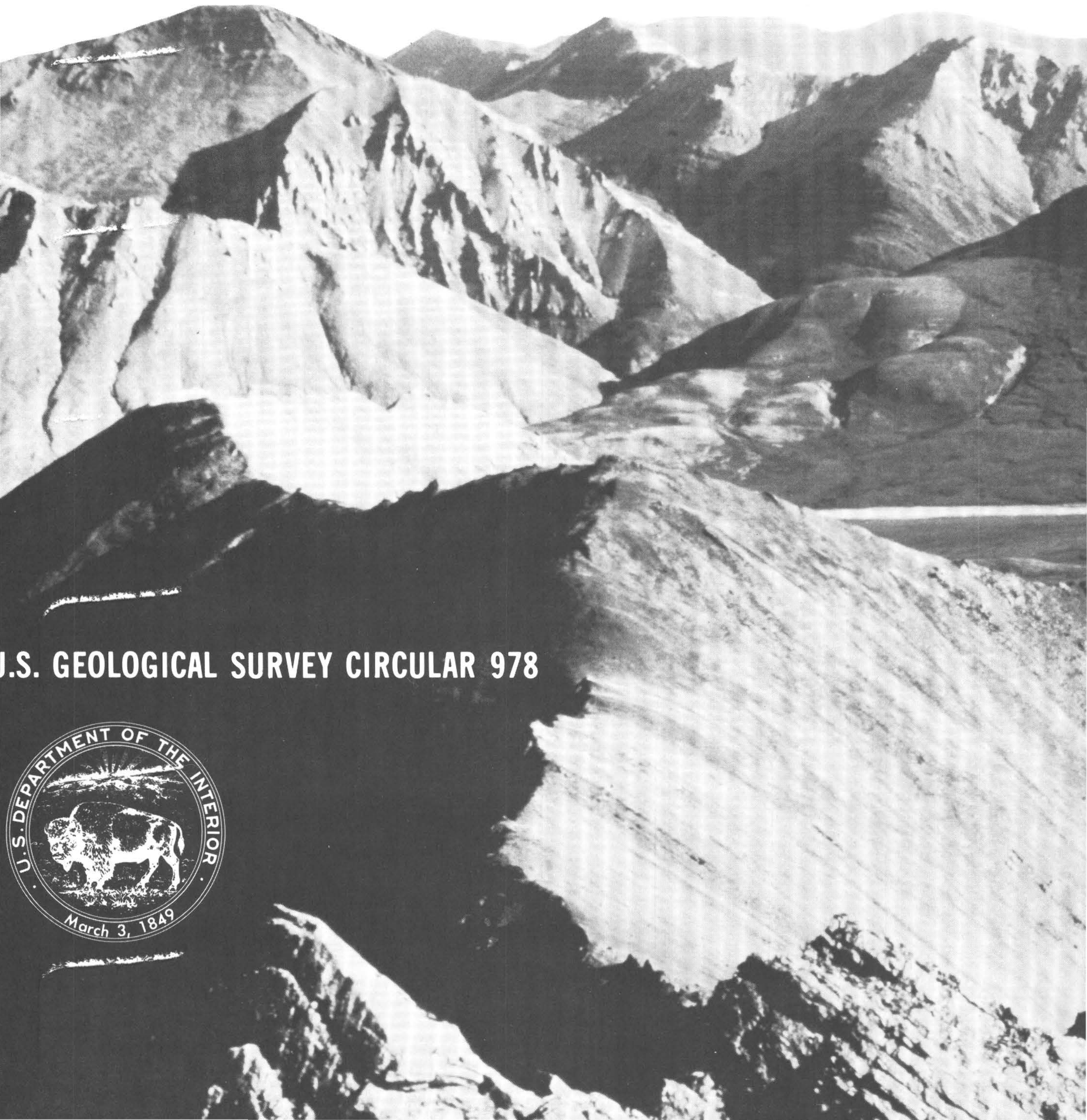
# GEOLOGIC STUDIES IN ALASKA

## by the U.S. Geological Survey

## during 1985



U.S. GEOLOGICAL SURVEY CIRCULAR 978

U.S. DEPARTMENT OF THE INTERIOR

March 3, 1849

## FRONT COVER

View to the south along the crest of an unnamed ridge in the Baird Mountains C-5 quadrangle (67°35'N., 161°51'W.) between the middle fork of the Squirrel River and the upper Agashashok River. The area is being studied under the Alaska Mineral Resource Assessment Program (AMRAP) (See Schmidt and Folger, p. 19, and Winkler and Grybeck, p. 3, this volume). The ridge is composed of brown-, buff-, or orange-weathering platy argillaceous limestone of Ordovician age that is locally bioturbated and contains diagenetic pyrite. A few kilometers to the south is a contact between these rocks and fine-grained light-gray dolostone that is also of Ordovician age and is the probable host rock for the lead-zinc Powdermilk prospect 25 km to the southeast. Preliminary results from geochemical analyses of stream-sediment samples indicate that other mineralized deposits may occur nearby. The ridge is north of the Arctic Circle and tree line and, like most areas underlain by carbonate rocks in the Squirrel River basin, is devoid of vegetation.

## BACK COVER

Status of the Alaska Mineral Resources Assessment Program (AMRAP) in January 1986, showing 1:250,000-scale quadrangles for which studies are completed, in progress, or scheduled for study in the next five years. Regions of the State that are covered by summary mineral resource folios at scales of 1:600,000 to 1:1,000,000 are outlined by bold lines.

# GEOLOGIC STUDIES IN ALASKA

# BY THE U.S. GEOLOGICAL SURVEY

# DURING 1985


Susan Bartsch-Winkler

and

Katherine M. Reed,

Editors



U.S. Geological Survey Circular 978


Short papers describing results of
recent geologic investigations
and lists of published reports


1986

DEPARTMENT OF THE INTERIOR
DONALD PAUL HODEL, Secretary

U.S. GEOLOGICAL SURVEY
Dallas L. Peck, Director

**Library of Congress Catalog Card Number 76-608093**

Free on application to Distribution Branch, Text Products Section,
U.S. Geological Survey, 604 South Pickett Street, Alexandria, VA 22304

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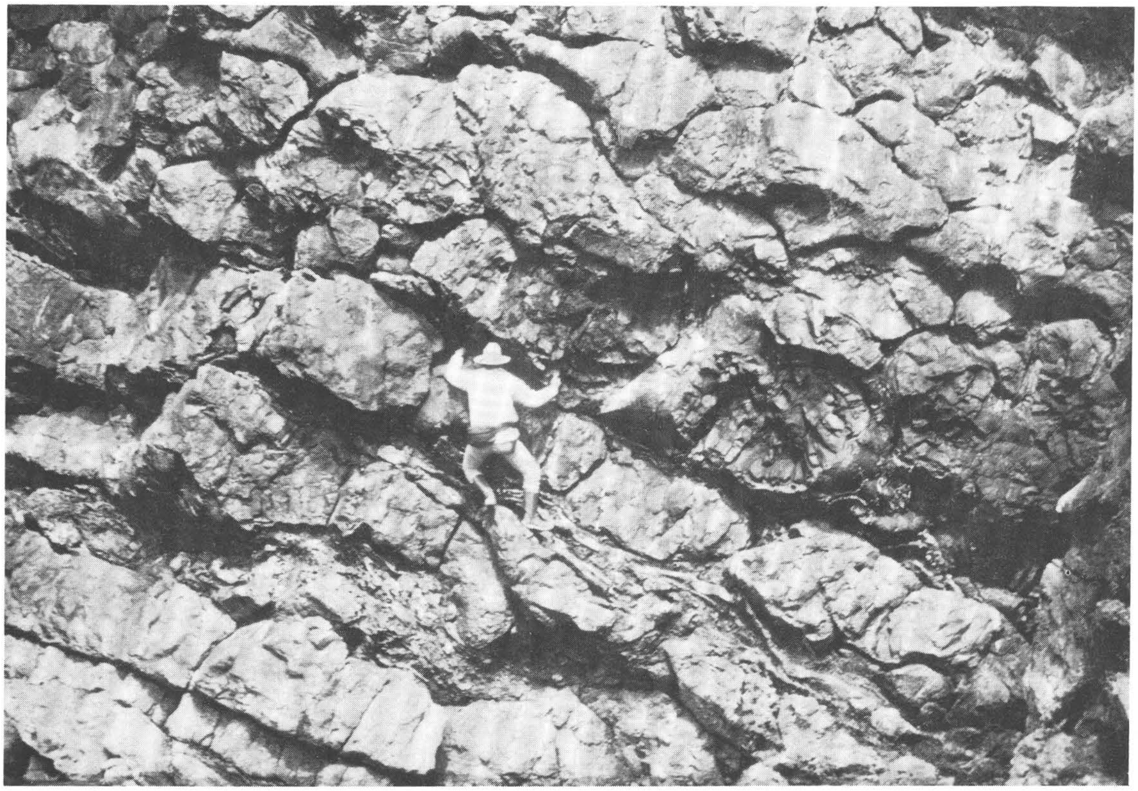

Pillow basalt near the mouth of Sawmill Bay,
Valdez Arm, Prince William Sound

# GEOLOGIC STUDIES IN ALASKA BY THE U.S. GEOLOGICAL SURVEY DURING 1985

**Susan Bartsch-Winkler and Katherine M. Reed, Editors**

ABSTRACT

This circular contains short reports about many of the geologic studies carried out in Alaska by the U.S. Geological Survey and cooperating agencies in 1985. The topics cover a wide range in scientific and economic interest. Separate bibliographic listings of published reports are included. These listings are: (1) data releases and folio components derived from the Alaska Mineral Resource Assessment Program, (2) reports on Alaska released in U.S. Geological Survey publications in 1985, and (3) reports about Alaska by U.S. Geological Survey authors in various scientific journals in 1985.

## INTRODUCTION

The U.S. Geological Survey investigates geological aspects of the onshore and offshore areas of Alaska. These studies range in scope from topical to statewide and reconnaissance regional assessments and include a broad spectrum of topics in geology.

An author index of included reports and a topical cross reference will assist

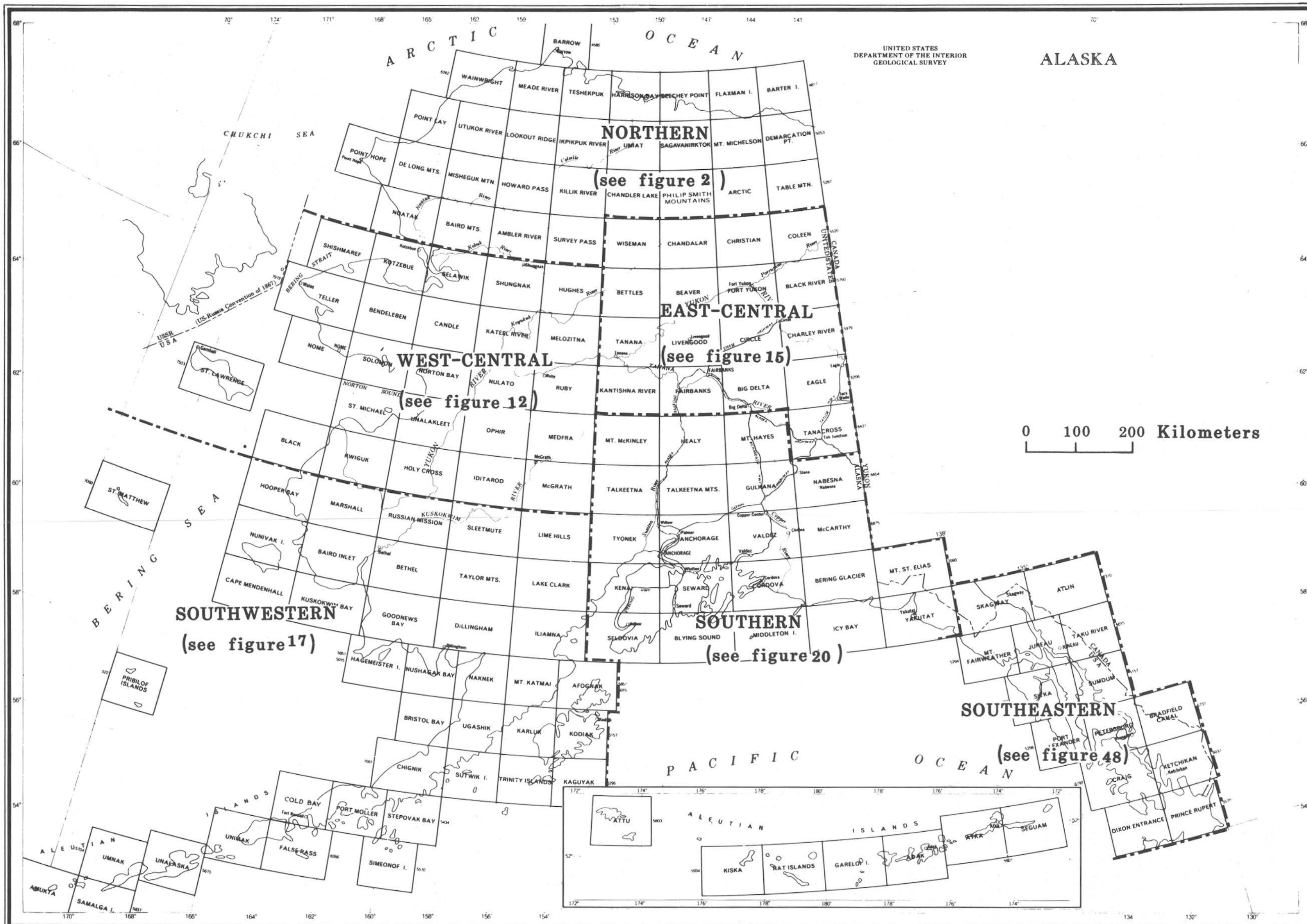


**Figure 1.**--Regions of Alaska used in this circular. Referenced figures indicate regional index maps that show areas discussed.

the reader in locating reports by specific authors or topics. Complete bibliographic listings are found at the back of this volume and include (1) folio components and data releases derived from the Alaska Mineral Resource Assessment Program, (2) reports about Alaska by the U.S. Geological Survey during the year, and (3) articles on Alaska written by U.S. Geological Survey geologists in various scientific journals published in 1985.

Reports in this circular are presented alphabetically by first author in the geographic order of occurrence: first, those investigations that are statewide in scope, and second, those that were undertaken in each of six regional subdivisions of Alaska from north to south. The subdivisions of Alaska are shown in figure 1. Index maps that show the individual study areas are included near the beginning of each regional section.

This volume is a continuation of the circular series published annually that was formerly titled "The United States Geological Survey in Alaska--Accomplishments during 19--".

# STATEWIDE

## THE ALASKA MINERAL RESOURCE ASSESSMENT PROGRAM IN 1985

**Gary R. Winkler and Donald J. Grybeck**

The Alaska Mineral Resource Assessment Program, "AMRAP", the prototypic U.S. Geological Survey (USGS) regional resource assessment program, enters its second decade with the continuing goal to provide comprehensive information on the mineral and energy resource endowment of Alaska to the public and private sectors, the academic community, and those concerned with national mineral policy. AMRAP was designed as a USGS policy response to the increased need for energy and mineral resource information engendered by passage of the Alaska Native Claims Settlement Act (ANCSA) of 1971. More recently, Section 1010 of the Alaska National Interest Lands Conservation Act (ANILCA) of 1980 recognizes AMRAP as the principal means by which the Secretary of the Interior will "assess the oil, gas, and other mineral potential on all public lands in the State of Alaska in order to expand the data base with respect to the mineral potential of such lands." AMRAP conducts four levels of study at progressively greater detail: Level I, statewide compilations (generally commodity-based) to identify the total mineral and energy resource base available for national needs; Level II, synoptic regional evaluations of geologic provinces (scale 1:600,000 to 1:1,000,000) with an analysis of probable sizes and grades of potential resources; Level III, interdisciplinary studies of key 1:250,000-scale quadrangles that are thought to have significant mineral or energy potential; and Level IV, investigations of specific mineral districts, deposits, and energy provinces, including detailed or technical studies that will aid in interpretation of their setting and character. These four interactive levels of AMRAP provide information in the short term for site-specific land-use decisions on wilderness and special purpose lands, and provide industry with relevant data to guide future exploration and development. In the long term, published products increase general knowledge of Alaska's geology and resources that will aid national strategic planning.

The so-called (d)(2) provision of ANCSA provided for about 325,000 $km^2$ of Federal lands throughout Alaska to be designated by 1979 as new or enlarged National Parks, Monuments, and Preserves; National Forests; National Conservation and Recreation Areas; Wildlife Refuges; and Wilderness Areas. These (d)(2) lands were identified tentatively soon after passage of ANCSA and were withdrawn from mineral entry. The mineral resources of these withdrawn lands were known so poorly, in general, that a derivative two-year program, the Regional Alaska Mineral Resource Assessment Program (now Level II of AMRAP), was developed to prepare syntheses of the mineral resources of almost all of Alaska prior to conveyance of most land selections by passage of ANILCA. Regional folios were published in 1977 and 1978 for all regions of the state except southeast-

**Table 1.**--List of categorized Alaska Mineral Resource Assessment Program Level-IV studies. Work is under way in Alaska on 39 research projects that support development of improved assessments in Alaska

| Category | Title | Category | Title |
| --- | --- | --- | --- |
| Areal mineral resource assessments | 1. Tin commodity studies<br>2. Placer gold deposit studies<br>3. Subjective probability estimations of Alaskan mineral resources<br>4. Mineral deposits, western Brooks Range<br>5. Metallogenesis, eastern Alaska Range<br>6. Mineral resources, interior Alaska<br>7. Geology and mineral resources, Yukon-Tanana Upland<br>8. Geology and mineral resources data summary, southwestern Alaska<br>9. Geology and mineral resources, Norton Bay-Unalakleet region<br>10. Tin and tungsten deposits, Circle district<br>11. Copper and gold deposits, Prince William Sound | Framework or process studies | 1. Yukon-Koyukuk crustal transect study<br>2. Structural analysis of interior metamorphic terranes<br>3. Mafic and ultramafic rocks of interior Alaska<br>4. Coastal sediments of upper Cook Inlet<br>5. Mesozoic stratigraphy of the Alaska Peninsula<br>6. Upper Cook Inlet-Nelchina area stratigraphic studies<br>7. Paleomagnetism of accreted terranes<br>8. Metamorphic facies map of Alaska |
| Areal energy resource assessments | 1. Oil and gas potential of interior basins<br>2. Geochemistry of sedimentary organic matter, crude oil, and natural gas in Alaska<br>3. Stratigraphy and depositional history of Jurassic and Cretaceous sequences, North Slope<br>4. Reservoir characteristics of the Lisburne Group, arctic Alaska<br>5. Coal resources of northern Alaska<br>6. Coal studies in the Nenana basin<br>7. Uranium potential of Alaskan basins | Exploration geochemical studies | 1. Gold amalgamation studies in streams<br>2. Evaluation of Department of Energy National Uranium Resource Evaluation geochemical data |
| | | Exploration geophysical studies | 1. Gravity studies (including detailed studies of Red Dog zinc-lead deposit, iron-rich Haines gabbro deposit, and chromite-bearing Red Mountain dunite)<br>2. Geophysics of Yukon-Koyukuk basin and its borderlands<br>3. Mining geophysics of central Alaska |
| Biostratigraphic studies | 1. Brooks Range and Arctic Slope studies<br>2. Paleozoic and Mesozoic Radiolaria<br>3. Brachiopod and conodont paleogeography<br>4. Cenozoic molluscan biostratigraphy, southern Alaska<br>5. Mesozoic dinoflagellate biostratigraphy, southern Alaska | Isotopic and radiometric studies | 1. K-Ar studies and radiometric age file<br>2. Zircon geochronology, interior Alaska<br>3. Lead- and oxygen-isotope studies (Seward Peninsula, Brooks Range, Alaska Range, Prince William Sound, Alaska Peninsula)<br>4. Geochemical characterization of accreted igneous arcs, southern Alaska |

ern Alaska; that report was published in 1984. The regions are outlined on the back cover, and references to the regional reports are included in the bibliography of folios produced to date by the program. (See bibliography I, p. 110.)

The current focus of the program is on Level III work. To date, fieldwork is finished in 29 quadrangles, and complete folios (including data releases, geologic, geochemical, and geophysical maps, a mineral and energy resource assessment, a summary circular, and derivative reports) are available for 19 of these quadrangles. Many individual publications are available for the other 10 quadrangles, and investigations continue in 17 quadrangles. (See back cover and bibliography I, p. 110.) More than 40 percent of the quadrangles intended for completion in the program are finished or are currently under study.

After completion of key quadrangles in a large region or geological province (such as the Seward Peninsula), a Level II synthesis is begun or updated. A regional mineral resource assessment for southeastern Alaska has been completed recently and has drawn upon detailed Level III studies. A regional assessment of the Alaska Peninsula will begin next year, and previous regional assessments of the Brooks Range and the eastern part of southern Alaska will be updated soon. Synoptic mineral and energy resource assessments will provide current information on the potential for these regions and will show where additional information is needed.

Level IV (research) investigations have always been an integral part of AMRAP. These studies include a broad spectrum of topical or detailed geological mapping, geochemical and isotopic studies, chronostratigraphic and biostratigraphic interpretation, geophysical studies, and evaluation of specific mineral or energy resource occurrences or commodities. Work is currently under way on 39 projects (table 1) that will improve mineral deposit models, appraise specific resource potential, and provide background data to aid exploration and development. Examples of these Level IV projects include: (1) detailed and comprehensive geologic, geophysical, and isotopic studies of the large, high-grade zinc-lead deposits of the western Brooks Range (including analysis of cores that have been made available by cooperating industry scientists); (2) studies of the unusual characteristics of the mineral deposits of the southern Ambler district, where the copper-base metal occurrences are rich in cobalt; (3) an evaluation of tectonic environments of terranes that host numerous base and precious metal deposits of the Delta district in the eastern Alaska Range; and (4) an analysis of geochemical environments and gold nucleation in streams of Alaskan placer districts. In addition, the future importance of Alaska's apparently large resources of such strategic commodities as tin and chromium is being addressed through detailed multidisciplinary studies on the Seward Peninsula, the southern Alaska Range, and the uplands of interior Alaska.

# NORTHERN

(Figure 2 shows study areas described.)

## AGE REVISIONS FOR THE NANOOK LIMESTONE AND KATAKTURUK DOLOMITE, NORTHEASTERN BROOKS RANGE

**Robert B. Blodgett, James G. Clough, J. Thomas Dutro, Jr., Allen R. Ormiston, Allison R. Palmer, and Michael E. Taylor**

The Katakturuk Dolomite and the probably unconformably overlying Nanook Limestone of the Shublik and Sadlerochit Mountains form a thick, dominantly platform-carbonate succession in the northeastern Brooks Range. Although age-diagnostic fossils were known previously only near the top of the Nanook, a large part of the sequence was tentatively assigned to the Middle Devonian as well (Dutro, 1970). During field studies of the region by the Alaska Division of Geological and Geophysical Surveys (ADGGS) in July 1985, fossils were found in the uppermost unit of the Nanook, unit 8 of Dutro (1970), that require significant revisions in the

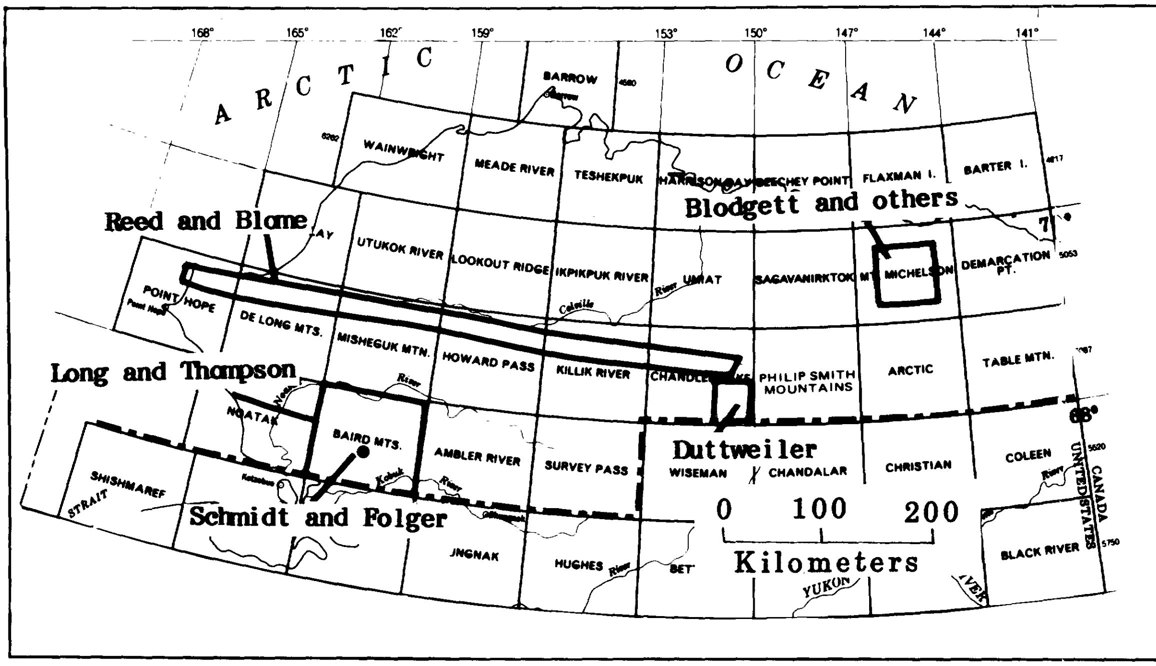


**Figure 2.**--Map showing areas in northern Alaska discussed in this circular.

age assignments for both formations. We here present preliminary results of a cooperative study of the newly discovered faunas by geologists of the ADGGS, the USGS, and other independent organizations.

All fossil localities are from exposures in the Shublik Mountains (fig. 3), which includes the type sections of both formations. A newly measured stratigraphic section of unit 8 of the Nanook Limestone is shown schematically in figure 4. This complete section is 561 m (1,841 ft) thick. The lower part of the section was measured along an unnamed south-flowing tributary of upper Nanook Creek in the Mt. Michelson C-3 and C-4 quadrangles. Its base is at the north end of the section line, and the upper part was measured up the hillside to the east of the stream (fig. 3).

Devonian strata are now recognized in unit 8 only towards the eastern end of the Nanook Limestone outcrop belt in the Shublik Mountains; farther west, Devonian strata are absent due to erosion predating the deposition of the Early Mississippian Kekiktuk Conglomerate. Devonian rocks below the unconformity thicken toward the eastern end of the outcrop belt where they are as thick as 107 m (350 ft). Abundant fossils include: rugose and tabulate corals, stromatoporoids, brachiopods, trilobites, tentaculitids, and crinoid ossicles. The crinoid ossicles include two-holed forms (Gasterocoma? bicaula) that reached their acme in rocks of late Emsian to Eifelian age in western and Arctic North America (Johnson and Lane, 1969). Rugose and tabulate corals from four localities in this unit were listed by Oliver (Oliver and others, 1975, table 10 and p. 28). Several paleoenvironments are represented in this interval, including Amphipora-rich lagoonal deposits, coral-stromatoporoid biostromes, and rough-water banks with pentameroid brachiopods. Parts of this interval are lithically similar to coeval parts of the Salmontrout Limestone of the Porcupine River area, east-central Alaska. Although analysis of the Devonian megafauna is in its preliminary phase, the fauna is of late Early (Emsian) and (or) early Middle (Eifelian) Devonian age. This Devonian sequence, separated from the thicker, pre-Devonian part of unit 8 by an angular unconformity, may necessitate the adoption of a separate formational name

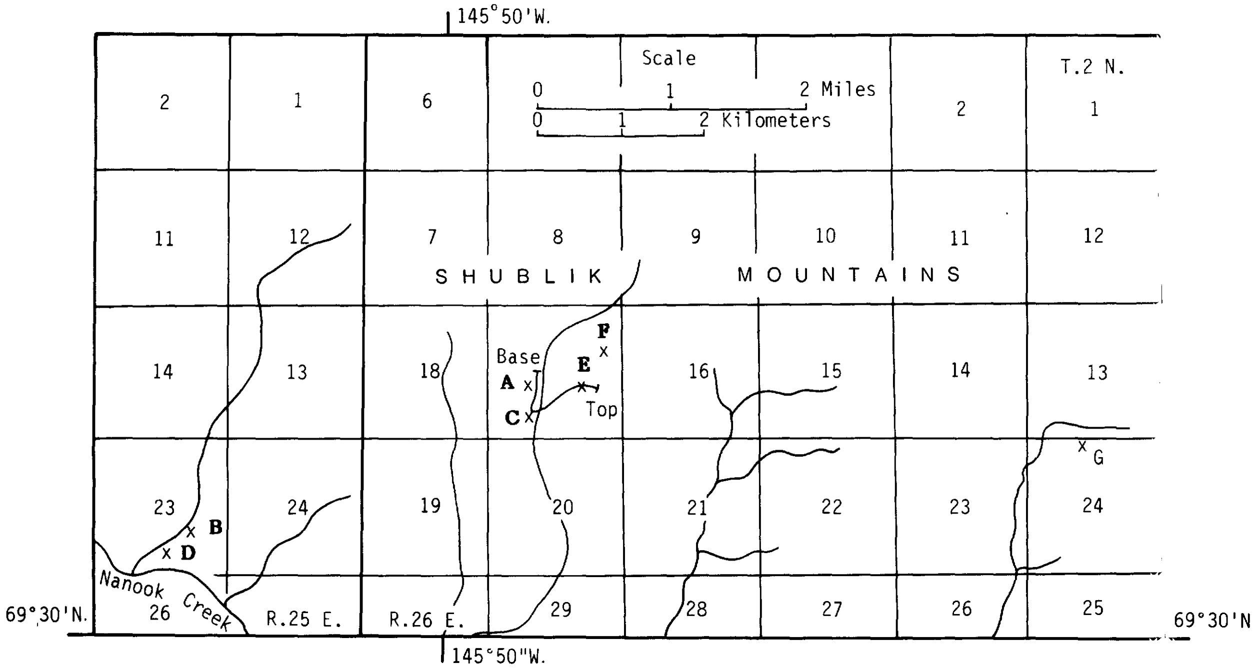


**Figure 3.**--Location of fossil localities and measured stratigraphic section of figure 4. Parts of Mount Michelson C-3 and C-4 quadrangles.

after further study and detailed geologic mapping.

