraces extends for 52 km between Icy Cape and Yakataga along the mainland coast of the Gulf of Alaska (Taliaferro, 1932; Heusser, 1960). These terraces were studied during 1975, 1979, and 1980 as part of a project to elucidate the uplift history of the area and to deduce the nature and recurrence interval of paleoearthquakes that may have produced the terraces. The tectonic and seismic history of the Yakataga area is of special current interest because the area lies in the seismic gap between the 1964 Alaska earthquake focal region to the west and the 1958 Lituya Bay earthquake focal region to the east (Lahr and others, 1979; Lahr and Plafker, 1980; Page, 1974). Structurally, the area lies within the disturbed Pamplona zone of late Cenozoic folds and faults that connects the east end of the Aleutian Trench with the Fairweather transform fault (Plafker and others, 1978; Lahr and Plafker, 1980).

Clear-cut logging operations since 1976 have made the part of the terrace sequence east of Munday Creek near Icy Cape the most readily accessible and best exposed, and our study of the terraces and related late Holocene glacial and alluvial sequences was concentrated in this area in 1979 and 1980 (Plafker and others, 1980). A more comprehensive report on the entire area will be prepared after the results of carbon-14 dating of organic samples collected in 1980 become available.

The study area consists of a low foreland, ranging in width from 0.5 km at Munday Creek to 6 km at Icy Cape. It is backed by a rugged mountain front, as much as 1,500 m high, made up of late Cenozoic marine strata (fig. 68). The foreland in the Icy Cape area is made up of a complex of neoglacial fluvial, lagoonal, beach, and terrace deposits that extend inland from the coast to about 60 m above sea level. The three marine terraces have average beach-angle elevations above mean higher high water (MHHW) of approximately 52, 24, and 16 m. Beach-angle elevations are estimated to be accurate within about 3 m for the high terrace (stage I), 2 m for the intermediate terrace (stage

II), and 1 m for the low terrace (stage III). Elevation uncertainties result from the extensive mantling of the beach angles by talus, slopewash, and alluvial deposits together with initial variations in the beach angles. The stage III terrace surface is the only one that is backed by a continuous, well-developed sea cliff. The higher terraces are considerably more dissected, and their associated sea cliffs are commonly degraded. Discontinuous, irregular benches of uncertain origin also occur locally along the mountain front at elevations ranging from 65 to 200 m.

Terrace development and modification in the area are intimately related to neoglacial fluctuations of Guyot and Beare Glaciers and to accompanying major variations in sediment supply to the coast. The absence of recognizable marine terraces east of Big River and the broad foreland in the Icy Cape area appears to result from the presence of extensive neoglacial morainal and fluvioglacial deposits that inhibit development of wave-cut abrasion surfaces and bury any older surfaces that may have been present. The extensive outer terminal moraine that extends northeastward from Icy Cape (fig. 68) is a marine glacial till that was deposited by Guyot Glacier and subsequently has been elevated as much as 13 m above present sea level. The inner terminal moraine shown on figure 68 is a pitted deposit as much as 14 m above sea level that was probably deposited

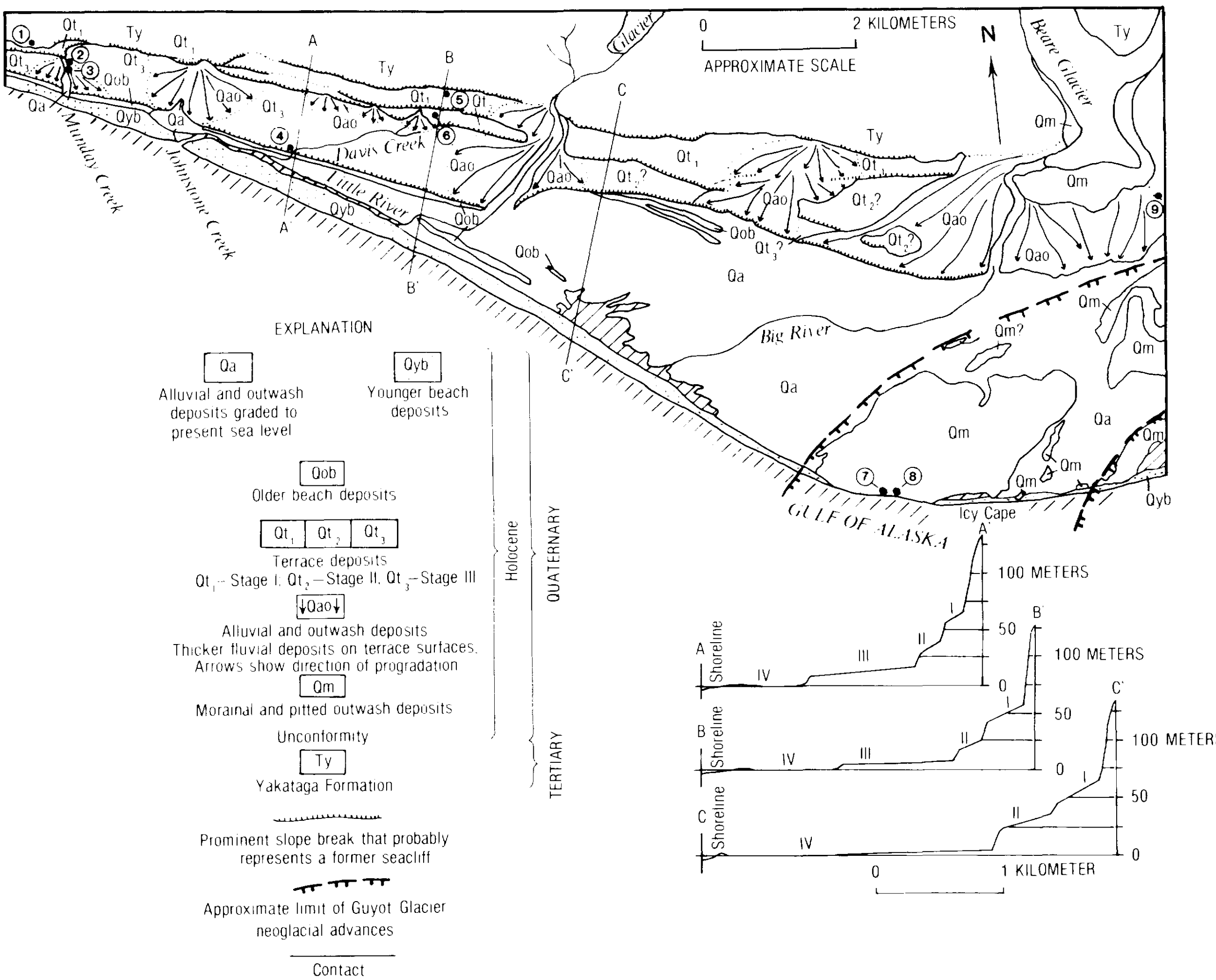

FIGURE 68.—Geological map and profiles of marine terraces and related features in Icy Cape area. Black dots with numerals refer to radiocarbon-dated samples shown in table 7 and Roman numerals refer to marine terrace stages described in text.

subaerially. It marks the position of the most recent neoglacial advance of Guyot Glacier.

Data for selected radiocarbon-dated peat and wood samples collected prior to 1980 from the marine terraces and benches above the terraces, as well as from glacial and fluvioglacial deposits, are shown in table 7. Additional samples collected during the 1980 field season are being dated. Preliminary ages for the stage I, II, and III marine terraces are about 5,000, 2,500, and 1,300 years B.P., respectively. They were determined by radiocarbon dating of peat collected from silty clay near the base of the unconsolidated deposits that mantle the terrace surfaces to depths of as much as 15 m. The fine-grained sediment in which the peat occurs presumably was deposited in quiet water at or near the abrasion surface after emergence. Although the peat dates represent minimum ages for terrace emergence, they could be close to the actual terrace age. The high benches above the terrace have been dated with peat from the base of cores drilled through muskeg on the bench surfaces. The oldest of these peats are 10,800 years B.P. (Heusser, 1960) and 12,000 years B.P. at elevations of 70 m and 64 m, respectively, and they represent minimum ages for these benches. The age of the outer terminal moraine of Guyot Glacier is between 1,200 and 1,600 years, on the basis of dating of wood embedded within the marine glacial deposits of Icy Cape. Historic records indicate that Guyot Glacier was at or near the position of its inner terminal moraine between 1886 and 1910 (Plafker and Miller, 1958). Beach ridges close to present sea level that occur along the inner margin of the lowland that is graded to present sea level (fig. 68) have been approximately dated by growth ring counts of trees growing on the ridge surfaces. These ridges, which are 2 m or less above MHHW, support trees with maximum ages of approximately 90 years east of Little River and roughly 65 years west of Little River.

Figure 69 shows the ages of the oldest radiometrically dated samples from the three marine terraces, average shoreline elevations, and the generalized stratigraphic sequence at the sample localities. Episodic emergence of these terraces during a period when eustatic sea level was either rising or was at its present level requires that the terraces be primarily tectonic in origin. Their heights, however, may be accentuated in the Icy Cape area by an isostatic component related to deglaciation.

Geologic data from the Icy Cape area lead to the following tentative conclusions:

1) Indicated average uplift rate has been about 1.05 cm/yr during the past 5,000 years. Approximately the same uplift rate would be indicated for the two samples from the benches above the terraces if their elevations are corrected for the postglacial eustatic sea level rise (about 57 m in the past 12,000 years). This rate is remarkably close to that of the Middleton Island and Lituya Bay terrace sequences (Plafker and Rubin, 1978; Hudson and others, 1976). The rate is an order of magnitude less than the uplift rate in the Icy Cape area as deduced from study of tree ages on the terraces by Beavan and others (1979).

2) The time between uplift of the stage II and III terraces is about 1,150 years, and the uplift step is roughly 8 m.

3) The interval of approximately 25 m between the stage I and II terraces probably

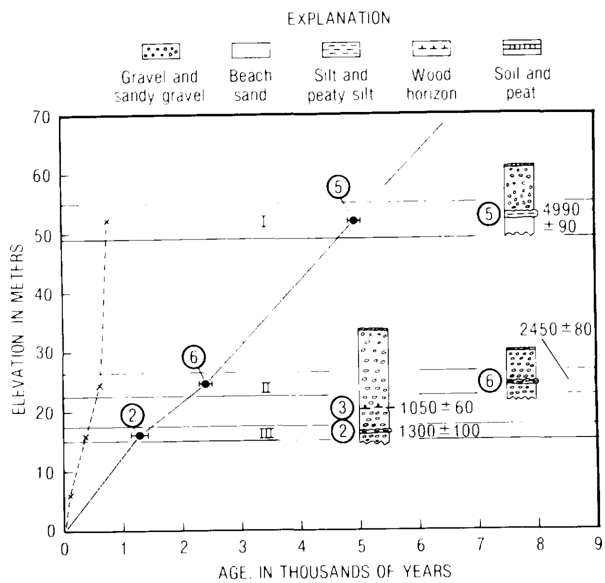

FIGURE 69.—Diagram showing terrace height and ages of dated samples. Locations of numbered samples shown in figure 68; sample age data presented in table 7. Columns show generalized stratigraphic sequence at sample localities indicated. Elevations relative to present MHHW; ages in radiocarbon years. Solid line indicates approximate average uplift rate indicated by this study; dashed line is rate postulated by Beavan and others (1979).

Table 7.--*Data for selected radiocarbon-dated samples from the vicinity of Icy Cape*

Map sample number (fig. 68)	Laboratory sample number	Age (radiocarbon yr B.P.)	Elevation in meters (above MHHW)	Year collected	Collector	Material dated	Comments
1	I(AGS-10)	10,820±420	70	1956	Heusser	Peat	Peat from base of muskeg (Heusser, 1960) on high bench.
2		1,300	16	1980	Rubin	Peat	Peat from near beach angle on stage III terrace.
3	W-369	1,050±160	20	1951	Miller	Wood	Logs in outwash on stage III terrace (Rubin and Alexander, 1958).
4	W-4512	1,210±70	6.5	1979	Plafker	Peat	Oldest peat interbedded with 3-m-thick silt section on stage III terrace.
5	W-4195	4,990±90	52	1979	Plafker	Peat	Peat overlying marine deposits on stage I terrace.
6	W-4485	2,450±80	24	1979	Plafker	Peat	Peat overlying possible marine deposits on stage II terrace.
7	W-374	1,200±160	0	1951	Miller	Wood	Wood in outer Guyot Glacier moraine (Plafker and Miller, 1958).
8	W-4510	1,630±70	0	1979	Plafker	Wood	Wood in outer Guyot Glacier moraine.
9	W-3164	12,010±350	64	1974	Hudson	Peat	Peat from base of muskeg on high bench.

represents more than one uplift step, but the intermediate terraces either have not been recognized or have been removed by erosion.

4) There is no evidence of measurable post-stage III uplift that could be associated with the 1899 earthquake as was reported by Beavan and others (1979). Abandonment of the sea cliff that marks the inner margin of the foreland between Johnstone Creek and Big River predates the 1899 earthquake east of Little River. It is likely that progradation of the foreland resulted primarily from an increase in sediment supply corresponding to the younger advance of Guyot Glacier which culminated by 1886. More recently, the coast has become relatively starved of sediment with the retreat of Guyot Glacier since 1904 (Plafker and Miller, 1958). As a result, most of the coast from Icy Cape westward is undergoing rapid erosion, and shoreward retreat of Icy Cape amounts to almost 1 km in the 30-year interval from 1948 to 1978.

5) The last uplift step of about 16 m that formed the stage III terrace occurred about 1,300 years ago. Thus, if the 1,150 years between the stage II and III terraces represents an approximate recurrence interval for major tectonic uplift steps—and presumably large earthquakes—the data suggest that the next uplift event may be overdue.

6) This study, together with previous terrace data from Middleton Island (Plafker and Rubin, 1978), suggests that tectonic uplift rates in both the Icy Cape and Middleton areas lag behind the long-term average rates. Future uplift of these areas most probably would occur during one or more major earthquakes along the convergent plate boundary extending northeastward from the east end of the Aleutian Trench along the Pamplona zone to the Fairweather transform fault. Because major earthquakes along this boundary are also likely to be accompanied by large vertical displacements on the continental shelf, they could generate seismic sea waves capable of causing coastal damage far from the earthquake focal region.

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Depositional system approaching maturity at Portage Flats

By Susan R. Bartsch-Winkler and Holly C. Garrow

The area surrounding Portage was lowered into the intertidal area during the 1964 Alaska earthquake (Richter magnitude=8.5). Ground subsidence, resulting from tectonic lowering of the land and compaction of the silts at depth totaling 2.4 m (Ovenshine and others, 1976), generated a pulse of deposition in the intertidal zone. By 1973, the new, measurable deposit, the Placer River Silt (Ovenshine and others, 1976) covered an area of 18 km^3 and contained a minimum of 20×10^6 m^3 of sediment. Other workers have calculated that the sedimentation cycle begun by the earthquake will be 98 percent complete by the year 1989 (Ovenshine and Kachadoorian, 1976); extrapolation of the sedimentation curve to 1980 shows that the sedimentation cycle is approximately 91 percent complete (R. Kachadoorian, oral commun., 1980). Field observations reveal that the vegetation thriving on Portage Flats at the time of the 1964 earthquake is rejuvenating and reseeding in the area. Our assumption is that these effects are primarily due to depositional processes; however, possible effects from tectonic rebound of the area cannot be identified with data currently available.

A sedimentological study has been underway at Portage Flats since 1973. One part of this study produced a textural map of the silt that covered the area in 1974 (Bartsch-Winkler and others, 1978, fig. 5). This textural map was used in comparing samples collected and analyzed in 1980 to see what changes in the texture of the sediments could be discerned. In June 1980, samples of the upper 1 cm of sediment were collected from 16 Portage Flats sites on the seaward side of the railroad and highway embankments southwest, west, and northwest of the junction of the Alaska Railroad-Seward Line with the Whittier Spur and from the 1974 profile east of the Seward Highway on the west edge of the deposit (fig. 70). The samples were weighed, sieved through a 0.0625-m screen, and reweighed, and the percentage of sand in each sample was calculated. The results show that all of the samples collected in 1980 contain less than 5 percent sand, in contrast to similar samples collected at the same locations in 1974 which contained in excess of 10 percent sand. The sediment analyzed in 1980 is texturally equivalent to the finest sediment deposited in the Portage area in 1974, then found only in areas landward of the highway and railroad embankments.

Perhaps the most convincing evidence that stabilization is nearly complete may be found in the vegetation now growing in the area. By 1974, grasses and a few willows began to return to the Portage Flats area (Ovenshine and Kachadoorian, 1976). In 1980 we returned to the sedimentation sites monitored in 1974 and observed that grasses covered areas that were actively intertidal and barren of vegetative matter in 1974. Also, cottonwoods that were thought killed by salt-water incursion and buried in intertidal silt have regenerated from the bark of the older cottonwood trees, especially from the basal parts of the trunks that were buried and therefore were able to retain moisture. Large, mature trees (estimated to be 10 m tall) that are located on silt between the junction of Placer and Portage creeks seaward of the highway embankment and were thought dead after 1964 have, in places where bark was retained, sprouted new growth on branches 6 to 9 m up (fig. 71). In addition, young spruce trees have begun to appear, probably from spruce cones buried in the silt in 1964 or carried to the new deposit by birds and animals. The spruce trees are not yet as numerous as the grasses, willows, and cottonwoods.

We conclude that growth of vegetation, begun by the salt-tolerant grasses, has accelerated deposition of fine-grained sediment by entrapping it in roots, stems, and leaves. Once the grasses began to flourish, Portage Flats was in the final, rapid reconstruction phase of the sedimentation cycle.

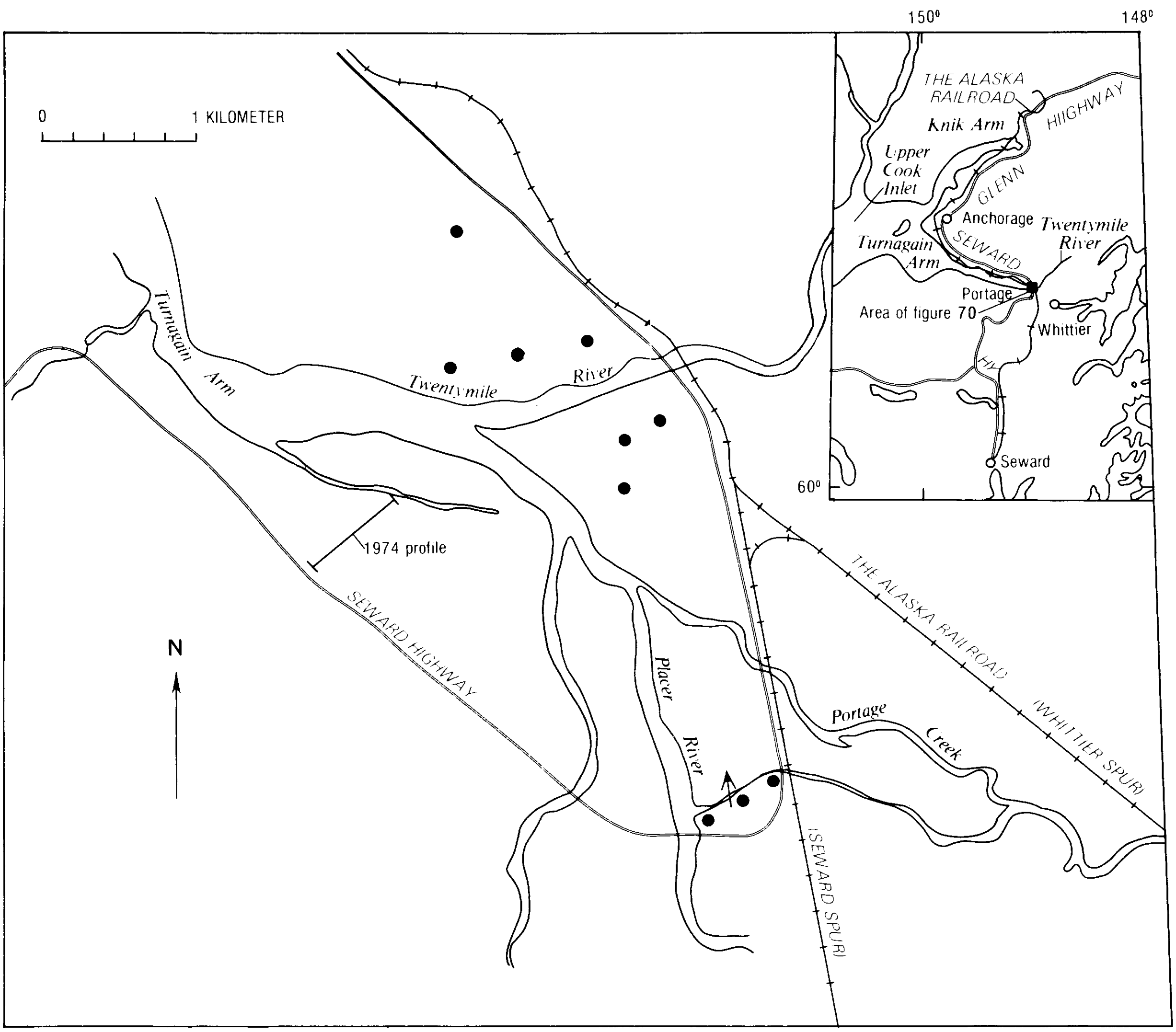

FIGURE 70.—Location of Portage at head of Turnagain Arm (inset map) and locations of samples analyzed in 1980 (dots). Arrow indicates approximate location of cottonwood trees in figure 71.