The pre-Devonian part of unit 8 is as much as 534.4 m (1,753 ft) thick. The lower and thicker part of the unit consists of peloidal limestones containing Late Cambrian and Early Ordovician trilobites. Late Cambrian trilobites were recovered at 108.2 m (355 ft) above the base of unit 8 in the newly measured stratigraphic section (locality D3455-CO, A in fig. 3) and belong to a single species identified as Plethometopus armatus (Billings). At another locality (D3458-CO, B in fig. 3), two poorly preserved Late Cambrian trilobites were identified as Plethometopus? sp. Locality D3458-CO is from near the base of unit 8 on the east side of the unnamed creek along which Dutro's (1970) section B was measured. Early Ordovician trilobites were collected at five horizons in unit 8. In the measured section (fig. 4), trilobites occur in three samples (shown collectively as locality C in fig. 3) from oldest to youngest:

--Locality D3454-CO: cf. Hystricurus? sainsburyi Ross, 188 m (617 ft) above the base of unit 8.

--Locality D3453-CO: cf. Hystricurus? sainsburyi Ross, Hystricurus species undetermined; 199.6 m (655 ft) above the base of unit 8.

--Locality D3452-CO: cf. Hystricurus? sainsburyi Ross, Hystricurus cf. H.? sp. E of Ross (1951); 201.2 m (660 ft) above the base of unit 8.

A correlative stratigraphic interval is represented by two closely spaced collections (shown collectively as locality D in fig. 3) from near Dutro's (1970) section B. Collection D3457-CO yielded cf. Hystricurus? sainsburyi Ross and Hystricurus cf. H.? species E of Ross (1951). Collection D3456-CO from 2.4 m (8 ft) higher in the section contains cf. Hystricurus? sainsburyi Ross, Hystricurus cf. H.? species E of Ross (1951) and another undetermined species of Hystricurus. The collections from localities C and D (fig. 3) are all Early Ordovician (Canadian) in age. The Early Ordovician samples also contain unidentified low- and moderately high-spired gastropods, and conodonts.

The paleobiogeographic affinity of the Late Cambrian fauna is with shallow, warm-water carbonate platform sites around the North American craton. The Early Ordovi-

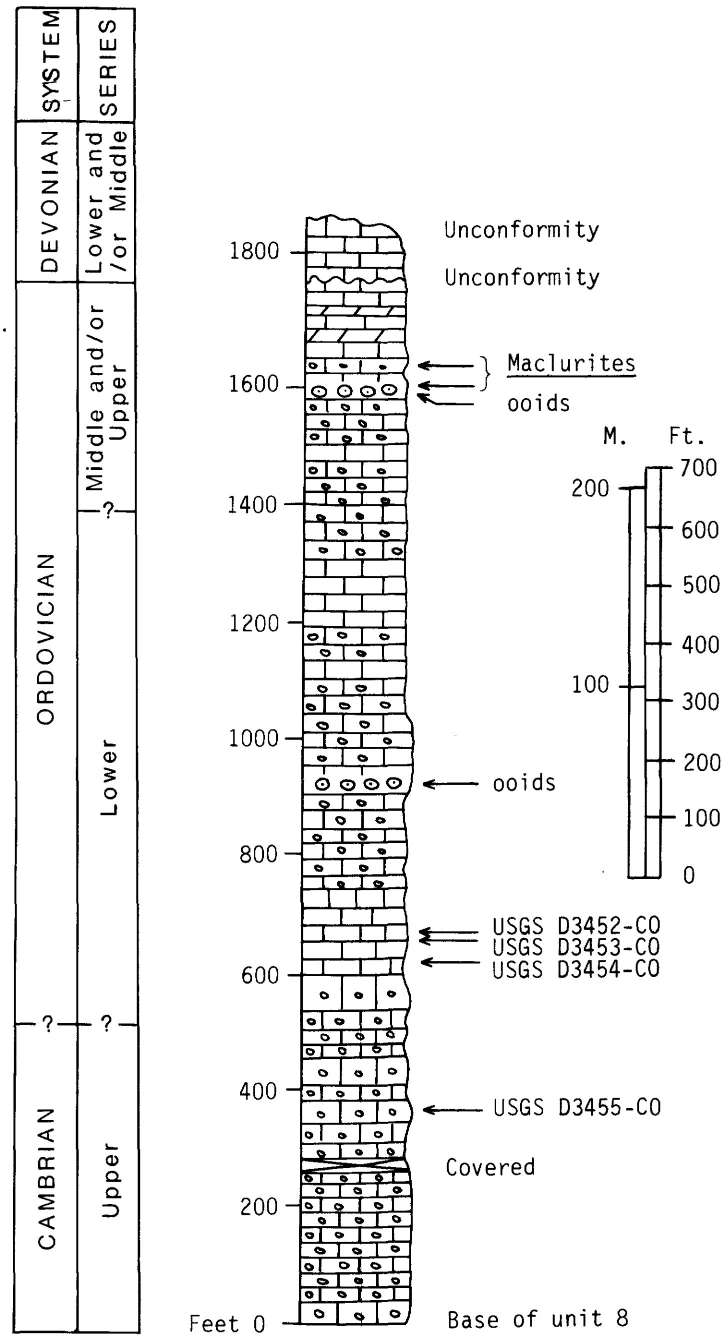


**Figure 4.**--Stratigraphic section of unit 8 of Nanook Limestone measured by ADGGS, July 1985. Section located in sec. 17, T. 2 N., R. 26 E; see figure 3 for location. Carbonate rock types: 0-260 ft, medium-bedded peloidal pack/grainstone; 278-600 ft, medium- to thick-bedded peloidal pack/grainstone; 600-725 ft, non-peloidal lime mudstone/wackestone; 725-1,650 ft, medium-bedded peloidal pack/grainstone, two prominent ooid levels at 925 ft and 1,600 ft, and peloids absent in several intervals (about 1,080 ft, 1,150 ft, 1,200-1,300 ft, and 1,475 ft); 1,650-1,750 ft, interbedded dolomite and lime mudstone; 1,750-1,763 ft, lime mudstone.

cian trilobite faunas have closest affinities with the Bathyurid Province, which occupied low-paleolatitude Early Ordovician shallow-water platform sites in North America, Greenland, northeastern USSR, and Kazakhstan (Whittington and Hughes, 1972). Early Ordovician Bathyurid Province faunas are also present in the Lost River area of the Seward Peninsula, Alaska (Ross, 1965), and in undescribed USGS samples listed by Brabb (1967) and Palmer (1968) from the Jones Ridge Limestone of east-central Alaska.

The gastropod genus Maclurites, indicative of either Middle or early Late Ordovician age, was found at two localities associated with oolitic beds just below the occurrence of dolomite in the upper part of the pre-Devonian part of unit 8. These two localities are (1) two collections of the genus in the measured section at locality E (fig. 3), 487.7 m (1,600 ft) and 495.3 m (1,625 ft) above the base of unit 8; and (2) Maclurites in moderate abundance at locality F (fig. 3). At locality G (fig. 3), Lower Silurian beds may also be present beneath the pre-Devonian unconformity. Ormiston identified three trilobites species, two of which (Platylichas aff. P. sp. C Lane and Encrinurus aff. E. sp. F Lane) closely resemble species described by Lane (1979) from the Lower Silurian (Llandovery) of northern Greenland.

No body fossils were recovered from any of the lower units of the Nanook Limestone. The report of possible tabulate corals and Amphipora? in units 1 and 2 (Dutro, 1970; Reiser and others, 1970) was in error. Regional correlations suggest that these beds may be of older Cambrian or Late Proterozoic age. Unit 1 of the Nanook, the red-marker unit of Dutro (1970), contains maroon and green shale, a very common lithology in strata of Late Proterozoic and Early Cambrian age in the Yukon Territory and east-central Alaska. Thus, we regard the overall age of the Nanook Limestone to be Late Proterozoic or Early Cambrian to Early and (or) Middle Devonian and consider the Katakturuk Dolomite, which is as much as 2,500 m thick in the Sadlerochit Mountains, to be Proterozoic in age on the basis of its stratigraphic relations with the overlying Nanook Limestone. The Katakturuk, deposited in a shallow-water platform setting, includes cross-bedded oolitic grainstone, herringbone cross-bedded tidal channel conglomerate, cryptalgal laminite, zebra dolomite, and dolomitic mudstone with mudcracks. Stromatolite morphologies found in the Katakturuk include: discrete, vertically-stacked columns and hemispheroids, laterally-linked large mounds, spheroids (oncolites), and irregular to flat mats (Clough, 1986).

These age revisions significantly modify our interpretations of the early depositional history of northern Alaska. For example, this is the first record of Ordovician shelly fossils from northern Alaska. In addition, future paleogeographic reconstructions must take into account the presence of a major Proterozoic to middle Paleozoic carbonate platform in the northeastern Brooks Range. Coeval strata to the south and east in the Mt. Michelson and Demarcation Point quadrangles consist of slope and basin deposits, including graywacke, slate, phyllite, and volcaniclastic rocks, as well as some carbonate rocks (Reiser and others, 1971, 1980). These age revisions for the carbonate sequence in the Shublik and Sadlerochit Mountains conform to the general model of a northern source area during the Paleozoic (Dutro, 1981). This new information extends the time frame of this depositional model backward into the late Precambrian.

Deposition was punctuated by at least two major unconformities during the early and middle Paleozoic, as recognized earlier in the Demarcation Point and Arctic quadrangles (Reiser and others, 1980; Brosgé and others, 1981). The younger (pre-Kekiktuk) unconformity is partly correlative with the Ellesmerian orogeny of the Canadian Arctic Islands and may represent a westward continuation of this major tectonic disturbance. The older (pre-Devonian) event may reflect a Taconic or early Caledonian orogenic phase. Three Precambrian unconformities were also recognized in the Demarcation Point quadrangle to the east (Reiser and others, 1980).

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## SULFIDE OCCURRENCES IN THE ITKILLIK RIVER REGION, SOUTHEAST CHANDLER LAKE QUADRANGLE, BROOKS RANGE


**Karen A. Duttweiler**


A geochemical investigation of the Chandler Lake quadrangle, northern Brooks Range, has been under way since 1981 as part of the Alaska Mineral Resource Assessment Program. Results of analyses of stream-sediment and nonmagnetic heavy-mineral-concentrate samples have been published in Barton and others (1982) and Sutley and others (1984). As a result of this geochemical work, three areas of sulfide mineral occurrences were identified in the Itkillik River region in the southeast part of the quadrangle (fig. 5). The areas were first identified from heavy-mineral-concentrate samples that contain anomalous amounts of lead, zinc, copper, and silver, as well as abundant pyrite, arsenopyrite, chalcopyrite, sphalerite, and galena. Detailed follow-up work has revealed the nature of the mineralization.

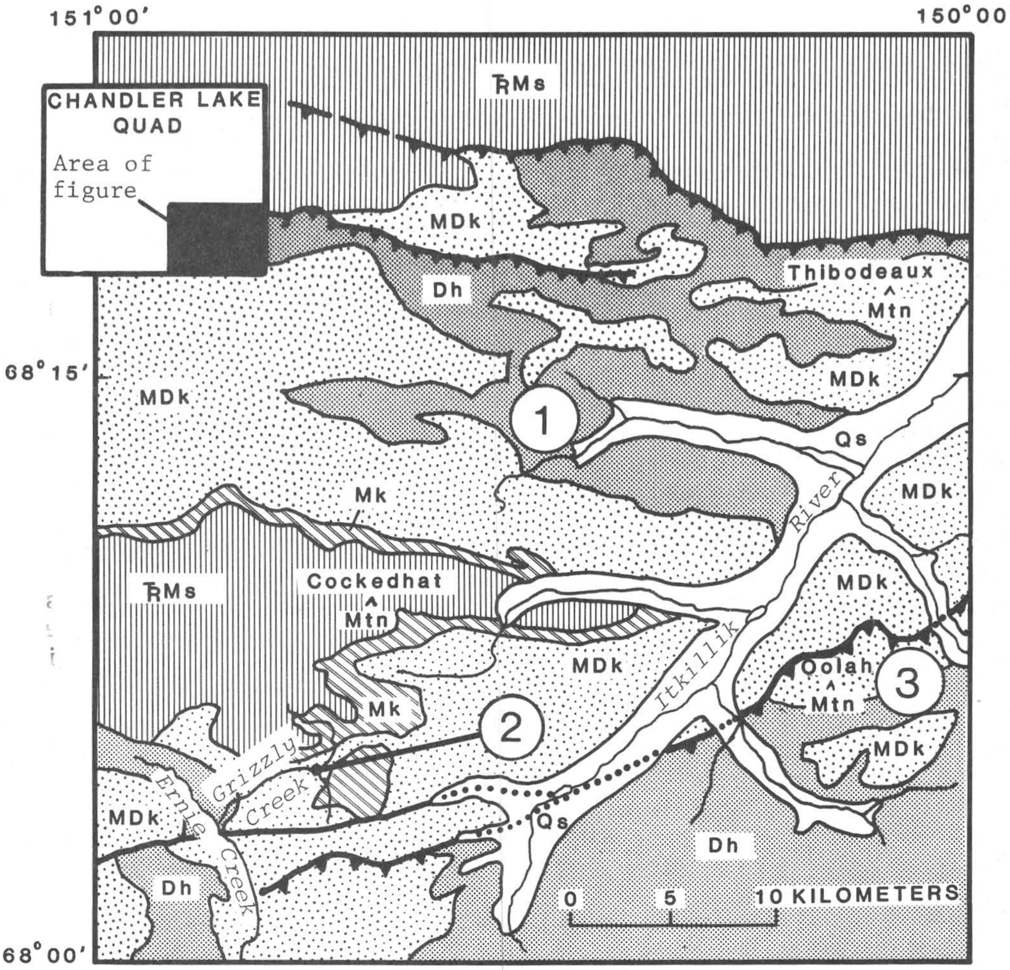
The Itkillik River region is underlain by Devonian to Triassic sedimentary rocks; these rocks are unmetamorphosed to weakly metamorphosed and intensely folded and thrust faulted (Brosgé and others, 1979). The Upper Devonian Hunt Fork Shale consists of a dark-gray shale with light-brown-weathering calcareous sandstone and siltstone interbedded with a gray-green limonitic wacke. The marine strata of the Hunt Fork Shale are overlain by a nonmarine sequence, the Kanayut Conglomerate. In most places, the Late Devonian and Early Mississippian(?) Kanayut Conglomerate (Nilsen and Moore, 1984) contains a coarse middle member of interbedded conglomerate and sandstone; the upper and lower members consist of sandstone, siltstone, and shale. All three members represent deposition in a deltaic environment. Overlying the Kanayut Conglomerate are fossiliferous marine strata of the Lower Mississippian Kayak Shale, which consist of shale, siltstone, and shaley sandstone interbedded with argillaceous and ferruginous limestone. Above that unit are thick gray limestone and dolomite of the Lisburne Group (Mississippian) and minor amounts of siliceous shale and siltstone of Mississippian to Triassic age. More detailed descriptions of these geologic units are found in Brosgé and others (1979), Kelley (1984), and Nilsen and others (1979).

Geochemical and mineralogical data from heavy-mineral-concentrate samples were useful in identifying three occurrences of sulfide minerals in the Itkillik River region. Concentrate samples from each of the three localities contain between 1,000 and 15,000 ppm Pb, 700 and 20,000 ppm Zn, 500 and 1,500 ppm Cu, and 1 and 20 ppm Ag (Sutley and others, 1984). These values are within the highest 5 percent of those recorded and are considered anomalous. As much as 1,500 ppm As is reported in many, but not all, of the anomalous samples. A high barium content (and corresponding barite occurrence) accompanies the lead, zinc, copper, and silver in nearly all samples, but is ubiquitous in the entire region and, therefore, is not a good indicator of possible mineralization.

The mineralogical data support the geochemical data in that the concentrates with anomalous Pb, Zn, Cu, and Ag also contain pyrite (up to 50 percent), chalcopyrite (trace), pale-yellow sphalerite (1-5 percent), and galena (trace to 1 percent). Many samples also contain arsenopyrite. A strong correlation between silver and lead in the geochemical data indicates that the silver is probably contained in the galena.

Follow-up studies revealed that galena, chalcopyrite, pyrite, and sphalerite are hosted in Upper Devonian sedimentary rocks, occurring either in small veins or disseminated in porous sandstones and conglomerates. At locality 1 (fig. 5) exposures are primarily in the Hunt Fork Shale. Galena, chalcopyrite, and pyrite occur in small discordant quartz-calcite veins that cut the Hunt Fork Shale units. In spite of its presence in concentrates from this locality, sphalerite is not observed in the veins. Other vein minerals include siderite and locally abundant limonite after siderite. The veins are generally 1-2 cm wide but reach widths of 5-6 cm. Several episodes of veining are evident by the offset and crosscut relations of the veins in their host rocks.

Locality 2 is south of Cockedhat Mountain at the headwaters of Grizzly Creek (fig. 5). Sulfide minerals occur in veins similar to those at locality 1, except minor sphalerite occurs with galena and pyrite in a gangue of quartz, calcite, and siderite. In addition to the vein occurrences here, sulfides are widely disseminated in coarse conglomerate and sandstone of the Kanayut Conglomerate. Oxidation of the sulfides has resulted in discontinuous red-weathering layers in the Kanayut Conglomerate. The distinctive oxidized layers are common throughout most of the area of locality 2 and to the west, in-




| | |
|---|---|
| Qs | Surficial deposits (Quaternary)--Alluvial, lacustrine, and glacial deposits |
| TRMs | Sedimentary rocks (Triassic to Mississippian)--Limestone and dolomite of the Mississippian Lisburne Group and Mississippian to Triassic shale and siltstone |
| Mk | Kayak Shale (Mississippian) |
| MDk | Kanayut Conglomerate (Mississippian(?) and Devonian) |
| Dh | Hunt Fork Shale (Devonian) |

Contact

Fault--dotted where inferred

Thrust fault--sawteeth on upper plate. Dashed where approximately located; dotted where inferred.

**Figure 5.**--Generalized geologic map of the Itkillik River region, modified from Brosgė and others (1979). Circled numbers are locations referred to in text.

cluding the area west of Ernie Creek. Several rock samples containing disseminated sulfides were crushed, sieved, and panned to determine which sulfide minerals were present, but pyrite and minor arsenopyrite were the only sulfides identified. However, bulk samples of the rocks contain 80-200 ppm Zn, with accompanying high values of Ag, As, Cu, and Ba. These geochemical data suggest the presence of other sulfides, such as chalcopyrite and sphalerite.

Locality 3, east of the Itkillik River and Oolah Mountain, is approximately at the contact between the Hunt Fork Shale and the Kanayut Conglomerate. Numerous small quartz veins 1-2-cm wide contain abundant pyrite (partly oxidized) and minor chalcopyrite. The shale beds in the Hunt Fork Shale host very fine grained disseminated pyrite. The coarse sandstone and conglomerate strata in the Kanayut Conglomerate here contain the same disseminated sulfide minerals as at locality 2.

Sulfide mineralization in the Itkillik River area may be the result of low-grade metamorphism of the shale and sandstone lithologies in the Hunt Fork Shale. The thick and porous strata of the overlying Kanayut Conglomerate are excellent permeable host rocks for upward-migrating fluids derived from compaction of the Hunt Fork Shale. Vein mineralization in the Itkillik River region does not seem to be localized in shear zones or systematically related to any major structural features. More detailed study of the polymetallic sulfide occurrences is required before the characteristics of the mineralization are fully understood.

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Reviewers:
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## AUDIO-MAGNETOTELLURIC RESISTIVITY TRAVERSES IN THE BAIRD MOUNTAINS QUADRANGLE


**Carl L. Long and Bill Thompson**


Audio-magnetotelluric (AMT) soundings in the Baird Mountains quadrangle were made during two weeks in July 1985 during the course of a mineral resource study. Sixty AMT soundings were made along three traverses, for a total distance of about 120 km at an average spacing of approximately 2 km (fig. 6). During this time, helicopter-borne aeromagnetic lines also were flown coincident with and parallel to

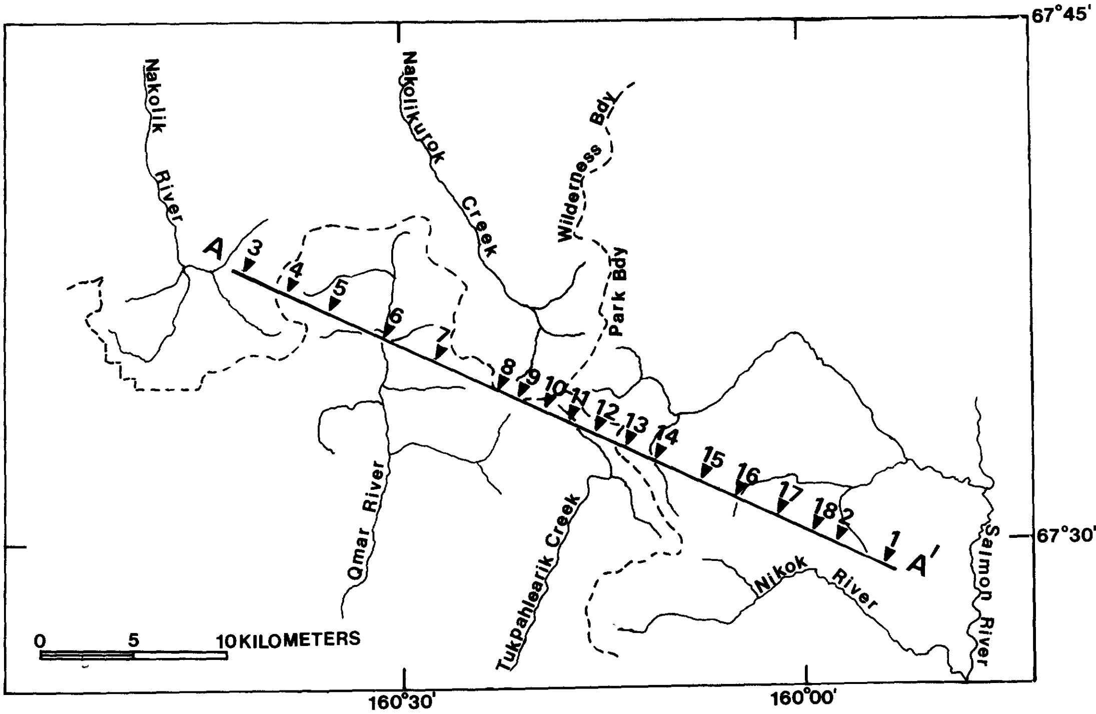


**Figure 6.**--Map showing station numbers, location of resistivity section, and coincidental aeromagnetic line A-A'.

the first two traverses, and gravity stations were established at or near each AMT station on all traverses.

The object of the AMT traverses was to define electrical resistivity changes laterally and vertically and to relate those changes to lithology and structure. Gravity and magnetic data from a large part of the area traversed suggest only small variations in density and magnetic properties; therefore, the AMT sounding method was utilized to help resolve the complex mapped geologic structures.

The AMT system used in this survey was designed and built by the USGS (Hoover and others, 1976) for use in reconnaissance geothermal and mineral surveys. Natural electromagnetic (EM) signals (sferics) generated by distant lightning storms are used to measure apparent resistivity as a function of frequency, from which conductance (resistivity-thickness product) of a layered-Earth model may be determined (Cagniard, 1953). The USGS system uses 50-m dipoles for the electric field sensor and a ferrite-core coil for the magnetic field sensor. At each station, simultaneous soundings are made with the electric field sensors oriented north-south and east-west and with the corresponding magnetic sensors oriented east-west and north-south, respectively. The naturally occurring EM fields are measured at 16 frequencies ranging from 4.5 Hz to 27 kHz, corresponding to an exploration depth range, for this survey, of roughly 0.2-6 km. In most instances, 10 scalar apparent resistivity values are averaged for each frequency, and a standard deviation is computed. Details of the data processing and interpretation are given by Hoover and others (1978) and Long (unpub. computer programs). Theory and applications of the AMT method to mineral exploration have been described by Strangway and others (1973), Hoover and others (1978), and Long (1983).

Interpretation of AMT data depends on the resistivity contrast between rock units, the geometry of the units, and the frequencies employed. Earth resistivities

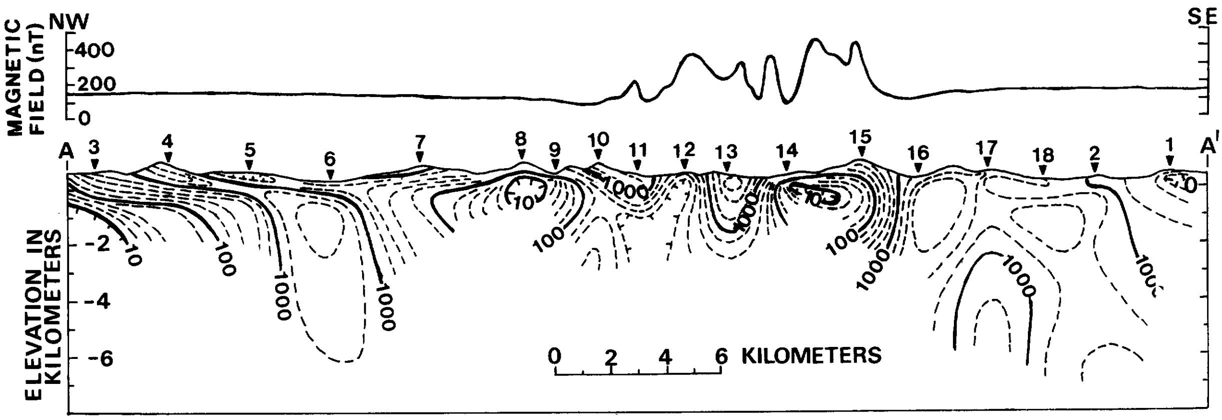


**Figure 7.**--Magnetic profile (above) and resistivity section (below) along line A-A' shown in figure 6. AMT station locations are indicated by solid numbered triangles. Resistivity contours are five even intervals per decade (i.e., 10, 16, 25, 40, and 63) in ohm-meters. Only the primary contours are labeled. Horizontal and vertical scales are equal on the resistivity section.

interpreted from AMT soundings reflect bulk rock resistivities averaged over large volumes of material. The primary factors affecting rock resistivity are bulk porosity of the rock, including intergranular pores as well as microscopic and macroscopic joints, and salinity of the pore fluids. Rocks containing large percentages of clay commonly have low resistivity because porosity increases and surface conduction effects become dominant. In low-porosity rocks, such as some igneous and metamorphic rocks, electrical conduction is mainly by ionic conduction within fluids filling interconnected joints, cracks, and faulted zones. Resistivities are high where such structures are absent. Rocks containing a significant percentage of metallic sulfide minerals (massive sulfide deposits) may have low resistivity due to electronic conduction. A more complete description of rock resistivities is given by Keller (1982) and Telford and others (1976).

Figure 7 is a resistivity cross-section of AMT data collected in this study. The section is derived from inversions of the AMT sounding curves using the Bostick (1977) computer algorithm. The inversion process produces a horizontally layered resistivity-depth model whose calculated effects match the data recorded at each station to an acceptable degree of fit. Resistivity models were computed for the NW-SE profile A-A' (fig. 6) and were hand contoured and then plotted along with the data from the coincident aeromagnetic profile (fig. 7).