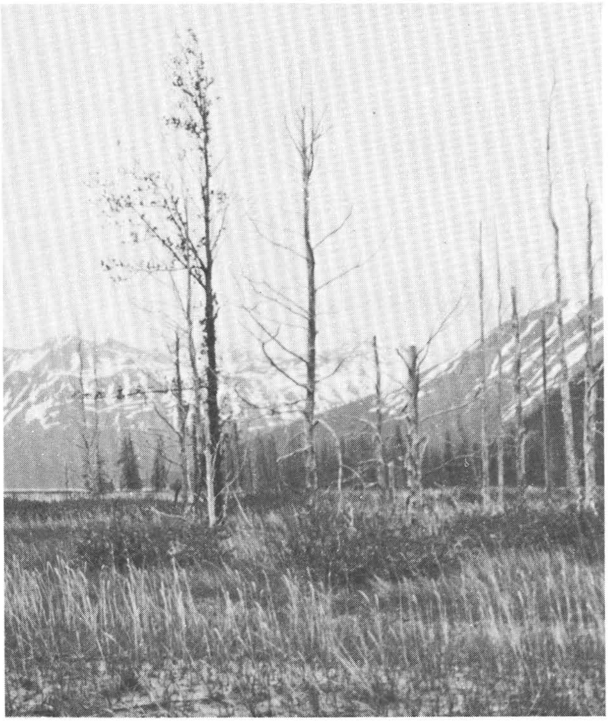

FIGURE 71.—Example of rejuvenating vegetation on Portage Flats, 1980. Note low clumps of alders and new growth on largely bare, 10-m tall cottonwood trees.

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Water resources studies of the Anchorage area

By Derrill J. Cowing and Richard Emanuel

A systematic study of the water-quality characteristics of streams within the Municipality of Anchorage was begun in 1980. The first stream being studied is Campbell Creek; others that will be studied include Chester, Ship, and Rabbit Creeks. Data collected on Campbell Creek show that water flowing from the non-urbanized headwater part of the basin is of excellent quality. As the stream traverses the urbanized part of the basin, water quality is degraded by urban runoff and carries increased concentrations of suspended sediment, bacteria, nutrients, and other materials. Data collected for this program are being used to calibrate and verify rainfall-runoff water-quality models that the Municipality of Anchorage has developed for use in their long-range water-quality management program.

Hydrogeologic studies to provide information for land-use planning in the Anchorage area are continuing. Investigations of the geology and hydrology of the Potter Creek area are almost finished. The Potter Creek studies complete coverage of the Hillside area of Anchorage begun by Dearborn and Barnwell (1975). Similar studies of hydrology for land-use planning in the Peters Creek area began in 1980.

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SOUTHEASTERN ALASKA
(Figure 72 shows study area discussed)

New paleontologic age determinations from the Taku terrane, Ketchikan area, southeastern Alaska

By Norman J. Silberling, Bruce R. Wardlaw, and Henry C. Berg

The Taku terrane (fig. 73*B*) is one of several major tectonostratigraphic terranes recognized in southeastern Alaska (Berg, Jones, and Coney, 1978), and is host to numerous undated stratabound massive sulfide deposits (Berg, 1979). For most of its extent, it is fault bounded on the west by Upper Jurassic and Cretaceous flysch and volcanic rocks of the Gravina-Nutzotin belt which separates the Taku from the Alexander terrane. On the east the Taku is bounded by the metamorphic and plutonic crystalline complex of the Tracy Arm terrane. Although grossly dissimilar in its geologic and structural history from adjoining terranes, the Taku is difficult to characterize stratigraphically. Its heterogeneous rocks are complexly deformed, pervasively metamorphosed, and sparsely fossiliferous. Heretofore, only general-

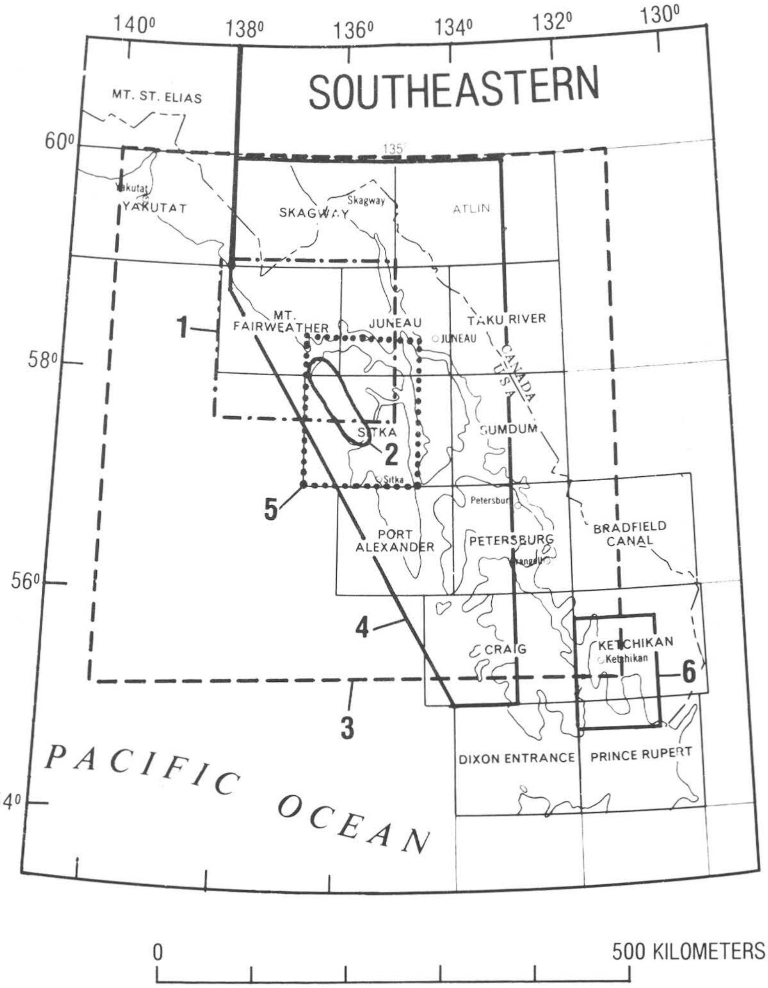

FIGURE 72.—Areas and localities in southeastern Alaska discussed in this volume. Authorship and applicable figures and tables relating to the numbered areas are listed below. 1, Decker and Plafker, figures 74 and 75, table 8; 2, Hessin; 3, Hudson, Dixon, and Plafker, figure 78, tables 10 and 11; 4, Hudson, Plafker, and Dixon, figure 77; 5, Karl, S. M., Decker, and Johnson, figure 76, table 9; 6, Silberling, Wardlaw, and Berg, figure 73.

ized late Paleozoic and Triassic ages have been obtained from the southern part of the Taku terrane near Ketchikan.

In July 1980, more definitive collections of Permian conodonts and brachiopods and of Triassic mollusks were made from the Taku terrane (fig. 73*A*) by H. C. Berg, N. J. Silberling, D. L. Jones, and P. J. Coney. Conodonts from black crinoidal marble intercalated with phyllite and felsic metatuff at locality 80-S-323, about 0.5 km south of Coon Cove, include *Neogondolella idahoensis* (Youngquist, Hawley, and Miller) and *Hindeodus* sp. of Leonardian (late Early Permian) age. In keeping with this age, poorly preserved brachiopods from this same locality can be assigned to *Stenocisma* sp. and *Neospirifer*? sp.

About 4.5 km south of Coon Cove, at locality 80-S-332, ammonites and fragments of halobiid bivalves (probably *Daonella*) are preserved as crushed, but undistorted molds in small black, carbonaceous and siliceous concretions within foliated limestone and slate[9]. The ammonites include *Lobites* cf. *L. pacianus* McLearn, *Joannites* sp., and *Meginoceras*? sp. diagnostic

[9]Field studies in 1981 (H. C. Berg, unpub. data) disclosed an occurrence of these fossiliferous Middle Triassic strata about 0.5 km south of Coon Cove. There, dark gray concretionary limestone and slate together with pyritic phyllite and rusty-weathering felsic metatuff structurally underlie Permian(?) crinoidal marble.

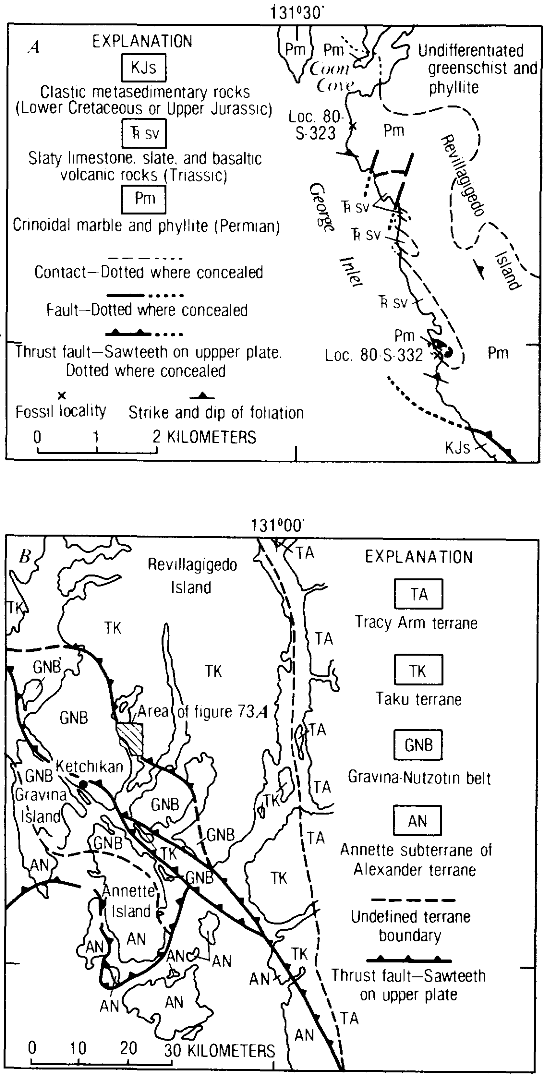

FIGURE 73.—Maps showing fossil localities and stratigraphic terranes discussed in text. *A*, Geologic sketch map showing fossil localities within Taku terrane. *B*, Index map showing location of map area among the tectonostratigraphic terranes in the Ketchikan area (modified from Berg, Jones, and Coney, 1978).

of a late Ladinian (latest Middle Triassic) age. The provisional identification of the Late Triassic genus *Halobia* from this locality by Silberling *in* Berg, Elliot, and others (1978) is incorrect. Extrapolation of a Triassic age to all of the slaty limestone and basalt unit (Ŧsv) seems warranted, even though the fossiliferous Triassic beds are only a few meters thick and separated along the shoreline from basaltic rocks by a klippe of limestone like that of the crinoidal marble-phyllite unit from which the Permian fossils were obtained.

Although the stratigraphy of the Taku terrane is still imperfectly known, these new age determinations reinforce the pronounced stratigraphic differences between the Taku terrane and the Annette subterrane of the Alexander terrane as portrayed by Berg, Jones, and Coney (1978). Briefly stated, in the Annette subterrane Upper Triassic strata, ranging from early Norian to middle Norian, rest unconformably on Devonian and older metamorphic and plutonic rocks; strata correlative with those of the two fossiliferous localities within the Taku terrane are not represented.

The new fossil discoveries also have potentially significant metallogenic implications for the stratabound mineral deposits in the Taku terrane. Many of these deposits are localized in rusty-weathering quartz-muscovite-calcite-pyrite schist similar to the felsic metatuff intercalated with the Permian and Triassic strata near Coon Cove. The original age of these metamorphosed syngenetic deposits and host rocks is unknown, but the similarity in lithology suggests that it may be Permian or Triassic. If so, many, and perhaps most, of the numerous stratabound massive sulfide deposits in the Taku terrane may be parts of a heretofore unrecognized metamorphosed and structurally dismembered Permian or Triassic metallogenic province that stretches for more than 300 km from Ketchikan to Juneau.

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Correlation of rocks in the Tarr Inlet suture zone with the Kelp Bay Group

By John E. Decker and George Plafker

Independent field examinations during 1980 by Plafker and by Decker of critical outcrops on northern Chichagof Island, the Inian Islands, and in Glacier Bay National Monument have led us to conclude that: 1) the Kelp Bay Group on Chichagof Island correlates with rocks of the Tarr Inlet suture zone in Glacier Bay National Monument, and 2) the Border Ranges fault delineates the northeastern boundary of the Tarr Inlet suture zone. We use Kelp Bay Group in the restricted sense suggested by Karl and others (following paper in this volume), referring specifically to the melange of western Chichagof and Baranof Islands.

Recently published correlations between rock units of Chichagof Island and Glacier Bay National Monument along with the correlations suggested in this report are shown schematically in figure 74. Historically, the Tarr Inlet suture zone was defined by Brew and Morrell (1978) based on four criteria: 1) east of the eastern limit of homogeneous turbidites, 2) its structural complexity, 3) west of the western limit of foliated Cretaceous granitic rocks, and 4) west of the western limit of thick Paleozoic carbonate rocks. Rock types reported by Brew and Morrell (1978) within the Tarr Inlet suture zone consist mainly of Permian(?) phyllite, slate, conglomerate, and chert in units that are continuous on kilometer scale, but have many internal discontinuities of individual lithic units; subordinate greenstone, greenschist, and other metavolcanic rocks that are apparently more coherent internally; and minor gray marble in lenses that range in size from a few meters to as much as 5 km long and several tens of meters thick. These country rocks are intruded by elongate bodies of highly foliated biotite-hornblende quartz diorite of inferred Cretaceous age (Brew and Morrell, 1978). Based on Brew and Morrell's brief descriptions, Berg and others (1978) correlated the suture zone rocks with lithologically similar rocks of Wrangellia (Jones and others, 1977). Assuming the Wrangellia-Tarr Inlet suture zone correlation, Berg and others (1978) and Plafker and Campbell (1979) placed the Border Ranges fault along the southwestern margin of the Tarr Inlet suture zone (fig. 74*A*). Brew and Morrell (1979), also assuming a Wrangellia-Tarr Inlet suture zone correlation and applying their fourfold criteria, extended this correlation to include

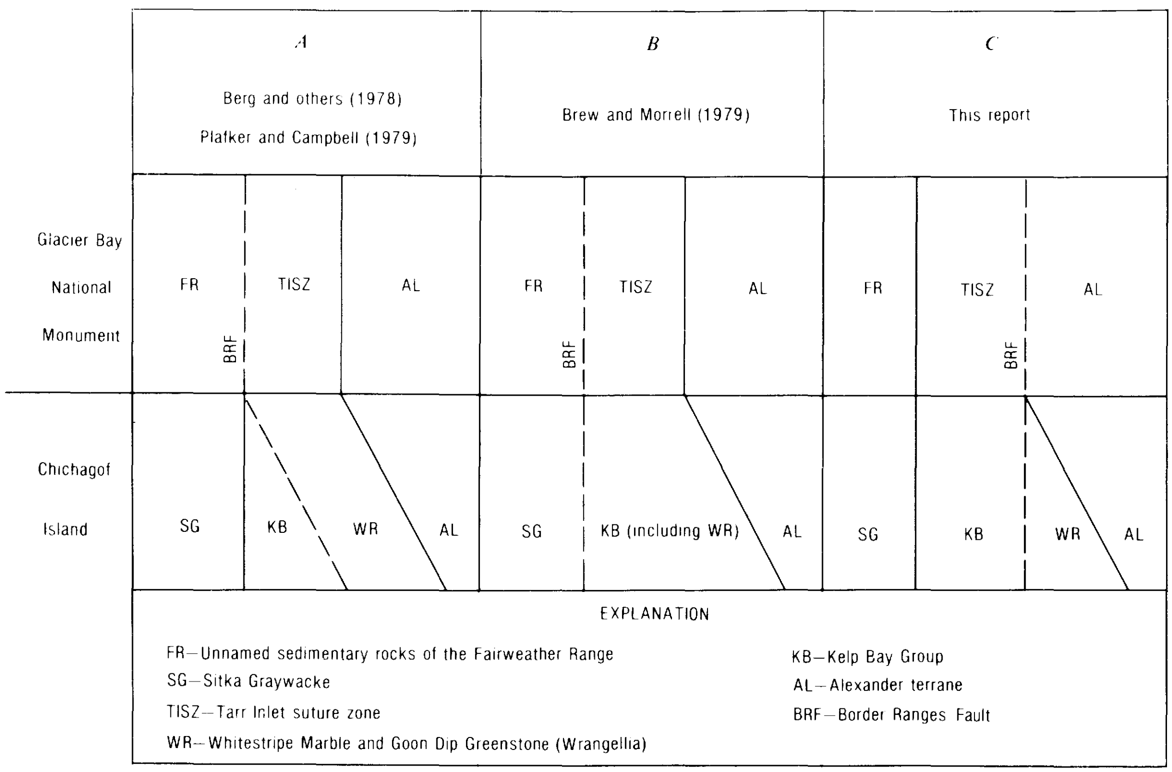

FIGURE 74.—Schematic correlation diagrams showing previous interpretations and interpretation proposed in this report.

the Kelp Bay Group on Chichagof and Baranof Islands (fig. 74*B*).

In this report, we suggest that rocks of the Tarr Inlet suture zone are correlative with only the Kelp Bay Group on Chichagof Island, and that the Wrangellia section is not present in Glacier Bay National Monument but does occur south of Lisianski Inlet (fig. 74*C*, 75). Furthermore, we suggest that the Border Ranges fault follows the northeastern margin of the Tarr Inlet suture zone where it juxtaposes Early Cretaceous melange to the southwest against Paleozoic rocks of the Alexander terrane to the northeast. The extension of the Border Ranges fault through Glacier Bay National Monument along the northeastern margin of the Tarr Inlet suture zone (fig. 74*C*, 75) is, thus, in contrast to the southwest position proposed by Berg and others (1978), Plafker and Campbell (1979), and Brew and Morrell (1979).