Preliminary results of AMT resistivity-depth models suggest that major high and low resistivities and varying resistivity gradients are indicative of changes in lithologies and of geologic structures. The magnetic profile from AMT stations 3 to 10 indicates little change in magnetic character of the rocks, but the resistivity cross-section has several notable features. Resistivities from stations 3 to 5 in the first 2 km of the crust suggest horizontal layering. At station 5, the near-surface, flat, low resistivity unit overlying the enclosed higher resistivity unit at about 0.5 km depth suggests that two different lithologies are present there. The broad, closed, high-resistivity anomaly below station 6 extends downward approximately 6-7 km and has a shallow lateral arm that extends northwestward to the area below station 4; this arm is possibly a continuous geologic unit from stations 4 to 6. The resistivity high is bounded by steep gradients between stations 5 and 6 and 6 and 7, indicating possible near-vertical faults or steeply dipping lithologic contacts at depth on both sides of the anomaly. The significant 10-ohm-m closed resistivity low under station 8 may be related to several possible features, including mineralized or altered rock, the occurrence of shale, and presence of clay, and also illustrates that the depth of exploration by the AMT method decreases as resistivity decreases.

Magnetic anomalies from AMT stations 10 to 16 are well defined. Therefore, a combination of magnetic and resistivity data may be used in this part of the section to determine the geologic structure. Between stations 10 and 11, the 1,000-ohm-m contour indicates a dipping unit bounded on either side by lower resistivities. Similarly, the nearly vertical gradient on either side of station 13 defines a resistive unit about 2 km wide, which may contain at least two different lithologies (according to the magnetic signature). From stations 14 to 15, a 10-ohm-m low is nearly flat lying and indicates lithology similar to that at station 8. Magnetic and resistivity data have contrasting characteristics from stations 16 to 1. Magnetic contrast in the rocks is low, and the resistivity contours are more widely spaced, suggesting a more uniform lithology is present in this part of the section. A variation in lithology probably occurs between stations 15 and 16, where resistivity contours are closely spaced.

These preliminary results show the AMT method to be effective in reconnaissance mapping of subsurface resistivity. The method provides another means of defining complex geologic structures.

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## USE OF RADIOLARIAN BIOSTRATIGRAPHY IN STRATIGRAPHIC PROBLEMS IN THE OTUK FORMATION


**Katherine M. Reed**
**and Charles D. Blome**


Radiolarian-bearing chert is widely distributed in Triassic strata on the North Slope. Our studies of more than ten sections (Blome and others, 1986) along the front of the Brooks Range show that radiolarians can provide closely controlled age information for much of the siliceous part of the Otuk Formation (Early Triassic to Middle Jurassic; Mull and others, 1982). The age of the chert-rich part of the formation varies from section to section and reflects local facies variation and may indicate diagenetic changes.

Late Triassic radiolarians have been well studied, and zonations have been established (Nakaseko and Nishimura, 1979; DeWever and others, 1979; Pessagno and Blome, 1980; Kishida and Sugano, 1982; Blome, 1984; Blome and others, unpub. data, 1985). Although preservation of radiolarian faunas in North Slope strata is poor relative to that of coeval faunas in Canada, the western United States, and Japan, several distinctive and robust taxa are present. Precise stratigraphic location of samples collected by individual workers at some of the sections is complicated by the gradational contacts of the members of the Otuk (fig. 8) and the use of different reference points in measuring the sections. The four sections discussed below illustrate the usefulness of Triassic radiolarian biostratigraphy in constraining the ages of the members of the Otuk Formation.

At Monotis Creek (fig. 9), radiolarians make it possible to refine the age of an exposure of the Otuk Formation that was previously dated by sparse megafossils. Species of _Halobia_ and _Monotis_ are widely distributed through, and locally abundant in, the informally designated chert and limestone members of the Otuk Formation, but these fossil pectens are relatively scarce at the Monotis Creek section. Early Triassic (Anisian) _Posidonia_ has been collected from the informally designated shale member, and the Late Triassic (early late Norian) _Monotis subcircularis_ Gabb was reported in a bed about 10 m above the base of the chert member (I.L. Tailleur, written commun., 1982; Patton and Tailleur, 1964). _Halobia_ sp. occurs near the top of the chert member. This genus ranges from early Karnian to early middle Norian, and, according to N.J. Silberling (USGS, written commun., 1985), the age range of North Slope halobiids may prove to be even greater. The overlying limestone member contains abundant _M. subcircularis_. Seven radiolarian samples from the chert member indicate an orderly progression from late Ladinian through late Karnian and into the early middle Norian at its top. The radiolarian data also suggest that the "_Monotis_" at the base of the chert member was misidentified and probably is _Daonella_ (C.D. Blome, written commun., 1985).

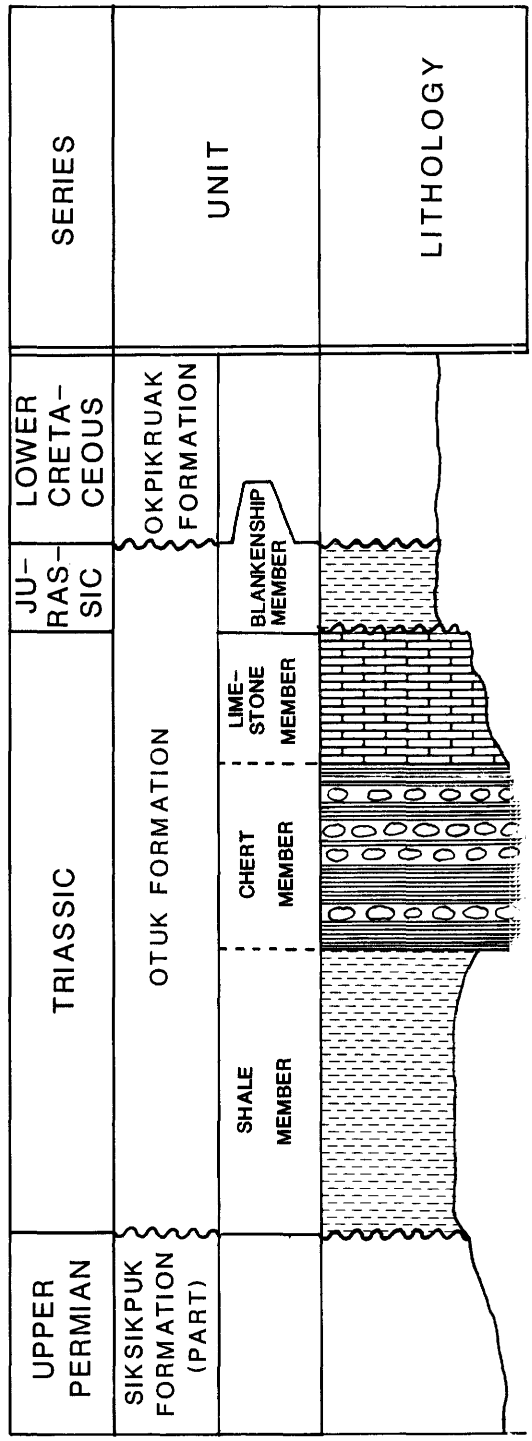


**Figure 8.**--Diagram of the Otuk Formation.

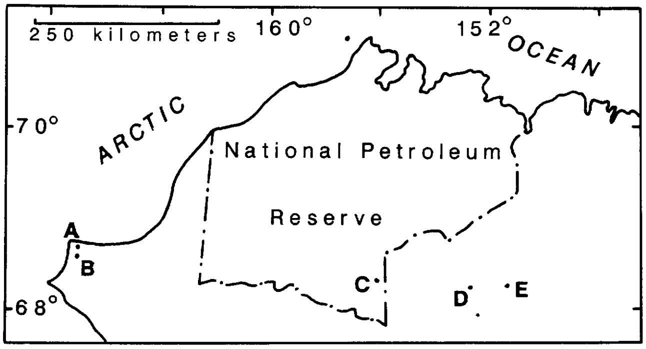


**Figure 9.**—Locations of sections of the Otuk Formation discussed in text. A, Eeneegiksook Creek; B, Ayugatak Creek; C, Otuk Creek; D, Monotis Creek; E, Tiglukpuk Creek.

Apparently discordant megafossil and radiolarian data suggested a repeat of Ladinian or Karnian strata at Eeneegiksook Creek on the Lisburne Peninsula (fig. 9), where about 30 m of the Otuk Formation is exposed. In this section, the Middle Triassic (middle Ladinian) Daonella frami Kittl occurs near the top of the shale member. Above it, in apparently normal succession, are two horizons containing the Karnian Halobia zitteli Lindstrom/ H. ornatissima Smith (Silberling, written commun., 1963, 1985), and in the limestone member, near the top of the section, is a coquina-like bed of early late Norian Monotis subcircularis and M. ochotica (Keyserling). A poorly preserved radiolarian assemblage occurs in a chert-rich bed above the Halobia-bearing beds; it contains (?)Staurocontium and Eptingium manfredi Dumitrica. These radiolarian genera are known only from late Ladinian and early Karnian assemblages. If the halobiid is H. ornatissima, a late Karnian age for those horizons is indicated, an age that conflicts with that of the radiolarian fauna and might suggest a repeat. However, and more simply, if zitteli is present, a late early Karnian age is suggested. The radiolarians support that early Karnian age and identification and suggest that the age range of zitteli might extend down into the earliest part of the Karnian.

As noted, radiolarians constrain the ages of the members of the Otuk Formation fairly well. Some variations in these ages, for example, at Otuk and Tiglukpuk Creeks (fig. 9), indicate that the highly siliceous nature of some of the beds may not be time dependent. At the Otuk Creek section, Bodnar (1984) reported several late Karnian radiolarian genera (including Pachus) in beds at the base of his informally designated chert member, which he shows as containing some shale beds. Halobia zitteli (late early Karnian) or H. superba Smith (?late Karnian) were collected by Mull and others (1982) from the highest beds in the shale member. In the same section, Swain (1981) assigned a late Karnian age to radiolarians in the uppermost beds of what she informally called the shale member, which might be equivalent to the lowest beds in Bodnar's chert member. Swain's sample 1.54 was taken from a position that appears to be at the base of the chert member; we infer from her identifications that the fauna in that sample might be as old as late Karnian but is not younger than late middle Norian. The contact between these members is, thus, of approximately late Karnian age at Otuk Creek. In contrast, at Tiglukpuk Creek, the contact between the shale and chert members is probably in the Anisian part of the section. There, Bodnar (1984) reported fossils that indicate that the shale member is Early Triassic (?Griesbachian to ?Spathian) in age and that the lower 20 m of the chert member is Anisian, based on sparse and possibly reworked conondonts, and middle Ladinian, based on a bed containing Daonella frami. Radiolarian faunas confirm the Ladinian age of this part of the section. However, radiolarians from other Otuk sections, for example at the widely separated Monotis Creek and Ayugatak Creek sections (fig. 9), show that chert predominates lithologically at those localities in late Ladinian or early Karnian strata. The variation in the age of the contact between these two lower members may reflect local depositional conditions, as Bodnar notes (1984, p. 133). On the other hand, variations in age of the shale-chert contact at many exposures could indicate diagenetic alterations that may have influenced the ratio of shale and chert.

Application of Triassic radiolarian age data to the Otuk Formation has revealed several complexities in lithostratigraphy and biostratigraphy. Our studies suggest that the members of the formation may not be entirely coeval along the range front, and that radiolarians permit more precise age control for the chert-rich beds than is possible using megafossils.

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Reviewers:
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## PB-ZN-AG MINERALIZATION IN PALEOZOIC DOLOSTONES, POWDERMILK PROSPECT, BAIRD MOUNTAINS B-4 QUADRANGLE


**Jeanine M. Schmidt and Peter F. Folger**


A previously unknown occurrence of Pb-Zn-Ag mineralization in dolostone host rocks was located on July 4, 1985, in the Baird Mountains as part of the Alaska Mineral Resource Assessment Program. Two outcrop exposures of sulfide minerals and adjacent areas of mineralized float are probably connected; the area of surface mineralization is approximately 0.5 $km^2$(fig. 10). This occurrence is herein informally named the Powdermilk prospect. The Powdermilk area lies along the east bank of an unnamed major tributary of the Omar River in the Baird Mountains B-4 quadrangle (fig. 11). The Bureau of Land Management presently administers this area within the Squirrel River drainage basin.

Rock sample 84ADn131, collected in 1984 by J.A. Dumoulin, USGS, from a knoll (K, fig. 11) along the east bank of the Omar tributary, was later identified as a vein of coarse-bladed barite with crystals as much as 3 cm long, intergrown with crustiform, dark-brown sphalerite. Disseminated to podiform sphalerite and galena in dolostones were discovered several hundred meters north of this knoll during the 1985 field season. A chain-and-compass baseline was established along the east bank of the river (R, fig. 11),

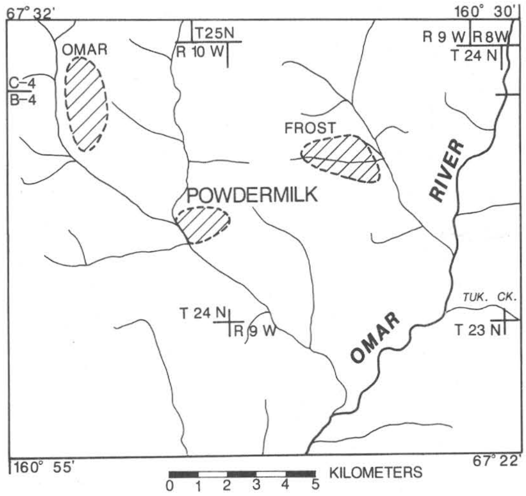


**Figure 10.**--Location of the Powdermilk prospect relative to the Omar River and nearby Omar and Frost mineralized areas (Degenhart and others, 1978), Baird Mountains B-4 quadrangle.

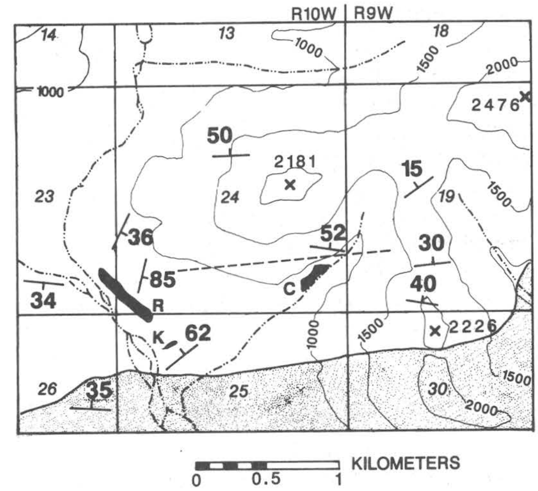


**Figure 11.**--Sketch map of the Powdermilk area with creek (C), riverbank (R) and knoll (K) exposures shown in black. Stippled area is platy brown argillaceous limestone. White area is fine-grained gray dolostone of Ordovician(?) age (J.A. Dumoulin, USGS, oral commun., 1985). Dashed line is a faint photolinear observed on false-color infrared photographs. Section numbers are indicated. Contours and spot elevations are in feet above sea level.

and 39 soil samples, as well as rock, stream-sediment, and panned-concentrate samples were taken along this line. Subsequent geologic mapping in the area to the east of the original exposure revealed additional mineralization along a southwest-flowing, unnamed creek draining into the tributary river (C, fig. 11). The strike of bedding in the host dolostones is northeast, and bedding attitudes change from north- to south-dipping at the river exposure, where beds are nearly vertical. This suggests a synformal structure. Mineralization at the creek is exposed along a projected northeast strike from the river, and a faint east-west photolinear (fig. 11) connects the two exposures.

Host rocks in the area are fine-grained, buff-weathering, gray dolostones. Minor gray limestone occurs east of the mineralized area, and the dolostones are in contact to the south with a distinctive, platy-weathering argillaceous limestone unit. Conodonts from dolostones within a few kilometers of the Powdermilk area suggest an Ordovician age for this unit (J.A. Dumoulin, USGS, oral commun., 1985).

Mineralization at the Powdermilk prospect consists of fine to coarse disseminations and clots (less than 2.5 cm diameter) of sphalerite, galena, and rare pyrite in light-gray dolostone. Hand samples contain as much as 20 percent sphalerite, 8 percent galena, 5 percent pyrite, and 30 percent total sulfide. The sphalerite is coarse grained, locally zoned, and medium to dark brown and red brown. Surficial weathering is variably developed and consists of rare Fe-oxide staining and local clay alteration.

Rock samples from the more oxidized eastern creek exposure (C, fig. 11) contain as much as 280 ppm Pb, 230 ppm Zn, 1,500 ppm Cd, and 21 ppm Ag (quantitative atomic absorption analyses). Semiquantitative emission spectrographic analyses indicate that these rocks contain moderate amounts of Fe and have low concentrations of Cu (less than 60 ppm), As, Ba, Bi, and Sb. Highly erratic Cd and Ag values and high contents of these elements in samples that contain only minor sphalerite and galena suggest that a discrete Ag- and Cd-bearing mineral may be present.

Soil samples from the western riverbank exposure (R, fig. 11) at Powdermilk contain as much as 1,600 ppm Pb, 420 ppm Zn, 76 ppm Cd, and 7.2 ppm Ag (atomic absorption analyses). Semiquantitative emission spectrographic analyses indicate low amounts of Fe, As, Cu, Bi, and Sb, and highly variable Ba content to several thousand parts per million. The irregularly spaced soil samples extend over an area approximately 90 x 700 m but do not cover the original vein discovery on the knoll. No analytical results of rocks from the riverbank exposure are available.

Mineralization at the Powdermilk prospect is dissimilar to that at the nearby Omar and Frost prospects. At Omar, Cu-sulfide minerals with high contents of Co and minor Zn and Ag occur as stockwork veins and disseminations in lower Devonian (Emsian) dolostones (Degenhart and others, 1978). At the Frost prospect, a large vein or shear zone composed of barite contains minor galena and sphalerite and crosscuts dolostone of uncertain age (Degenhart and others, 1978). At the Powdermilk prospect, mineralogy and host rocks are similar to many Mississippi Valley-type deposits, although no evidence of solution collapse or large accumulations of sulfides has yet been found. The sulfide minerals are epigenetic with respect to the host dolostone, and the mineralization may be related to structural control of the synform.

Studies planned for 1986 on the Powdermilk prospect include petrography to determine sulfide-gangue textural relations, staining of carbonates to determine mineralogy, detailed mapping, reconnaissance of the area between the riverbank and creek exposures, and additional geochemical sampling.

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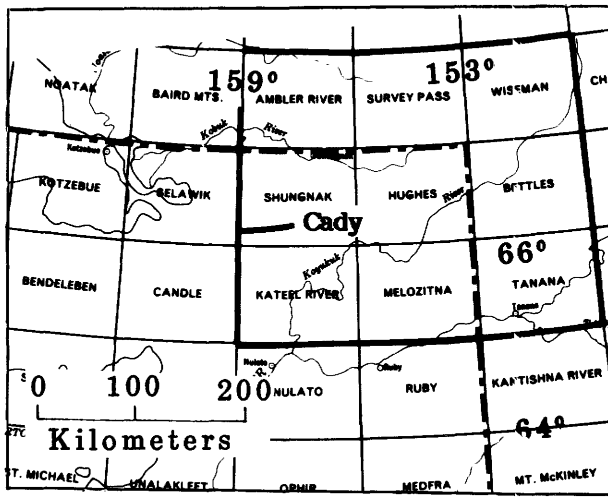



**Figure 12.**--Map showing area in west-central Alaska discussed in this circular.

# WEST-CENTRAL

(Figure 12 shows study area described.)

## GEOPHYSICS OF THE YUKON-KOYUKUK PROVINCE

**John W. Cady**

Digital gravity and aeromagnetic data are being compiled and quantitative analyses and modeling of these data are under way to determine geologic structure in the Yukon-Koyukuk province and Tozitna terrane (fig. 13). Seismic refraction and deep reflection studies will be made in the next few years as part of the Trans-Alaska Crustal Transect (TACT) program. Gravity and magnetic data and their interpretation will help plan optimal locations for TACT seismic profiles. This paper presents a preliminary interpretation of the subsurface geology and crustal structure of the Yukon-Koyukuk province. Data compilation is still incomplete; therefore, an index map is used in place of gravity and aeromagnetic maps in this report. Previously published geophysical maps of the area include the Alaska Bouguer anomaly map of Barnes (1977) and the Alaska aeromagnetic map of Godson (1984).

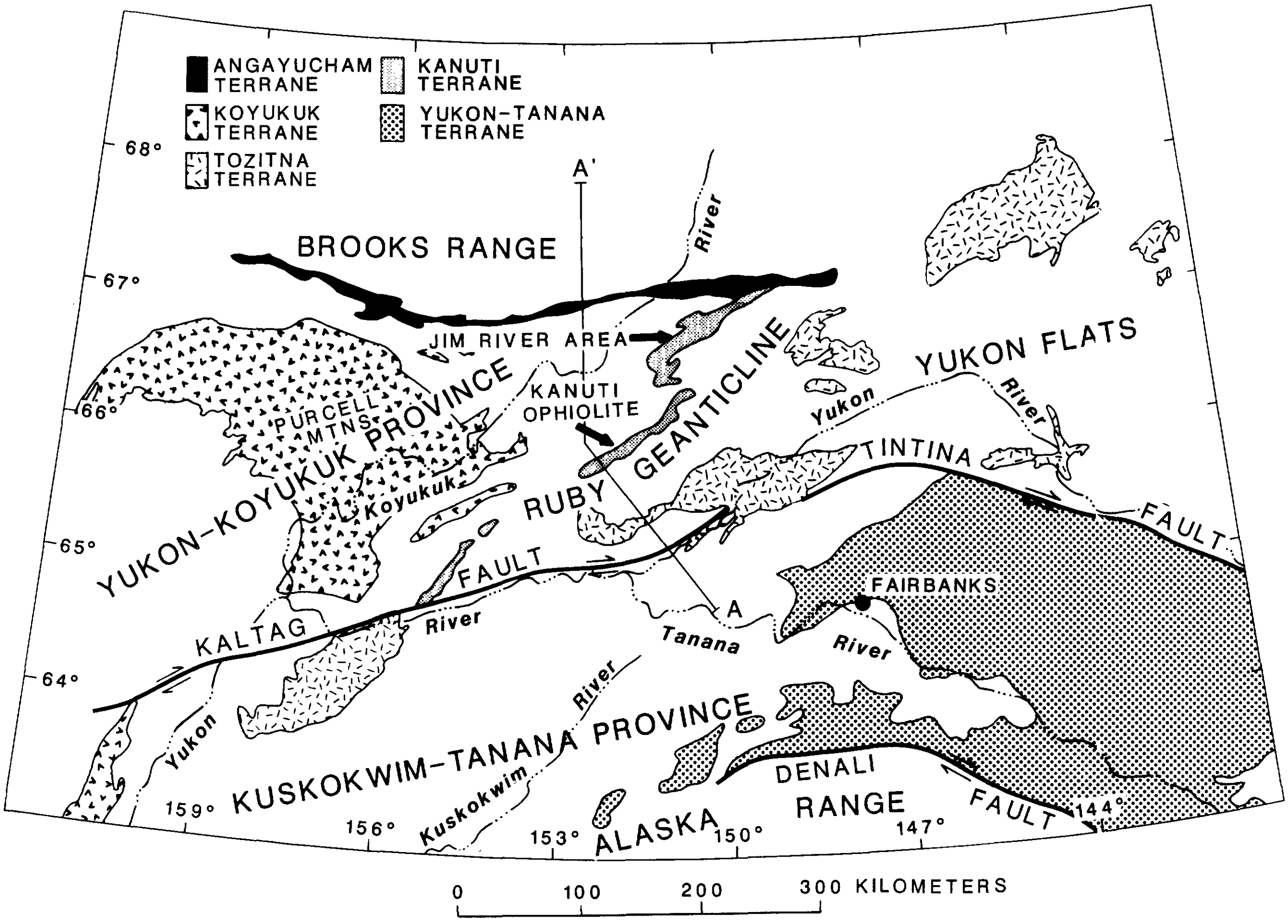



**Figure 13.**--Geologic index map of interior Alaska (after Beikman, 1980; Jones and others, 1984). Line A-A' is the gravity model section line shown in figure 14.

Rocks of the oceanic Angayucham, Koyukuk, and Tozitna terranes (Jones and others, 1984; Howell and others, 1985) and an area herein referred to as the Kuskokwim-Tanana province produce Bouguer gravity anomaly highs. Bouguer anomaly lows occur over old continental crust of the Brooks Range and most of the Yukon-Tanana terrane and tectonically thickened rocks of the Alaska Range. Gravity gradients are steep and coincide with terrane boundaries along the south flank of the Brooks Range and along the northwest-trending part of the Tintina fault. Such boundaries are interpreted to be relatively steep and to separate regions of differing crustal structure. In contrast, along the northwest boundary of the Yukon-Tanana terrane west of Fairbanks and along the Denali fault, gravity gradients are not present at terrane boundaries, suggesting that terranes have subhorizontal boundaries or do not contrast in density structure. A relatively minor gravity low over the Ruby geanticline indicates that it is only a continental sliver.

Gravity and magnetic highs form a V open to the southwest, coincident with the northern and southeastern margins of the Yukon-Koyukuk province (YKP). These anomalies are asymmetrical, with steeper gradients to the outside of the province, and can be modeled by assuming dense, magnetic sources that dip 30-70 degrees inward beneath the YKP. Oceanic rocks of both the northern and southeastern margins of the V are assigned to the Angayucham terrane by Jones and others (1984). Geologically and geophysically, however, the arms of the V are different, suggesting that they should be assigned to different lithotectonic terranes. In this discussion, the term Angayucham terrane refers to the northern arm of the V, whereas the term Kanuti terrane refers to the southeast margin. The position of the boundary between the Kanuti and Angayucham terranes is not yet determined, but it is tentatively shown north of the Jim River area.

The predominant rock type of the Angayucham terrane is pillow basalt formed on an oceanic plateau and (or) oceanic island setting (Pallister, 1985). The amplitude of the magnetic anomaly over the Angayucham terrane is generally less than 500 nT. In contrast, the Kanuti terrane contains a complete dismembered ophiolite sequence. Both the cumulus mafic-ultramafic rocks and ultramafic tectonite of the informally designated Kanuti ophiolite cause a magnetic anomaly of more than 2,000 nT. Magnetic modeling shows that rocks equivalent to the Kanuti ophiolite cannot be present at depth beneath the Angayucham terrane.

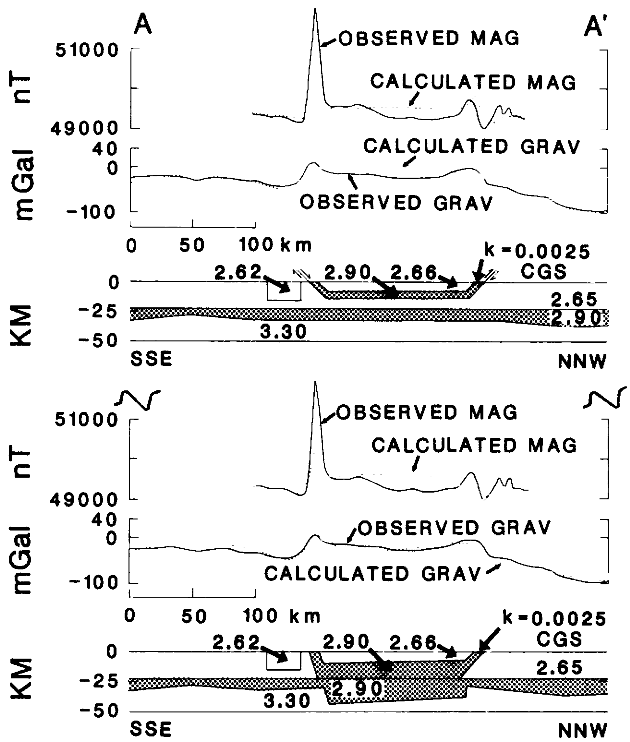
In contrast to the nearly continuous gravity high over the Angayucham terrane, the gravity high associated with the Kanuti terrane is interrupted by lows over granitoids that have intruded both the Kanuti ophiolite and the mafic-ultramafic rocks exposed en echelon to the north along the Jim River. The magnetic high, however, is continuous over the Kanuti ophiolite and the plutons that intrude it. These plutons of the Ruby geanticline, which are normally non-magnetic S-type granites (Miller, 1985), are magnetic where they have intruded mafic and ultramafic rocks. Perhaps the plutons are magnetic because they were contaminated by the mafic crust that they assimilated or from which they were derived.

Within the YKP are U-shaped gravity and magnetic highs open to the southwest produced by an Early Cretaceous intraoceanic island arc (Box and Patton, 1985) labeled Koyukuk terrane (fig. 13). This island arc divides middle Cretaceous clastic rocks of the YKP into two subbasins marked by gravity and magnetic lows: an outer, remnant-forearc basin and an inner, remnant-backarc basin (Box and Patton, 1985). Gravity lows also occur over I-type (Miller, 1985) granitoids of the Purcell Mountains that have summit elevations of about 1,000 m, or about 800 m above valley floors. Thus, the YKP contains varied tectonic elements, including thrust sheets of ophiolite, abundant pillow basalts, a magmatic arc containing both volcanic and granitic rocks, and two graywacke basins. The elements have distinctive topographic relief and geophysical signatures. Although the entire assembly has often been termed the Yukon-Koyukuk "basin", the term Yukon-Koyukuk province implies more tectonic variability and is preferred.

Modeling for deep crustal structure is ambiguous in the absence of seismic refraction or deep seismic reflection data. An unreversed seismic refraction profile (Hanson and others, 1968) indicates continental crust thinning from 48 km in the Alaska Range to 32 km near Fairbanks. Marine seismic studies off Kuskokwim Bay southwest of figure 13 show oceanic crustal velocities of 6.7-6.9 km/s at a depth of 10 km, but mantle velocities are not observed (Cooper and Marlow, 1984). Oceanic rocks, including oceanic crust and upper mantle of the Kanuti ophiolite, are exposed at the surface in the YKP. Since the area is above sea level, the crust of the YKP must be thicker than oceanic crust and have a lower mean density. How has the crust been thickened?

Figure 14 shows two regional crustal models across the eastern YKP (fig. 13) that satisfy the gravity data and illustrate contrasting hypotheses about how the crust was thickened. Model A shows allochthonous oceanic crust thrust over attenuated continental crust. Although Model A satisfies the gravity data, isotopic data (Arth, 1985) argue against the presence of old continental crust beneath the YKP. Preferred model B shows thick, rooted crust of oceanic affinity. Because the crust has relatively high density, it must have a root protruding into the mantle to maintain isostatic equilibrium. The magnetic data were used only to constrain the near-surface configuration of

MODEL A. ALLOCHTHONOUS OCEANIC CRUST OVERLYING ATTENUATED CONTINENTAL CRUST




MODEL B. THICK, ROOTED CRUST OF OCEANIC AFFINITY

**Figure 14.**--Gravity models across the eastern Yukon-Koyukuk province along line A-A' (fig. 13). Densities in g/cm$^3$. Magnetic susceptibility (k) is in the cgs system. Patterned area with density of 2.90 g/cm$^3$ includes informally designated Kanuti ophiolite (in south), Angayucham terrane (in north), and mafic lower crust. Area with density of 2.66 g/cm$^3$ is graywacke and volcanic rocks of Yukon-Koyukuk province. Area with density of 2.62 g/cm$^3$ is granite. Upper crust and mantle have densities of 2.65 and 3.30 g/cm$^3$, respectively.

dense, magnetic rocks. Note that if a similar source is used to explain anomalies over both the Kanuti ophiolite and Angayucham terrane, the calculated anomaly is too small over the Kanuti ophiolite.

Geophysical data are compatible with the interpretation (Box, 1985) that the YKP contains oceanic crust that remained unsubducted when an island arc collided with an irregular continental margin in Early Cretaceous time and was oroclinally bent. Further tightening of the orocline accompanied by probable imbricate thrusting occurred when terranes were accreted from the southeast in Late Cretaceous time. Thus, the YKP is underlain by oceanic crust that has been thickened by island arc magmatism and sedimentation and by probable imbricate faulting; only at the province margins has oceanic crust been thrust over continental crust.

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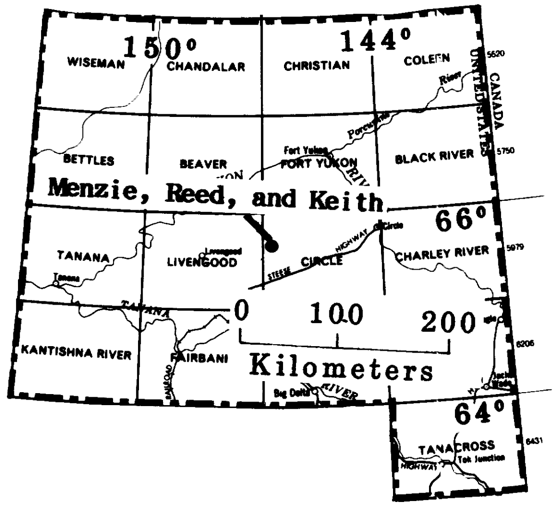


**Figure 15.**--Map showing area in east-central Alaska discussed in this circular.

# EAST-CENTRAL

(Figure 15 shows study area described.)

## LIME PEAK--AN EVOLVED GRANITE WITH TIN-ENRICHED ALTERATION

**W. David Menzie, Bruce L. Reed, and Terry E.C. Keith**

Anomalously high levels of tin and associated elements (Ag, B, Bi, Mo, Pb, Zn) are present in pan-concentrate, stream-sediment, and rock samples from near the Lime Peak pluton in the northwest part of the Circle quadrangle (Menzie and others, 1983; Burton and others, 1985). Intrusive relations indicate the pluton is composed of two main phases: (1) an early coarse-grained equigranular biotite granite, and (2) a later chiefly porphyritic biotite granite with a fine-grained groundmass. Small areas of quartz-feldspar porphyry may represent a minor third phase, and sparse intermediate to mafic dike rocks are also present. The pluton intrudes metasedimentary rocks, including quartzite and argillite of late Precambrian and (or) early Paleozoic age. Epizonal depths are suggested by the presence of miarolitic cavities in the granite. A

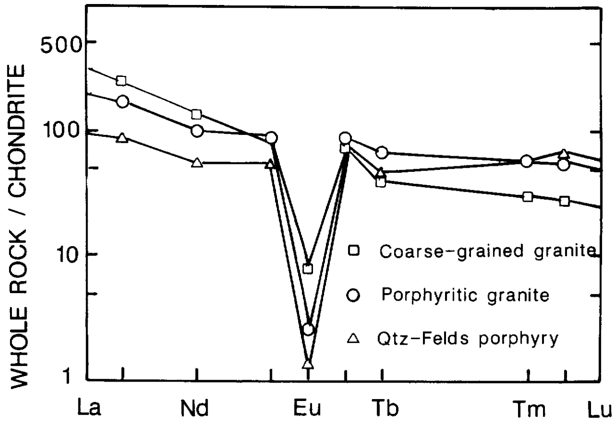


**Figure 16.**—Mean chondrite-normalized rare-earth-element patterns for the three lithologic phases characteristic of the Lime Peak pluton: early coarse-grained equigranular biotite granite (3 samples); later porphyritic biotite granite (3 samples); quartz-feldspar porphyry (2 samples), which may represent a minor third phase.

single potassium-argon age determination on biotite yielded an age of 56.7±0.95 m.y. (Wilson and Shew, 1981).

The range (and average), in weight percent, for some oxides of the main phases of granitic rocks are:

| | | |
|---|---|---|
| $SiO_2$ | 73.6-76.8 | (75.6) |
| $Al_2O_3$ | 11.6-13.4 | (12.5) |
| MgO | <0.1-0.4 | |
| CaO | 0.61-1.02 | (0.75) |
| $Na_2O$ | 2.72-3.1 | (2.86) |
| $K_2O$ | 4.74-5.58 | (5.13) |
| FeO(total) | 1.00-1.85 | (1.54) |

The granites are peraluminous (molar $Al_2O_3/CaO+Na_2O+K_2O$ = 1.06 to 1.1), contain normative corundum, and have $Fe_2O_3/FeO$ ratios less than 0.5, suggesting that they belong to the ilmenite series granitoids (Ishihara, 1981). Differentiation indices range from 90.5 to 94.2 and average 92.4.

On a ternary Q-Ab-Or diagram, the granites form a confined group close to minimum-melt composition. Rare-earth-element patterns are typical of highly evolved granites associated with tin deposits. The first phase shows uniform negative slopes ($Ce/Yb_N$ = 7-14) and negative Eu anomalies (Eu/Eu* =0.045-0.212); later porphyritic biotite granite phases have relative LREE depletion, relatively flat REE patterns ($Ce/Yb_N$ = 1.4-5.6) and well-developed negative Eu anomalies (Eu/Eu* =0.01-0.04). The quartz-feldspar porphyry is strongly depleted in LREE and has extreme negative Eu anomalies (fig. 16).

The pluton locally shows the effects of deuteric (late magmatic) and hydrothermal alteration. Deuteric alteration consists of veinlets, breccias, and pods of black tourmaline. Hydrothermally altered rocks show evidence of two stages of alteration. An early stage resulted in extensive chloritization of biotite, partial sericitization of feldspars, and deposition of hydrothermal quartz. Much of the chlorite has the structure of septachlorite, which generally forms at temperatures less than 200°C. Lack of epidote associated with this stage of alteration suggests that temperatures must be less than approximately 240°C. Iron and manganese were mobilized and deposited as oxides and hydroxides concentrated in and adjacent to fractures and shear zones in the granitic rocks, especially in zones in the southern and southeastern parts of the pluton. These zones trend west-northwest and are several hundred feet in length. The chloritized, Fe-, Mn-stained rocks of the first alteration stage contain a median of 100 ppm and a maximum of 1,500 ppm tin.

A later alteration is superimposed locally on the earlier stage and is characterized by the alteration of biotite (chlorite?) to a fine-grained mixture of chlorite and white mica and alteration of feldspars to white mica. The latest hydrothermal minerals are muscovite, green tourmaline, and fluorite (in paragenetic sequence), which have been deposited in thin veins and vugs.

The field relations and alteration mineral assemblages suggest that the exposed alteration zones represent the outer margins of a late-stage, tin-rich mineralized system.

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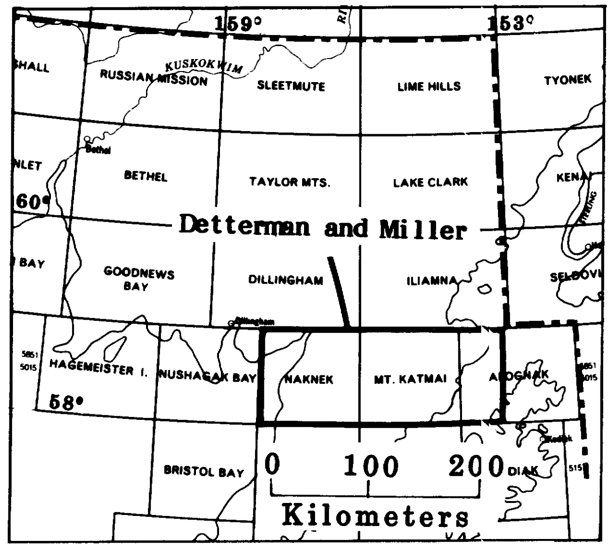


**Figure 17.**--Map showing area in southwestern Alaska discussed in this circular.

# SOUTHWESTERN

(Figure 17 shows study area described.)

## A WIDESPREAD CATASTROPHIC EVENT IN THE NAKNEK FORMATION, ALASKA PENINSULA

**Robert L. Detterman and John W. Miller**

Evidence for the cause of a widespread faunal catastrophy in the Naknek Formation (Upper Jurassic) is preserved in strata on the north shore of Hallo Bay, Katmai National Park (fig. 18). The evidence consists of a coquina layer about 0.5 m thick, composed of the pelecypods Buchia concentrica, Pleuromya sp., and Oxytoma sp. and the ammonites Phylloceras sp. and Lytoceras sp., and a volcanic graywacke 2 m thick that overlies the coquina horizon (fig. 19). The section underlying the coquina contains an abundant megafauna characteristic of the Naknek Formation elsewhere on the Alaska Peninsula. Dark siltstone with a few thin sandstone and limestone interbeds overlies the volcanic graywacke. The basal 18 m of this siltstone are devoid of megafauna remains, and 100 m of section overlying that contains a sparse megafauna. Above 118 m, the megafauna is of normal abundance and diversity and is characteristic of the Naknek Formation on the Alaska Peninsula.

Eight sections in the Naknek Formation, including the Hallo Bay section, have been measured during stratigraphic investigations on the Alaskan Peninsula as part of the Alaska Mineral Resource Assessment Program. Many other sections have been studied in lesser detail. In all sections where the Buchia concentrica biozone is exposed, an unfossiliferous sequence of beds has been noted within that biozone. In most places, the barren interval is much thicker than at Hallo Bay. The Hallo Bay section represents a condensed sequence deposited in a starved basin. The Buchia concentrica, B. rugosa, and B. mosquensis biozones (Miller and Detterman, 1985) are found through 250 m of section at Hallo Bay. Elsewhere, however, the three biozones are found throughout 700-1,000 m of strata.

The Hallo Bay section provides a possible clue for the cause of the barren interval. The strata above and below the coquina and graywacke are mainly thin-bedded siltstone with a few thin intercalated

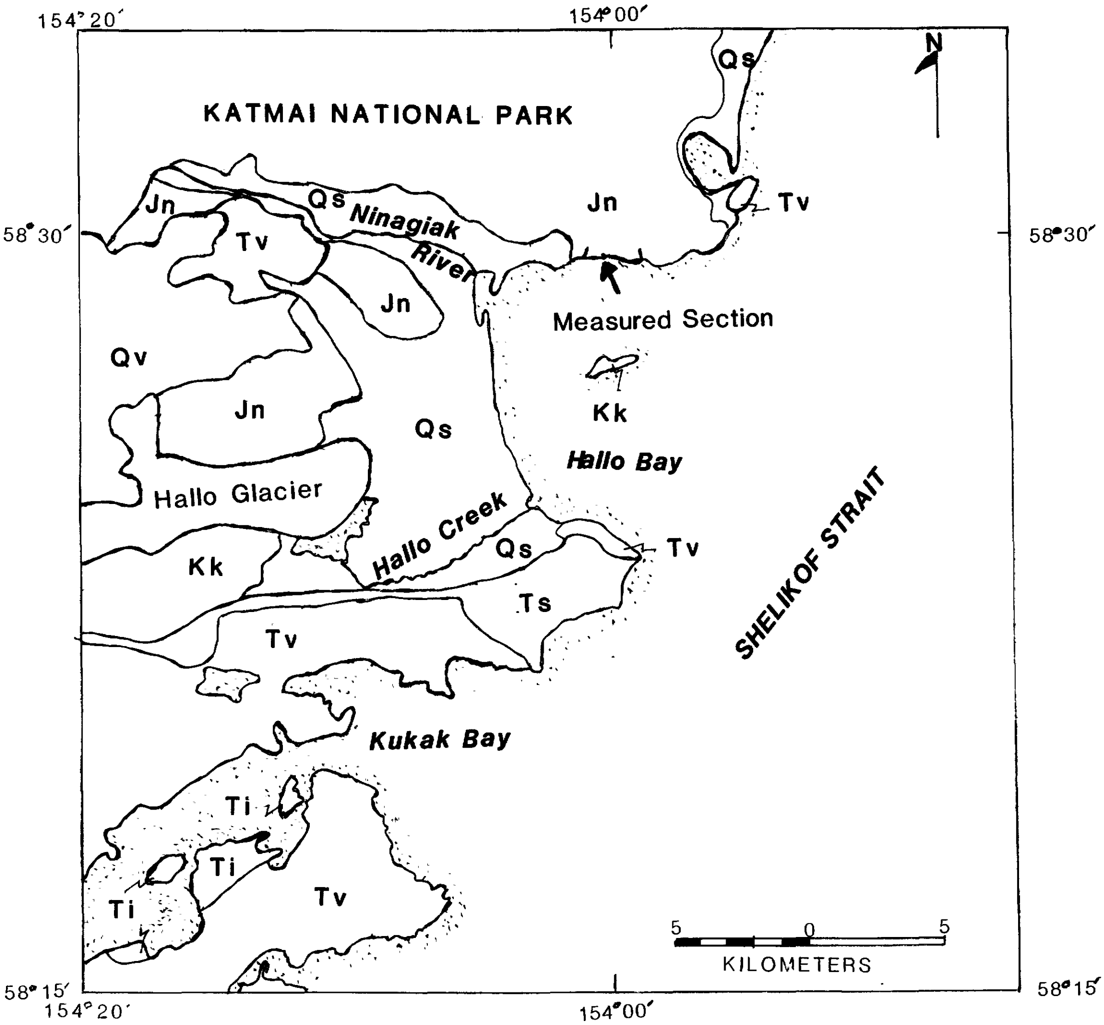


**Figure 18.**--Location of measured section and distribution of geologic units. Qs, surficial deposits (Quaternary); Qv, volcanic rocks (Quaternary); Ts, sedimentary rocks (Tertiary); Tv, volcanic rocks (Tertiary); Ti, intrusive rocks (Tertiary); Kk, Kaguyak Formation (Upper Cretaceous); Jn, Naknek Formation (Upper Jurassic).

sandstone and limestone beds. These beds are typical of a low-volume depositional environment. The thin sandstones are well-sorted arkosic arenites. In contrast, the bed overlying the coquina horizon is a massive, poorly sorted, volcanic graywacke. The lack of sorting in this rock would indicate rapid deposition. In thin section this rock is composed of about 50 percent plagioclase, 5-10 percent orthoclase, 5-8 percent quartz, about 12 percent mafic minerals (mainly hypersthene), 8-10 percent rock fragments mainly of volcanic origin, and about 15 percent matrix. Mineral grains are angular to subangular and range in size from 0.05 to 0.75 mm.

The Hallo Bay section includes the Buchia concentrica, B. rugosa, and lower part of the B. mosquensis biozones. Based on the zonation of Buchia (Imlay, 1959; Miller and Detterman, 1985), the section would range from late Oxfordian to early Tithonian (Late Jurassic), an interval of approximately 8x10$^6$ yrs. Assuming a constant rate of deposition and exclud-

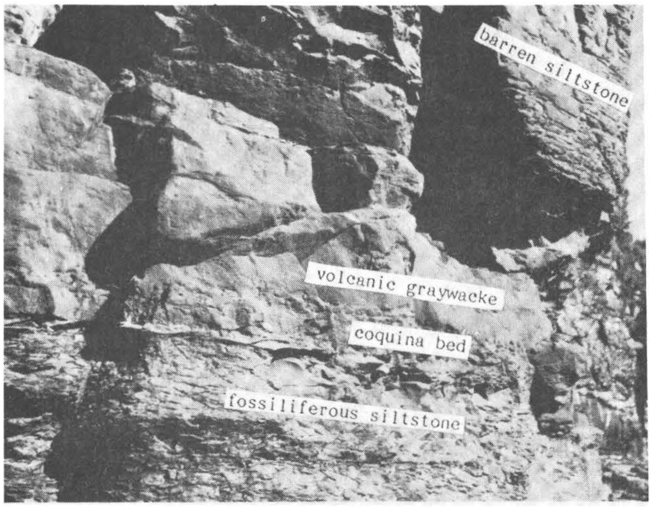


**Figure 19.**--Photograph of north shore of Hallo Bay, showing coquina horizon and overlying volcanic graywacke.

ing the volcanic graywacke, the Hallo Bay section was deposited at a rate of about 3 cm/1,000 yr. Thus, the 18-m-thick barren interval represents about $6x10^5$ yrs before the fauna was re-established. This interval of time is not constrained by radiometric age dates and should be considered only as an approximation.

The most likely cause for the barren interval was the rapid deposition of volcanic graywacke with attendant loss of habitat. Similar volcanic graywacke units underlying barren intervals in the B. concentrica zone are present for at least 600 km along the peninsula, indicating that this was a widespread event. The loss of both habitat and a viable reproducing population would explain why it apparently took $6x10^5$ yrs to re-establish the fauna at Hallo Bay.

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# SOUTHERN

(Figure 20 shows study areas described.)

## CHANNEL MIGRATION IN TURNAGAIN ARM

Susan Bartsch-Winkler

The Turnagain Arm of Cook Inlet (fig. 21) is subject to extremes of climate and large tide range and has experienced a tectonically induced lowering of the sediment surface. These factors result in widespread redistribution of estuarine sediment and major migration of tidal channels. In this setting, rapid erosion is a natural process that represents a potential hazard to the construction of permanent structures in or adjacent to the intertidal zone. Prior experience indicates that erosion by tidal channels could affect the Seward Highway and The Alaska Railroad, both of which skirt the northern shore of Turnagain Arm, as well as buried utility lines that cross the Arm (Teel, 1981). Thus, the migration of tidal channels in Turnagain Arm must be considered in any transportation or utility construction plan for this intertidal zone.

Neither the frequency nor the extent of channel shifting in Turnagain Arm has been documented previously. This report discusses the extent of intertidal channel shifting in the upper part of Turnagain Arm as shown in aerial photographs taken from 1950 through 1982 and compares the locations of channels before and after the great Alaskan earthquake of March 27, 1964.

The estuary is an ancient fjord that has been filling with tidal sediment for at least the past 14,000 years (Bartsch-Winkler and others, 1983). Depth to bedrock is unknown anywhere in the central part of the bay (Alaska Department of

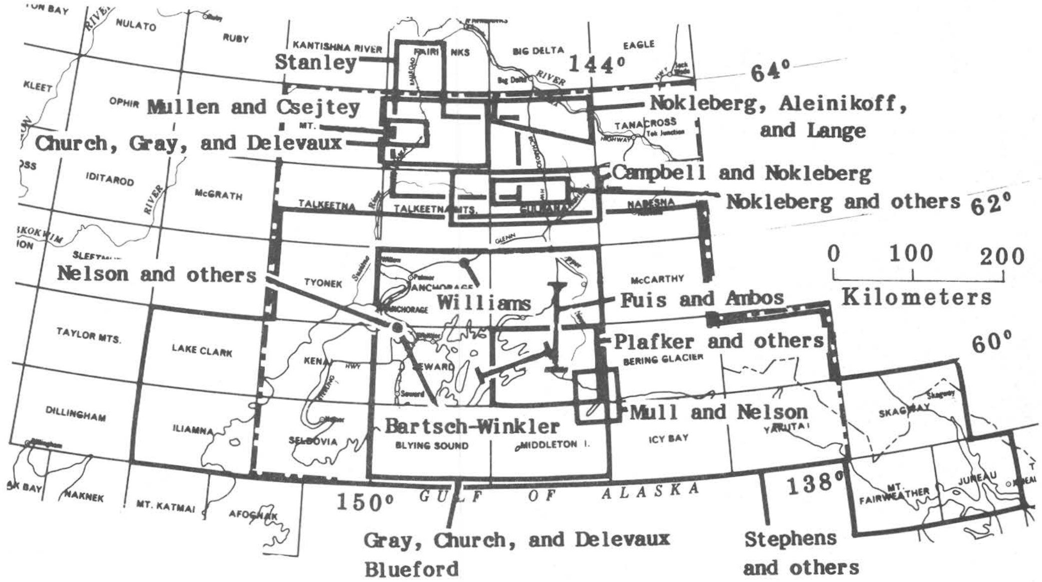


**Figure 20.**--Map showing areas in southern Alaska discussed in this circular.

Highways, unpub. report, 1968; Bartsch-Winkler and others, 1983), but along the shores, the bedrock contact with unconsolidated sediment is steeply dipping and irregular (Alaska Department of Highways, unpub. report, 1968). At low tide the unconsolidated estuarine sediment is exposed, except where tidal channels of unknown depth and morphology scour into it (Bartsch-Winkler, 1985; Bartsch-Winkler and Ovenshine, 1984). Turnagain Arm has a high tide range of about 11.5 m and current velocities that reach 4 m/s (Bartsch-Winkler and Ovenshine, 1984).

Available topographic maps, vertical aerial photographs, and hand-held aerial photographs (for 1982 only) were used to locate channels on a base map of the part of Turnagain Arm extending from Bird Point to Portage (figs. 21 and 22). In all instances except 1982, mapping was accomplished using a PG-2 stereo-plotter. Most of the maps show channel locations at low-tide stages, but the map for 1951 shows the onset of flooding in the lower, distal parts of the estuary, and the map for 1957 shows the channel conditions at mid-ebb-tide stage. For some dates, coverage was incomplete due to cloud cover or limited aerial views in photography intended to be of land surfaces only. Photographs taken at low tide in 1973, 1974, 1981, and 1982 were to be used in an ongoing study of the

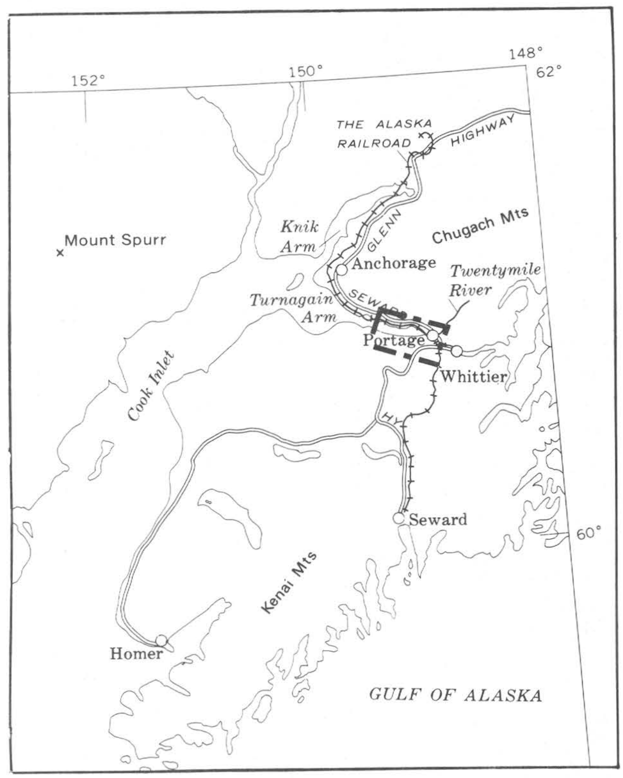


**Figure 21.**--Index map of Turnagain Arm and vicinity. Outline shows study area.

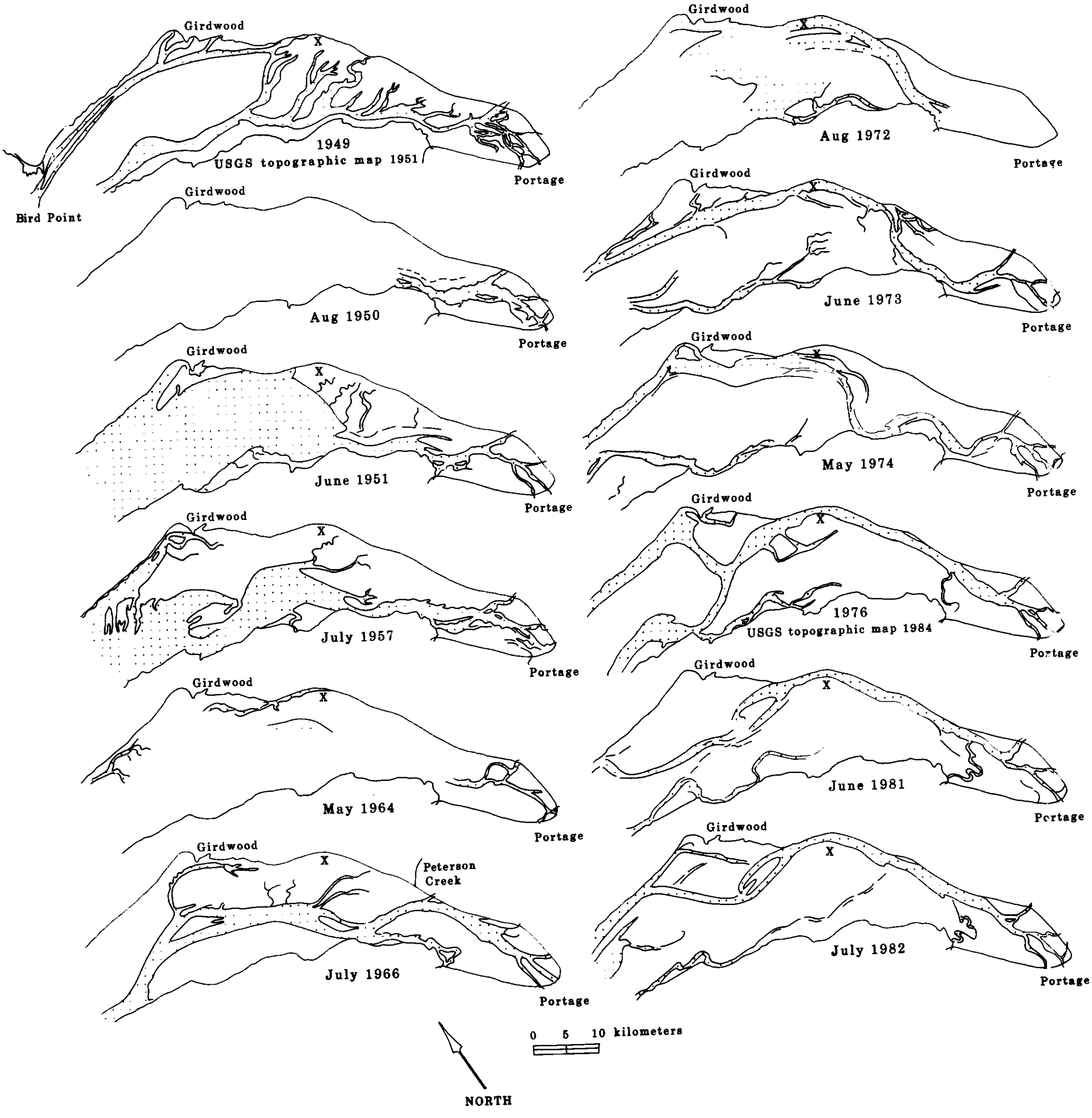


**Figure 22.**--Locations of tidal channels in upper Turnagain Arm (vicinity of Bird Point to Portage) from 1949 to 1982. Dotted areas indicate channels. Base map drawn from 1973 photographs.

estuary by this author and others (Bartsch-Winkler, 1985; Bartsch-Winkler and Ovenshine, 1975, 1984; Bartsch-Winkler and others, 1975, 1978, 1983; Ovenshine and others, 1976). Other photography used in this study was obtained from aerial-photo archives.