Our field examinations of shoreline exposures in Glacier Bay National Monument and on northern Chichagof Island indicate that: 1) the Tarr Inlet suture zone consists largely of rock types that are characteristic of the Kelp Bay Group and that they are unlike rocks comprising

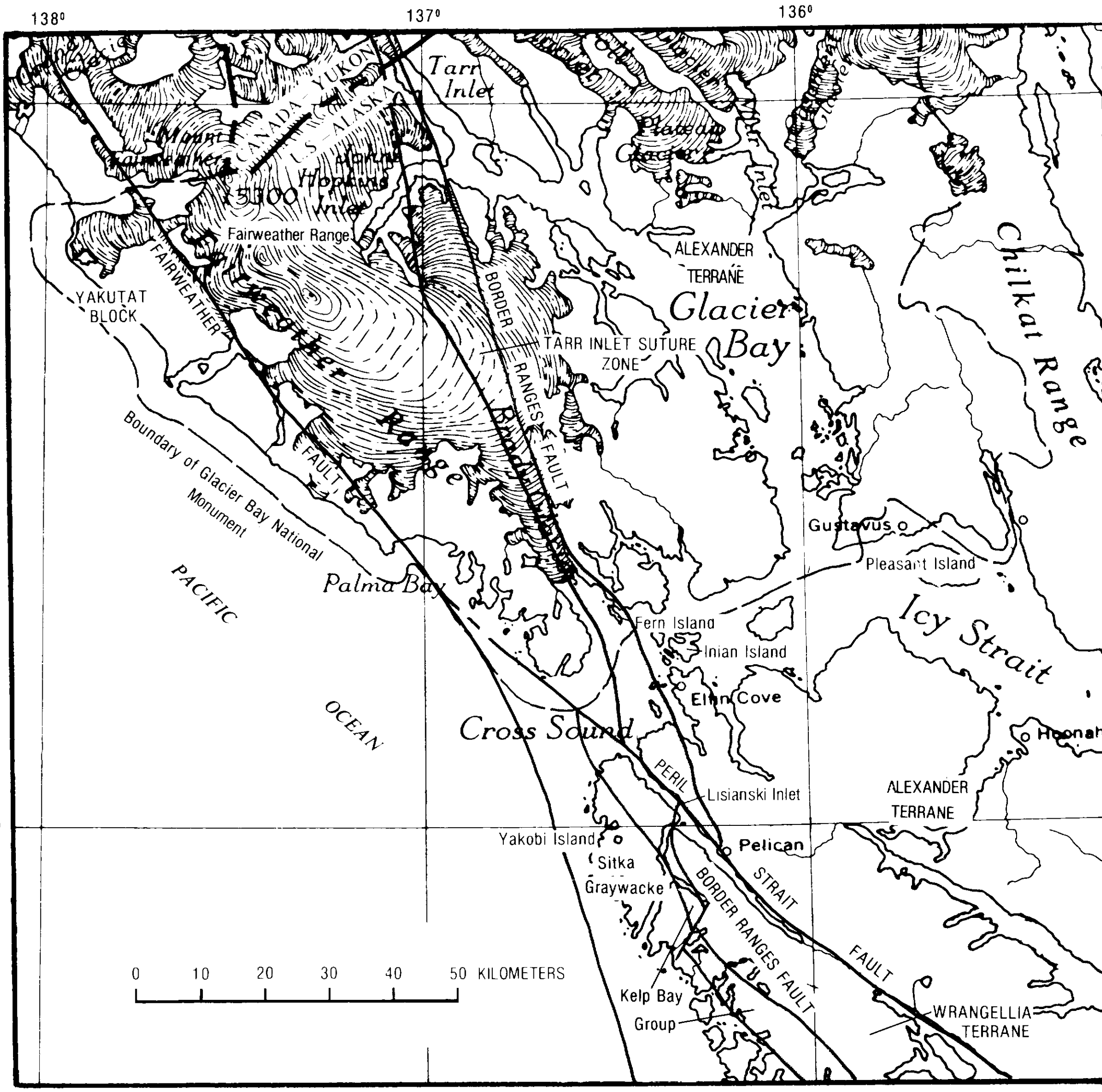

FIGURE 75.—Generalized geologic map of northern Chichagof Island and Glacier Bay National Monument, showing distribution of major rock units or terranes, position of Border Ranges fault, and correlation proposed in this report.

the Wrangellia section on Chichagof Island, and 2) the easternmost exposures of the Kelp Bay Group can be traced continuously through northern Chichagof and nearby smaller islands, trending across Cross Sound toward Fern Island, which was mapped by Brew and others (1978) as part of the Tarr Inlet suture zone (fig. 75). Key rock types used in tracking the Kelp Bay Group into the Tarr Inlet suture zone in Glacier Bay National Monument are thinly interlaminated or chaotically intermixed green metatuff and black argillite that locally enclose blocks of bedded radiolarian chert, greenstone, or graywacke. Fern Island is underlain almost entirely by these rock types, particularly tuff, argillite, and bedded chert. At Johns Hopkins Inlet in northern Glacier Bay National Monument, intermixed tuff and argillite is present locally but is volumetrically subordinate to coherent zones of argillite and graywacke, massive to weakly foliated greenstone, sheared, highly altered, garnet-bearing diorite, that has a potassium-argon age of 96.7±5.5 m.y. (no.1, table 8), and probably intrusive fresh biotite-bearing diorite with a potassium-argon age of 58.6± 5.2 m.y. (no. 2, table 8). Although calcareous siltstone and thin marble lenses were observed along the south side of Johns Hopkins Inlet, thick or continuous sections of carbonate or amygdaloidal greenstone, typical of Wrangellia, were not observed anywhere along shoreline exposures of the Tarr Inlet suture zone.

On northern Chichagof Island, the Border Ranges fault has a steep, near vertical, westward dip, as is the case elsewhere on Chichagof Island (Decker and Johnson, 1981). In Glacier Bay National Monument, the trace of the Border Ranges fault, as here defined, is almost entirely covered by glacier ice, but the fault is nearly vertical where it crosses Johns Hopkins Inlet. Rocks northeast of the fault on western Chichagof Island (Wrangellia) consist of undated but inferred late Paleozoic to Triassic, massive to weakly foliated, commonly amygdaloidal, greenstone, and bedded siliceous metasediments, that are intruded by foliated diorite and quartz diorite-tonalite plutons of probable mid-Cretaceous age (Loney and others, 1975).

Rocks northeast of the Border Ranges fault in Glacier Bay National Monument (Alexander terrane) consist of hornfelsed pelitic and semipelitic rock, marble, greenstone, and amphibolite, of inferred middle Paleozoic age based on scanty fossil evidence. The bedded rocks are extensively intruded by a variety of Cretaceous and Tertiary granitic plutons (Brew and Morrell, 1978). One highly altered foliated quartz-bearing diorite sample from a large pluton at the southeastern entrance to Johns Hopkins Inlet yielded a potassium-argon age of 119±8 m.y. (no. 3, table 8). Rocks immediately northeast of the fault on northern Chichagof Island and on the Inian Islands consist of a complex assemblage of hornfelsed bedded rocks, hornblende gneiss, and migmatite of uncertain original age. The metamorphic rocks are pervasively intruded by massive to foliated, commonly altered, granitic rocks that are dominantly of dioritic or quartz dioritic composition. A fresh diorite sample from Inian Island has a potassium-argon age of 113±7 m.y. (no. 4, table 8) and hornblende in the schist it intrudes has been reset to 31.8±2.9 m.y. (no. 5, table 8). Although these rocks could correlate with strata of either Wrangellia or the Alexander terrane, the thin even-bedded nature of hornfelsed country rock, the presence of thick marble beds, and the pervasive Early Cretaceous plutonism of intermediate composition suggest they belong to the Alexander terrane.

Two conclusions that can be drawn from the correlations presented here are: 1) rocks of the Tarr Inlet suture zone comprise part of the melange facies of the Chugach terrane of Early Cretaceous age (Plafker and others, 1977) that extends in a nearly continuous belt from Kodiak Island through the Chugach Mountains and into southeastern Alaska, and 2) the Tarr Inlet suture zone probably formed by subduction of oceanic crust during a mid-Cretaceous subduction event (Decker and others, 1980).

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Table 8.--*Location and potassium-argon data for rock samples from Johns Hopkins Inlet and Inian Island*

[Potassium-argon analyses and calculations on hornblende separates by Geochron Laboratories. Constants used in calculation of ages: $\lambda_\beta = 4.72X10^{-10}$ yr^{-1}, $\lambda_\varepsilon = 0.585X10^{-10}$ yr^{-1}, and $^{40}K/K = 1.22X10^{-4}$ g/g.]

Number	Sample number	Rock description	K_2O (percent)	K_2O average (percent)	^{40}Ar rad (ppm)	^{40}Ar rad average (ppm)	^{40}Ar rad / ^{40}Ar total	Apparent age (m.y.)
1	80APr29	Sheared and altered diorite	0.392 .401	0.396	0.002823 .002795	0.002809	0.544 .495	96.7±5.5
2	80APr32	Biotite-bearing diorite	.177 .182	.179	.000785 .000740	.000763	.122 .094	58.6±5.2
3	80APr33	Foliated diorite	.194 .194	.194	.001696 .001691	.001694	.284 .355	119±8
4	80APr22	Diorite	.315 .314	.314	.002693 .002551	.002622	.363 .348	113±7
5	80APr23	Hornblende schist	.288 .286	.187	.000738 .000576	.000657	.246 .253	31.8±2.9

80APr29. South shore Johns Hopkins Inlet (lat 58°54′N., long 137°01′W.).

80APr32. South shore Johns Hopkins Inlet (lat 58°54′N., long 136°57′W.).

80APr33. Prominent point at south side of entrance to Johns Hopkins Inlet (lat 58°15′N., long 136°20′W.).

80APr22. Point southwest of cabin at head of Inian Cove (lat 58°15′N., long 136°20′W.).

80APr23. Point southwest of cabin at head of Inian Cove (lat 58°15′N., long 136°20′W.).

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Discrimination of Wrangellia and the Chugach terrane in the Kelp Bay Group on Chichagof and Baranof Islands, southeastern Alaska

By Susan M. Karl, John E. Decker, and Bruce R. Johnson

The Kelp Bay Group contains components of both Wrangellia and the Chugach terrane of Chichagof and Baranof Islands, southeastern Alaska, which has contributed much confusion to the definition of the group. The purpose of this article is to summarize the evolution of our understanding of the components of the Kelp Bay Group, and to report on the results of a field examination of these rocks in September 1980 by a group of Geological Survey and outside scientists who have worked extensively in these rocks and similar rocks elsewhere.

The Kelp Bay Group, as defined by Loney and others (1975) (fig. 76 and table 9) consists of five formations: the Goon Dip Greenstone, Whitestripe Marble, Pinnacle Peak Phyllite, Waterfall Greenstone, all of Triassic(?) age, and the Khaz Formation of Triassic and (or) Jurassic age. The Goon Dip Greenstone consists of a continuous belt of massive, locally amygdaloidal greenstone, which is faulted against, or stratigraphically overlain by the massive to weakly foliated Whitestripe Marble. These two rock units extend for most of the length of Chichagof Island, and are characterized by their uniformity and lateral continuity. They lie east of the Border Ranges fault, and represent Wrangellia on Chichagof Island (Plafker and others, 1976; Plafker and others, 1977; Jones and others, 1977). West of the fault, the Pinnacle Peak Phyllite consists of generally subphyllitic to schistose fine-grained metasedimentary and metavolcanic rock locally intercalated or faulted against each other. The Waterfall Greenstone consists dominantly of massive or pillowed greenstone, with intercalated lenses of red ribbon chert locally containing radiolaria of Tithonian to Valanginian age, as well as tuffaceous argillite and coarse-grained volcaniclastic graywacke. The Khaz Formation is a melange of fault-bounded blocks of graywacke turbidite, massive *Buchia*-bearing graywacke, tuff and volcanic breccia, greenstone, chert, limestone, phyllite, greenschist, and glaucophane-bearing greenschist in a tuffaceous argillite matrix. Megafossils from sandstone blocks and radiolarians from chert blocks suggest a Tithonian to Valanginian depositional age for those blocks.

West of the Kelp Bay Group lies the Sitka Graywacke, a broken formation composed of proximal turbidites (Decker and others, 1979) compositionally similar to the graywackes found in the Kelp Bay Group (Decker, 1980b).

All of the units west of the Border Ranges fault are characterized by their disrupted nature, and together represent a portion of the Chugach terrane of Berg and others (1972; 1978) which has also been referred to as the accretionary flysch and melange terrane of southern Alaska (Plafker and others, 1977). Rocks of the Chugach terrane on Chichagof and Baranof Islands record a complicated history of slope and trench sedimentation, with associated volcanism and plate margin tectonism.

Previous work

Most geologic mapping and related studies on Baranof and Chichagof Islands prior to 1976 (Reed and Coats, 1942; Rossman, 1959; Berg and Hinckley, 1963; Loney and others, 1963; 1975) treated the components of the Kelp Bay Group as a conventional stratigraphic sequence.

Jones and others (1971) and Berg and others (1972) considered the rocks on western Chichagof Island to be part of a regional upper Mesozoic accretionary complex along the southern Alaska continental margin. MacKevett and Plafker (1974) defined the Border Ranges fault as the major tectonic boundary marking the continentward edge of this upper Mesozoic accretionary terrane. Extending the Border Ranges fault into southeastern Alaska, Plafker and others (1976) suggested a lack of continuity within the Kelp Bay Group by placing the fault between the Goon Dip Greenstone and Whitestripe Marble to the east and the rest of the Kelp Bay Group to the west. Plafker and others (1976; 1977) correlated the Goon Dip Greenstone and the Whitestripe Marble with the Nikolai Greenstone and the well-dated Late Triassic Chitistone Limestone, respectively, of the Wrangell Mountains. These units were subsequently assigned to "Wrangellia," a distinctive Upper Triassic terrane (Jones and others, 1977; Berg and others, 1978). Plafker and others (1976; 1977) correlated the Kelp Bay Group west of the Border Ranges fault with the lithologically similar, highly deformed upper Mesozoic units around the southern Alaskan margin, namely,

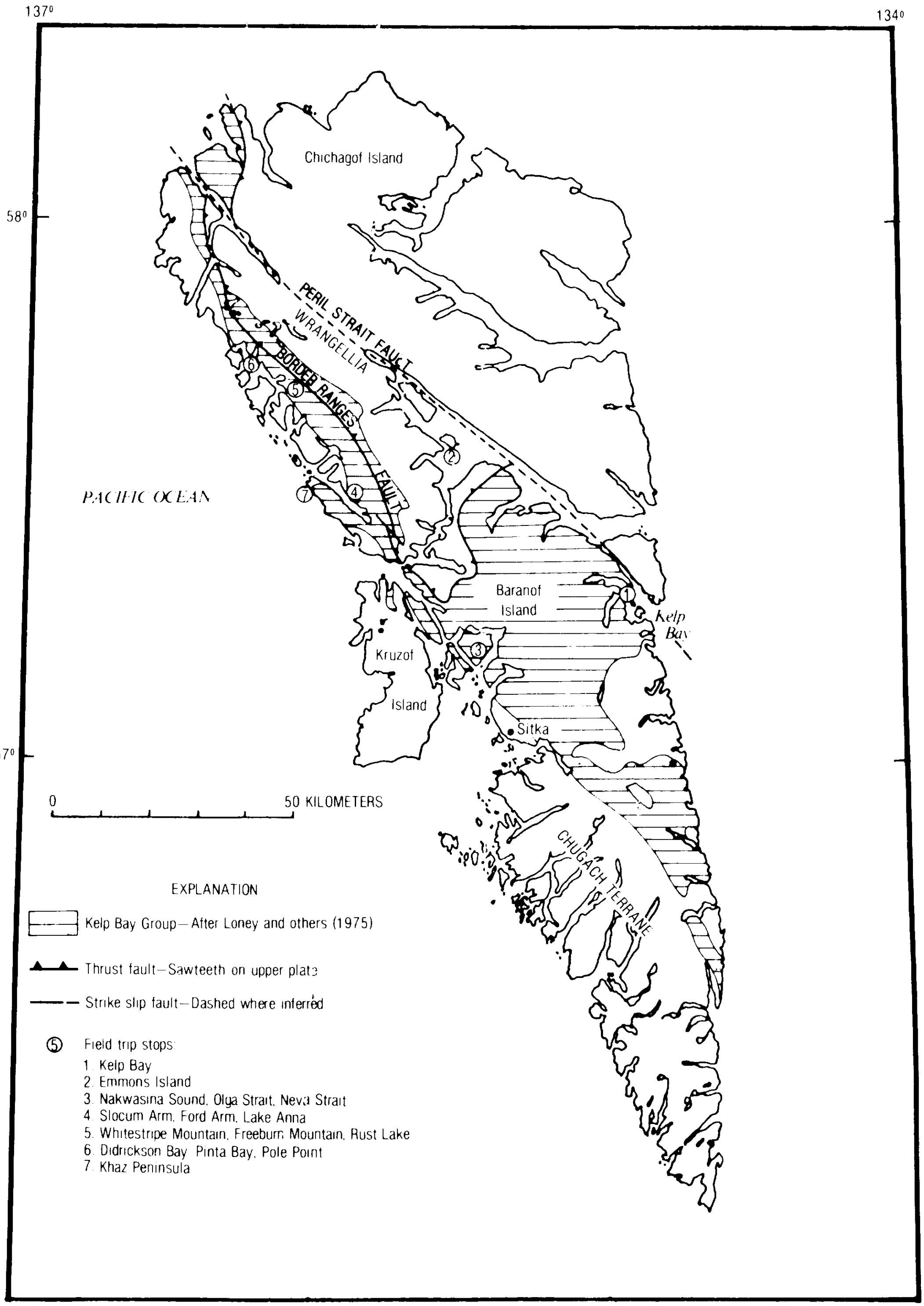

FIGURE 76.—Map showing areas visited on northern Baranof and western Chichagof Islands.

Table 9.--*Stratigraphic nomenclature and age assignments in the Chichagof, Baranof, and Kruzof Islands area*

Stratigraphic nomenclature and age assignment							Area and reference
Undifferentiated metamorphic rocks (probably Mesozoic and older)						Graywacke (Upper Jurassic?)	Western Chichagof Island; Overbeck (1919, pl. 2)
Greenstone schist (pre-Triassic?)	Greenstone (Triassic?)	Limestone and marble (Triassic?)	Schist (Triassic?)			Graywacke (Cretaceous?)	Chichagof mining district; Reed and Coats (1941, pl. 3)
	Greenstone unit (Triassic?)	Marble unit (Jurassic?)	Schist unit (Jurassic?)			Graywacke unit (Cretaceous)	Northwestern Chichagof Island; Rossman (1959, pl. 12)
Sedimentary and metamorphic rocks, undivided (Paleozoic? and (or) Mesozoic?)	Kelp Bay Group (Triassic?)					Sitka Group (Jurassic and Cretaceous	Northern Baranof Island; Berg and Hinckley (1963, pl. 1)
	Goon Dip Greenstone (Permian? and (Triassic?)	Whitestripe Marble (Triassic?)	Pinnacle Peak Phyllite (Triassic?)	Waterfall Greenstone (Triassic?)	Kelp Bay Group (Triassic or Jurassic)	Sitka Graywacke (Jurassic and Cretaceous)	Chichagof, Baranof, and Kruzof Islands; Loney and others (1963)
Cherty limestone, sandstone, and greenstone (Paleozoic or Mesozoic)	Kelp Bay Group (Triassic and (or) Jurassic)					Sitka Graywacke (Jurassic and Cretaceous)	Chichagof, Baranof, and Kruzof Islands; Loney and others (1975)
	Goon Dip Greenstone (Triassic?)	Whitestripe Marble (Triassic?)	Pinnacle Peak Phyllite (Triassic?)	Waterfall Greenstone (Triassic?)	Khaz Formation (Triassic and (or) Jurassic)		
				Melange (Cretaceous)		Flysch (Cretaceous)	Chichagof and Baranof Islands; Plafker and others (1977)
	Goon Dip Greenstone	Whitestripe Marble	Kelp Bay Group (Cretaceous)			Sitka Graywacke (Cretaceous)	Proposed classification, (this report)
	(Triassic?)	(Triassic?)	Pinnacle Peak Phyllite (Cretaceous?)	Waterfall Greenstone (Cretaceous?)	Khaz Formation (Cretaceous?)		

the Uyak Complex, the McHugh Complex, and the melange facies of the Yakutat Group, which together they interpreted as the melange facies of a subduction complex accreted to the continental margin along the Border Ranges fault system.

Recent detailed mapping (Decker, 1980a; 1980b; Decker and Johnson, 1980) has confirmed or reinforced the general conclusions of the earlier workers through documented field relationships and additional age information. Radiolarian age determinations from the Kelp Bay Group west of the Border Ranges fault establish that some of these rocks are as young as Early Cretaceous (Plafker and others, 1976; Decker, 1980a) and isotopic age determinations on metamorphic minerals from units west of the fault establish the intermediate to high pressure metamorphism associated with accretion as being of mid-Cretaceous (106 to 91 m.y.) age (Decker and others, 1980). Finally, it has been emphasized by Plafker and others (1977) and Decker (1980b) that both the Goon Dip Greenstone and Whitestripe Marble east of the fault form continuous mappable belts, whereas components of the Kelp Bay Group west of the fault form a collage of kilometer-scale, fault-bounded blocks, consisting of metavolcanic rocks, metasedimentary rocks, and melange. Thus to all recent workers, there appeared to be a need for redefinition of the Kelp Bay Group. Decker (1980a; 1980b) suggested informal revisions and also reviewed the evolution of nomenclature concerning units of the Kelp Bay Group and Sitka Graywacke.

Compositional and textural characteristics confirm the fundamental affinities between the units in the Kelp Bay Group west of the Border Ranges fault. Thus, on Chichagof and Baranof Islands, the flysch facies of the Chugach terrane is represented by the Sitka Graywacke and the melange facies is represented by the disrupted units in the Kelp Bay Group west of the fault (Plafker and others, 1977).