Figure 22 shows locations of tidal channels determined from photographs taken from 1949 to 1982. The maps show that the main channel course ranged from gently or strongly sinuous (as in 1974) to relatively straight (as in 1981). Exact channel locations were not stationary from year to year (for example, 1973-74 and 1981-82), but major changes occurred during a 6-7-yr period when channels migrated from one side of the arm to the other. (Compare maps for 1966-1972 and 1973-1981.)

In addition to lowering the land surface an estimated 2.4 m near Portage (Ovenshine and others, 1976), the great Alaskan earthquake of 1964 may have tilted the basin such that the main channel moved to the northeast side of the estuary from the southwest side. This is corroborated by two types of evidence: In 1964-65, it was necessary to construct a protective berm along the highway embankment near Peterson Creek to prevent erosion by the migrating channel (Tony Barter, Alaska Department of Transportation and Public Facilities, oral commun., January 1984). From 1949 to 1966 no major channel lay close to the "X" in figure 22; however, since 1966, a major channel has been continually situated adjacent to this site. The two years from 1964 to 1966 may have been the lag time needed for natural flow adjustments after the earthquake.

Maps drawn from photographs of upper Turnagain Arm show that the channels can change from highly meandering to relatively straight and shift position from year to year, sometimes in response to seismic events. More extensive changes in channel location appear to take more time to accomplish, generally occurring over a period of at least 6-7 years; however, more data are needed to confirm that intertidal channels shift with this periodicity. Because the shifting of channels in the intertidal zone of Turnagain Arm is a major natural process, it must be considered in any plan for construction of transportation or utility structures in or adjacent to the estuary.

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## NEW RADIOLARIAN DATA FROM THE ORCA GROUP, PRINCE WILLIAM SOUND

**Joyce R. Blueford**

Geologic interpretation of the Orca Group, a deep sea fan and basalt complex in the Prince William Sound region, has been hindered by lack of reliable microfossil data. Minor siliceous limestone, limestone, mudstone, and chert in the Orca Group contain assemblages of radiolarians and other microfossils (Plafker and others, 1985). Present extraction techniques do not permit observation of the complete radiolarian assemblage in these well-indurated rocks because of uneven diagenesis. However, the faunas can be studied from thin sections of rock samples, even though structural details of some radiolarians may be obscured.

Examination of 110 samples from various localities in the Prince William Sound area showed that only those from Cape Martin, Fox Island, Jeanie Point, Neck Point, Wingham Island, and Kayak Island contained sufficient radiolarian assemblages for study (Plafker and others, 1985). In addition, the small number of samples, the poor preservation of the radiolarians, and the incomplete nature of the faunas from Jeanie Point and Wingham and Kayak Islands proved insufficient to date the enclosing rocks. The Cape Martin samples contained the most abundant and best preserved assemblages.

Samples from a measured section at Cape Martin, at the south end of Ragged Mountain (locality shown in Plafker and others, 1985, fig. 1), reveal that older radiolarians in the Orca Group are characterized by an assemblage that includes Cromydruppa ovoidea Lipman or C. regularis Lipman, C. concentrica Lipman, Amphibrachium sibiricum Lipman, and Porodiscus charlestonensis Clark and Campbell. Later deposition in that area is characterized by Amphibrachium gracilis Lipman, A. fusoideus Lipman, A. tshelkavensis Lipman, Spongodiscus russicus Lipman, S. maculatus Clark and Campbell, and Spongurus bilobatus Clark and Campbell. These assemblages show the same transitional characteristics as radiolarian assemblages in Late Cretaceous through Paleogene rocks in both the Soviet Union and California. Faunas of later deposits of the Orca Group at Cape Martin have greater affinity with Eocene deposits of central California than with earlier deposits in the study area. Faunas from Fox Island and Neck Point have a close affinity to the older radiolarian assemblage from Cape Martin.

Radiolarian assemblages from Orca Group rocks represent a probable time interval ranging from late Cretaceous to middle Eocene. The faunas are best compared with Paleogene assemblages of the Soviet Union; they differ significantly from the well-documented Cenozoic assemblages of the Pacific Tropics (Riedel and Sanfilippo, 1978) and North Atlantic (Bjorklund, 1976). Radiolarian zonation of the Paleocene and early to middle Eocene of central California has not been deciphered in sufficient detail for meaningful comparison (Clark and Campbell, 1942, 1945; Blueford and White, 1984; Blueford and Brunner, 1984), though components of the California fauna can be used to describe the Orca Group assemblage.

Radiolarian biostratigraphy of the Soviet Union is useful in deciphering the age of the Orca Group. Sedimentary rocks in the Soviet Union of Paleogene and Late Cretaceous age are abundant and well documented (Lipman, 1952, 1960, 1971, 1972; Krasheninnikov, 1960; Borisenko, 1958). Radiolarian zonations have been established for the Russian Platform, the Northern Caucasus, the East Urals, Western Siberian Lowlands, and North Kazakhastan, and these have been used for correlation of sections within large regions of the Soviet Union. Comparison of the Russian biostratigraphy with that of the Orca Group reveals many similarities, but lack of both complete translation of Russian taxonomic literature and poor photographs in existing monographs prevents positive identification of many of the Russian taxa. The many similarities of some of the Russian and Orca Group faunas indicate that they are in part coeval. Affinities with faunas from central California seem apparent in younger Alaskan sections, in part because the stratigraphy of middle Eocene rocks in California is the best documented. However, lack of information about the first and last occurrences of radiolarian species in the Paleogene deposits in California prevents exact dating of these rocks. Therefore, precise placement of the Orca faunas within the Eocene is not possible.

Faunas of the Orca Group are dominated by spongodiscids, porodiscids, and actinommids, implying a cool or cold water environment most likely occurring in areas of upwelling (Blueford and White, 1984). Modern warm-water faunas typically contain relatively more Nassellaria and armed spongodiscids, that are not common forms in the Orca assemblage.

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# MAGNETIC MODEL OF A PROFILE ACROSS NORTHERN COPPER RIVER BASIN, NORTHEASTERN GULKANA QUADRANGLE

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The northern Copper River basin contains several strong west-trending magnetic highs (Andreasen and others, 1964). From south to north, these are informally named the West Fork feature, the Alphabet hills high, the Media high, and the Upper Tangle Lakes high (fig. 23). Andreasen and others (1964) named and discussed the West Fork feature and described the other three highs collectively as the "Maclaren-Gulkana anomalies". The continuity, direction, and length of these magnetic features indicate that the source rocks are also continuous, west-striking, near-vertical, and probably tabular in this area. As part of field studies in the Gulkana quadrangle for the Trans-Alaska Crustal Transect program and the Alaskan Mineral Resource Assessment Program, the area of these magnetic highs was mapped and sampled (Nokleberg and others, 1986) and resurveyed with a helicopter-borne magnetometer in an attempt to identify probable source rocks for the magnetic highs.

From south to north, three major belts of bedrock occur in the northwestern Gulkana quadrangle: the Peninsular terrane, the metamorphic complex of Gulkana River (GRC), and the Wrangellia terrane (Nokleberg and others, 1986, fig. 38, p. 71). Nokleberg and others (1986) describe the lithologies and structures of these three belts. The West Fork fault system (WFFS, Nokleberg and others, 1986) occurs between the Peninsular terrane and the GRC. The West Fork magnetic feature lies along the WFFS and may reflect rocks of either the Peninsular terrane or the GRC. The Alphabet hills high, the Media high, and the Upper Tangle Lakes high occur over rocks of the GRC and south of the Wrangellia terrane.

The West Fork feature is a prominent magnetic high in the northern Copper River basin. It has a typical amplitude of about 1,200 nT, a width of 6-10 km, and a length of more than 100 km. The West Fork feature occurs along the WFFS and underlies the valleys of both the West Fork of the Gulkana River and the Tyone River. Its source rocks are covered along most of its length by glacial-lake sediments. Csejtey and Griscom (1978) report a small outcrop of probable amphibolite, altered diorite, or gabbro intrusive rocks at the far western end of the West Fork feature (fig. 23). The susceptibility of these rocks has not been measured; consequently, we cannot evaluate them as a possible source rock for the West Fork feature.

North of the West Fork feature are outcrops of schistose, medium-grained, granitic plutons of the southern part of the GRC (Nokleberg and others, 1986) that have moderate susceptibilities, typically about 0.0006 (cgs). Typically, they would

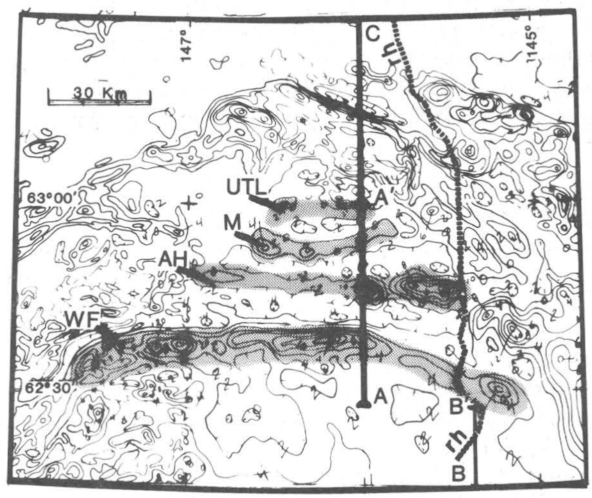


**Figure 23.**--Part of the aeromagnetic map of Alaska (Godson, 1984) showing the four major elongate magnetic highs (shaded) that are discussed in the text. Contour interval = 200 nT; numbers on the contours indicate hundreds of nT with respect to an arbitrary zero level. WF, West Fork feature; AH, Alphabet hills high; M, Media high; UTL, Upper Tangle Lakes high. A-A', location of profile shown in figure 24. A'-C, location of southern end of the profile of Campbell and Nokleberg (1984). B-B', location of northern end of the profile of Campbell and Barnes (1985). rh, Richardson Highway.

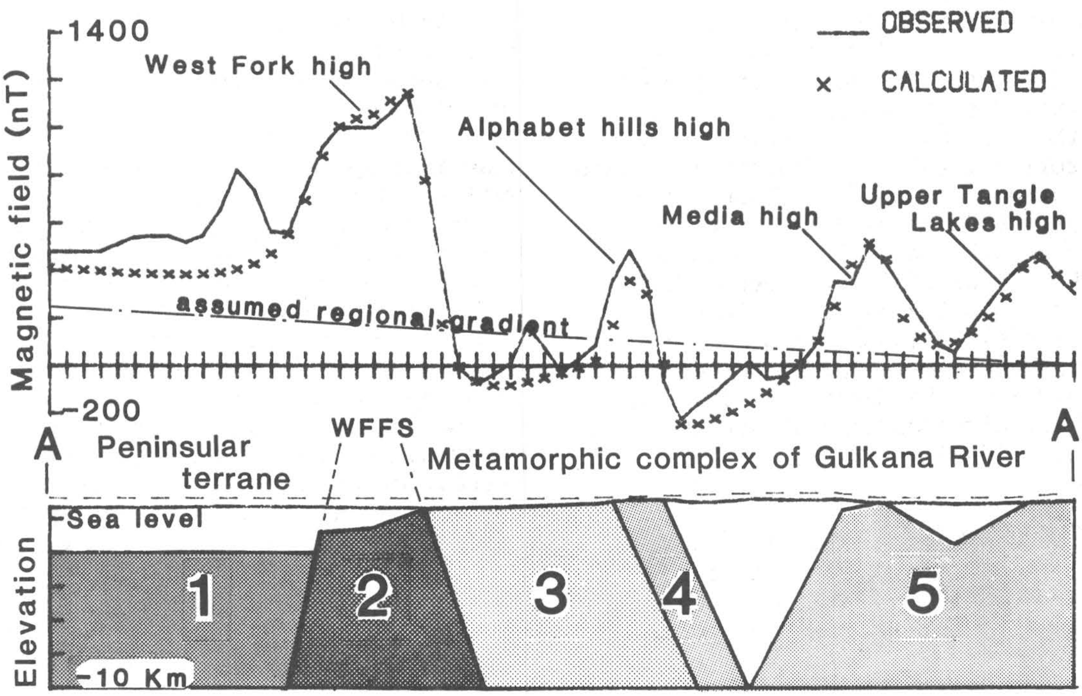


**Figure 24.**--Model of magnetic bodies along profile A-A', northern Copper River basin. The cross-section in the lower part of the figure shows the assumed magnetic structures. The upper part of the figure shows a plot comparing observed magnetic field (solid curve) and calculated magnetic field (line of x's) that results from these structures. Properties of the assumed magnetic units are given in table 2. On the cross-section, darker shading indicates magnetic units having higher susceptibilities, and the unshaded areas represent non-magnetic rocks or rocks having very low susceptibility. No vertical exaggeration. WFFS, West Fork fault system.

produce a discernible magnetic high, but here they occur in a region of relatively low magnetic fields that lies between the West Fork and Alphabet hills highs (fig. 23). Apparently, the belts of rocks to the north and south are even more magnetic, obscuring the magnetic effect of the schistose granitic rocks.

The Alphabet hills high is thinner and shorter than the West Fork feature; it has a typical amplitude of about 300 nT. It extends from approximately the Richardson Highway to the west about 75 km. Its source rocks are massive, schistose mafic volcanic and plutonic rocks of the northern part of the GRC. Magnetic susceptibilities range from 0.0001 (cgs) for strongly schistose samples, to more typical values of about 0.0034 (cgs) for less metamorphosed samples.

The Media and Upper Tangle Lakes highs are wider and shorter than the Alphabet hills high and have amplitudes of about 600 nT. These two highs appear to be connected on the east and west, forming a 30-km-long oval whose southern (Media high) and northern (Upper Tangle Lakes high) parts could be due to the same source rocks. No samples were collected from the Upper Tangle Lakes high. Schistose mafic volcanic rocks, metagabbro, and lesser granodiorite of the northern part of the GRC crop out along the Media high. Samples of these outcrops generally have low susceptibilities, about 0.00002 (cgs), though one magnetite-bearing schistose andesite sample registered 0.008 (cgs). Because the rocks in outcrop generally have such low susceptibilities, they may not represent the source rocks for the anomaly.

**Table 2.**--Parameters for interpreted bodies shown in figure 24. Y1 and Y2 are distances the bodies are assumed to extend east and west of the plane of the figure, respectively. Susc., effective magnetic susceptibility used in making the models. These values reflect induced magnetizations in the source rocks that give rise to the "true" susceptibilities reported in the text, but also may include additional permanent magnetizations in the source rocks that are oriented parallel to the Earth's field. GRC, Gulkana River complex of Nokleberg and others (1986)

| Body | Y1 (km) | Y2 (km) | Susc. (cgs) | Rock type or feature represented |
|---|---|---|---|---|
| 1 | 55 | 80 | 0.0040 | Basement rocks--Peninsular terrane. |
| 2 | 55 | 80 | .0064 | Mafic or ultramafic buttress. |
| 3 | 32 | 20 | .0014 | Schistose granitic body of the southern part of the GRC. |
| 4 | 32 | 40 | .0022 | Mafic metavolcanic unit of the northern part of the GRC. |
| 5 | 25 | 10 | .0019 | Tangle Lakes intrusive(?) body. |

Figure 24 shows a model of possible magnetic structures along profile A-A', which crosses the West Fork, Alphabet Hills, Media, and Upper Tangle Lakes magnetic highs (fig. 23). Observed magnetic fields along this profile were taken from Andreasen and others (1958). All bodies are presumed to be magnetized in the direction of the Earth's present field, of 56,400 nT magnitude, 76° inclination, and 28° easterly declination. Magnetic susceptibilities used for the bodies are listed in table 2; higher susceptibilities are indicated on figure 24 by darker shading. Unshaded parts of the section represent nonmagnetic (very low susceptibility) rocks. The bodies shown were assumed to extend unchanged perpendicular to the plane of the section for varying strike-length distances Y1 and Y2 (table 2) and then to be cut off vertically. Y1 and Y2 were chosen to match strike lengths of corresponding anomalies on the aeromagnetic map.

Body 1 represents a thick magnetic slab that apparently comprises the basement of most of the Peninsular terrane and floors much of the Copper River basin. This unit is the same as that modeled by Campbell and Barnes (1985) along profile B-B'. The bases of all bodies were set at -10 km to match the abrupt decrease in seismic velocity recorded at approximately this depth by Fuis and others (1985).

Body 2, the source of the West Fork feature, is very magnetic, having the susceptibility of mafic or ultramafic rocks. An unpublished seismic profile crossing the southern boundary of the high 30 km to the east (Robert Shaffer, AMOCO Production Co., oral commun., 1984) shows a near-vertical fault. The sharpness and linearity of the West Fork feature argue for a similar near-vertical fault along its northern edge. Therefore, body 2 may represent a fragment of magnetic oceanic crust that was caught during accretion of the Peninsular terrane against the GRC along the WFFS. Alternatively, it may represent a very long horst of Peninsular terrane basement formed during Tertiary normal(?) faulting in the Copper River basin.

Body 3 is presumed to be composed of the schistose granitic rock of moderate susceptibility of the southern part of the GRC that crops out south of the Alphabet hills high. Its only effect on the profile is to moderate the deep polarity low just north of the West Fork feature. Body 3 probably extends only about 20 km west of the profile; beyond that point, the West Fork polarity low becomes a few hundred nT deeper.

Body 4 represents mafic volcanic and plutonic rocks of the northern part of the GRC that crop out along the Alphabet hills high. Its steeply north-dipping attitude matches the regional structure reported by Nokleberg and others (1986).

Body 5 represents an intrusive body, possibly metagabbro, that crops out in the area. This interpretation indicates that both the Media and the Upper Tangle Lakes highs arise from similar, if not identical, source rocks. This part of the magnetic profile can alternatively be modeled by a series of tabular source bodies that sheathe the exterior of body 5. Campbell and Nokleberg (1984) model the northernmost edge of body 5 in this way. Using that interpretation, the source rocks could be metasomatic concentrations of magnetite near the intrusion, such as the magnetite-bearing schist whose susceptibility was reported above.

General features on figure 24 are similar to features shown on the geologic cross section of Nokleberg and others (1986), with which it should be compared. Magnetic modeling, of which this is an example, can provide information about attitudes and depth of particular magnetic units on such cross sections. In regions of sparse outcrop, such as this part of the Gulkana quadrangle, it can also help to more precisely locate geologic boundaries, as for the WFFS, and to clarify their geometry.

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## USE OF PB-ISOTOPIC SIGNATURES FOR GEOCHEMICAL EXPLORATION IN THE HEALY QUADRANGLE, EASTERN ALASKA RANGE


**Stanley E. Church, John E. Gray, and Maryse H. Delevaux**


Studies of the Pb-isotope ratios from several mineral deposits of differing types and ages in the Healy quadrangle provide distinctive Pb-isotope signatures. These ratios can be used to fingerprint a deposit. The ratios and baseline geologic and geochemical data provide a basis for deriving a model that can

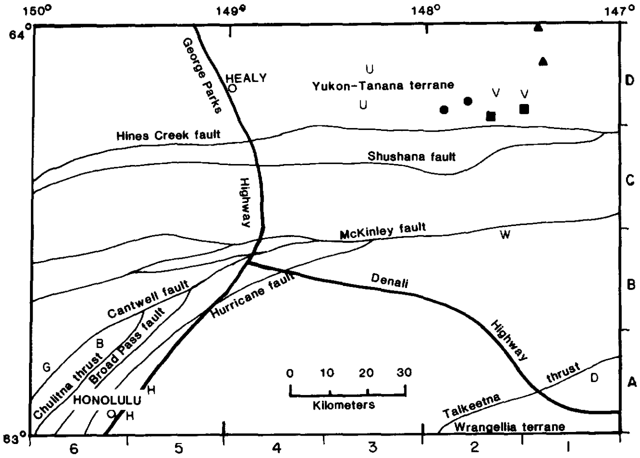


**Figure 25.**--Locations of samples from the Healy quadrangle: ●, Anderson Mountain and Virginia Creek; ■, Cirque and West Fork Little Delta River; ▲, Dry Creek and Snow Mountain Gulch; V, hydrothermal veins; W, an oxidized-Cu deposit in the West Fork Glacier area; G, greisen of Ohio Creek, Chulitna district; B, Golden Zone breccia pipe, Chulitna district; H, hydrothermal veins from the Honolulu district; U, deposits of unknown type. Sample locations from Clark and Cobb (1972); some samples and field data from D.P. Cox (USGS, written commun., 1985) and J.M. Kelly (Getty Mining Company, written commun., 1985). Letters on the right side of the figure and numerals along the bottom designate 15' x 30' subdivisions of the Healy quadrangle.

identify type and relative age of mineralization. This technique can be readily applied in mineral exploration studies in several areas in Alaska, and it is a cost-effective method for evaluation of areas of probable mineralization. The method may prove especially useful in complex tectonostratigraphic terranes where there has been metamorphic overprinting.

We have measured the Pb-isotope compositions of several deposit types in the quadrangle (fig. 25): (1) Kuroko massive sulfide deposits (KMS) and hydrothermal stibnite and polymetallic veins located along faults in the northeastern corner of the Healy quadrangle (D-1 to D-3) north of the Hines Creek strand of the Denali fault; (2) Oxidized-Cu mineralization within the Triassic limestones south of the McKinley strand of the Denali fault; (3) Hydrothermal veins in the Honolulu district that are associated with early Tertiary intrusive activity, and greisen and sulfide mineralization in the Chulitna district that is associated with Tertiary igneous activity in the southwestern corner of the quadrangle (B-6, A-6, and A-5), both south of the McKinley strand of the Denali fault; and (4) the Denali basaltic-Cu deposit (D.P. Cox, USGS, written commun., 1985) in the Nikolai Greenstone from the southeastern part of the quadrangle (A-1) in the Wrangellia terrane south of the Talkeetna thrust. Each of these groups of deposits has a distinctive Pb-isotopic signature (fig. 26).

Interpretation of the Pb-isotope data from KMS mineralization is complicated by the complex geology (Wahrhaftig, 1970a, b). Signatures of the Anderson Mountain and Virginia Creek deposits (●) plot within the field of Pb-isotope data from KMS deposits in the Jarvis Creek Glacier terrane of Aleinikoff and Nokleberg (1984) (LeHuray and others, 1985) of Devonian age (Aleinikoff and Nokleberg, 1983) in the Mount Hayes quadrangle immediately to the east. The Cirque and West Fork Little Delta River occurrences (■) associated with rhyolitic volcanic rocks in the southern part of the D-1 and D-2 quadrangles give a slightly different signature, but also reflect the high $^{207}Pb/^{204}Pb$ values in the Jarvis Creek Glacier terrane. The Snow Mountain Gulch and Dry Creek occurrences (▲) from the Mystic Creek Member of the Totatlanika Schist of Mississippian(?) age (Wahrhaftig, 1970a) reflect a high $^{238}U/^{204}Pb$ and $^{232}Th/^{238}U$ source. The high $^{207}Pb/^{204}Pb$ values in these Paleozoic KMS deposits suggest an old source terrane with a high $^{238}U/^{204}Pb$, reflecting the protolith of the Yukon-Tanana terrane.

The Pb-isotope composition of the oxidized-Cu deposit (D.P. Cox, USGS, written commun., 1985) in the Triassic limestones just west of the West Fork Glacier (B-2) and south of the McKinley strand of the Denali fault also reflects the high $^{238}U/^{204}Pb$ of the Yukon-Tanana terrane.

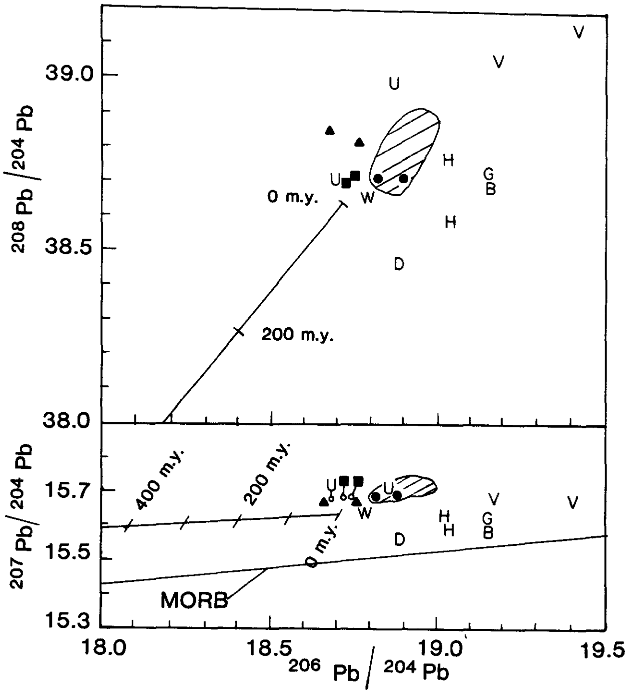


**Figure 26.**--Lead-isotope diagram showing the field of data for the massive sulfide deposits from the Jarvis Creek Glacier terrane in the Mount Hayes quadrangle (hachured field) and the Denali basaltic-Cu deposit (D, LeHuray and others, 1985). Data from the Anderson Mountain and Virginia Creek deposits (●) lie in the Jarvis Creek Glacier terrane data field. Data from both the Cirque and West Fork Little Delta River areas (■) plot outside this field. The Dry Creek and Snow Mountain Gulch KMS deposits (▲) also have distictive Pb-isotope compositions. Pb-isotope signatures of younger deposits, including the Cretaceous hydrothermal vein deposit (V), the oxidized-Cu deposit near West Fork Glacier (W), and deposits from Chulitna (G, B) and Honolulu (H) districts, are also shown. Two samples of unknown deposit type and age (U) collected from drainage basins shown to be anomalous during the reconnaissance geochemical studies have Pb-isotope compositions like the Paleozoic(?) KMS deposits. Growth curves (Stacy and Kramers, 1975) and the MORB regression line (Church and Tatsumoto, 1975) are included for reference.

These data support the hypothesis of Csejtey and others (1982) that the Triassic calcareous sedimentary rocks that occur south of the McKinley strand of Denali fault are part of or were derived from the Yukon-Tanana terrane.

Several hydrothermal events associated with igneous activity that occurred between 100 and 50 m.y. ago are distinguished in our Pb-isotope studies. Hydrothermal veins (V) deposited along the east-west trending faults in the D-1 and D-2 quadrangle are highly radiogenic, their isotope ratios plotting to the right of and above all others studied from this quadrangle (fig. 26). The Pb-isotope composition of the Golden Zone breccia pipe (B) and the Ohio Creek greisen (G) (D.P. Cox, USGS, written commun., 1985) lie along a lower $^{238}U/^{204}Pb$ growth curve, as do those of the hydrothermal vein samples analyzed from the Honolulu district.

The Denali copper deposit (D), a stratabound, basaltic-Cu sulfide deposit in the Nikolai Greenstone of Triassic age, is from the Wrangellia terrane south of the Talkeetna thrust. The Pb-isotope composition of the Denali deposit reflects a mafic or oceanic Pb-isotopic signature. Note that it plots close to the regression line for mid-ocean ridge basalts (MORB) (Church and Tatsumoto, 1975). The composition differs from that in deposits associated with the Yukon-Tanana terrane. This deposit, as well as those of the Slana River subterrane to the east in the Mount Hayes quadrangle (LeHuray and others, 1985), reflect the allochthonous nature of rocks south of this major structural boundary.

Application of Pb-isotope signatures to identification of geochemical anomalies can be demonstrated by examining the data from two anomalies in Paleozoic(?) schist in the D-3 quadrangle. There, Wahrhaftig (1970b) mapped Tertiary dikes that have intruded the Birch Creek Schist. Pb-isotope data from two mineral occurrences (U) plot near the data field defined by the KMS deposits rather than near that of any of the younger hydrothermal vein deposits. They thus reflect the high $^{238}U/^{204}Pb$ environment of the Yukon-Tanana terrane. If there is a sulfide deposit present in either of these two areas, it will be a Paleozoic KMS deposit and not hydrothermal veins associated with later igneous activity.

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## DEEP STRUCTURE OF THE CONTACT FAULT AND PRINCE WILLIAM TERRANE--PRELIMINARY RESULTS OF THE 1985 TACT SEISMIC-REFRACTION SURVEY

**Gary S. Fuis**
**and Elizabeth L. Ambos**

In 1984, the Trans-Alaska Crustal Transect (TACT) program collected about 500 km of seismic-refraction data in the Copper River Basin and Chugach Mountains of southern Alaska. In 1985, the TACT program extended coverage southward to the Pacific coast and also northeastward with the collection of another 350 km of data (fig. 27). Major results from analysis of our 1984 data include the following: (1) The Border Ranges fault (BRF), the boundary between the Chugach and Peninsular-Wrangellia (composite) terranes (CGT and PET-WRT), appears to be truncated at a relatively shallow depth (10 km) by a horizontal reflector-refractor of regional (100-km) extent; thus the CGT and PET-WRT appear rootless. (2) A series of layers with alternating low (5.5?-7.0? km/s) and high (7.6-8.1 km/s) velocities, each several kilometers thick, is observed beneath the CGT and southernmost PET-WRT; these layers dip shallowly (4-13°) northward from a minimum depth of 12 km under the central CGT. Because of their high velocities, indicating ultramafic rocks, and their dipping configuration, these layers are postulated to be subduction assemblages consisting of slices of oceanic crust and mantle, of which at least the upper two pairs appear to be now subjoined to the North American plate. These results influenced us to design the 1985 seismic-refraction experiment to (1) trace the shallow regional re-

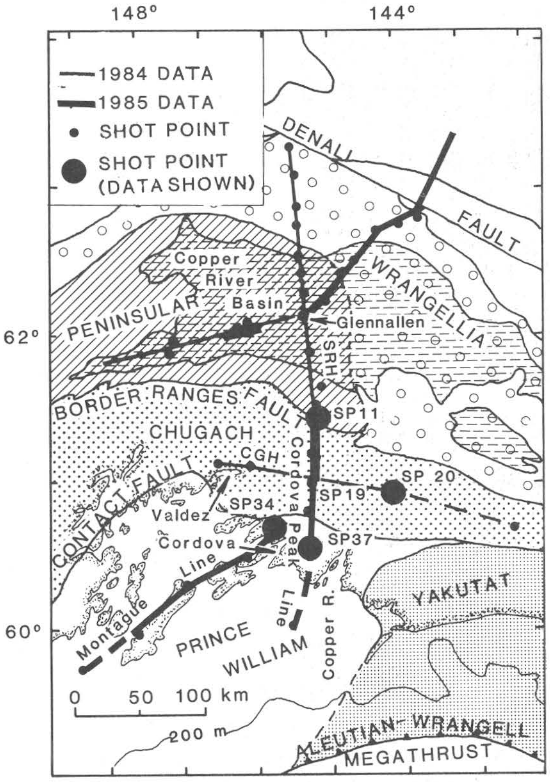


**Figure 27.**—Index map of southern Alaska showing terranes, faults, and seismic-refraction lines for 1984 (including SRH, South Richardson Highway line, and CGH, Chugach line) and 1985. Instruments were deployed along solid lines; linking of offset shotpoints to a particular profile is indicated by dashed lines. Data for shotpoints 20 and 34 are shown in figure 28 and for shotpoints 11 and 37 in figure 29. SRH overlaps the Cordova Peak line between shotpoints 11 and 19. Terranes (and patterns) are Peninsular (diagonal), Wrangellia (open circle), Prince William (clear), Chugach (large stipple), and Yakutat (small stipple). Copper River basin is indicated by dashed pattern.

flector-refractor southward to see if it also truncated the deep projection of the Contact fault (CF), which is the boundary between the Prince William terrane (PWT) and the CGT, and (2) trace the series of low- and high-velocity layers up-dip to the south to see what, if any, relation they bore to the CF or the modern subduction zone.

In 1985, data were collected along two seismic-refraction profiles in the coastal area of southern Alaska: the Cordova Peak line, which extended northward 110 km from the Copper River delta to a point on the Richardson Highway northeast of Valdez, and the Montague line, which extended 135 km northeastward from Hanning Bay on Montague Island to the mainland north of Cordova. The Cordova Peak line is perpendicular to the gross structural grain of the CGT and PWT and crosses the CF; the Montague line is parallel to the gross structural grain of the PWT. Instruments (discussed in Healy and others, 1982) were spaced approximately 1 km apart on both lines; shotpoint spacing averaged 27 km on the Cordova Peak line and 45 km on the Montague line. Shots, including two ocean shots, were also fired 50 km off the ends of both lines to increase the coverage of deeper layers, and fan shots were fired from a shotpoint west of Valdez into both lines to observe how reflectors and refractors changed laterally. The Cordova Peak line overlapped the South Richardson Highway line of 1984 by 55 km (fig. 27). Except along the Richardson Highway, instruments were deployed by helicopter, and shots were fired in glacier-terminus and other lakes.

Preliminary analysis of the 1985 data indicates that (1) the velocity structure of the PWT is quite different from that of the CGT, especially the southern part of the CGT, and (2) unlike the BRF, the CF is a profound seismic boundary apparently extending more than 20 km into the crust and truncating the regional shallow (10-km-deep) reflector-refractor, as well as the series of north-dipping low- and high-velocity layers underlying the CGT and PET-WRT.

The Montague line has not been modeled in detail yet, but a direct comparison of the data from a shotpoint at its northeast end with that from a shotpoint at the east end of the Chugach line (fig. 27), a "strike" line in the CGT, indicates major differences in velocity structure (fig. 28). More refractors are seen on the Chugach line, and velocity is higher at a given depth than velocity on the Montague line. (Depth to a refractor with a given velocity is proportional to the time-

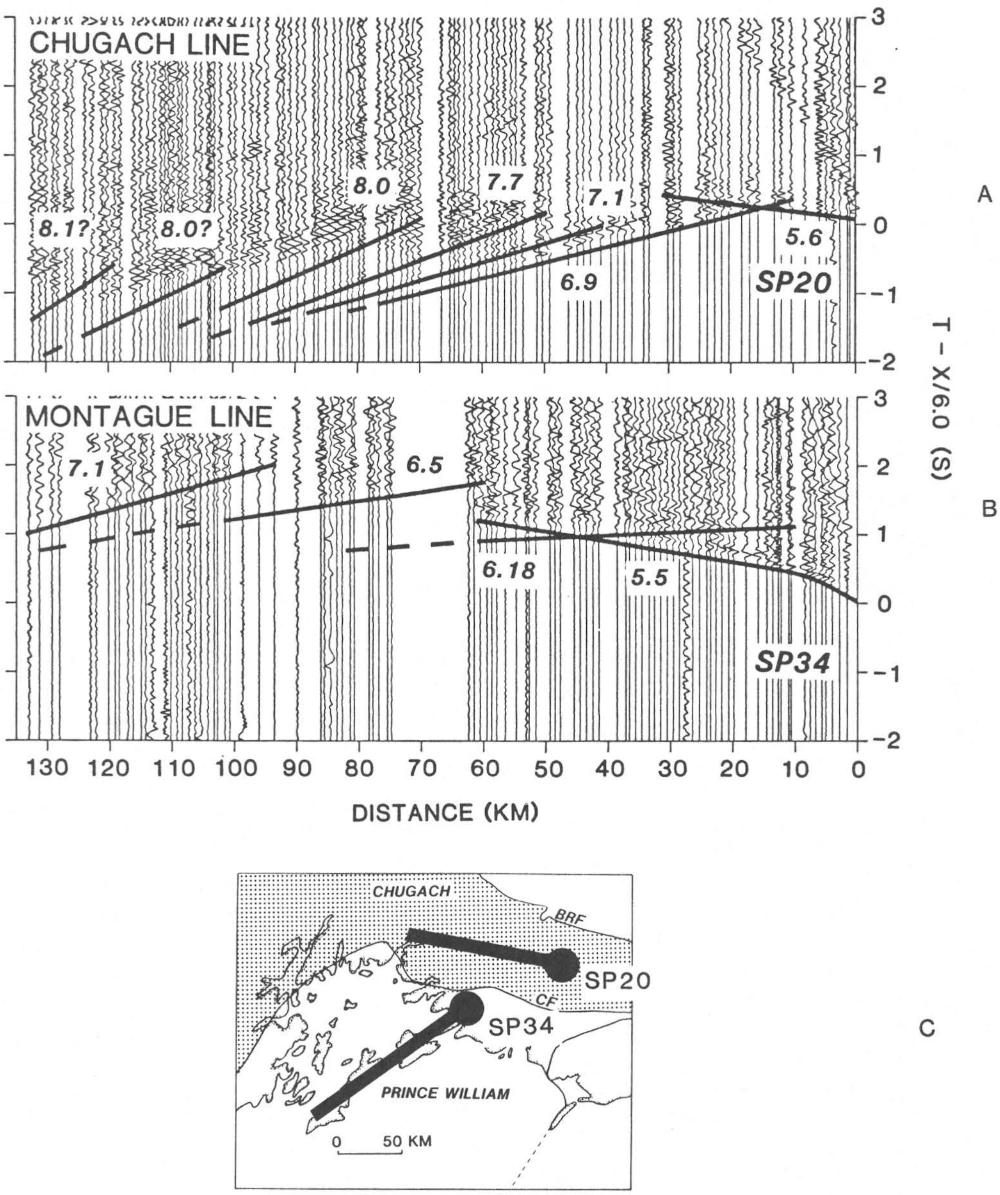


**Figure 28.**--Seismic record sections for shotpoint 20 (A) on the Chugach line and shotpoint 34 (B) on the Montague line. (C) Index map of Prince William Sound region showing Chugach terrane (shaded), Prince William terrane (south of Chugach terrane and west of dashed line), Border Ranges fault (BRF), Contact fault (CF), Chugach and Montague lines (heavy lines; see fig. 27), and shotpoints 20 and 34 (dots). Record sections are reduced by 6 km/s. Solid lines are refraction branches (associated reflections not delineated), and numbers are model velocities (Chugach line) and apparent velocities (Montague line), in km/s.

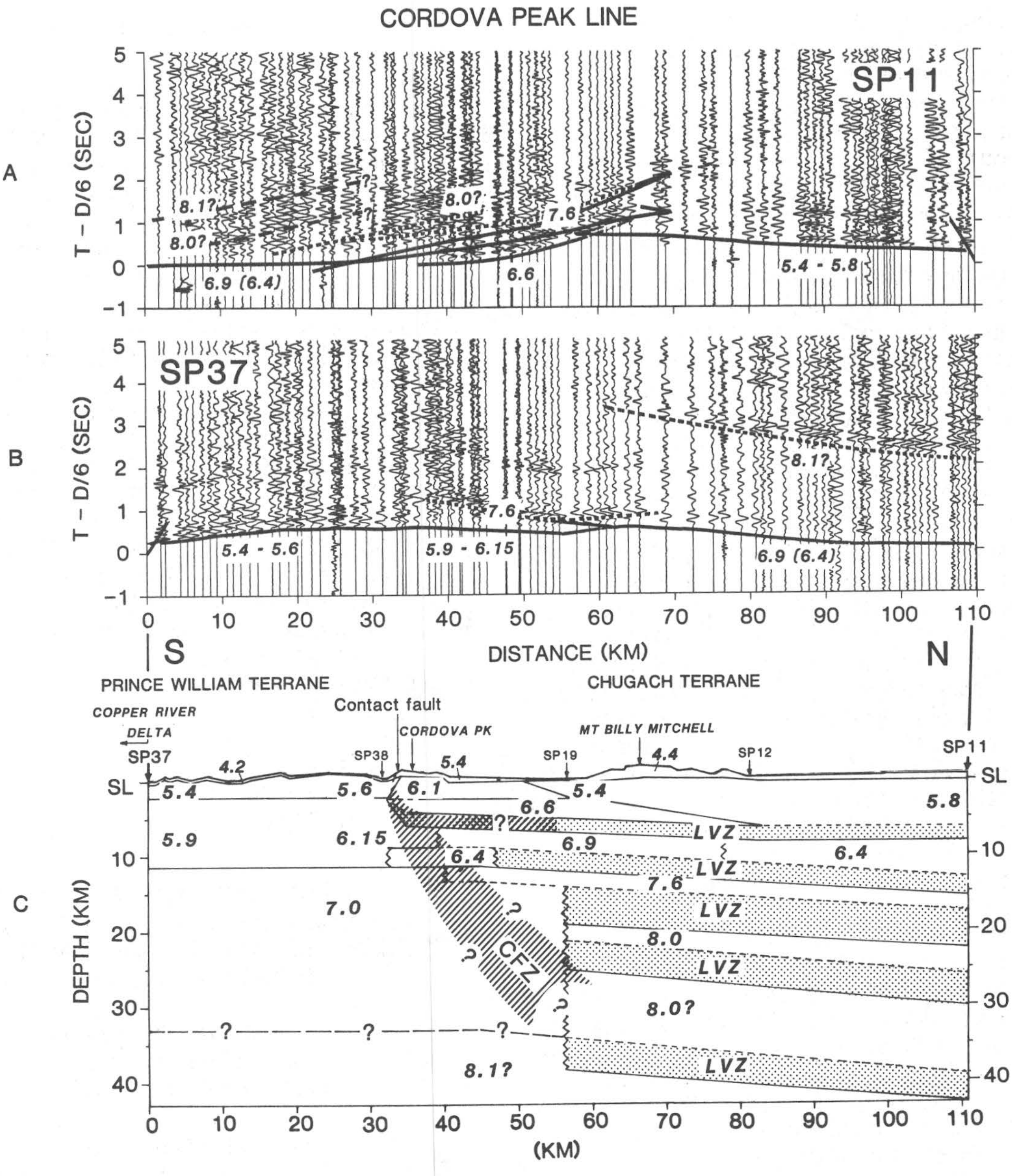


**Figure 29.**--Record section for shotpoint 11 (A) and shotpoint 37 (B) on Cordova Peak line and velocity model (C), with no vertical exaggeration. Record sections are reduced by 6 km/s. Heavy lines are model-generated traveltime curves for refractors (solid; long-dashed and queried where uncertain) and reflectors (short-dashed). Numbers on record sections and model are layer velocities, in km/s. 8.1?-km/s branch in (B) is not model-generated curve. Low-velocity zones (LVZ's) in (C) are stippled. Contact fault zone (CFZ; diagonally lined) is postulated, poorly defined zone of lateral velocity discontinuity separating the Prince William terrane from the Chugach terrane. (Branches of CFZ may sole into various model layers to north; one such branch is shown in uppermost LVZ.)

intercept, at 0-km distance, of the travel-time branch for that refractor.) In particular, no velocities greater than 7.5 km/s, characteristic of ultramafic rocks, are seen on the Montague line. The die-out of refraction branches on both the Chugach and Montague lines (indicated by dashed lines in fig. 28) indicates intervening low-velocity zones (LVZ's).

On the Cordova Peak line, velocity-structure differences between the CGT and PWT are also apparent. Data from a shotpoint in the CGT at the north end of the line (SP11, fig. 29) are characterized by a multiplicity of reflectors and refractors corresponding to the numerous layers with low (LVZ) and high (6.4-6.9, 7.6, 8.0, 8.0?, 8.1? km/s) velocities. In contrast, data from a shotpoint in the PWT at the south end of the line (SP37, fig. 29) indicate relatively simple travel-time curves. The preliminary model for this line, which is fairly well constrained down to 20-km depth by these data and the data from three other shotpoints (fig. 29), agrees with previous modelling between SP11 and SP19 (South Richardson Highway line; Fuis and others, 1985; Page and others, 1986) but shows marked lateral change southward toward the CF. In particular, the regional reflector-refractor at about 10-km depth beneath the CGT, which is overlain by a LVZ, is truncated or shoals below the trace of the CF. Also, deeper low- and high-velocity layers appear to terminate as they near the CF. South of the CF, the PWT consists of a simple stack of layers of increasing velocity, with no evidence of velocity greater than 7.5 km/s in the upper 20-30 km. No LVZ's are observed on the Cordova Peak line, as on the Montague line. Thus, the CF appears to correlate with a poorly defined, moderately northward-dipping zone of lateral velocity discontinuity.

In summary, the PWT exhibits (1) velocities that are lower at all depths than in the CGT, (2) no velocities characteristic of ultramafic rocks, and (3) less velocity layering than the CGT. In contrast to the BRF, the CF is a profound discontinuity in velocity structure apparently extending to more than 20-km depth.

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Reviewers:
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## LEAD-ISOTOPE RESULTS FROM GOLD-BEARING QUARTZ VEINS FROM THE VALDEZ AND ORCA GROUPS, CHUGACH NATIONAL FOREST

**John E. Gray,**
**Stanley E. Church,**
**and Maryse H. Delevaux**

We have initiated a regional Pb-isotope study of Au-bearing quartz veins hosted in the turbidite sequences of the Valdez and Orca Groups in order to (1) characterize and compare the Pb-isotope compositions of Au-bearing quartz veins in the Valdez and Orca Groups, (2) evaluate the known age constraints on the duration of the mineralization, and (3) identify possible sources of the Pb in these deposits. Galena and mixed sulfide minerals from Au-bearing quartz veins within the Moose Pass, Port Wells, Port Valdez, Hope-Sunrise, McKinley Lake, and Girdwood

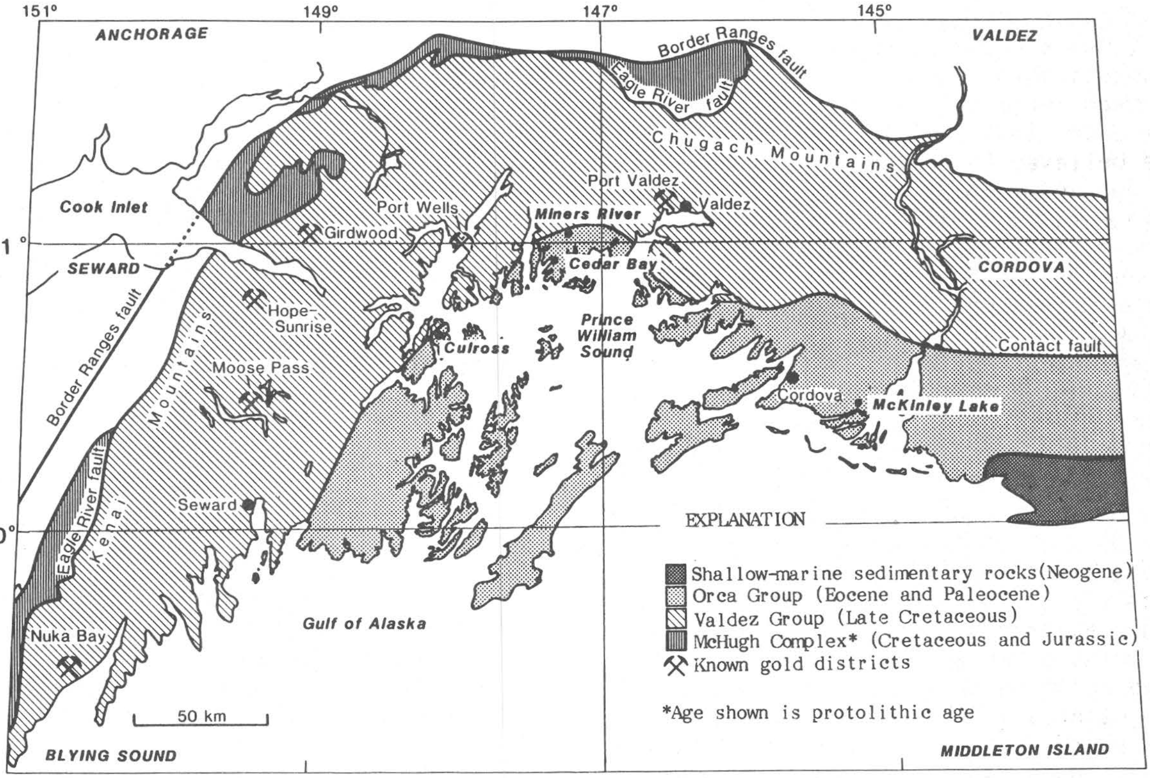


**Figure 30.**--Geologic map showing the various mining districts and mines in the Orca and Valdez Groups in the study area; geology from Winkler and others (1984).

mining districts, as well as from the Miner's River, Cedar Bay, and Culross mines have been studied (fig. 30).

The study area includes parts of the Late Cretaceous Valdez Group of the Chugach terrane and the Paleocene and Eocene Orca Group of the Prince William terrane (Jones and others, 1981). Both of these turbidite sequences are composed primarily of graywacke, siltstone, and mudstone with minor intercalated mafic volcanic rocks and pebble conglomerates (Tysdal and Case, 1979; Nelson and others, 1984). The Valdez Group was accreted to the continental margin during the Late Cretaceous or early Tertiary (Plafker and others, 1977; Tysdal and Case, 1979). The Orca Group is juxtaposed against, and in fault contact with, the Valdez Group south of the Contact fault system (Plafker and others, 1977) and was accreted to the Chugach terrane during the Paleogene (Winkler and Plafker, 1975). Both the Valdez and Orca Groups range in metamorphic grade from prehnite/pumpellyite to amphibolite grade (Nelson and others, 1986). Metamorphism of the Valdez Group followed or overlapped penetrative deformation during accretion to the continental margin (Hudson and others, 1977).

Two episodes of intrusive activity have been identified in the study area. Early Eocene intrusive rocks, ranging in composition from granite to tonalite, form part of the Sanak-Baranof plutonic belt, which extends from the Sanak Islands to Baranof Island in southern Alaska (Hudson and others, 1979), and crop out in the McKinley Lake district (Winkler and Plafker, 1981) (fig. 30). Dikes of Eocene age also occur in the Hope-Sunrise district (Mitchell and others, 1981), the Girdwood and Moose Pass districts (Nelson and others, 1985), and the Port Valdez district (Winkler and others, 1984). Oligocene intrusions range in composition from gabbro to granite and crop out in the Port Wells district and represent the youngest dated igneous activity in the Chugach terrane (Lanphere, 1966; Tysdal and Case, 1979). Eocene magmatic events are believed to represent crustal anatexis of the sedimentary prism after tectonic thickening and deformation (Hudson and others, 1979). Hudson and his coworkers showed that the K-Ar age of amphibolite-grade metamorphic rocks of the Valdez Group northeast of McKinley Lake is also Eocene and suggested that the anatectic melts were coeval with regional metamorphism. A similar origin has been proposed for the granitic Oligocene intrusions (Mitchell and others, 1981).

Two groups of Au-bearing quartz veins are present in the Valdez and Orca Groups; one group of Au-bearing quartz veins is proximal to the intrusive bodies, whereas other Au-bearing quartz veins are hosted in metasedimentary rocks where no intrusive activity has been identified (Nelson and others, 1984). Most Au-bearing quartz veins are hosted in the Valdez Group; however, those exposed at McKinley Lake, Miner's River, Cedar Bay, and Culross Island are hosted in the Orca Group. In a regional study of the Au-bearing quartz veins in the Valdez Group, Goldfarb and others (1986) demonstrate a spatial association of many Au-bearing quartz veins with medium-grade greenschist facies metamorphic rocks. Their evidence from geologic and fluid-inclusion data suggests that the mineralizing fluids were derived during regional metamorphism. However, localized studies in the Hope-Sunrise (Mitchell and others, 1981) and Port Valdez districts (Pickthorn and Silberman, 1981) suggest that mineralization may be associated with hydrothermal convection of meteoric water leaching metals from the surrounding metamorphic rocks.

Lead isotopes vary systematically over time, and because Pb commonly occurs in many ore deposits, examination of the variation of Pb isotopes in ore deposits provides potentially useful data for resolving the timing of mineralization and determining the source of metals found in deposits. This study contrasts the Pb-isotope signatures from well-studied deposits in the Valdez Group with those of the Orca Group. Our data from all the deposits studied lie along a steeply dipping linear array (fig. 31) and represent a homogeneous, well-mixed source having variable $^{238}U/^{204}Pb$. This type of Pb-isotope array is typical of volcanic arcs associated with the active continental margin of western North America. Similarly, studies of the Pb-isotope data from the Sierra Nevada batholith (Doe and Delevaux, 1973) and from the Cascade Range (Church, 1976) have shown that calc-alkaline rocks and the sediments derived from them reflect the process of crustal recycling.

The Valdez and Orca Groups have been considered as separate metallic mineral provinces by Tysdal and Case (1982). Volcanogenic-massive sulfide deposits appear to be hosted in the Orca Group, whereas the Au-bearing quartz vein mineralization is more typical of the Valdez Group. However, recent studies by Nelson and others (1984) suggest that the Orca and Valdez Groups do not represent separate mineral provinces because both deposit types commonly occur within both groups. Our data tend to support this hypothesis in that the Pb-isotope signatures from various Au-bearing quartz veins in both the Valdez and Orca Groups are very similar and in most instances analytically overlap. Furthermore, it should also be noted that the Au-bearing quartz veins hosted in the Orca Group tend to lie to the right of the regression line (fig. 31). Conversely, one sample from the Bruno-Agostino mine in the Girdwood district also lies to the right of the regression line but is hosted in the Valdez Group.

Comparison of the relative positions of individual samples on the $^{206}Pb/^{204}Pb$--$^{207}Pb/^{204}Pb$ diagram (fig. 31 A) with the existing age data indicates that mineralization took place over a relatively short time, probably 20 m.y., which is approximately equivalent to the resolution in this method. The precise age of Au mineralization in the deposits of the various districts is still somewhat uncertain, primarily because mineralized veins in the Port Valdez, Moose Pass, and Hope-Sunrise districts exhibit cross-cutting relations with Eocene intrusions, as do veins in the

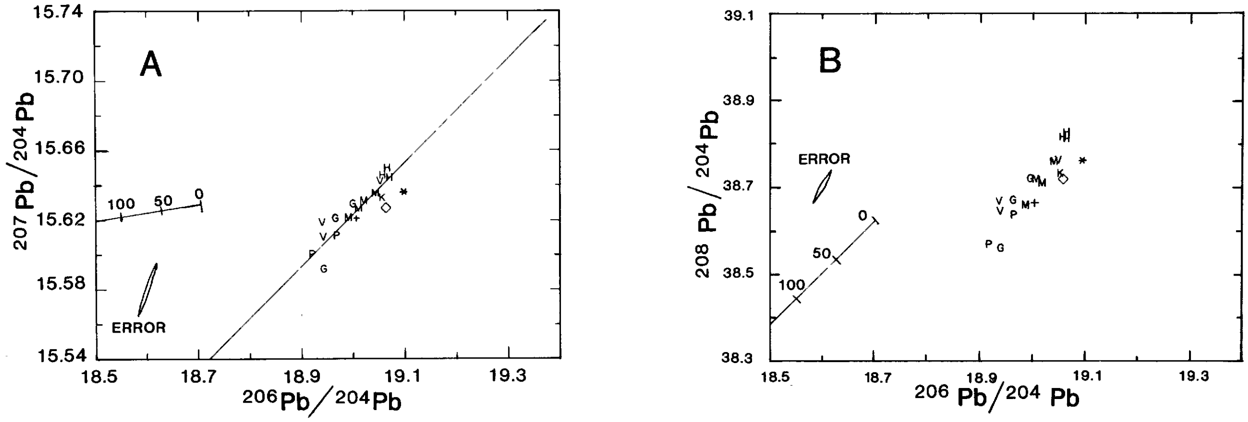


**Figure 31.**--Lead-isotope diagram showing the array of data from the Au-bearing quartz veins from rocks of the Orca and Valdez Groups. (A) ratio of $^{207}Pb/^{204}Pb$ to $^{206}Pb/^{204}Pb$, and (B) ratio of $^{208}Pb/^{204}Pb$ to $^{206}Pb/^{204}Pb$. The $^{207}Pb/^{206}Pb$ model age of the array is 3.47 b.y.±395 m.y. Growth curve (Stacy and Kramers, 1975) is also shown for reference. Data from mines within districts, and from individual mines are shown as follows: M, Moose Pass district; V, Port Valdez district; H, Hope-Sunrise district; P, Port Wells district; G, Girdwood district; K, McKinley Lake district; +, Culross Mine, *, Cedar Bay, and ◇, Miner's River. Composition of lead from the mantle, as represented by the mid-ocean ridge basalt regression line (MORB; Church and Tatsumoto, 1975), would lie along a line with much lower slope (1.6 b.y.) at $^{206}Pb/^{204}Pb$ = 18.45 and $^{207}Pb/^{204}Pb$ = 15.45, below the plot shown here.

Port Wells district with nearby Oligocene intrusions.

The Pb-isotope data from this study provide important constraints on the origin of the metals in these Au-bearing quartz veins of the Valdez and Orca Groups. The $^{207}Pb/^{204}Pb$ values and the relative position of the Pb-isotope array suggest that the source for the protolithic sediments preserved in the Orca and Valdez Groups was part of an active continental margin. This conclusion is consistent with studies of the sedimentary petrology of these rocks (for example, Dumoulin, 1984). There is no need to invoke a mantle component in these deposits because the composition of Pb from the mantle would plot substantially below the Pb-isotope array (A, fig. 31). If we assume that the Pb-isotope data from the mid-ocean ridge basalts are representative of Pb from the mantle, then our studies indicate that there are no mantle-derived metals in the Au-bearing quartz veins in the Orca and Valdez Groups.