Present Investigation

In September 1980, the U.S. Geological Survey sponsored a 7-day field trip on Chichagof and Baranof Islands to focus attention on the nature, origin, and style of deformation of the Kelp Bay Group (fig. 76). Important conclusions of earlier work confirmed during the field trip included the following:

(1) The Border Ranges fault is located between the Goon Dip Greenstone and Whitestripe Marble to the east and the rest of the Kelp Bay Group to the west as Plafker and others (1976) and Decker and Johnson (1981) have suggested.

(2) The Goon Dip Greenstone and Whitestripe Marble should be excluded from the Kelp Bay Group (Plafker and others, 1976; Decker, 1980b).

(3) The Whitestripe Marble and Goon Dip Greenstone are strikingly similar to the Chitistone Limestone and Nikolai Greenstone in the Wrangell Mountains, and should be considered part of Wrangellia (Plafker and others, 1976; Jones and others, 1977).

(4) The Kelp Bay Group west of the Border Ranges fault is part of the accretionary Chugach terrane which is of late Mesozoic age (Plafker and others, 1977) rather than Triassic age as previously inferred.

(5) Field relationships suggest that within the accretionary complex, a dominantly volcanogenic, sediment-poor regime during Late Jurassic and Early Cretaceous time was succeeded by a sediment-rich depositional regime in Early(?) and Late Cretaceous time.

(6) The volcanic rocks in the accretionary complex, which are chemically mid-ocean ridge tholeiites (Decker, 1980b), are compatible with typical oceanic spreading ridge or leaky transform tectonic models.

(7) The proximal sedimentary rocks in the accretionary complex appear to be trench deposits (Decker and others, 1979; Zuffa and others, 1980) compatible with typical subduction or continental borderland tectonic models.

(8) Much of the initial deformation within the accretionary portion of the Kelp Bay Group was ductile deformation which involved volcanic rocks and chert as well as clastic sedimentary rocks, and which locally may signify contemporaneous sedimentation, volcanism, and deformation affecting the continental margin sedimentary basin (Decker, 1980b).

(9) Metamorphic mineral ages and paleontologic data together indicate an initial(?) mid-Cretaceous subduction event (Decker and others, 1980) for the rocks in the Chugach terrane on western Chichagof Island.

The three main results of the field trip included a consensus that:

(1) The Code of Stratigraphic Nomenclature does not adequately provide for naming and classifying disrupted rocks such as those west of the Border Ranges fault presently designated as the Pinnacle Peak Phyllite, the Waterfall Greenstone, and the Khaz Formation.

(2) Some part of the tectonic process that deformed these rocks allowed the development of quartz segregation layers in metasedimentary rocks apparently without initial development of isoclinal folds and axial plane foliation. These segregation layers follow convolute layering in turbidite beds.

(3) The Kelp Bay Group may have developed initially along a transform margin, and may have subsequently been subducted and accreted to the continental margin. The observations and conclusions above are significant chiefly in terms of focusing more precisely on the problems concerned with unravelling the complex geologic history of the Kelp Bay Group on Baranof and Chichagof Islands.

Summary

The Kelp Bay Group presently includes parts of two distinct tectonostratigraphic terranes, Wrangellia and the Chugach terrane (Jones and others, 1977; Berg and others, 1978), with very different geologic histories, separated by a major crustal suture, the Border Ranges fault. The present nomenclature for rock-stratigraphic units as presented in Articles 6 through 9 of the Code of Stratigraphic Nomenclature, although technically adequate and applicable to the disrupted rocks of the Chugach terrane, implies to some geologists a consistently observable stratigraphic succession that is misleading when applied to the chaotic rocks of the Kelp Bay Group. However, there is no currently accepted alternative way of classifying these units. In addition, the Kelp Bay Group in its type area on Baranof Island does not contain the limestone and greenstone units assigned to Wrangellia that are found east of the Border Ranges fault and thus the name Kelp Bay Group should be restricted to rocks in the melange facies of the Chugach terrane on

Chichagof and Baranof Islands. The names Goon Dip Greenstone and Whitestripe Marble should be retained for independent formations representing Wrangellia on Chichagof Island (table 9).

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Horizontal offset history of the Chatham Strait fault

By Travis Hudson, George Plafker, and Kirk Dixon

The Chatham Strait fault segment of the Denali fault system has long been recognized as an important structural feature in southeastern Alaska. Estimates of offset on the Chatham Strait fault include dextral horizontal displacements of about 80 km (Brew, Loney, and Muffler, 1966), 190 km (Lathram, 1964), and 200 km (Ovenshine and Brew, 1972). Although very inconclusively demonstrated, Sonnevil (1981) suggests 100 to 180 km dextral displacement. Investigation of neotectonics in southeastern Alaska by personnel of the Alaska Geologic Earthquake Hazards project has involved geologic and marine geophysical studies along the Chatham Strait fault together with mapping along its onshore continuation, the Chilkat fault. Our data indicate a total dextral offset of about 150 km, during post-middle Cretaceous and pre-Holocene time.

Although previous workers have recognized the presence of significant dextral separation on the fault, determinations of the timing and amount of this displacement have been handicapped by: (1) later vertical faulting that has disrupted the geology in areas adjacent to both sides of the fault, (2) the lack of specific information about the nature and distribution of rock units along the west side of Lynn Canal, (3) incomplete knowledge of the age of several important regional units adjacent to the fault, and (4) extensive cover of the fault by waters of Lynn Canal and Chatham Strait. The west side of Lynn Canal has been studied only in reconnaissance fashion (Lathram and others, 1959) but, as Lathram (1964) pointed out, rock units are present there that have correlatives southeast of the Chatham Strait fault.

We mapped the shoreline geology along the west side of Lynn Canal and examined the major mapped units east of the Chatham Strait fault during parts of the summers of 1979 and 1980 to define correlative rock units across the fault. Two distinct regional units can be correlated across the fault, and two other such units place constraints on the offset history. These four units are: (1) a polydeformed metamorphic terrane, (2) a Silurian turbidite sequence, (3) Cretaceous flysch and melange of Baranof Island and (4) a Tertiary volcanic sequence. The distribution of these regional units is outlined in figure 77, and the nature and correlation of each unit are discussed below. Additional information relevant to the Cenozoic offset history of the Chatham Strait/Chilkat fault is provided by deformed early Tertiary sedimentary rocks within the Chilkat fault zone, and by Quaternary deposits on the floors of Lynn Canal and Chatham Strait.

Metamorphic terrane

A sequence of metasedimentary, metavolcanic, and metaplutonic rocks, characterized by two well-defined metamorphic foliations and typically of greenschist facies, is exposed along the west shore of Lynn Canal from near Haines south some 50 km. The sequence includes schistose marble, chlorite-muscovite-quartz schist, muscovite-calcite schist, actinolite-quartz-calcite schist, schistose felsite containing K-feldspar, feldspathic schist derived from tonalitic plutonic rocks, and actinolitic greenschist lenses and bands. Metapelitic rocks with greenish metavolcanic streaks and bands are common and best display the typical polydeformational character of the sequence; the principal foliation is crosscut by a second foliation developed along microfold axes. These microfolds produce a well-defined lineation on the principal foliation surface. The metamorphic terrane west of the Chatham Strait fault is believed to be the same as that making up at least parts of the Retreat Group (Barker, 1957) and the Gambier Bay Formation (Loney, 1964) on Admiralty Island, east of the fault.

The Retreat Group and Gambier Bay Formation are exposed for 130 km along the west side of Admiralty Island. At the south end of the

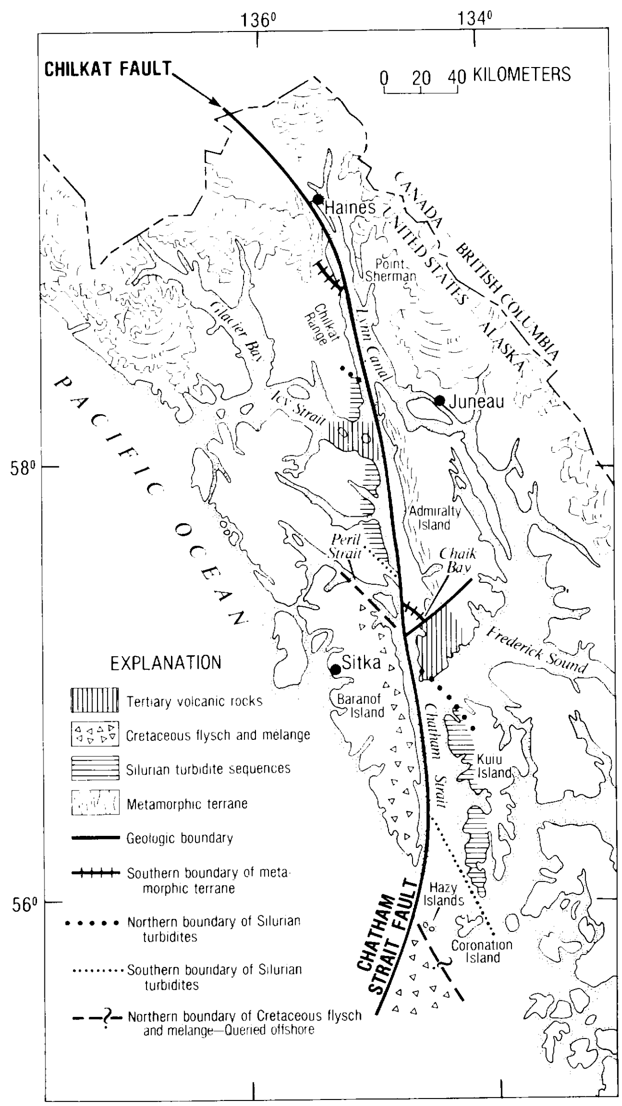

FIGURE 77.—Map showing distribution and correlation of regional units across Chatham Strait fault.

exposure a northeast-trending vertical fault (fig. 77) has displaced rocks on the southern part of the island relatively downward, thereby disrupting the regional trend of the Gambier Bay Formation and juxtaposing this formation against the much less deformed and metamorphosed Ordovician (Hood Bay Formation), Devonian and Mississippian (Cannery Formation), and Triassic (Hyd Formation) rocks. The southernmost exposures of the Gambier Bay Formation are on small islands at the mouth of Chaik Bay. In figure 77, a western boundary to this formation with a strike parallel to the regional strike of the unit is placed adjacent to these islands.

In our view the metamorphic terrane is a distinctive terrane that has been offset by the Chatham Strait fault, and it is not correlative with other less metamorphosed or deformed units in the area with which it locally is either in depositional or structural contact. The southern boundary of the metamorphic terrane as defined here is displaced dextrally, at least 148 km between the projected intersection on the west side of the fault opposite Point Sherman and the projected intersection on the east side opposite Peril Strait (fig. 77). Because displacement on the vertical fault at Chaik Bay may have broadened the exposed area of metamorphic rocks on southern Admiralty Island subsequent to at least some dextral displacement, the 148-km separation is a measure of the maximum dextral displacement on the fault. This displacement is younger than the age of the metamorphic terrane, which is known only to be Devonian or older (Loney, 1964).

Silurian turbidite sequence

The second regional unit that appears to be correlative across the fault is a turbidite sequence of known or suspected Silurian age. East of the fault, the sequence makes up the west side of Kuiu Island where it is locally fossiliferous and has been named the Bay of Pillars Formation (Muffler, 1967). A small area of clastic rocks in the vicinity of the southern tip of Admiralty Island may also be a part of this formation (Lathram and others, 1965). The sandstone in the Bay of Pillars Formation is rich in lithic detritus. Most of the lithic grains are angular fragments of igneous rocks including plagioclase porphyry, very fine grained and poorly crystallized tuff(?), and coarser dioritic grains. Fragments of several types of limestone are the next most abundant lithic grains. Polycrystalline quartz is not abundant, tectonized grains are absent, and clastic rock fragments are restricted to some fine sandstone grains. Mineral grains are dominantly angular to subhedral plagioclase and quartz with minor amounts of heavy minerals such as zircon, biotite, and indeterminate chloritized minerals. The modal composition of the sandstone suggests that it was derived from a dissected volcanoplutonic terrane of andesitic to dacitic composition that was built upon, or proximal to, a well-developed carbonate platform.

West of the fault, the Bay of Pillars Formation probably correlates with a deformed turbidite sequence on northern Chichagof Island named the Point Augusta Formation (Loney and others, 1975). The Point Augusta Formation is nonfossiliferous, but it is unconformably overlain by Devonian sedimentary rocks. We have mapped similar rocks on the west side of Lynn Canal for 20 km northward from Icy Strait. There, as on Chichagof Island, other rock units are present that are either in depositional or structural contact with the turbidites; the turbidites, however, do not reappear along the shoreline north of the boundary shown in figure 77.

The sandstone of the Point Augusta Formation and similar rocks north of Icy Strait are generally similar in modal composition to that of the Bay of Pillars Formation on western Kuiu Island. They are lithic-rich sandstone dominated by fragments of fine-grained igneous rocks. The most abundant lithic grains appear to be very fine grained poorly crystallized tuff(?), but porphyritic grains, various limestone fragments, plagioclase, and quartz are also important constituents. Although the turbidites west of the fault appear to have been derived from a similar but possibly less dissected or more distant source than those east of the fault, the lithology of sandstone of the Point Augusta Formation permits its correlation with sandstone of the Bay of Pillars Formation, as was suggested by Loney and others (1975). Correlation of the Silurian turbidites (Bay of Pillars Formation and Point Augusta Formation and similar rocks north of Icy Strait) across the Chatham Strait fault provides two additional measures of the lateral displacement—the northern and southern boundaries to the turbidite sequences. These boundaries, which are known to be faults in some places, are drawn in figure 77 with trends parallel to the regional trends of the turbidite sequence. The northern boundary is offset dextrally 133 km and the southern boundary about 111 km. Both are minimum displacements in that the placement of the boundaries could be revised to increase the measured offset by about 10 km for the northern boundary and 20 to 30 km for the southern boundary. However, increases greater than these amounts do not seem justified. Our placement of the southern boundary east of the fault differs from that of Ovenshine and Brew (1972) because we projected it parallel to the general strike of the turbidite sequence. The Silurian facies transition discussed by Ovenshine and Brew (1972) may exist to the south and east of Kuiu Island, but we do not see clear evidence for its presence west of the fault. As Churkin and Eberlein (1977, p. 783) noted, the presence of Silurian carbonate rocks to the north of the turbidite sequence on both sides of the fault suggests that the Silurian facies relations are more complex than depicted by Ovenshine and Brew (1972). In any case, dextral separation of Silurian rocks across the fault can be no more than about 150 km.

Cretaceous flysch and melange

Most of the rocks on Baranof Island were recently recognized to be part of a predominantly Cretaceous flysch and melange terrane that is present throughout the Gulf of Alaska region (Plafker and others, 1977). These rocks are exposed along the west side of the Chatham Strait fault, but they have no onshore correlatives east of the fault (fig. 77). The boundary of this terrane east of the fault must be seaward of the Hazy Islands and of Coronation Island, although its actual location offshore is not known. As shown in figure 77, the minimum dextral separation of this boundary is about 150 km; the maximum could be roughly 200 km if the terrane boundary were placed at the base of the continental slope. The minimum figure is considered the more likely one because of its similarity to the separation measured for the metamorphic terrane discussed above. The data require that all the dextral displacement on the Chatham Strait fault postdate accretion of the Late Jurassic(?) to Early Cretaceous melange and undated flysch of possible Late Cretaceous age that constitute this terrane.

Cenozoic units

The Admiralty Island Volcanics (Loney, 1964; Lathram and others, 1965) lie adjacent to the fault on southern Admiralty Island. These basaltic rocks of Eocene and Oligocene age may have correlatives west of the fault in Icy Strait, as suggested by Ovenshine and Brew (1972). We have examined the basaltic rocks from both areas and have sampled them for potassium-argon age determinations to test their correlation. The preliminary results indicate that they are lithologically similar and that samples from both areas are 25 to 27 m.y. old. If the two areas were originally contiguous in outcrop distribution, then roughly 100 km of post-Oligocene dextral displacement of the volcanic sequence is indicated.

Our mapping along the onshore extension of the Chatham Strait fault in the Chilkat River valley during 1978 has demonstrated that poorly preserved leaf-bearing continental strata of Paleogene (Eocene?) age along the fault zone have been intensely deformed as a result of large-scale dextral shear during the middle or late Cenozoic. Holocene features, however, have not been offset along the Chilkat fault. No detectable offset of reflectors within fiord-fill deposits could be identified on high-resolution and 80-KJ sparker profiles taken by the R/V *Sea Sounder* during 1978 at 14 crossings of the Chatham Strait fault in Lynn Canal and Chatham Strait (Paul Carlson and George Plafker, unpub. data). The marine data, together with studies along the onshore projection of the Chatham Strait fault, suggest that it is not presently active and that the most recent significant displacement is pre-Holocene and may be pre-Quaternary.

Discussion

The Chatham Strait fault is part of the Denali fault system, which extends northward through Canada into Alaska. Evidence concerning the displacement history for onshore parts of the Denali fault system, summarized recently by Lanphere (1978), suggests that 350 km of dextral displacement took place between 55 and 38 m.y. ago and no more than 40 km since about 38 m.y. ago. This offset history differs from that on the Chatham Strait fault both in total amount (350 km vs. 150 km) and possibly in timing. Much of the displacement on the Chatham Strait fault may have taken place in the middle and late Cenozoic, whereas elsewhere on the Denali fault system this seems to have been a time of more limited (±40 km) displacement. These discrepancies suggest several possibilities: (1) the timing and amount of displacement so far suggested for the onshore Denali fault system are incorrect; (2) another splay or splays of the Denali fault system are present in southeastern Alaska, such as the Coast Range megalineament (Lanphere, 1978); (3) late Cenozoic displacement in the central Alaska Range may have taken place on other faults besides the McKinley Strand of the Denali fault system (Stout and Chase, 1980), or (4) late Cenozoic displacement on the Chatham Strait fault is minor, and even though the Admiralty Island Volcanics are apparently correlative with the Tertiary basalts of the Icy Strait area, they may not have been offset from one another. We cannot presently evaluate all of these possibilities, but in our view the offset history of the remainder of the Denali fault system is now less well known than that of the Chatham Strait fault.

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Regional uplift in southeastern Alaska

By Travis Hudson, Kirk Dixon, and George Plafker

Residents of the Alexander Archipelago in southeastern Alaska have long recognized effects of gradual land emergence in some coastal areas with attendant newly exposed shoals, raised beaches, and wave-cut benches. Uplift of the region lying roughly between Petersburg and Yakutat was documented by Hicks and Shofnos (1965) with continuous sea level observations at five permanent tide stations and separated sea level observations during 1959-60 at an additional 28 temporary stations (table 10). Figure 78 summarizes these data by contouring total uplift for a 20-year period prior to 1959-60 with the assumption that average emergence rates (table 10) were constant during this time interval. Uplift affected a region at least 500 km long by 150 km wide that was centered in the Glacier Bay area. Maximum uplift rates as high as 4 cm/yr occurred at Bartlett Cove in a recently deglaciated part of Glacier Bay (loc. 7, fig. 78*A*). Level data along the White Pass Yukon Railway suggest that the uplifted area extends across the Coast Mountains from Skagway northeastward to Carcross (St. Amand, 1957).

During 1979 and 1980 personnel of the Alaska Geologic Earthquake Hazards Project measured the elevations of selected tidal bench marks relative to sea levels to determine whether emergence in the region still continues. During 1979, elevations of 15 tidal bench marks were measured in the area centered on Glacier Bay, Icy Strait, and Lynn Canal. In 1980, 13 of these tidal bench marks were reoccupied to evaluate the precision of the measuring technique, and an additional 8 were measured along Chatham Strait, Frederick Sound, and Taku Inlet (table 11).

The tidal bench mark elevations were measured by leveling from the observed high tide mark at the shoreline to the bench marks. High tide levels for the localities were obtained from U.S. Coast and Geodetic Survey tide tables, and the elevation of the bench marks relative to mean low water were then calculated. Two or more bench marks are commonly present at each locality. Independent measurements at these bench marks were commonly within 3 cm of one another. At the 13 localities where replicate measurements were obtained, 7 localities had a range of 10 cm for the measured elevations, 5 had a range between 16 and 21 cm, and 1 had a range of 31 cm between three separate measurements. The precision of the replicate measurements (approximately ±10 cm for more than 90 percent of the measurements) contrasts with the variation observed within the simultaneous independent measurements at any one locality (commonly 3 cm). The uncertainty in the measurements is dominated by differences in the observed high tide levels from those predicted by the tide tables; these differences are caused by meteorologic variations that can influence tide levels at any particular locality. Thus, our measurements at any given locality are conservatively estimated to have an uncertainty of ±10 cm—a value considerably less than the uplift observed in all but the western fringe of the region.