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Reviewers:
S.W. Nelson
M.L. Silberman

## ANOMALOUS THERMAL MATURITY DATA FROM THE ORCA GROUP (PALEOCENE AND EOCENE), KATALLA-KAYAK ISLAND AREA

**Charles G. Mull and Steven W. Nelson**

Organic geochemical analyses of 23 samples of rocks of Tertiary age collected in the Katalla-Kayak Island area (fig. 32) reveal apparently anomalous indications of thermal maturity of the Paleocene and Eocene Orca Group. Based on a small number of samples, our data seem to indicate that the Orca Group samples--the oldest rocks sampled in the study area--have a lower level of thermal maturity than do the younger rocks (fig. 33, table 3). The study was undertaken as part of a U.S. Geological Survey and U.S. Bureau of Mines resource assessment of the Chugach National Forest (Nelson and others, 1984).

The mudstone and shale samples analyzed were collected from the Orca Group at Point Martin near Ragged Mountain and from all the younger stratigraphic units in the nearby areas of Katalla, the Don Miller Hills, Kushtaka Mountain, Carbon Mountain, and on Kayak Island. In order to obtain an indication of the hydrocarbon source rock potential of the Katalla and Kayak Island area, the samples were analyzed by several techniques designed to evaluate the amount and type of organic material (kerogen) in the sedimentary rocks, the ratio of pyrolytic hydrocarbon to total organic carbon determined by thermal evolution analysis-flame ionization detector (TEA-FID), and three indicators of the thermal history of the sedimentary rocks. Analyses for thermal history were vitrinite reflectance ($R_o$), thermal alteration index (TAI), and temperature of maximum pyrolysis yield from TEA-FID. Analyses were carried out by Geochem Incorporated in an analytical program designed by L.B. Magoon and G.E. Claypool, U.S. Geological Survey. Descriptions of these analytical methods are given in Barker (1974), Claypool and Reed (1976), Magoon and Claypool (1979), and Nelson and others (1984, p. 7-8).

The three independent indicators of thermal history of the Tertiary rocks of the Katalla-Kayak Island area (fig. 33) show close similarities throughout the section. Taken together, the three sets of data suggest that the Yakataga, Redwood, and Poul Creek Formations in outcrop are thermally immature or approaching the threshold of maturity for generation of hydrocarbons. Where it is more deeply buried and has a longer burial history, part of the Poul Creek Formation, which has a high total organic content (OC) and a high ratio of pyrolytic hydrocarbon to total organic carbon (PHC/OC), may be thermally mature and capable of generating hydrocarbons. The data also suggest that the underlying Tokun, Kulthieth, and Stillwater Formations in outcrop are thermally mature and within the zone of hydrocarbon generation where sufficient amounts of organic material are present.

However, an anomaly is evident in the geochemical data for the Orca Group. The two samples from these rocks have low organic content (less than 0.5 percent) and a high ratio of pyrolytic hydrocarbon to organic carbon. All three indicators of thermal maturity for the Orca Group samples suggest a lower level of organic maturity than for the Stillwater, Kulthieth, Tokun, and other younger rocks. In a normal stratigraphic sequence older rocks tend have a higher level of thermal maturity than younger rocks. This trend is thought to be due to increased thermal effects resulting from greater age and usually greater depth of burial of the older rocks.

The apparently lower level of thermal maturity for the Orca suggests that the Orca near Point Martin has had a cooler burial history than the younger Stillwater, Kulthieth, and Tokun Formations in the area. However, the data set is limited, and the interpretation could be disproportionately influenced by only a few erroneous data points. Sample 80AMu66 is from the Stillwater Formation at a location close to the Ragged Mountain fault; it could have been subject to dynamic thermal affects resulting from the faulting. If the maturity of sample

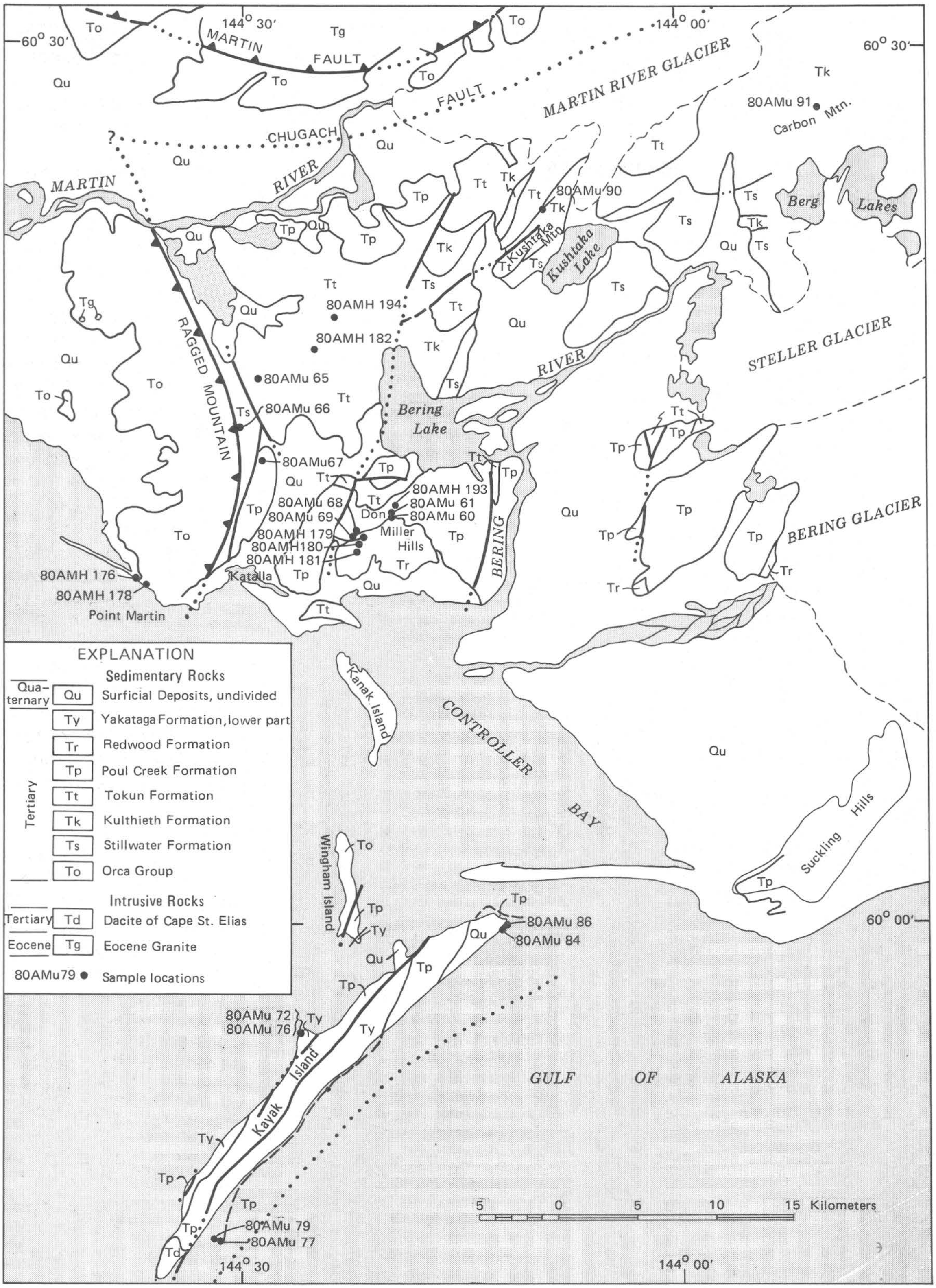


**Figure 32.**--Generalized geologic map of Katalla-Kayak Island area showing locations of samples analyzed in this study. Geology generalized from Winkler and Plafker (1981) and Nelson and others (1984).

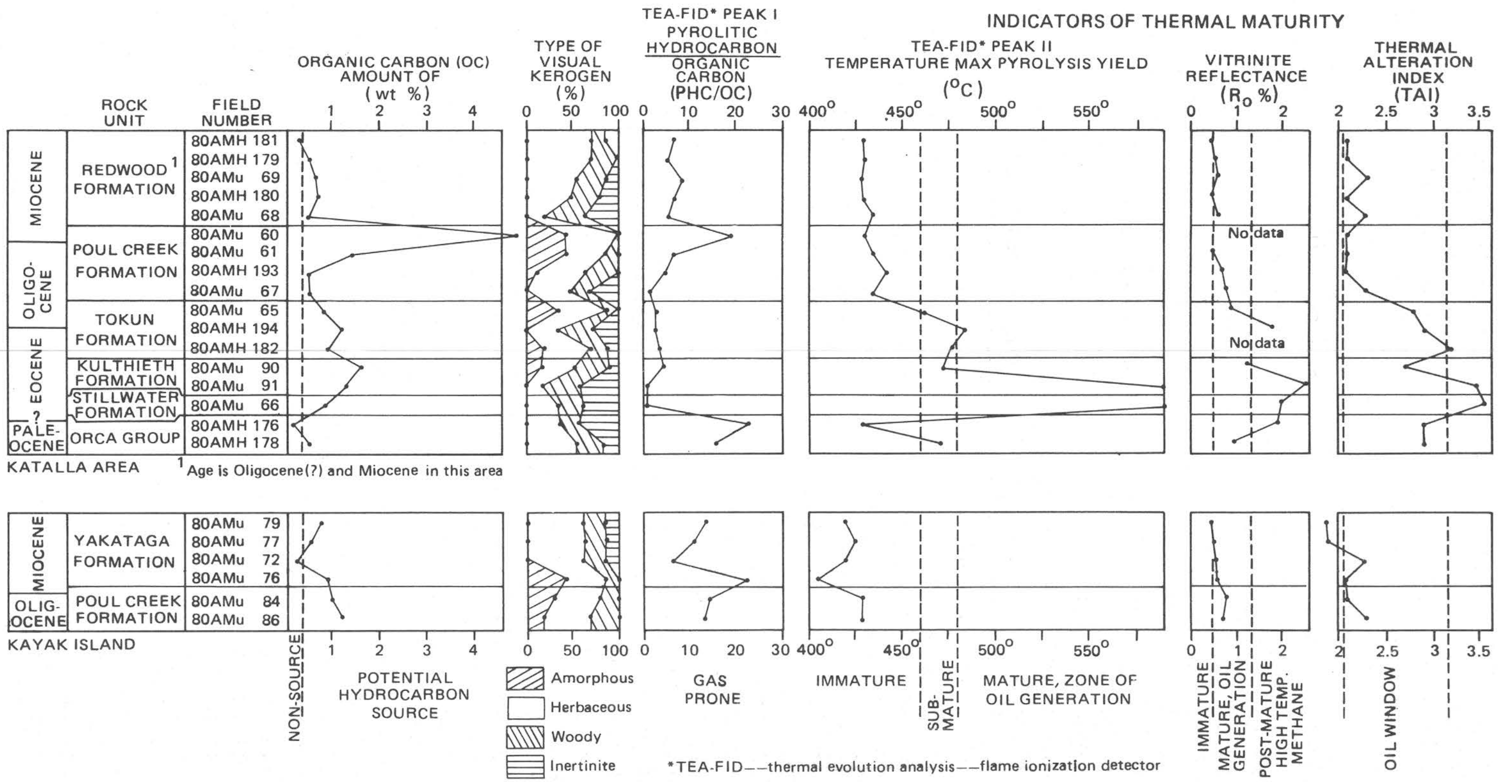


**Figure 33.**--Summary of organic geochemical data from Katalla-Kayak Island area. See table 3 for detailed analytical data.

80AMu91 from the Kulthieth Formation were erroneously high and the maturity of sample 80AMu176 or 178 from the Orca were erroneously low, some of the indicators of thermal maturity of the Orca would not be significantly different from those of the Kulthieth and Tokun Formations. Although our petrographic examinations of sandstones from Point Martin reveal no evidence of metamorphism, George Plafker (USGS, written commun., 1986) reports that a higher regional level of thermal maturity is suggested by the presence of two small stocks that intrude the Orca near the north end of Ragged Mountain and by zeolite facies metamorphism in the Ragged Mountain area. The high thermal values in the Kulthieth and Stillwater Formations in the Katalla area (table 3) are supported by petrographic observations that show zeolite facies metamorphism and quartz overgrowths (Winkler and others, 1976). Helwig and Emmet(1981) also report vitrinite reflectance ($R_o$) values ranging from 0.83 to 1.27 for ten Orca samples from Hinchinbrook and Montague Islands, 100 to 150 km west of Ragged Mountain. They also noted an unresolved discrepancy between low vitrinite reflectance values and the presence of prehnite-pumpellyite mineral assemblages reported in associated volcanic rocks (Tysdal and Case, 1979). However, the metamorphic characteristics of these volcanic rocks may be the result of seafloor metamorphism rather than an indication of a regional metamorphic imprint that also affects the sedimentary rocks of the Orca Group.

The anomalous geochemical data could result from weathering or an unexplained geochemical relation between the amount or type of organic material, the ratio of pyrolytic hydrocarbon to organic carbon, and the level of organic maturity. However, we cannot adequately evaluate these possibilities. Alternatively, the data may indicate that the Orca at Point Martin has had a shallower and/or shorter burial history than some of the younger sedimentary rocks in the Katalla area.

In the Katalla area, the contact between the Orca Group and younger rocks is a thrust fault--the Ragged Mountain fault on the west and the Chugach fault on the north. Along both faults the Orca is regionally emplaced over younger Tertiary sedimentary rocks; it has not been mapped stratigraphically beneath younger rocks anywhere in the Katalla area or in the Cape Yakataga area 110 km to the east. Regional geologic relations suggest that the Orca Group does not underlie the Stillwater and younger Tertiary rocks and was emplaced over them as a result of faulting.

The implications of the levels of organic maturity and faulting in this interpretation are compatible with regional tectonic reconstructions that indicate that the rocks of the Gulf of Alaska Tertiary Province lying seaward of the Chugach-Saint Elias fault and the Fairweather fault are part of a small tectonostratigraphic terrane--the Yakutat block (Plafker and others, 1978; Plafker, 1983). The similarity of the Eocene and younger stratigraphic succession (excluding the Orca Group) of the Katalla area to that of the Cape Yakataga and Yakutat areas to the east suggests that the Stillwater and younger rocks of the Katalla area are part of the Yakutat block. The Yakutat block is thought to have been translated northwestward along the Queen Charlotte-Fairweather fault system. Estimates of the magnitude of movement of the block range from 550 km (Plafker, 1983) to as much as 2,000 km (Keller and others, 1984) to about 3,000 km (Bruns, 1983). In the Katalla area, the Ragged Mountain and Chugach faults are probably part of the terrane boundaries on the west and north side of the Yakutat block. The Orca Group is thus part of a separate terrane, with a thermal maturation history different from that of the younger rocks on the Yakutat block.

Analyses of additional samples from the formations are in progress in an effort to confirm the apparent relations in thermal history indicated by this limited study.

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**Table 3.**--Organic geochemical data from Katalla and Kayak Island areas; see figures 32 and 33 for sample sites, formation from which the samples were taken, and age of the formation

| | | Thermal evolution analysis-flame ionization detector | | | | | Visual kerogen | | | |
|---|---|---|---|---|---|---|---|---|---|---|
| Sample number | Total organic carbon (wt%) | Volatile hydro-carbon (ppm)* | Pyrolytic hydro-carbon/ organic carbon (%) | Temp. of peak II in pyrolysis (°C) | Mean vitri-nite reflect-ance | Thermal alter-ation index (1-5)** | Amor-phous kerogen (%) | Herb-aceous kerogen (%) | Woody kerogen (%) | Inert-inite (%) |
| | | | Katalla area | | | | | | | |
| 80AMu60 | 4.83 | 3070 | 19.25 | 430 | --- | 2.1 | 43 | 57 | 0 | 0 |
| 80AMu61 | 1.42 | 184 | 7.15 | 434 | 0.5 | 2.1 | 44 | 44 | 11 | 0 |
| 80AMu65 | 0.91 | 96 | 3.34 | 464 | .9 | 2.8 | 38 | 50 | 13 | 0 |
| 80AMu66 | .87 | 151 | 1.71 | 592 | 2.1 | 3.6 | 0 | 36 | 27 | 36 |
| 80AMu67 | .58 | 171 | 2.22 | 435 | .8 | 2.3 | 0 | 50 | 20 | 30 |
| 80AMu68 | .50 | 98 | 6.04 | 435 | .6 | 2.3 | 0 | 22 | 44 | 33 |
| 80AMu69 | .62 | 136 | 9.10 | 429 | .6 | 2.3 | 0 | 56 | 33 | 11 |
| 80AMu90 | 1.61 | 209 | 4.29 | 472 | 1.3 | 2.7 | 0 | 36 | 36 | 9 |
| 80AMu91 | 1.30 | 114 | 1.51 | 590 | 2.5 | 3.5 | 18 | 20 | 40 | 40 |
| 80AMH193 | .50 | 128 | 5.44 | 442 | .7 | 2.1 | 0 | 56 | 33 | 0 |
| 80AMH194 | 1.22 | 196 | 3.31 | 484 | 1.8 | 2.9 | 11 | 36 | 36 | 27 |
| 80AMH176 | .20 | 246 | 22.70 | 429 | 1.9 | 2.9 | 0 | 40 | 20 | 40 |
| 80AMH178 | .47 | 345 | 16.06 | 470 | 1.2 | 2.9 | 0 | 57 | 29 | 14 |
| 80AMH179 | .49 | 108 | 5.96 | 432 | .6 | 2.1 | 0 | 71 | 29 | 0 |
| 80AMH180 | .67 | 92 | 7.58 | 430 | .5 | 2.1 | 0 | 50 | 30 | 20 |
| 80AMH181 | .32 | 78 | 6.94 | 434 | .5 | 2.1 | 0 | 71 | 14 | 14 |
| 80AMH182 | .96 | 114 | 4.02 | 477 | --- | 3.2 | 20 | 50 | 20 | 10 |
| | | | Kayak Island | | | | | | | |
| 80AMH72 | .37 | 131 | 7.27 | 422 | .51 | 2.3 | 0 | 63 | 25 | 13 |
| 80AMH76 | .91 | 274 | 22.26 | 407 | .50 | 2.1 | 44 | 44 | 11 | 0 |
| 80AMH77 | .56 | 255 | 11.52 | 427 | .47 | 1.9 | 0 | 63 | 25 | 13 |
| 80AMH79 | .75 | 261 | 13.85 | 418 | .44 | 1.9 | 0 | 63 | 25 | 13 |
| 80AMH84 | 1.09 | 642 | 14.29 | 431 | .71 | 2.1 | 30 | 50 | 20 | 0 |
| 80AMH86 | 1.20 | 568 | 13.60 | 430 | .67 | 2.3 | 20 | 50 | 30 | 0 |

* Peak I in pyrolysis

** 1 = unaltered; 5 = very mature

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Reviewers:
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George Plafker

## RECOGNITION OF A NIXON FORK TERRANE EQUIVALENT IN THE HEALY QUADRANGLE

**Michael W. Mullen and Béla Csejtey, Jr.**

A sequence of predominantly carbonate rocks in the Alaska Range, previously designated as the Dillinger tectonostratigraphic terrane, was reported to contain only slope and basinal deposits (Jones and others, 1981). Recent studies of this sequence in the west-central Healy quadrangle (fig. 34) indicate that it also contains well-developed shallow-water facies carbonate rocks and probably is related to the Nixon Fork terrane. This terrane was described by Patton (1978) and Dutro and Patton (1982) as consisting of a thick sequence of Ordovician through Devonian shallow-water platform carbonate rocks with an incursion of Silurian deep-water facies.

This sequence, which is shown as units DOs and Dls on the geologic map (fig. 35), crops out in a southwest-trending fault sliver that extends southwestward into the

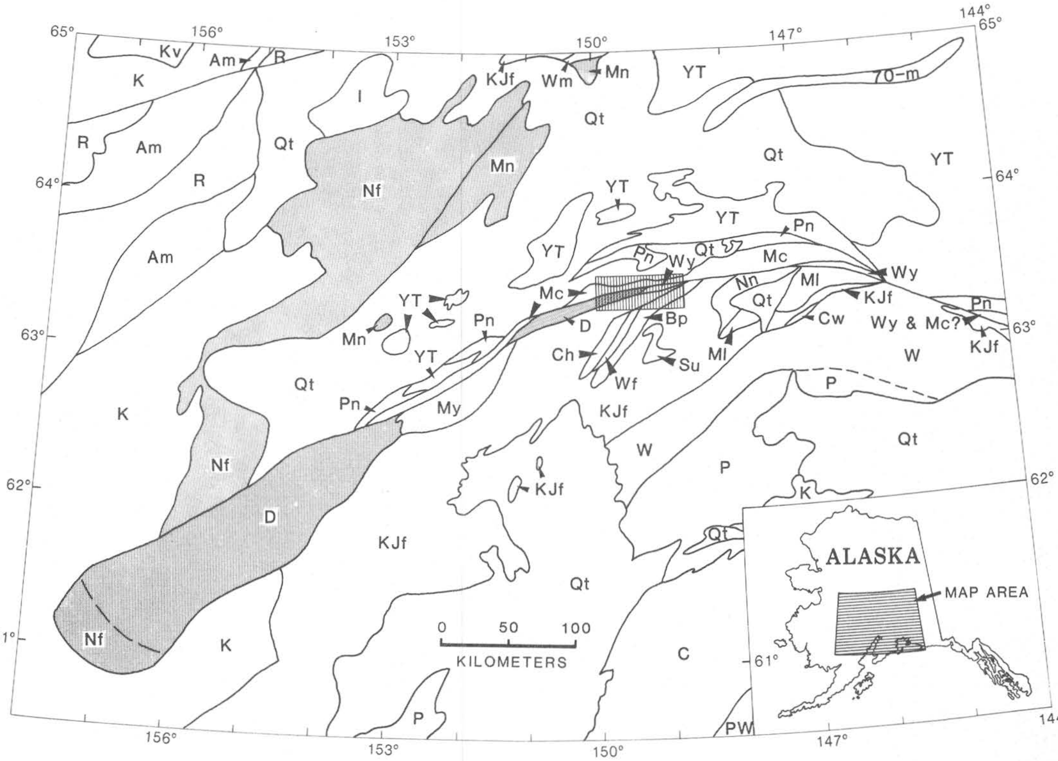


**Figure 34.**—Tectonostratigraphic terrane map of south-central Alaska (after Jones and others, 1981) showing the distribution of the Nixon Fork (NF), Dillinger (D), and Minchumina (Mn) terranes (shaded), and the location of the study area (vertical lines) in the west-central Healy quadrangle. Other terrane designations from Jones and others (1981).

Mt. McKinley quadrangle. These rocks form a complexly folded and faulted, regionally metamorphosed, but integrated marine sequence of slope and basinal turbidites and hemipelagic deposits with lesser amounts of shelf-type deposits. The thickness of the sequence cannot be accurately determined due to folding and faulting, but a minimum thickness of 1,000 m is postulated.

Lithologically, the sequence consists of: (a) About 300 m of interbedded medium- to dark-gray, medium- to fine-grained calcareous sandstone, siltstone, and dark-gray to black argillite (DOs) interpreted to be turbidite deposits; beds are as much as 1.5 m thick; (b) Intercalated thin-bedded, dark- to medium-gray limestone and dark-gray calcareous shale (DOs); (c) About 250 m of dark-gray to black, well-bedded lime mudstone to wackestone with rare argillite and chert interbeds (DOs); beds are as much as 0.2 m thick; and (d) About 20 m of massive to thick-bedded, light-gray, finely to medium-crystalline, partly dolomitic limestone (Dls). The sandstone in the turbidites contains grains of mica, feldspar, and minor pyroxene, in addition to abundant quartz and chert grains, suggesting a continental source. Outcrops of well-bedded lime mudstones and wackestones (locality 3, fig. 35) consist mainly of thinly laminated hemipelagic layers with intercalated thin-bedded turbidites, indicating an intercanyon slope or a basinal environment of deposition. In many places, poorly preserved radiolarians occur as concentrations along the more carboniferous laminae. Where its original texture is not obscured by recrystallization, the massive limestone unit (Dls) exhibits many characteristics indicative of deposition in a shallow-shelf to shore-

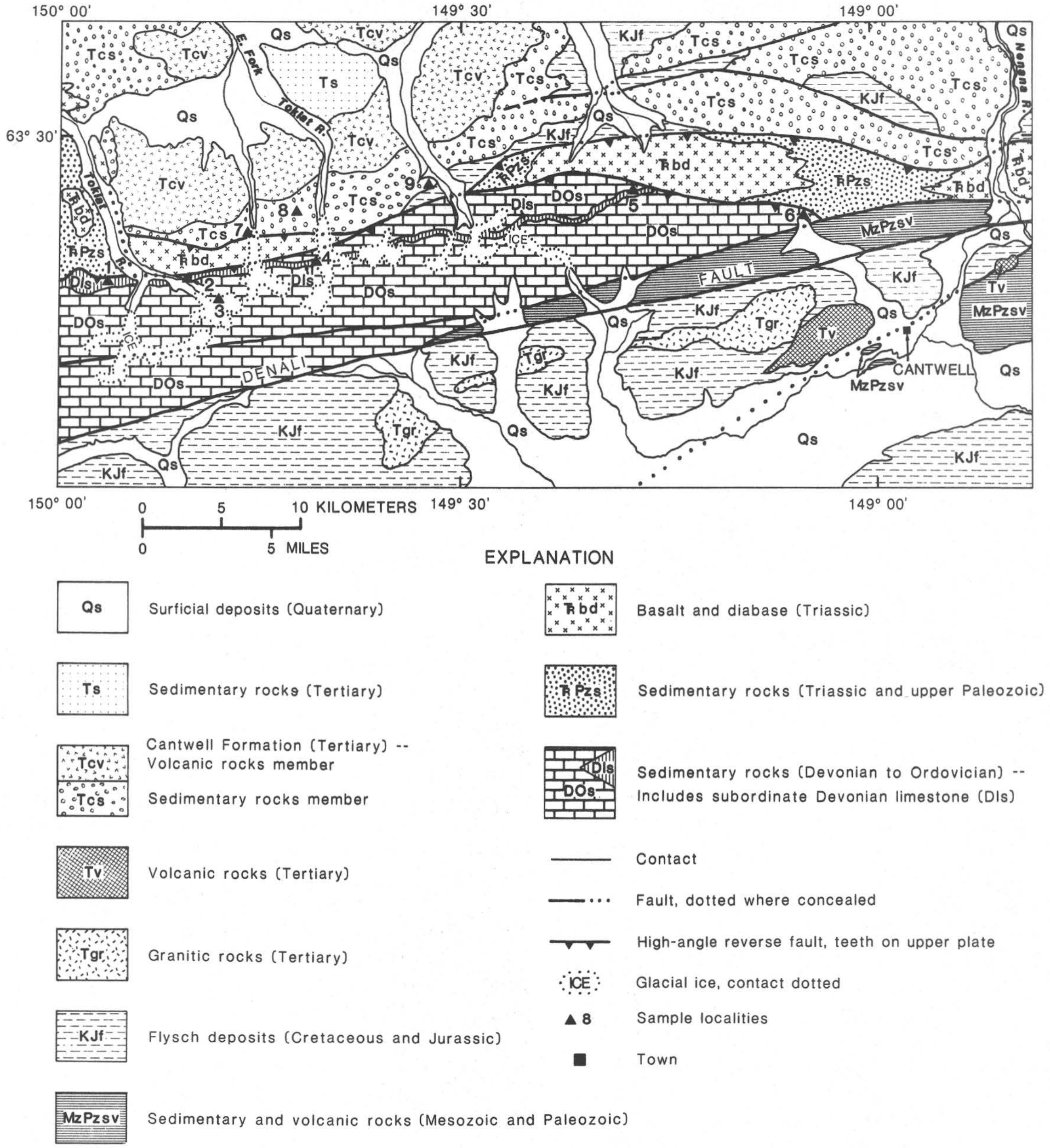


**Figure 35.**--Generalized geologic map of a part of the west-central Healy quadrangle showing fossil and carbonate rock sample localities shown in table 4. The geology is modified from Gilbert and Redman (1975), Jones and others (1983), and Csejtey and Mullen (unpublished data).

line environment. At localities 1 and 2 (fig. 35) the rocks consist mainly of algal bindstones (carbonate terminology of Embry and Klovan, 1971) and interbedded pelletal wackestone to packstone, as well as intraclastic packstone layers. These rocks also exhibit mudcracks and fenestral structures consistent with structures found in intertidal deposits. The intraclastic packstones at location 1 (fig. 35) consist of fragmented tabular algal bindstone clasts in a lime-mud matrix, indicating disruption of the tidal flat crust by periodic storm waves. At locality 4, the limestone consists of bioclastic packstone to wackestone containing some intraclasts and composite grains with micritic rinds. These characteristics indicate deposition in a shallow subtidal environment with minor wave agitation.