Figure 78*B* summarizes the results of the 1979-80 studies by contouring the total uplift since 1959-60 in the same 10-cm intervals as figure 78*A*. Comparison of these two data sets indicates that regional uplift has continued at rates comparable to those determined in 1959-60. This uplift is differential and includes an area of more rapid emergence that is roughly centered between Glacier Bay and Lynn Canal. However, some differences seem apparent between the results of the two studies. In 1959-60 the area of most rapid emergence (at rates of about 4 cm/yr) extended southward to Bartlett Cove in Glacier Bay. The 1979-80 data suggest that the area of most rapid emergence (at about 3 cm/yr) has shifted northward to the upper parts of Glacier Bay and the Chilkat Peninsula (stations 3, 4, 5, fig. 78*B*). The 1959-60 study showed that a smooth regional gradient of diminishing uplift characterized southeastern

Table 10.--*Uplift in centimeters for period 1939 to 1959 (modified from Hicks and Shofnos, 1965)*

[Station locations are shown on figure 78*A*]

Station number	Uplift	Station number	Uplift	Station number	Uplift
1	[1]15	11	32	21	[1]7
2	[1]36	12	28	22	14
3	45	13	38	23	15
4	70	14	[1]29	24	17
5	45	15	28	25	10
6	54	16	26	14	
7	79	17	27	27	16
8	46	18	5	28	4
9	39	19	8	29	3
10	32	20	9	30	[1]7

[1]Continuous series data.

Table 11.--*Uplift in centimeters for the period 1959 to 1979/1980*

[Station localities shown on figure 78*B*]

Station number	Uplift	Station number	Uplift	Station number	Uplift
1	[1]14	10	[2]32	19	[3]21
2	[1]30	11	[2]32	20	[3]29
3	[2]56	12	[3]38	21	[2]0
4	[1]58	13	[2]26	22	[3]40
5	[2]56	14	[2]28	23	[1]2
6	[2]36	15	[2]25	24	[3]12
7	[2]47	16	[1]29	25	[2]33
8	[2]32	17	[3]46	26	[3]21
9	[2]48	18	[3]48	27	[3]0

[1]Continuous series data.
[2]Separated series using average of 1979 and 1980 data.
[3]Separated series based on 1980 data only.

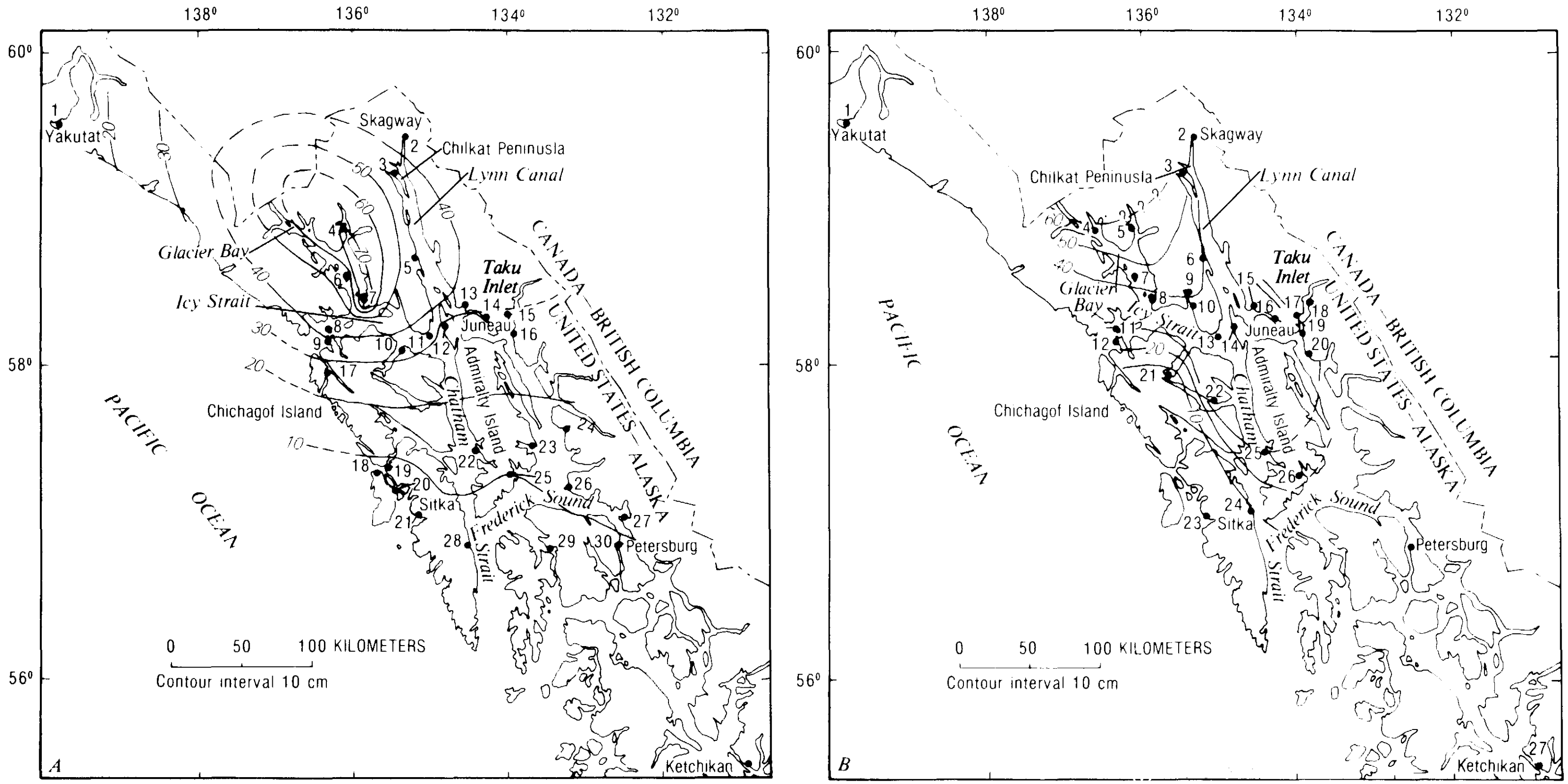

FIGURE 78.—Maps of Alexander Archipelago in southeastern Alaska showing contoured total uplift over 20-year period based on sea-level observations at numbered station locations: *A*, Stations and uplift data for period 1939 to 1959 modified from Hicks and Shofnos (1965), *B*, Stations and uplift data for period 1959 to 1979/1980 from the authors' observations in 1979 and 1980.

Alaska south of Glacier Bay—this relation was particularly well defined on Chichagof and Baranof Islands. The 1979-80 data, however, indicate more variability. The gradient on northern Chichagof Island appears to be steeper and to have shifted northward from that determined in 1959-60, and a broad region between Juneau and eastern Chichagof Island (including most of Admiralty Island) seems to be characterized by about 30 cm of uplift since 1959-60 whereas the regional gradient determined in 1959-60 was developed across this region.

The data are too incomplete to show details of the differential uplift clearly or to reveal discontinuities that may be present in the regional pattern of uplift. Nevertheless, comparison of figure 78*A* and *B* suggests the following tentative interpretations:

(1) The localized zone of most rapid uplift in Glacier Bay is due mainly to isostatic rebound related to ice unloading, as suggested by Hicks and Shofnos (1965). The northward migration of the locus of maximum uplift between the 1959-60 and 1979-80 studies seems to reflect rapid historic retreat of the major tidewater glaciers. It is not at all clear, however, why between the 1959-60 and 1979-80 studies this zone of maximum uplift should extend eastward to the vicinity of the Chilkat Peninsula (station 3).

(2) The uplift noted in the 1979-80 study at Taku Inlet does not seem to be attributable to unloading of ice because the Taku Glacier at the head of the inlet has been advancing slowly. The cause of this apparent anomaly is unknown and may be due simply to measurement errors in the 1980 reconnaissance study.

(3) The broad region of uplift that includes Admirality Island (as indicated by the 1979-80 data) includes topographically low nonglacial areas as well as higher areas that were not coverd recently by large ice sheets. Thus uplift in this region cannot be related to ice unloading. Assuming that the data are not systematically in error, this feature can be interpreted as a transient bulge of tectonic origin. Additional closely spaced tide-series measurements are required in this area to determine the configuration and uplift rate of this feature.

(4) The cause of the broad regional uplift outside of Glacier Bay remains enigmatic. Its distribution relative to mountain areas with extensive Holocene glaciers and icefields suggests that it cannot be soley attributed to unloading of ice, as suggested by Hicks and Shofnos (1965). However, shrinkage of glaciers certainly could be a factor in the northern part of the region. An alternative is that the uplift is tectonic in origin and is somehow related to strain buildup along the active transform boundary that is defined by the active Queen Charlotte-Fairweather transform fault system which lies roughly along the western margin of the uplifted region.

(5) The minimum area affected by uplift is roughly 500 km long by 200 km wide, and large areas have average uplift rates of 2 to 3 cm/yr. A combination of more accurate and closely spaced measurements of uplift throughout northern southeastern Alaska, together with an analysis of the effects of ice unloading, will be important to understanding this major active geodynamic system.

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Geochemical studies in the West Chichagof-Yakobi Wilderness

By Thomas D. Hessin

Several types of geochemical samples were collected from the West Chichagof-Yakobi Wilderness to complete a geochemical survey of the region and to detect possible anomalous mineralized areas. Filtered and unfiltered water and nonmagnetic heavy-mineral concentrates proved to be important sample media because of the favorable analytical results they provided

(Hessin and others, 1980). The unfiltered water samples were analyzed for sulfate and fluoride, and the filtered water samples (pressure squeezed through a 0.45-micrometer filter) were analyzed for copper, zinc, arsenic, molybdenum, and uranium. Both sulfate and fluoride are associates of nearly all types of mineral deposits, and anomalous values are therefore indirect indicators of possible metallic mineralization. Correlative anomalous amounts of base and precious metals in the nonmagnetic heavy-mineral concentrates and the filtered water samples are direct indicators of mineralization within the drainage source area.

Results of sulfate anion analyses are quite anomalous in the former mining districts of the Hirst-Chichagof and Apex-ElNido mines. In addition, anomalous amounts of flouride in the unfiltered water samples correlate well with anomalous amounts of uranium in the filtered water samples in the proximity of Deep Bay in the southwest quadrant of Chichagof Island.

Correlative anomalous amounts of metallic elements in filtered water and in the nonmagnetic heavy-mineral concentrates indicate the presence of several previously unknown mineralized areas. Moser Island on Hoonah Sound and the vicinity of Stag Bay on Chichigof Island near Yakobi Island are two of the more significant prospective areas indicated by anomalous amounts of metallic elements in the filtered water and concentrates.

The anomalies recognized from analyses of water correlate well with those from analyses of heavy-mineral concentrates. Although the Chichagof—Yakobi Wilderness is subject to heavy rainfall which dilutes surface waters, the discriminating range of values measured in the water samples warrants continued use of water as a geochemical sample medium.

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OFFSHORE ALASKA

(Figure 79 shows study areas discussed)

Preliminary interpretation of geophysical data over the projected offshore location of the Kaltag fault

By Michael A. Fisher

The Kaltag fault is a major right-lateral strike-slip fault in west-central Alaska. Along this fault, lateral separation of major geologic features by as much as 60 to 130 km occurred during the Late Cretaceous and early Tertiary, and Quaternary movement along the fault is suggested by scarps in Quaternary alluvium and by stream valleys that are offset by as much as 1 to 2.5 km (Grantz, 1966; Patton and Hoare, 1968). Various authors (Scholl and Hopkins, 1969; Scholl and others, 1970; Patton, 1973; Marlow and others, 1976; Grantz and Kirschner, 1976) propose that the Kaltag fault extends offshore to the southwest along its onshore strike and that the fault caused some structural disruption of the continental shelf beneath the Bering Sea. Fisher and others (1979) described the structure of Norton basin, which lies beneath Norton Sound along the north side of the projected location of the Kaltag fault, and attributed formation of Norton basin to extensional stress that may have developed along the fault. This report presents a preliminary interpretation of geophysical data obtained along the southern margin of Norton basin and over the projected location of the Kaltag fault, to determine whether these data show the location of the Kaltag fault (fig. 80).

Data used in this report consist of 24-fold seismic-reflection data that were obtained aboard the R/V *S. P. Lee* in 1978 and single-channel seismic-reflection data, obtained with a 176 in.^{3} air-gun array, and magnetic data that were collected aboard the *Lee* in 1980 (fig. 80). Presently, only seismic-reflection data are interpreted; magnetic data were interpreted only to the extent that a magnetic lineament was found to exist offshore.

The topmost area of figure 80 shows the structure of the southern part of Norton basin. Faults in this basin began to form in the Late Creta-

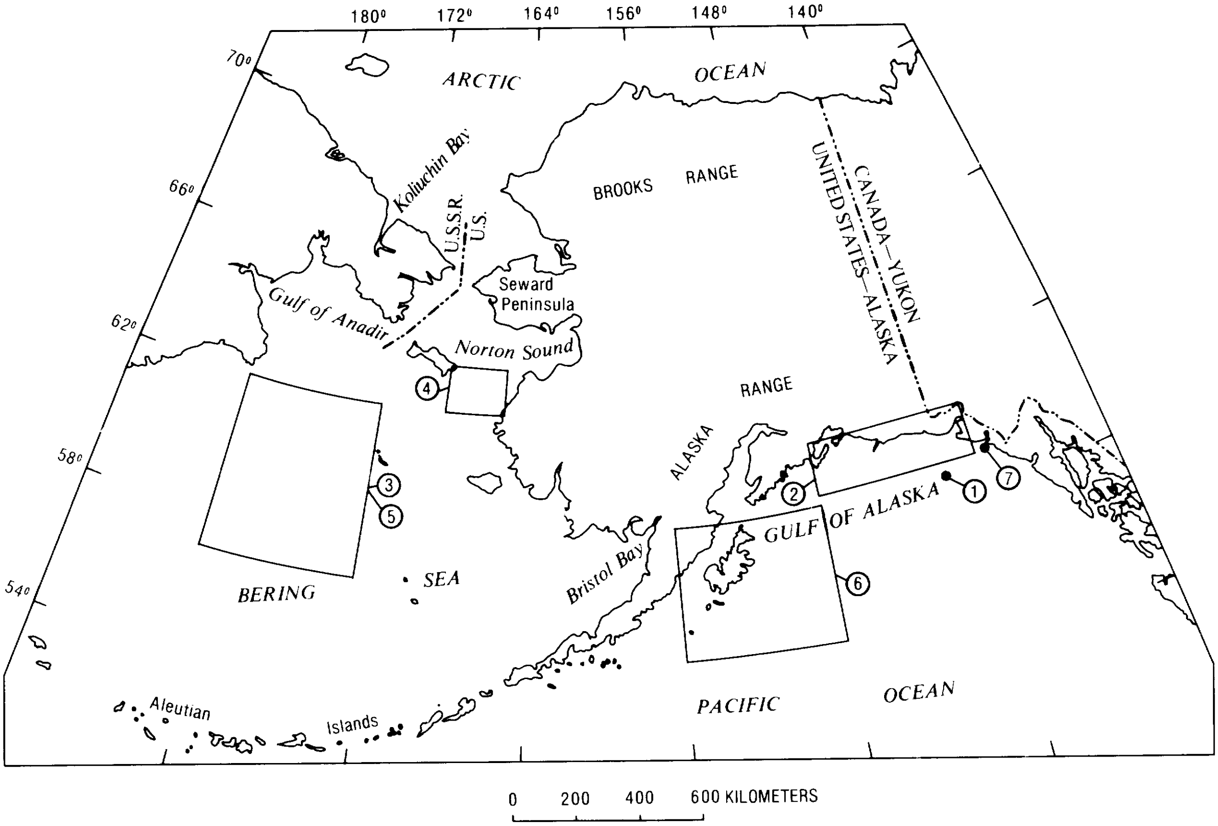

FIGURE 79.—Offshore areas discussed in this volume. Authorship and applicable figures relating to the numbered areas are listed below. 1, Barnes, Plafker, White, and Armstrong, figures 83 and 84; 2, Brouwers, figure 88; 3, Carlson, Karl, H., Johnson, and Fischer, figure 81; 4, Fisher, figure 80; 5, Karl, H., Carlson, and Lamb, figures 81 and 82; 6, McClellan, Fisher, von Huene, and Moore, figures 86 and 87; 7, Molnia and Rappeport, figure 85.

ceous or early Tertiary (Fisher and others, 1979). I place the southern boundary of the basin along the 63° N. parallel of latitude; south of the this parallel, faults are fewer and smaller than faults in the basin. In Norton basin (near the 167° W. meridian of longitude) lies a major normal fault that strikes about N. 35° W. and that has a throw locally as great as 2.5 km. This fault dies out along strike to the southeast.

South of Norton basin lie two normal faults of moderate throw (about 0.5 km). The time of movement along these faults is not known; however, the faults may have been active at about the same time as faults in Norton basin, and offset strata that are shallower than about 1 km and are of probable latest Tertiary and Quaternary age. The northern fault of the two strikes about N. 55° W.; the southern, about N. 80° W. Thus from north to south, the three large faults shown in figure 80 strike progressively more to the west. This rotation of strike, the decrease in throw of faults south of Norton basin, and the northeast strike of minor normal faults (near 168° W. and 62°30′ N.) are the only evidence from seismic-reflection data that geologic structure changes at all near the projected location of the Kaltag fault. No structures are evident that are typical of deformation along strike-slip faults, such as anticlines in echelon. All three normal faults strike obliquely to, and two faults cross, the projected Kaltag fault. No right-lateral offset of faults or of other structures has yet been determined from the grid of data used here.

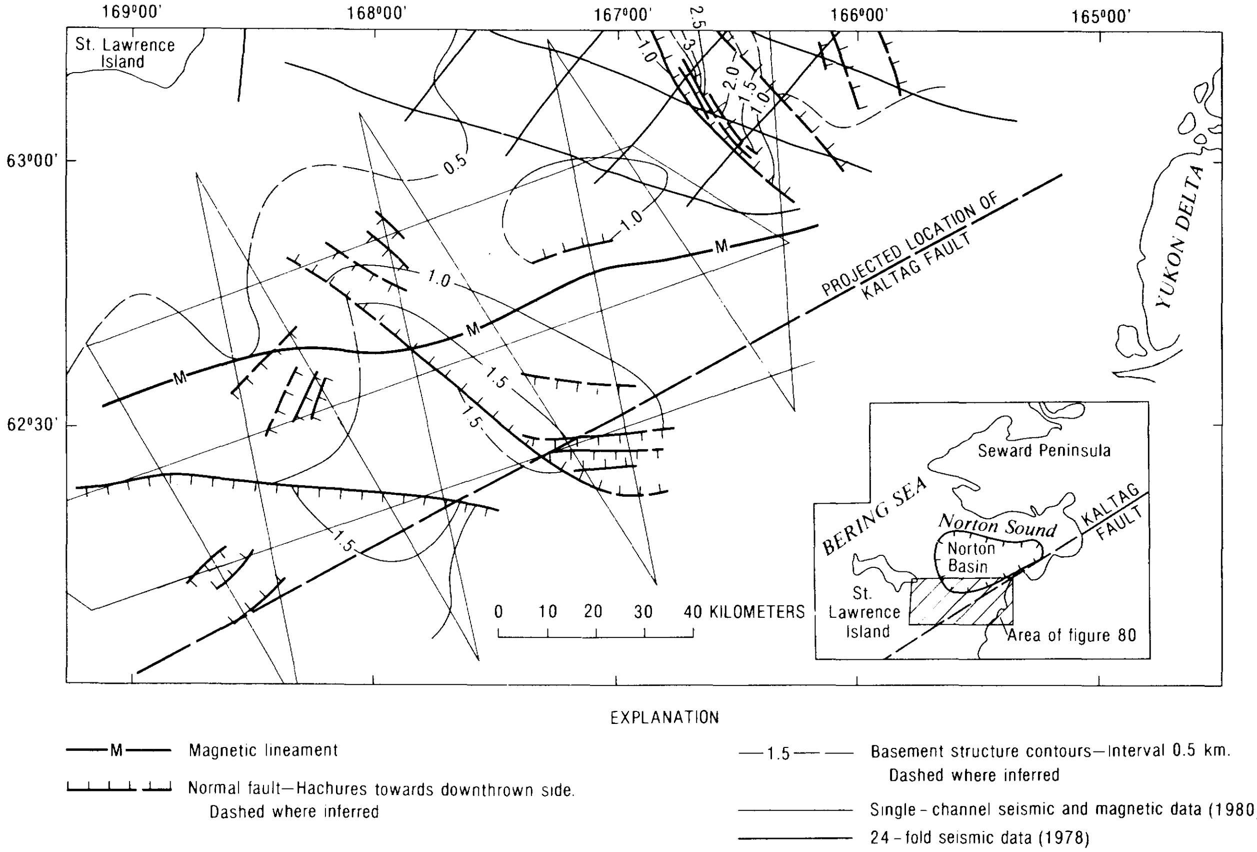

FIGURE 80.—Index map and map showing structure of the continental shelf near St. Lawrence Island and the Yukon delta. Numerous minor normal faults that could not be correlated among the seismic lines were deleted.

Magnetic data suggest that basement lithology changes north of the projected location of the Kaltag fault, along a magnetic lineament that strikes westward from the Yukon delta (fig. 80). The magnetic field north of the lineament is mostly smooth with only local 100- to 200-gamma anomalies, whereas south of the lineament, the field is made up of closely spaced 200- to 300-gamma anomalies that extend to the southern limit of the tracklines. The absence of consistent change in both geologic structure and the appearance of seismic reflections near the lineament suggests that the difference in magnetic character that causes the lineament is in basement rocks. Rocks that could be sources for the rough magnetic field south of the magnetic lineament include: (1) Jurassic or Cretaceous andesitic volcanic rocks that crop out locally near the Yukon delta (Hoare and Condon, 1966, 1971); (2) Late Cretaceous granitic rocks that crop out on St. Lawrence Island (Patton and Csejtey, 1980) and about 30 km southeast of the area shown in figure 80 (Hoare and Condon, 1968); and (3) late Tertiary and Quaternary volcanic rocks that crop out extensively on land areas in and around the Bering Sea. Because the young volcanic rocks would not be in the basement, I do not believe that these rocks are the source of the rough magnetic field.

Although the structure of the continental shelf between St. Lawrence Island and the Yukon delta does not show evidence of a throughgoing, active, strike-slip fault, magnetic data show that a lineament exists that is due to differences in the magnetic properties of basement rocks. The lineament, then, could mark the location of the Kaltag fault or of a splay of this fault; alternatively, the lineament could show the northernmost limit of Jurassic or Cretaceous andesitic volcanic rocks.

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Submarine canyons flanking Navarin Basin, Bering Sea

By Paul R. Carlson, Herman A. Karl, Kenneth A. Johnson, and Jeffrey M. Fischer

Some of the world's largest submarine canyons are incised into the eastern continental margin of the deep Bering Sea (fig. 81). Three canyons, Navarinsky, Pervenets, and Zhemchug, cut into the northeastern margin and provide windows into the large and thick accumulations of sediment that fill Navarin Basin. This basin, a potential petroleum province which contains as much as 15 km of Cenozoic sediment, is scheduled for leasing in early 1984. As a result of the heightened interest in this lease area, a USGS cruise was conducted in 1980 throughout the Navarin Basin province to identify possible geologic sea-floor hazards. Seismic-reflection profiles were collected along about 6700 km of trackline over the basin province.

Preliminary bathymetric data across the Beringian margin, including profiles from several previous cruises (Marlow and Cooper, 1979), show that Navarinsky, Pervenets, and Zhemchug Canyons head in water depths less than 150 m and debouch onto extensive deep-sea fans at depths of about 3000 m (fig. 81). Navarinsky and Zhemchug Canyons are similar in several respects: each canyon consists of two main branches landward of the shelf break and are underlain by large trough-shaped basins; both canyons are about 220 km long, each canyon is about 100 km wide near the shelf break, and both canyons have axial gradients of less than 1°. Pervenets is shorter (125 km), has a steeper axial gradient (1°15′), and is narrower near the shelf break (30 km). Whereas Navarinsky and Zhemchug Canyons generally trend northeast-southwest, Pervenets Canyon trends nearly east-west. Wall relief of the three canyons at the shelf break ranges from 400 m for Navarin, to 600 m for Pervenets, to a spectacular 2600 m for Zhemchug. These canyons are incised into Neogene and older, more-lithified rocks (Paleogene?) that make up much of Navarin Basin province (Marlow and others, 1979). The canyons are structurally controlled, the structures dating back at least into the Paleogene (Scholl and others, 1975).

The major cutting of the canyons probably occurred when glacially lowered sea levels exposed most of the Bering shelf and allowed large rivers such as the Yukon and the Anadyr to carry large amounts of sediment to the outer shelf. Although the rivers may have been responsible for the cutting of the extreme headward parts of the canyons, slumping and resulting turbidity currents were the most likely canyon-cutting agents. Slumping of unconsolidated sediment that had accummulated near the shelf edge may have been triggered by large storm waves, internal waves, and tsunamis. Seismic-reflection profiles collected across, as well as down, the canyon axes show numerous slumps along the surface of the modern canyons. Buried slumps and well-developed cut-and-fill structures are present throughout the canyon systems. Graded sand layers have been cored in the canyon fill and on the deep-sea fans at the

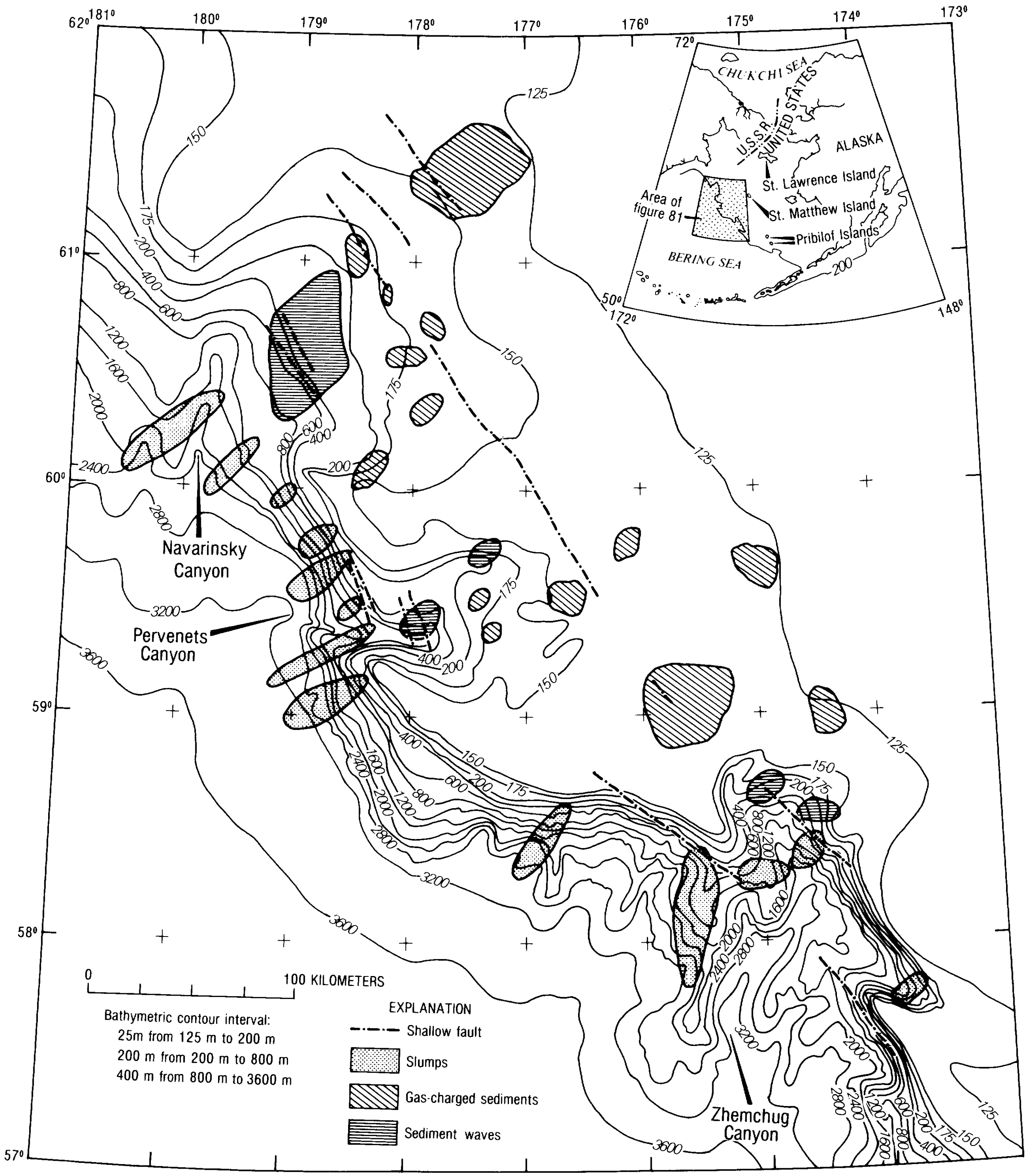

FIGURE 81.—Preliminary bathymetric and seafloor hazards map of Navarin Basin province.

base of the slope seaward of these three canyons. Seismic-reflection profiles across the fans show leveed channels and parallel, flat-lying reflectors that are typically ascribed to turbidity current deposits.

In addition to the submarine slumps, other geologic hazards present along the Beringean margin include near-surface faults, shallow zones of gas-charged sediment, and fields of large bedforms (fig. 81). The faults generally trend northwest-southeast parallel to the shelf break. Some faults form the bedrock flanks of large, trough-shaped basins that form the upper parts of Navarinsky and Zhemchung Canyons. Zones of shallow, gas-charged sediment (0-50 m subbottom) are most prevalent in the northern part of Navarin basin. A few of the cores analyzed for hydrocarbons contain concentrations of methane and ethane 5-10 times higher than background values (Vogel and others, 1981). Gas in the sediments and the resulting increase in pore pressure may be a cause of some of the sediment instability. Fields of large bedforms are located near the heads of three canyons in water depths of 200 to 400 m. These bedforms are well-developed sediment waves that have wavelengths of about 600 m and heights of 10-15 m (see Karl and others, this volume).

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Sediment waves in the head of Navarinsky, Pervenets, and Zhemchug Canyons, northwestern Bering Sea

By Herman A. Karl, Paul R. Carlson, and Beth Lamb

Reconnaisance geophysical profiling of Navarin Basin province, northwestern Bering Sea, revealed large sediment waves in the heads of Navarinsky, Pervenets, and Zhemchug Canyons (fig. 81). Several grab samples and short cores recovered fine sand from each sediment wave site. These sediment waves were not observed elsewhere in the study area. If active, such large bedforms could represent a sea-floor hazard; for this reason, we surveyed the sediment waves (fig. 82) in the head of Navarinsky Canyon at a closer track-line spacing than the 30-km reconnaissance grid in order better to understand their distribution, geometry, and history.

Most of the sediment waves are contained within a 600- to 700-km^2 area between the 215- and 450-m isobaths at the head of Navarinsky Canyon. By measuring the apparent wavelengths along three seismic lines which intersect in the southeastern corner of the sediment wave field, we determined that the bedforms strike approximately north (N. 5°E.), have an average wavelength of about 600 m, and have an average height of about 8 m and a maximum of about 15 m. Both symmetrical and asymmetrical sediment waves have been observed. The steeper face of asymmetrical bedforms faces east. Internal trough cross-stratification apparently dips toward the east. The bedforms are not only expressed on the surface, but also are remarkably well defined in the subsurface. The stratigraphic unit containing the sediment waves has developed over a flat-lying reflector and attains a maximum thickness of about 120 m in the sediment wave field. This crossbedded unit thins to 10-15 m toward the northwestern boundary of the field, thins to about 70-90 m toward the southeastern margin, and appears to wedge out at the extreme southeastern corner of the field. The sediment waves grade into flat beds toward the northeast (shoreward). At least seven sets can be recognized in the thickest part of the section. Individual crossbedded pods within sets are as thick as 20 m. The number of sets decreases as the unit thins. The superposed sediment waves climb at an apparent angle of

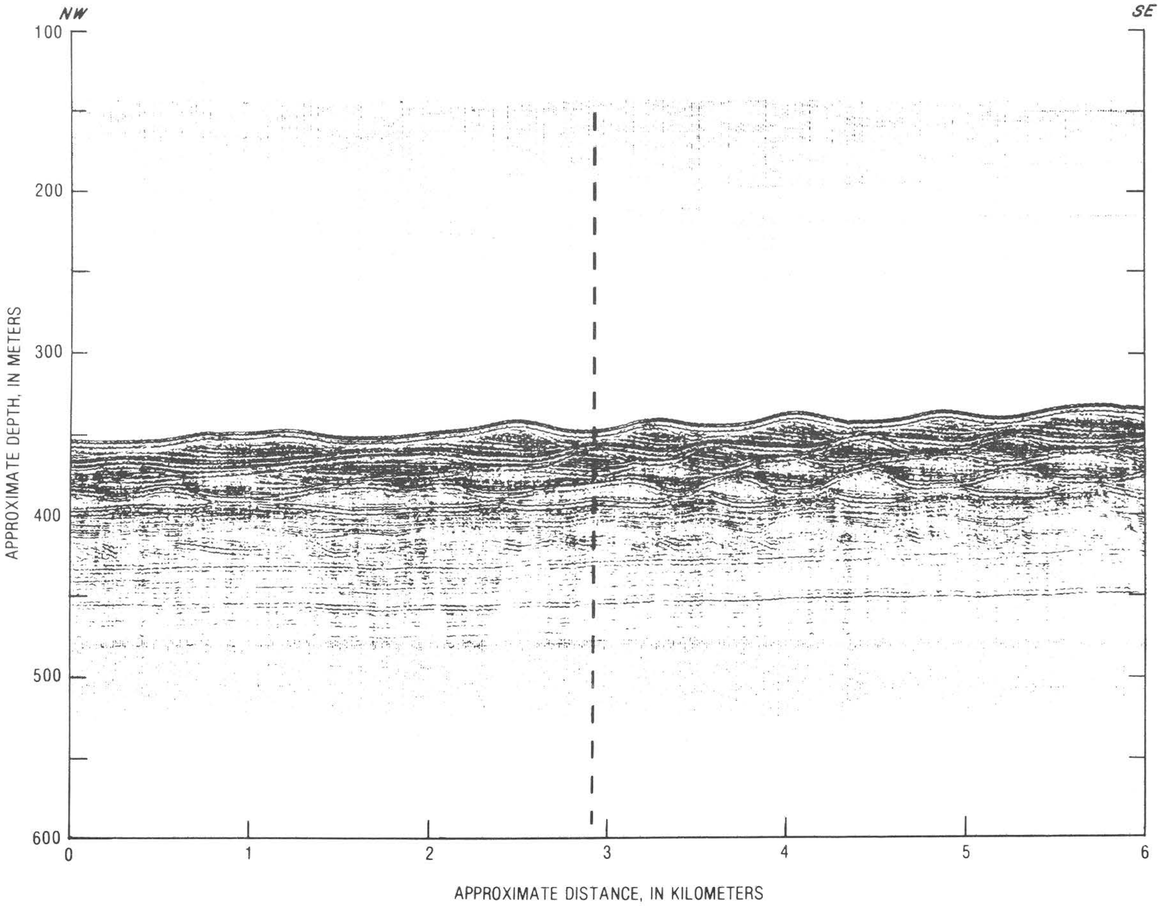

FIGURE 82.—Segment of minisparker seismic record showing sediment waves at the head of Navarin submarine canyon.

about 2½° toward the east. In many ways these large sediment waves resemble enormous climbing ripples. Alternatively, the sets may simply be stacked one on the other.

We can only speculate on the history of these bedforms at this stage of our investigation. We do not know if they are relict or presently active, nor are we certain if they indicate continuous sedimentation or discrete episodes of activity and nonactivity. The association with each of the submarine canyons is significant. Submarine canyons intensify internal waves and tides (Shepard and others, 1974) and affect flow on the adjacent shelf (Inman and others, 1976; Knebel and Folger, 1976; Karl, 1980). The sediment waves are at depths at which internal waves may form, and migration of the bedforms upcanyon indicates upcanyon flow, unless the bedforms are antidunes. The superposition and upsection migration of the bedforms suggest aggradation of sediment from suspension coupled with either a unidirectional flow or a very long period oscillatory motion combined with, or giving the effect of, a unidirectional flow. The hydrodynamic interpretation of these sediment waves, however, requires more detailed analysis of existing data and further field investigation.

The internal stratification, stratigraphic thickness, sediment size, and depositional setting combine to make these sediment waves unique among the large bedforms heretofore observed on the continental margins and in the ocean basins (Swift and Ludwick, 1976; Bouma and others, 1977; Damuth, 1979; Normark and others, 1980).

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Potential natural gas in the Gulf of Alaska indicated by calcite depleted in carbon-13

By Ivan Barnes, George Plafker, Lloyd D. White, and Augustus K. Armstrong

Massive banded calcite dredged from a sequence of Eocene basalt and sedimentary rocks along the continental slope of the northeastern Gulf of Alaska is depleted in Carbon-13 (^{13}C). This unusual material is an indicator of possible natural gas in the Gulf of Alaska region.

Occurrences of carbonate mineral depleted in ^{13}C relative to normal marine carbonates have been reported as aragonite cement in Quaternary sandstone on the Atlantic continental shelf (Hathaway and Degens, 1968; Allen and others, 1969) and high-magnesium carbonates that occur as authigenic beds, replacement zones, or cement in sedimentary strata of Late Cretaceous and Cenozoic age in the deep Bering Sea (Hein and others, 1979). The ^{13}C-depleted carbonate has been interpreted as due to the oxidation of methane (Hathaway and Degens, 1968; Allen and others, 1969; Hein and others, 1979). The occurence of ^{13}C-depleted carbonate is of special interest because it has been suggested that these unusual minerals may indicate proximity to petroleum reservoirs (Donovan and others, 1974) and because the isotopic composition of carbonates can provide clues to their conditions of formation. Many carbon and oxygen isotope compositions of limestones have been reported (for example, Keith and Weber, 1964).

As part of a petroleum resource investigation of the eastern Gulf of Alaska, a dredging program on the continental slope was carried out during 1977, 1978, and 1979 (Plafker and others, 1978; 1979; 1980a, b). Samples of ^{13}C-depleted calcite were recovered in dredge cast S10-79-EG-CHAN 33 from water depths between 3,020 and 1,780 m (fig. 83). Coordinates for start and finish of the dredge cast (No. S10-79-EG-CHAN-33) are lat 58°40.61′ N., long 140°59.85′ W. to lat 58°42.42′ N., long 150°56.63′ W. Roughly 100 kg of rock believed to represent outcrop was recovered. Most of the material in the dredge cast is altered amygdaloidal basalt and basaltic lapilli tuff. Associated with the volcanic rocks are minor amounts of sandstone and brown siltstone.

The five blocks of calcite or rocks containing

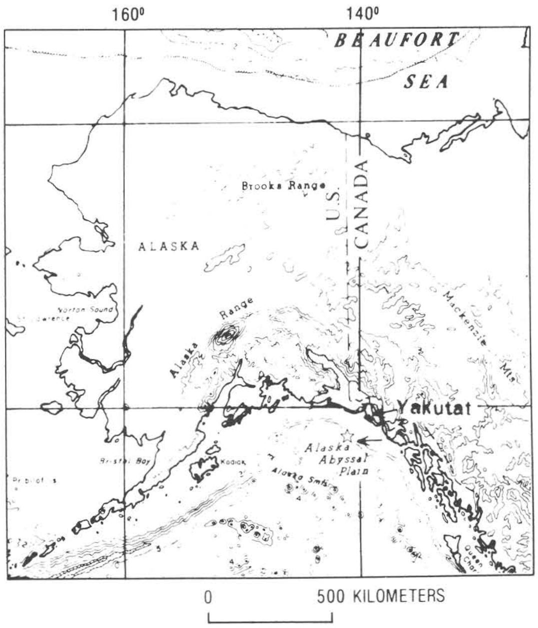


FIGURE 83.—Index map (Mercator projection) showing location approximately 115 km southwest of Yakutat where calcite samples were dredged from Gulf of Alaska (star).

massive calcite recovered in the dredge cast are: (1) banded vuggy calcite, green altered tuff, and calcite-cemented angular basaltic rock fragments (No. S10-79-EG-CHAN 33B, dimensions 35×30×22 cm); (2) interlayered banded vuggy calcite, green altered tuff, and angular basaltic rock fragments shown in figure 84 *A* (No. S10-79-EG-CHAN 33C, dimensions 15×10×10 cm); (3) basalt fragments cemented with calcite veinlets (No. S10-79-EG-CHAN 33D, dmensions 10×10×15 cm); (4) calcite with undulating centimeter-scale bands (No. S10-79-EG-CHAN 33E, dimensions 8×6×6 cm); and (5) irregularly banded vuggy calcite (No. S10-79-EG-CHAN 33F, dimensions 5×5×5 cm). The calcite is mostly salmon, grayish orange pink, or amber brown with some bluish-gray millimeter-scale bands and crosscutting veinlets. It ranges from very fine grained to coarsely crystalline. Individual scalenohedral crystals as long as 15 mm are oriented perpendicular to the bands, and crystal terminations commonly line vugs within the rock. The banded and vuggy appearance of these rocks suggest travertine or cave deposits; however, their association indicates that they probably formed as fracture and void fillings in basalt and fragmental basalt.

In thin section, the coarse calcite has a distinct banding caused by concentric or linear zones of differing crystal size and inclusion density (fig. 84 *B*, *C*). Each band has at its apparent base a zone of micritelike cement of possibly vadose origin overlain by progressively larger crystals of calcite. The coarser sparry calcite is of the type that forms in freshwater-influenced sediments (Folk, 1974). The coarseness may indicate a slow rate of precipitation without interference by magnesium or other ions. Under the cathode ray luminoscope, these samples have strong orange luminescence that characterizes manganese-bearing nonmarine carbonate rocks.

Although no fossils were found in the sedimentary rocks of this dredge cast, data from numerous dredge samples taken nearby suggest that the calcite is from a predominantly volcanic rock sequence at least 800 m thick that has been mapped along the lower part of the continental slope between long 138° W. and 142.5° W. (Plafker and others, 1978; 1979; 1980a, b). This sequence consists mainly of basalt and fragmental basaltic rocks with minor amounts of

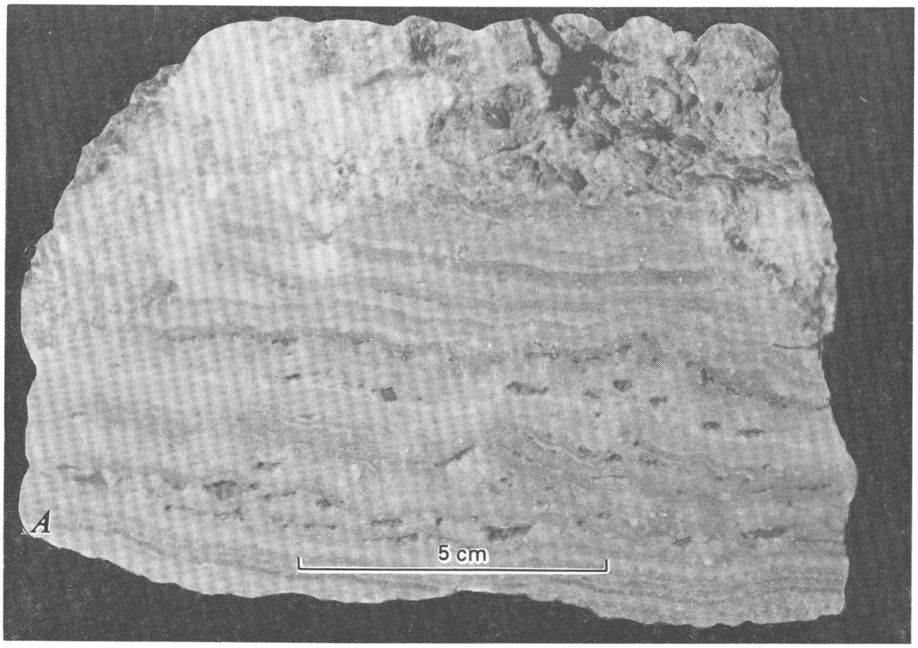


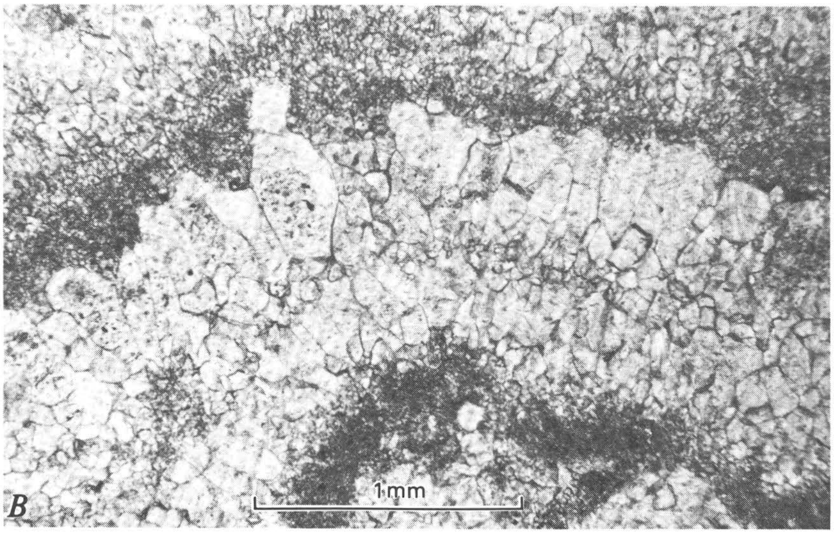
FIGURE 84.—Dredged sample of coarsely crystalline ^{13}C-depleted calcite and photomicrographs of a thin section of banded calcite: *A*, Interlayered banded vuggy calcite, green altered tuff, and angular basaltic rock fragments (No. S10-79-EG-CHAN 33C, dimensions 15×10×10 cm). *B*, Photomicrograph — plain light, and *C*, Photomicrograph — crossed nicols.

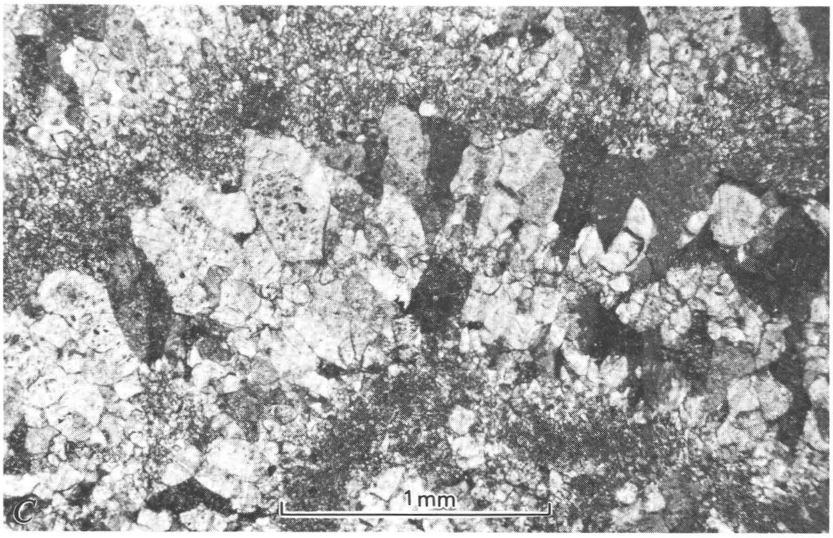
intercalated sandstone and kerogen-rich shale. Some of the sandstone in the sequence is made up of well-rounded basaltic rock and mineral fragments and contains a rich shallow-water biota that suggests a beach environment adjacent to a tropical carbonate reef.

Determinations of the ^{13}C and ^{18}O compositions of calcite in sample S10-79-EG-CHAN 33C yielded -53.2 permil PDB and +21.2 permil SMOW, respectively. The ^{13}C composition of the calcite is in the range of the methanes reported by Schoell (1980) for Eocene natural gases (-52.6 to -58.0 permil PDB) and is compatible with the early Eocene age of the dredge samples inferred from paleontologic data.

The ^{18}O composition of the calcite is not in the range reported by Keith and Weber (1964) for Eocene or younger marine limestones. It is, however, in the range for Eocone or younger freshwater limestones. If the calcite formed from seawater at isotopic equilibrium with seawater, the temperature would have been 70°C (Friedman and O'Neil, 1977).

From the carbon isotopic composition, we conclude that the calcite probably was produced by complete oxidation of Eocene methane or incomplete oxidation of methane less depleted in ^{13}C. The ^{18}O composition of the calcite indicates that oxidation may have occurred during a period of emergence because the ^{18}O is more depleted than in marine limestone. Other evidence supporting a subaerial origin includes the absence of marine organisms in the calcite, the characteristic coarse crystals and cathode





ray luminescence of freshwater carbonate, and the nature of the associated clastic sedimentary rocks, which contain a typical carbonate reef biota in beach sand.

The ^{13}C composition of the calcite indicates that the early Eocene marine sediments contain, or did contain, a source of natural gas.

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Mosaic of pockmarked sea-floor area near the Alsek River, northeastern Gulf of Alaska

By Bruce F. Molnia and Melvyn L. Rappeport

In May and June 1980, a 20-km^2 sea floor area offshore of the mouth of the Alsek River that had been previously identified as containing pockmarks, slumps, and other related sediment-failure features (Molnia and others, 1978, Molnia, 1979) was mosaicked as part of a detailed multisystem investigation of the northeastern Gulf of Alaska continental shelf. The multisystem survey of the area was run at a 100-m line spacing utilizing Mini-Ranger for navigation, 3.5- and 12-kHz echo sounders, 400 to 800-J minisparker, and 5- to 25-in^3 air-gun acoustic systems. In addition, a digital-recording and processing side-scan sonar system with slant-range correction was used to compile the 100-percent-overlap sea-floor mosaic of the area (fig. 85). Sediment samples were collected by Van Veen grab samplers and small corers from within the study area. Uniboom seismic profiles were made between the study area and the coastline.

A complete picture of the sea floor in the 10×2 km mosaicked area was made by assembling 21 speed-corrected, digitally processed, side-scan sonar lines (fig. 85). Analysis of this mosaic and related bathymetric and seismic data has delineated five sea-floor zones: (1) a northwest zone of minimal sediment disturbance characterized by isolated pockmarks and a single large slump, (2) a north-central zone of medium-density slumping; (3) a northeastern zone containing slumps and pockmarks; (4) a central zone of intensive and massive sediment disturbance characterized by blocky failures, pockmarks, bottleneck depressions, and multiple scarps; and (5) a southern area of large slumps, accumulation debris, and numerous flow lobes.

The upper few meters of sediment are failing as a result of dewatering and degassing induced by the action of one or more of the following processes: cyclic wave loading, earthquake ground shaking, rapid sedimentation, or saturation of the sediment by biogenic methane gas. Additional factors that may contribute to the sediment instability include high pore-water content and the possible presence of a slip

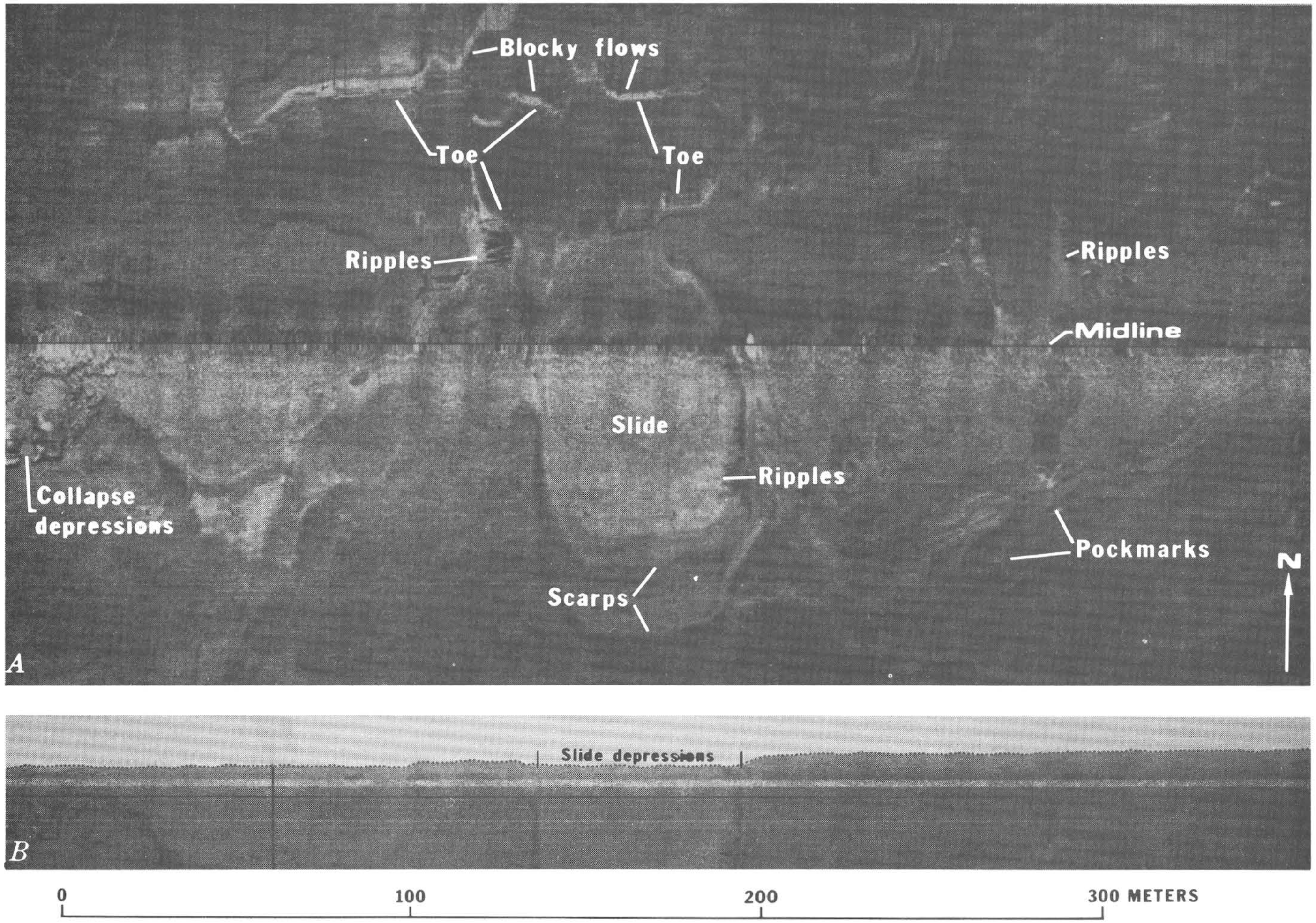

FIGURE 85.—Detail of side-scan sonar mosaic (*A*) and midline profile of mosaic (*B*) of continental shelf near Alsek River, northeastern Gulf of Alaska. Besides a prominent slide, other features that can be seen include pockmarks, ripples, and collapse depressions. Mosaic is digitally corrected to remove any lateral and slant-range distortion.

surface between the present day Alsek River sediment and an underlying, dewatered, older silty-sand and clayey-silt layer. The pockmarks, slumps, and other sediment-failure features occur on slopes as gentle as 0.4° and in water depths of 35 to 80 m. Sedimentological evidence from the cores and grab samples suggests that the regional stratigraphy in the mosaic area consists of a veneer of sand less than 1 m thick overlying a 2- to 4-m thickness of underconsolidated clayey silt with a high water content. The silt, which contains thin sand lenses, overlies a much thicker dewatered clayey silt. Minisparker and air-gun seismic data indicate that the total thickness of Holocene sediment in the mosaic area, and in the area adjacent to the mosaic, ranges from 40 to 120 m and unconformably overlies an older lithified unit. The boundary between the two units is characterized by rounded glacially eroded features and many small U-shaped channels.

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Summary and discussion of microfossil biostratigraphy in the western gulf of Alaska

By Patrick H. McClellan, Michael A. Fisher, Roland von Huene, and George W. Moore

This report summarizes the biostratigraphic results of dredging and dart-coring by the U.S. Geological Survey in the western Gulf of Alaska and discusses the tectonic significance of those results. The sampling program is aimed at documenting the ages and lithologies of deformed Neogene and Quaternary strata that crop out along the continental margin south and east of Kodiak Island (fig. 86). The program to date has included two cruises, in 1978 and 1979, aboard the R/V *Sea Sounder*, during which 42 dredge hauls and 64 dart cores were recovered from the Kodiak shelf and upper slope. Those recoveries yielded 119 datable rock samples that were analyzed for microfossils by J. A. Barron (diatoms), R. E. Arnal (foraminifers), S. Kling (radiolarians), and D. Bukry (coccoliths). The results of the analyses, published in detail elsewhere (McClellan, Arnal and others, 1980a, McClellan, Fisher and others, 1980), are the basis of the present report.

The primary target of our investigation is a shelf-break uplift about 80 km south of Kodiak Island, at the seaward margin of a bathymetric high called Albatross Bank. Records of multichannel and high-resolution seismic lines crossing the uplift (Fisher and von Huene, 1980; McClellan, Arnal and others, 1980) show it to be an anticlinal structure that trends parallel to the Aleutian trench and that is cut by a series of high-angle faults near and subparallel to the axis of the uplift. If the Aleutian trench is correctly hypothesized to be a zone of tectonic convergence between the North American and Pacific plates (for example, Atwater, 1970; von Huene and others, 1979), then the deformed strata under the Kodiak shelf, such as those near Albatross Bank, may contain a record of events related to subduction. Our study provides evidence for a chronology of deformational events that may be so related. This evidence is summarized below.

On the basis of the microfossils present in our samples, diatoms and foraminifers seem to have the greatest value to Neogene biostratigraphy in shelf and upper slope areas in the western Gulf of Alaska. Neogene rocks there are about as likely to contain diatoms (63 percent) as foraminifers (62 percent), less likely to contain radiolarians (29 percent), and least likely to contain coccoliths (7 percent). Diatoms and foraminifers, moreover, are about as likely to occur together (as in 33 percent of the samples) as they are separately (diatoms in 30 percent, foraminifers in 29 percent).

Biostratigraphic results of the fossil analyses indicate that the oldest rocks sampled on the Kodiak shelf are exposed in the breached and faulted axial zone of the shelf-break uplift at the seaward edge of Albatross Bank. Those rocks are at least middle Miocene in age and may be as old as late early Miocene on the basis of concordant diatom ages from three rock samples dredged from the upper continental slope in that area. We infer from the geographic distribution of dated sample sites near Albatross Bank and from seismic-reflection records collected there

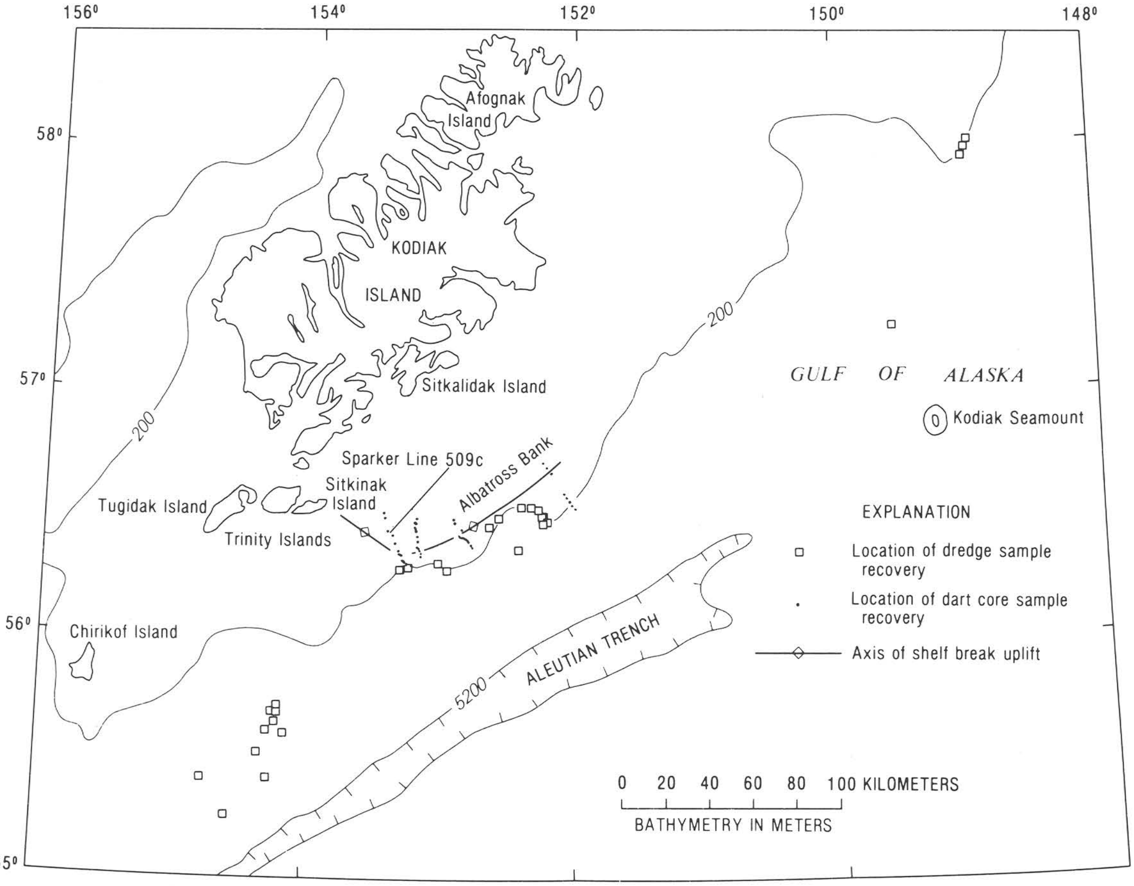

FIGURE 86.—Areas sampled in the western Gulf of Alaska showing location of seismic sparker line 509C (sparker record shown in fig. 87).

(Fisher and von Huene, 1980) that the middle Miocene strata form the basal unit of a northwest-dipping Miocene to Quaternary sequence underneath the shelf, beneath and landward of Albatross Bank (fig. 87). The middle Miocene strata also form the core of the shelf-break uplift, which is an elongate northeast-plunging structure.

Multichannel-seismic data (Fisher and von Huene, 1980) show that the axis of the shelf-break uplift has risen at least 3 km since the late Miocene. To help resolve the tectonic history of the uplift, seven dart cores from near the axis of the uplift were analyzed for paleobathymetry on the basis of foraminiferal paleoecology, by R. E. Arnal. Those cores yielded the oldest foraminifers from the Kodiak shelf. Because the depth distribution of living foraminifers in the Gulf of Alaska is poorly known, Arnal compared the fossil foraminiferal assemblages from the Kodiak shelf to living assemblages off the California coast, where a depth zonation has been established (Bandy, 1953; Bandy and Arnal, 1957, 1960; Ingle, 1967; Arnal and Vedder, 1976). To apply the depth zonation of living foraminifers off California to our high-latitude Neogene assemblages requires two assumptions: (1) that the bathymetric tolerances of the fossil species are accurately represented by the tolerances of the modern species used in the comparison—a uniformitarian assumption common to most approaches to paleoecology, and (2) that the development and fluctuation of the late Cenozoic trans-latitudinal ocean-temperature gradients did not affect the depth distribution of high-latitude foraminifers independently of the distribution of foraminifers off the California coast. We are presently investigating the validity of the second assumption through a study of modern foraminiferal ecology

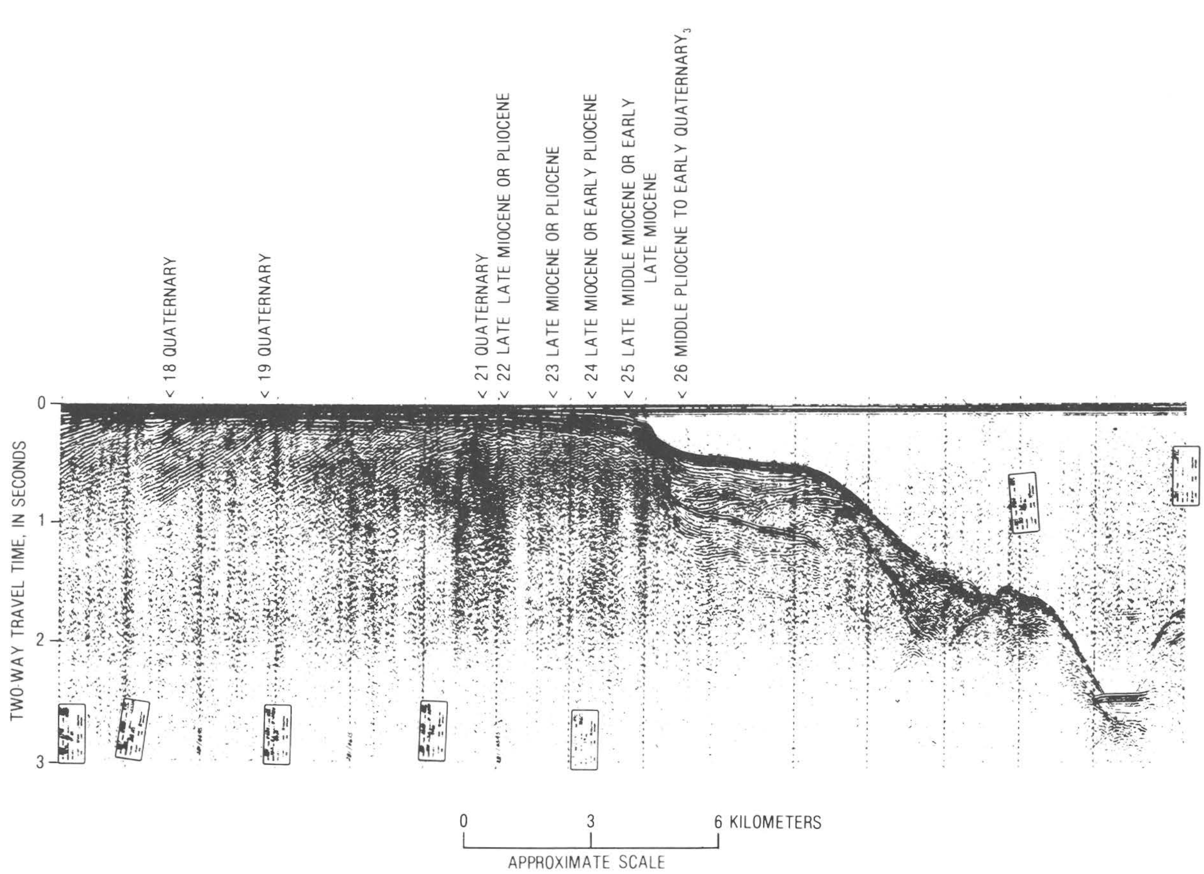

FIGURE 87.—Record of seismic sparker line 509C across Albatross Bank showing ages of microfossils recovered from dart cores. Area near site 25 is also probable source of Upper Miocene rocks dredged at shelf break (McClellan, Fisher and others, 1980). Middle (or possibly upper Lower) Miocene rocks were dredged from shelf break 10.2 km east of line 509C.

in a series of grab-samples recently collected in the Gulf of Alaska.

On the basis of the two preceding assumptions, we conclude from the paleobathymetric results (McClellan, Arnal and others, 1980) that middle Pliocene strata presently exposed near the crest of the shelf-break uplift at a water depth of about 50 m initially were deposited 3 to 5 m.y. ago at a depth of less than 500 m. Those strata subsequently subsided and were buried by about 2000 m of sediment (McClellan, Arnal and others, 1980, fig. 6) probably until the late Pliocene or Quaternary, 1 to 3 m.y. ago. The average rate of subsidence, therefore, was 2000 m over not more than 4 and not less than 1 m.y., or between 500 and 2000 m/m.y.. Sometime since the late Pliocene or early Quaternary, probably during the last 1 to 3 m.y., rocks now on Albatross Bank at the crest of the shelf-break were elevated more than 3000 m. The average rate of uplift of the shelf-break structure, therefore, was between 1000 and 3000 m/m.y.

In summary, the Miocene and later history of Albatross Bank has been characterized by high rates of vertical tectonism, including as much as 2 km of subsidence 1 to 5 m.y. ago and 3 km of uplift subsequently. If that history is related to tectonic convergence of the Pacific and North American plates along the Aleutian trench, then late Cenozoic plate convergence there may have involved changes in rate, the subduction of large topographic features such as seamounts or ridges, or other unknown processes.

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Preliminary report on ostracode assemblages from the northeast Gulf of Alaska continental shelf

By Elisabeth M. Brouwers

This study is part of a U.S. Geological Survey program to determine regions and processes of possible environmental concern to resource development in the Gulf of Alaska. A baseline datum is being established of the environmental factors that significantly control or contribute to the distributional patterns of modern ostracode species occurring on the continental shelf of the eastern gulf, from Montague Island (148°W.) to Yakutat Bay (140°W.).

The Gulf of Alaska today consists of a variety of habitats, defined by chemical and physical parameters such as water temperatures, depth and depth-related factors, salinity, sediment influxes, bathymetric features, and current and wave patterns. Biofacies are characterized by distinct assemblages of ostracode species which respond to a particular set of physical-chemical conditions. Preliminary results delineate four biofacies in the eastern Gulf of Alaska: a) middle to outer neritic, b) Icy Bay, c) Pleistocene lag, and d) shallow nearshore sand (see fig. 88).

Typical of the middle to outer neritic biofacies are the assemblages found east of Pamplona Ridge, where water depths are 86-300 m. The lithology typifying this biofacies consists of olive-green mud overlying gray mud, corresponding to the glacial marine clayey silt and silty clay facies of Molnia and Carlson (1980). The ostracodes that define this biofacies include: "*Acanthocythereis*," *Cytheropteron* spp., *Palmanella*, *Krithe*, "*Buntonia*", *Eucytherura*, *Munseyella*, *Cytheromorpha*, and *Robertsonites*. Associated organisms include benthic and planktic foraminifers (dominated by *Cassidulina* and planktic forms), pelecypods, agglutinated and proteinaceous worm tubes, echinoderm fragments, and some siliceous material (sponge spicules, diatoms, and radiolarians). At progressively greater depths, the amount of siliceous material increases and the amount of material larger than 75 μm decreases.

Three lithologies characterize Icy Bay: a) Holocene morainal material with coarse gravel, sand, and stiff clay, b) fine sand, and c) Holocene glacial marine greenish-gray mud. The morainal material is found in water depths of 23-34 m at the mouth of and outside of Icy Bay. Fine sand occurs in water depths of 31-34 m, just west of Point Riou. The greenish-gray mud occurs inside of the bay mouth in water depths of 41-67 m, and includes abundant diatoms and worm tubes; this mud represents the sediment type that is presently being deposited in Icy Bay. A distinct assemblage of species occurs in Icy Bay. The ostracode biofacies, defined by the following forms, is nearly the same in all three lithologies: *Cytheropteron* spp., *Loxoconcha*, *Semicytherura*, and *Hemicytherura*. Associated organisms include rare benthic foraminifers (*Elphidium*, *Quinqueloculina*, *Florilus*, and *Epistominella*), pelecypods, cheilostome bryozoans, and diatoms. Cirriped plates, and echinoderm spines, are also present.

Pleistocene lag deposits occur on bathymetric highs, where currents and wave action winnow out the fine-grained glacial sediments leaving rounded cobbles, gravel, and shell debris, with some sandy silt. The ostracode assemblage occurring in these lag deposits, reflecting a mixture of mild temperate to subfrigid climates and faunas, includes: *Sclerochilus*, "*Australicythere*," *Finmarchinella* spp., *Pectocythere* spp., *Bythoceratina*, *Cytheropteron* spp., *Hemicytherura*, *Aurila*, *Acuminocythere*, *Hemicythere*, *Munseyella*, *Cythere* spp., *Loxoconcha* spp., *Semicytherura*, *Xestoleberis*, *Krithe*, *Argilloecia*, *Bairdia*, *Eucythere*, *Robertsonites*, "*Acanthocythereis*", *Palmanella*, *Baffinicythere*, *Pseudocythere*, *Eucytherura*, and *Cytheromorpha*. Associated organisms include abundant encrusting and erect cheilostome bryozoans, brachiopods, pelecypods, gastropods, cyclostome bryozoans, benthic foraminifers, some planktic foraminifers. Sponge spicules and echinoderm fragments are also present. The area around Middleton Island, which lies in shallow water (43-63 m depth), is typical of the lag deposit facies.

Samples of shallow nearshore sand occurring offshore of the Copper River barrier islands yield a characteristic ostracode fauna. This is a region of active longshore drift where the sediments consist of well-sorted, dark, fine sand (the littoral and nearshore sand facies of Molnia and Carlson, 1980). Water depths above these sediments range from 18-34 m. Ostracodes characteristic of this biofacies include: *Hemicythere*, *Cythere*, *Pectocythere*, *Cytheromorpha*, *Eucythere*, *Loxoconcha*, *Aurila*, *Cytheropteron*, and "*Cytheretta*." Associated organisms include pelecypods, and occasional benthic foraminifers. Agglutinated worm tubes, plant debris, and echinoderm fragments are also present. The benthic foraminifer fauna, which is typical of shallow-water regions, includes *Elphidium*, *Quinqueloculina*, *Florilus*, *Cassidulina*, and *Biloculina*.

Recognition of biofacies and their primary limiting physical-chemical conditions is important because the appearances and local extinctions that record changes in environment (at the biofacies level) in Neogene and Quaternary sediments in the gulf of Alaska must be

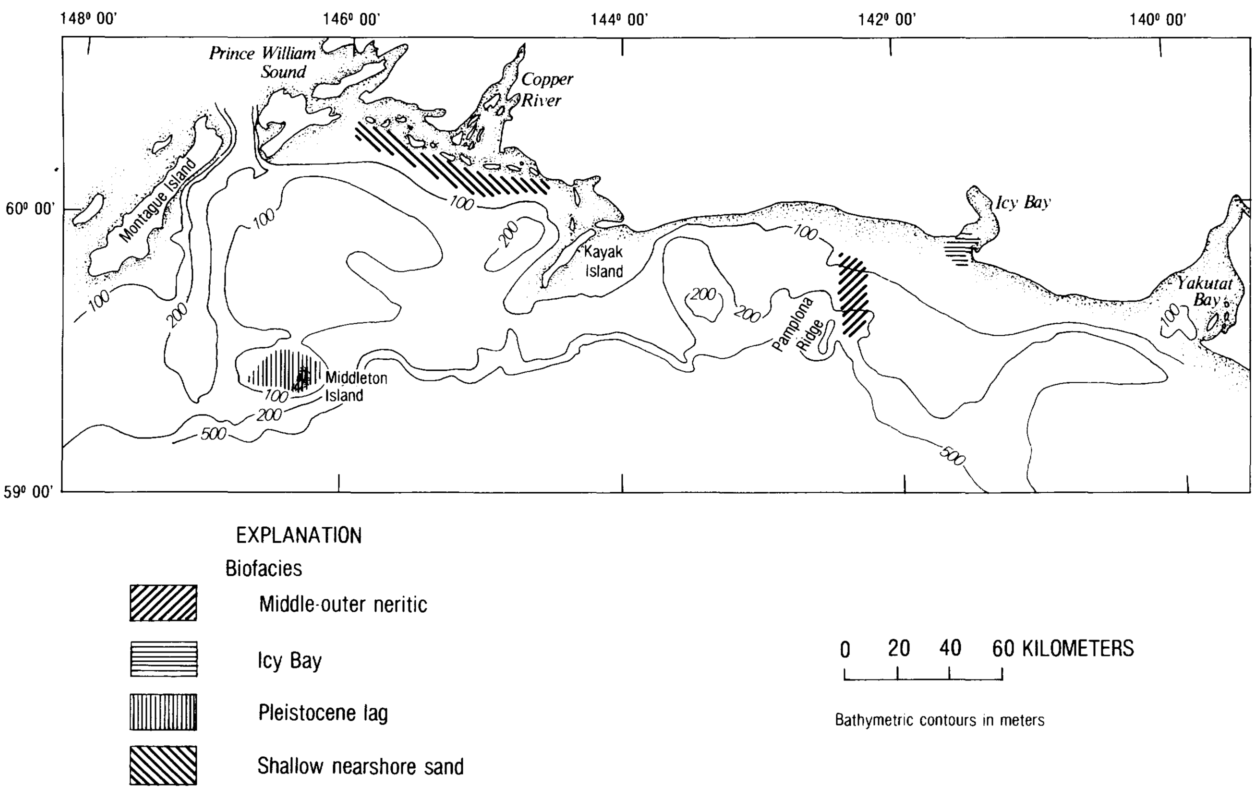

FIGURE 88.—Map of northeast Gulf of Alaska illustrating ostracode biofacies (modified from Carlson and others, 1977).

distinguished from phylogenetic changes (that is, first appearance, evolution, and final extinction).

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[Page number underscored indicates listed name is other than senior author; number in parentheses indicates listed name has authorship in more than one report on that page]

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BACK COVER

The back cover is a simplified small-scale version of a tectonostratigraphic terrane map of Alaska that is briefly described by D. L. Jones and others in the first article in this volume (p. 1-5).

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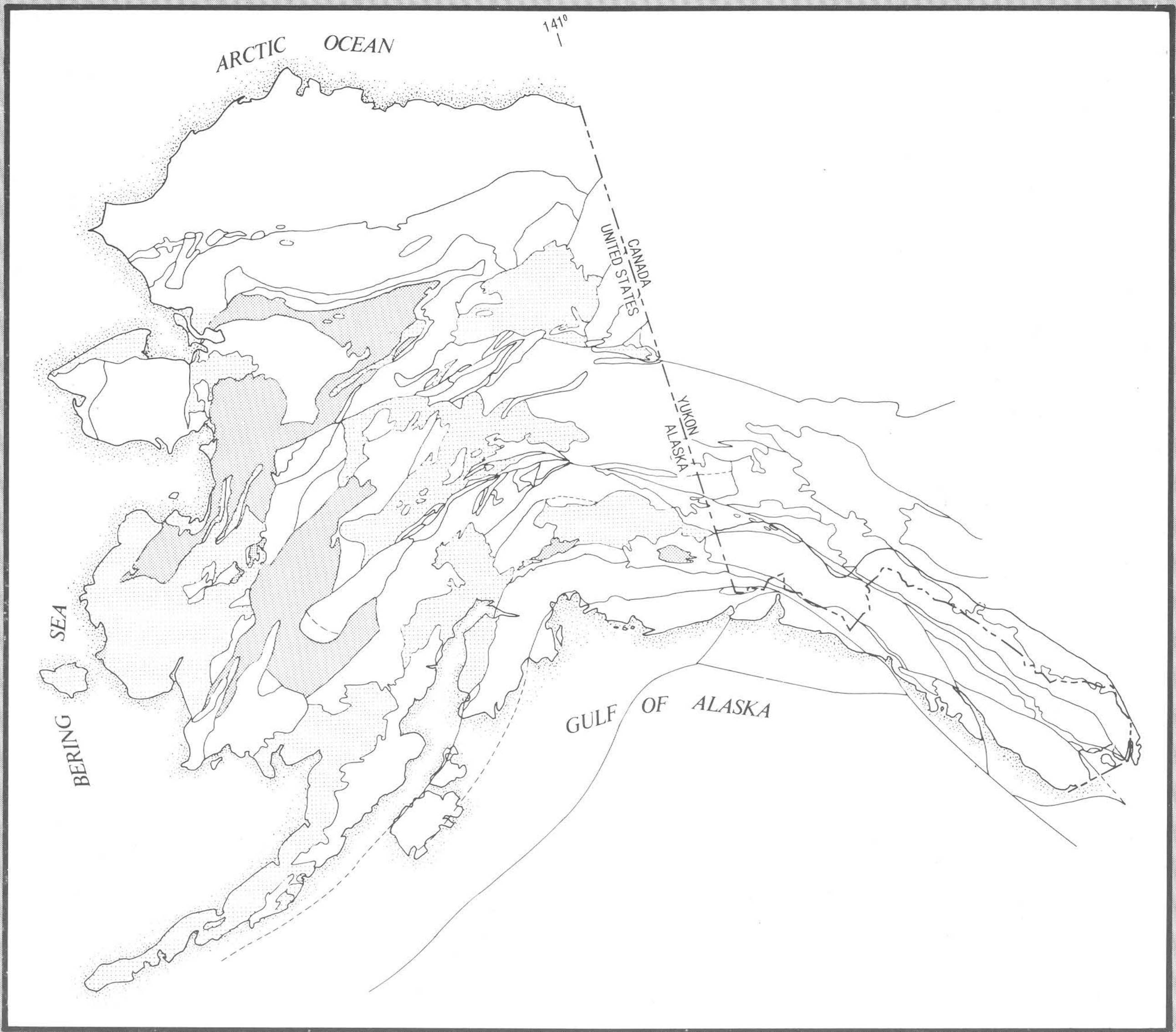
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