Carbonate clasts collected from nearby exposures of the overlying Cantwell Formation (localities 6, 7, and 8, fig. 35) exhibit shallow-water depositional characteristics. Cross-bedded, bioclastic grainstone clasts from locality 8 contain ostracodes, brachiopods, bryozoans, echinoids, foraminifers, tentaculites, girvanellid algal nodule fragments, and composite grains in sparry calcite cement. These components are indicative of a shallow-water high-energy environment, such as a sand shoal or tidal channel. The general trend of the lithologies in the dolomitic limestone unit (Dls) suggests an increase in water depth to the east.

The age of this sequence ranges from Ordovician through Devonian on the basis of: (a) Middle Devonian corals collected by Capps (1932, p. 255; and locality 5, fig. 35; table 4), (b) Ordovician to Devonian gastropods reported by Jones and others (1983; and loc. 6, fig. 35; table 4), and (c) Silurian to Devonian conodonts (loc. 4, fig. 35; table 4). Carbonate clasts collected from the nearby, stratigraphically overlying Cantwell Formation (locs. 7, 8 and 9, fig. 35; table 4) contain conodont assemblages ranging in age from Ordovician to Devonian. Because many of these clasts are essentially identical to outcrop specimens in the Dls and DOs units, they were probably derived from this carbonate sequence.

The overall lithologic character of this sequence suggests a stratigraphically upward-shallowing section having a distinct shallow-shelf facies near its top. The predominant slope and basinal deposits and the upper thinner shelf unit indicate a depositional site near the continental margin where shelf deposits prograded over deeper water deposits during Devonian time. A similar Devonian progradational event was noted by Patton and others (1984) in the Minchumina terrane. This progradation may have been due to local tectonic uplift or a global lowering of sea level, and it represents a depositional interface between the Nixon Fork shallow-water carbonate platform deposits and its deep-water equivalents.

The lithologic characteristics and age range of this sequence in the Healy quadrangle, along with our paleoenvironmental interpretations, strongly suggest a close spatial relation with strata of the Nixon Fork terrane. Churkin and others (1980) suggested that the Dillinger lithologies represent the deep-water or "shale-out" facies of the Nixon Fork carbonate platform, and Dutro and Patton (1982) provisionally considered them to be at least partly correlative with deep-water facies equivalents of the Nixon Fork and Minchumina (previously East Fork of Dutro and Patton, 1982) terranes. Blodgett and Clough (1985), Blodgett and others (1984), and Blodgett and Gilbert (1983) concluded that the Nixon Fork and Dillinger terranes were actually part of the same early Paleozoic continental margin and have undergone minor, if any, displacement with respect to the North American craton. We also conclude that the so-called Dillinger terrane in the Healy quadrangle is depositionally related to the Nixon Fork terrane and represents a tectonically fragmented segment of the Nixon Fork continental margin. Finally we conclude that because the Nixon Fork, Dillinger, and Minchumina terranes were all part of a single depositional sequence, they should be included under one term. We suggest Kuskokwim Mountains superterrane as the new term because many of the outcrops occur in the Kuskokwim Mountains and the drainage network of the Kuskokwim River.

**Table 4.**--Lithologies and fossil ages for carbonate rock sample localities in the west-central Healy quadrangle. Numbers refer to localities shown in figure 35

| Map No. | Map Unit | Field No. | General Lithology | Fossils | Fossil Age | Reference |
|---|---|---|---|---|---|---|
| 1 | Dls | 83AMM-2 | Massive to thick-bedded dolomitic lime mudstone and crystalline limestone | Algae | Devonian(?) | This paper |
| 2 | Dls | 83AMM-3 | Sheared, massive crystalline limestone with relict algal laminations | -- | -- | do |
| 3 | DOs | 83AMM-5 | Thin-bedded to finely laminated carbonaceous lime mudstone and wackestone | Radiolarians | Ordovician-Devonian(?) | do |
| 4 | Dls | 83AMM-6 | Massive to thick-bedded bioclastic wackestone and packstone | Conodonts | Silurian-Devonian | do |
| 5 | Dls | Fossils No. 3 | Massive to thick-bedded limestone | Corals | Early-Middle Devonian | Capps, 1932, p. 255 |
| 6 | DOs | 80-S-503, 80-S-504 | Thick-bedded limestone | Gastropods, bivalves | Ordovician-Devonian | Jones and others, 1983 |
| 7 | Tc | 83ACy-6 | Various carbonate clasts | Conodonts | do | This paper |
| 8 | Tc | 83AMM-7 | do | Conodonts | do | do |
| 9 | Tc | 83ACy-5 | do | Conodonts | do | do |

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Reviewers:
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## LATE PALEOZOIC AND EARLY JURASSIC FOSSIL AGES FROM THE McHUGH COMPLEX,

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Late Mississippian through Early Pennsylvanian and Early Jurassic microfossils were collected from the type locality of the McHugh Complex (Clark, 1973) along the Seward Highway southeast of Anchorage (fig. 36). Radiolarians collected indicate that some of the oceanic rocks in the McHugh Complex are Early Jurassic in age. Conodonts extracted from a conglomerate clast are of Late Mississippian through earliest Pennsylvanian age. The conodont-bearing clast may have been derived from the Strelna Formation.

The radiolarians occur in chert exposed in a road cut (sec. 15, T. 11 N., R. 3 W., Anchorage A-8 quadrangle) 0.7 km southeast of the Early Cretaceous locality described by Karl and others (1979). The chert is 10 m thick, red and gray weathering, and ribbon bedded. The chert forms a lens in greenstone. Bagotum modestum Pessagno and Whalen, Canoptum sp., Droltus(?) hecatensis Pessagno and Whalen, ?Wrangellium sp., and Zartus sp. are present. This fauna occurs in Pliensbachian rocks in Oregon and British Columbia and extends into some Toarcian strata. The age of the fauna is thus considered late Early Jurassic (late Pliensbachian to Toarcian).

Identified conodonts occur in a limestone clast from a polymictic cobble conglomerate 2 km southeast of the viewing area at Beluga Point, on the north side of the Seward Highway (sec. 32, T. 11 N., R. 2 W., Anchorage A-8 quadrangle; fig. 36). Other clasts in the conglomerate include diorite, gray argillite, sandstone, amphibolite, volcanic porphyry, and white quartz clasts. The clast yielded one Pa element of Gnathodus bilineatus (Roundy), one Pa element of Gnathodus cf. G. girtyi (Roundy) and three indeterminate bar and blade fragments. The indicated age is late Meramecian through very earliest Morrowan. The color alteration index (CAI) of 5 to 5.5 indicates that the host rock reached at least 300° to 350°C, probably before incorporation in the McHugh Complex. The presence of laumontite-, prehnite-, and pumpellyite-bearing rocks in the McHugh Complex (Clark, 1973, 1981) indicates very low-grade conditions of metamorphism, and unrecrystallized radiolarians suggest temperatures of metamorphism less than 250° C.

The McHugh Complex is a mélange of low-grade metamorphosed sedimentary and volcanic rocks and includes irregularly

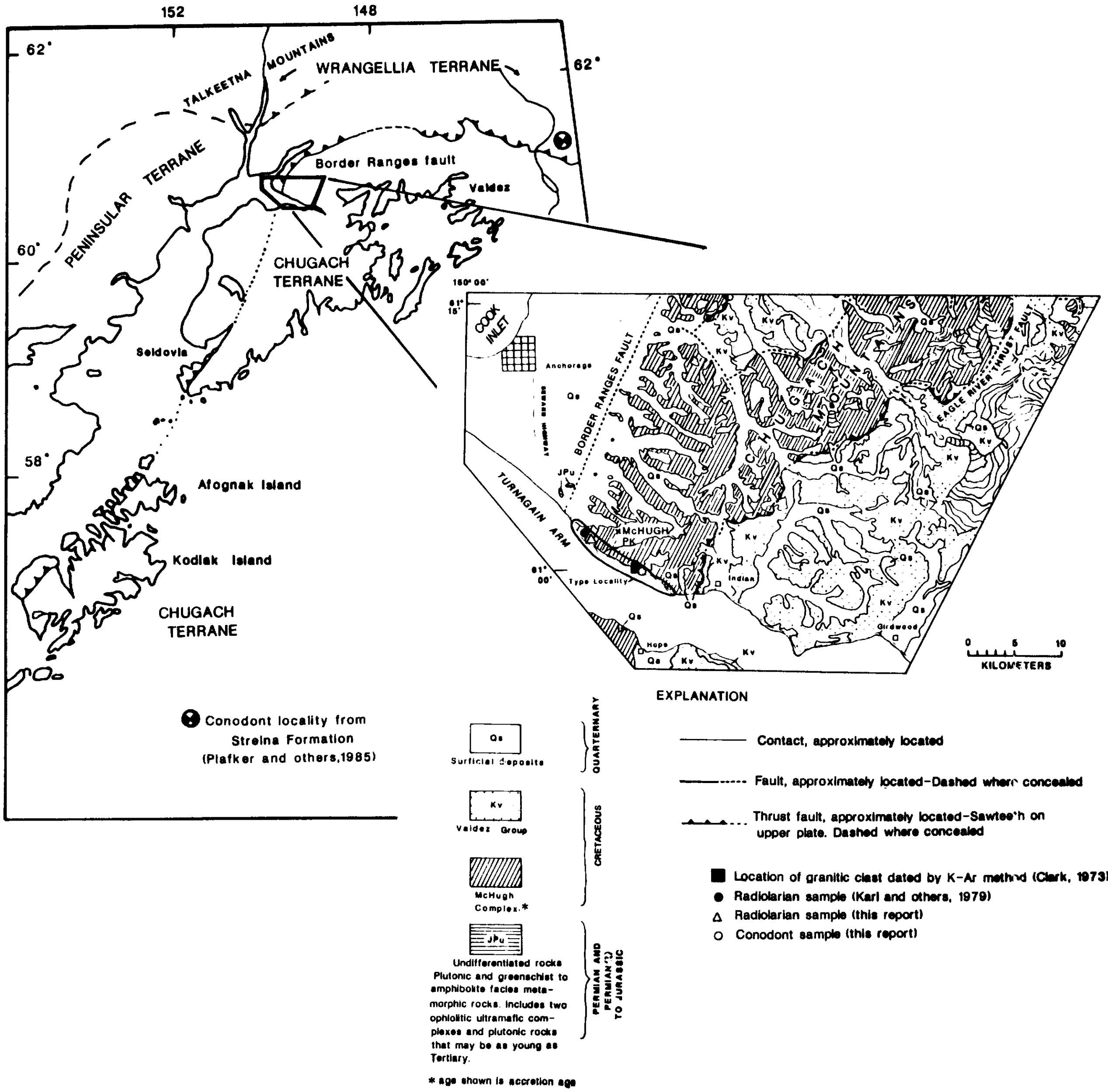


**Figure 36.**--Sketch maps showing sample localities and locations discussed in text (after Clark, 1973, and Wilson and others, 1985).

sized blocks and clasts of chert, marble, plutonic, volcanic, and ultramafic rocks (Clark, 1973). Clark (1973) recognized two fault-bounded, disrupted lithologic sequences. To the north is an oceanic sequence consisting of dark-gray argillite, with wispy lenses of green tuff, gray siliceous argillite and chert, and minor pillow(?) basalt. To the south, a largely clastic sequence of greenish-gray graywacke occurs that generally coarsens southward to a cobble conglomerate.

Plafker and others (1976) included the McHugh Complex in the mélange facies of the accretionary Chugach terrane. This terrane is part of a 1,600-km-long belt of marine flysch and mélange extending from Chatham Strait in southeastern Alaska through the Chugach Mountains to the Shumagain and Sanak Islands along the Alaska Peninsula (Plafker and others, 1977).

The age of accretion of the McHugh Complex is considered the age of the youngest rocks incorporated in the mélange, which has been reported as mid(?)-Cretaceous (Plafker and others, 1985). In contrast, older fossil ages provide clues as to possible source terranes for the various rock units found in the mélange. The allochthonous Peninsular terrane of Jones and Silberling (1979) is considered by Winkler and others (1984) a likely source for some of the fragments in the mélange. Limestone phacoids in McHugh Complex correlative rocks on Kodiak Island contain Spongiomorpha, which is strikingly similar to a Late Triassic (Norian) hydrozoan from the Alaska Peninsula (N. J. Silberling, in Connelly, 1978). Clark (1973) obtained a K-Ar date of 150 ± 7.2 Ma on hornblende from a granitic clast in McHugh conglomerate; this date corresponds with dates from Jurassic plutons in the northern Alaska Peninsula and southern Talkeetna Mountains (Reed and Lanphere, 1972; Csejtey and others, 1978; Hudson, 1983). Winkler and others (1981) described and dated by K-Ar methods large tectonic inclusions of layered gabbro (185 ± 19 Ma), amphibolite (267 ± 8 Ma), and blueschist-greenschist (152-175 Ma) in the McHugh Complex in the Valdez quadrangle. Analogous rocks occur on Kodiak and Afognak Islands and near Seldovia (Carden and others, 1977). In the northern Chugach Mountains, rocks similar to those described by Winkler and others (1981) and Carden and others (1977) form the basement for, and early plutonic phases of, the northern part of the Peninsular terrane (Pavlis, 1983; Winkler, 1983).

The previously reported paleontologic age limits for the McHugh Complex in the Anchorage area were determined from late Paleozoic (Pennsylvanian?) fusulinids (Clark, 1973) to Early Cretaceous radiolarians (Karl and others, 1979). Paleontologic ages from correlative rocks elsewhere in southern Alaska (Plafker and others, 1976; Decker, 1980; Karl and others, 1982), in the Chugach Mountains (Winkler and others, 1981), and on Kodiak Island (Connelly, 1978) span a similar range. Middle Cretaceous (Albian to Cenomanian) radiolarians from the Valdez quadrangle are the youngest fossils yet recovered from the McHugh Complex (Winkler and others, 1981). The presence of both late Early Jurassic (this report) and Early Cretaceous fossils (Karl and others, 1979) from the same outcrop area suggests that Lower Jurassic and Lower Cretaceous rocks may have been mixed by geologic processes taking place on the ocean floor and (or) at the continental margin during accretion.

New conodont ages reported herein allow consideration of the source area for debris in the clastic sequence of the McHugh Complex. Winkler and others (1984) suggest that the Peninsular terrane (Jones and Silberling, 1979) was a likely source for part of the McHugh Complex based on faunal similarity and a radiometric date from a clast in the conglomerate unit at the type locality. The Peninsular terrane, however, as originally defined (Jones and Silberling, 1979; Silberling and Jones, 1984) and redefined (Wilson and others, 1985), is not recognized as containing rocks older than Permian (R.L. Detterman, USGS, oral commun., 1985); hence, another source terrane is required for some of the older rocks in the McHugh Complex. Plafker and others (1985) report an early Early Pennsylvanian fauna from the Strelna Formation in the Chitina Valley (fig. 36). Conodonts of the Declinognathodus noduliferous-Neognathodus bassleri symmetricus zones of early

Morrowan age occur in the Strelna Formation. The CAI of 5.5 indicates that the host rock there reached at least 350° C.

The Strelna Formation also contains upper greenschist to lower amphibolite facies metavolcanic rocks, schist, chert, and marble that have been intruded by various plutons of Late Pennsylvanian, Early Jurassic to Early Cretaceous, and late Cenozoic ages (Plafker and others, 1985). The Strelna Formation is tentatively considered by Plafker and others (1985) as part of the Wrangellia terrane of Jones and others (1977), on the basis of similar lithologies and ages.

The Strelna Formation (or Wrangellia terrane) may have been the source of clasts in the McHugh Complex although other terranes, either unrecognized at present or missing, cannot be ruled out. This hypothesis is supported by similar ages for fossils collected in McHugh Complex conglomerate and fossils collected in the Strelna Formation, similar metamorphic temperatures (CAI), and the presence in the McHugh Complex of Jurassic plutonic clasts of similar type and age as some plutons that intrude the Strelna Formation.

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## CRETACEOUS DEFORMATION AND METAMORPHISM IN THE NORTHEASTERN MOUNT HAYES QUADRANGLE, EASTERN ALASKA RANGE

**Warren J. Nokleberg, John N. Aleinikoff, and Ian M. Lange**

The northeastern Mount Hayes quadrangle, north of the Mount Gakona fault, is composed, from north to south, mainly of the Lake George, Macomb, and Jarvis Creek Glacier tectonostratigraphic terranes (fig. 37) (Nokleberg and others, 1983; Nokleberg and Aleinikoff, 1985), which occur in the southern part of the Yukon-Tanana terrane of Jones and others (1984). To the south, the Hayes Glacier and Windy terranes occur between the Mount Gakona and Denali faults (fig. 37) (Nokleberg and others, 1983; Nokleberg and Aleinikoff, 1985). Locally extensive granitic plutons of mainly Late Cretaceous age occur in the area (fig. 37). The five terranes occur on the south limb of a major east-west-striking antiform whose axis is north of the area shown in figure 37; this major antiform is partly defined by the progressive change in attitude of a prominent schistosity (fig. 37). Major faults between all of the terranes approximately parallel this prominent schistosity (fig. 37). The Lake George terrane, which dips gently to the south, occurs near the core of the antiform and is interpreted to be the deeper level of a Devonian igneous arc (about 370 m.y.) (Aleinikoff and Nokleberg, 1985a, b; Nokleberg and Aleinikoff, 1985). The Macomb, Jarvis Creek Glacier, and Hayes Glacier terranes are present at successively shallower structural levels on the south limb of the antiform and are interpreted as successively higher levels of

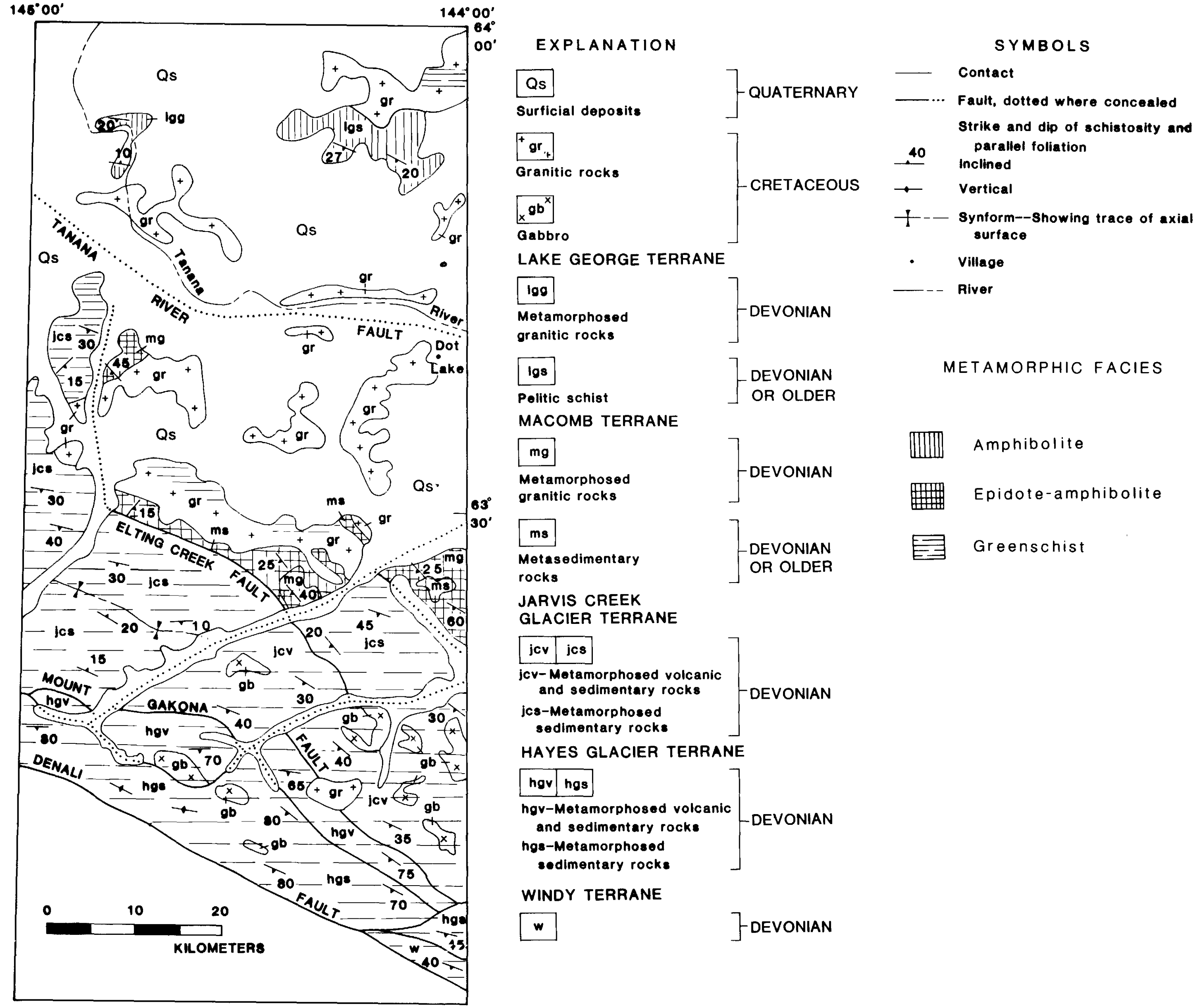


**Figure 37.**--Generalized bedrock geologic map of the northeastern Mount Hayes quadrangle, eastern Alaska Range, showing distribution of metamorphic facies.

the arc; dips steepen progressively to the south and are nearly vertical along the Denali fault (fig. 37).

Two generations (G1 and G2) of penetrative minor structures occur in the area. Because younger G2 structures predominate, they are described first. The prominent (G2) schistosity on figure 37 defines the major antiform and is an intensely developed fluxion structure that occurs throughout the area. Other G2 structures consist of sparse to rare, small- to moderate-size folds and a strongly developed lineation that parallels fold axes. G2 folds, which occur mainly in metasedimentary and metavolcanic rocks, vary from tightly appressed to isoclinal, with axial planes parallel to schistosity. G2 folds are strongly asymmetrical and exhibit a nearly constant south-southwest vergence (Nauman and others, 1980; and this study). G2 schistosity and lineation are nearly equally well developed in all metamorphic rocks. G2 lineations are constant in orientation and plunge shallowly to the west-northwest or east-southeast, approximately parallel to the strike of G2 planar structures.

G2 structures deform older G1 structures, which consist principally of well-developed schistosity and parallel compositional layering. Other G1 structures are sparse isoclinal folds with axial planes that parallel G1 schistosity and lineations that parallel G1 fold axes. G1 planar structures have variable orientations due to being folded around G2 fold axes and transposed into attitudes that parallel G2 schistosity. No asymmetry could be determined for G1 folds. G1 fold axes, where well-exposed, appear to be subparallel to G2 fold axes. G1 structures, which are not easily recognized, are more common to the north in the postulated deeper structural levels of the Devonian igneous arc. To the south, intense development of G2 schistosity and lineation has obliterated G1 structures and associated metamorphic minerals, particularly in the Hayes Glacier terrane.

Two major regional metamorphic events occur in the area. Minerals produced during the older metamorphic event crystallized along G1 schistosity and lineation; those produced during the younger metamorphic event crystallized along G2 schistosity and lineation. Minerals from the younger metamorphic event replace those from the older event and are therefore the most common, since G2 structures are more commonly developed. The later metamorphism decreases in grade systematically from north to south, with middle or upper amphibolite facies minerals occurring in the deeper structural levels in the north, and low greenschist facies minerals occurring in the shallower structural levels to the south (fig. 37; table 5). Metamorphic effects of the older event are most easily recognized in the Lake George terrane where G1 schistosity, defined by amphibolite facies minerals, is refolded around G2 minor folds and crosscut by G2 axial plane schistosity. However, in this area G2 schistosity also exhibits amphibolite facies minerals. From north to south, abrupt lowering of the metamorphic grade of the younger event occurs at major faults between terranes (fig. 37). In addition, in the Jarvis Creek Glacier terrane, hornblende, biotite, and garnet are successively replaced by chlorite towards the south. These relations suggest that during the younger metamorphic event, high-grade assemblages at deep structural levels were partly to completely replaced by lower grade assemblages at shallower structural levels.

Several isotopic studies in this area and in the Yukon-Tanana terrane farther to the north have investigated the age of what we interpret to be the younger event in this study. The results of these studies include: (1) A Rb-Sr internal isochron age of 102 m.y. for splits of plagioclase, K-feldspar, biotite, and whole rock from a Devonian metagranodiorite, metamorphosed at amphibolite facies, in the Lake George terrane (Aleinikoff and Nokleberg, 1985a, b); (2) A Rb-Sr internal isochron age of 115 m.y. for splits of plagioclase, K-feldspar, white mica, apatite, and whole rock from a Mississippian augen gneiss metamorphosed at amphibolite facies north of the study area (Dusel-Bacon and Aleinikoff, 1985; Aleinikoff and others, 1986); (3) A K-Ar age of 119 m.y. for white mica in amphibolite facies schist of the Yukon-Tanana terrane in the

**Table 5.**--Metamorphic minerals of the youngest (Early Cretaceous) metamorphic event, facies of Devonian igneous-arc terranes, and Late Cretaceous granitic rocks, northeastern Mount Hayes quadrangle, eastern Alaska Range

| Unit | Metamorphic facies | Major metamorphic minerals |
|---|---|---|
| Lake George terrane | Middle to upper amphibolite facies | Sillimanite, garnet, hornblende, plagioclase, biotite, K-feldspar, muscovite, quartz |
| Macomb terrane | Epidote-amphibolite facies | Hornblende, biotite, quartz, plagioclase, K-feldspar, andalusite, muscovite, clinozoisite, wollastonite |
| Jarvis Creek Glacier terrane | Upper to lower greenschist facies (north to south) | Muscovite, quartz, albite, andalusite, actinolite, epidote, chlorite to south; hornblende, biotite, garnet to north, successively replaced by chlorite towards south |
| Hayes Glacier terrane | Low greenschist facies | Muscovite, quartz, muscovite, actinolite, albite, andalusite, epidote, chlorite, biotite (mostly replaced by chlorite) |
| Late Cretaceous granitic rocks | Low greenschist facies | Muscovite, quartz, albite, chlorite, epidote |

western Tanacross quadrangle (Foster, 1970; Wilson and others, 1985); (4) A Rb-Sr internal isochron age of 110 m.y. for splits of biotite, plagioclase, and whole rock in the Macomb terrane (this study); and (5) K-Ar ages of 112-114 m.y. for white mica in low greenschist facies rocks in the southern part of the Jarvis Creek Glacier terrane, near the Denali fault (Turner and Smith, 1974). These isotopic data show a remarkably narrow age range of about 102-119 m.y. from what we interpret to be younger metamorphic event minerals (ranging from amphibolite to low greenschist facies) and suggest an Early Cretaceous age for this younger regional metamorphic event and associated penetrative deformation. This age is one of two interpretations discussed by Dusel-Bacon and Aleinikoff (1985) and Aleinikoff and others (1986). The age of the older metamorphic event is constrained only by the Mississippian age of the augen gneiss and the Early Cretaceous age of the younger event (Aleinikoff and others, 1985b).

Subsequent to the younger deformation and metamorphism in the Early Cretaceous, the intense metamorphic fabric of the Devonian igneous-arc terranes was locally crosscut during intrusion of Late Cretaceous granitic plutons in the northern terranes (fig. 37). The granitic plutons in this and nearby areas have K-Ar hornblende and biotite ages of about 90-97 m.y. (Wilson and others, 1985) and a U-Pb zircon age of 90 m.y. (this study). Local penetrative deformation and metamorphism sporadically occur in these Late Cretaceous plutons in the northern part of the study area and are defined by lower greenschist facies minerals occurring along G2 schistosity, or along parallel cataclastic shear zones (Nokleberg and Aleinikoff, 1985). The granitic plutons also generally exhibit narrow contact metamorphic aureoles. Together, these observations suggest the granitic plutons intruded during the waning stages of the younger deformation and metamorphic event.

The above data and interpretations show that G2 structures and metamorphic minerals formed during an intense, widespread Early Cretaceous deformation that affected large areas of the eastern Alaska Range north of the Denali fault and adjacent areas of the Yukon-Tanana Upland. This deformation was characterized by: (1) development of a G2 schistosity, lineation, and south-verging, tightly appressed to isoclinal folds, with schistosity more intensely developed at shallower structural levels; and (2) regional metamorphism ranging from upper amphibolite facies at the deeper and hotter structural levels to low greenschist facies at the shallower and cooler structural levels. The association of low greenschist facies minerals at the shallowest structural levels with south-verging, G2 folds suggests that the Devonian igneous-arc terranes in this region were being overthrust from the north by units subsequently removed by erosion, and (or) were being underthrust from the south. After the younger event and associated deformation, G2 planar structures were rotated, and near-vertical dips occur in these rocks to the south near the Denali fault.

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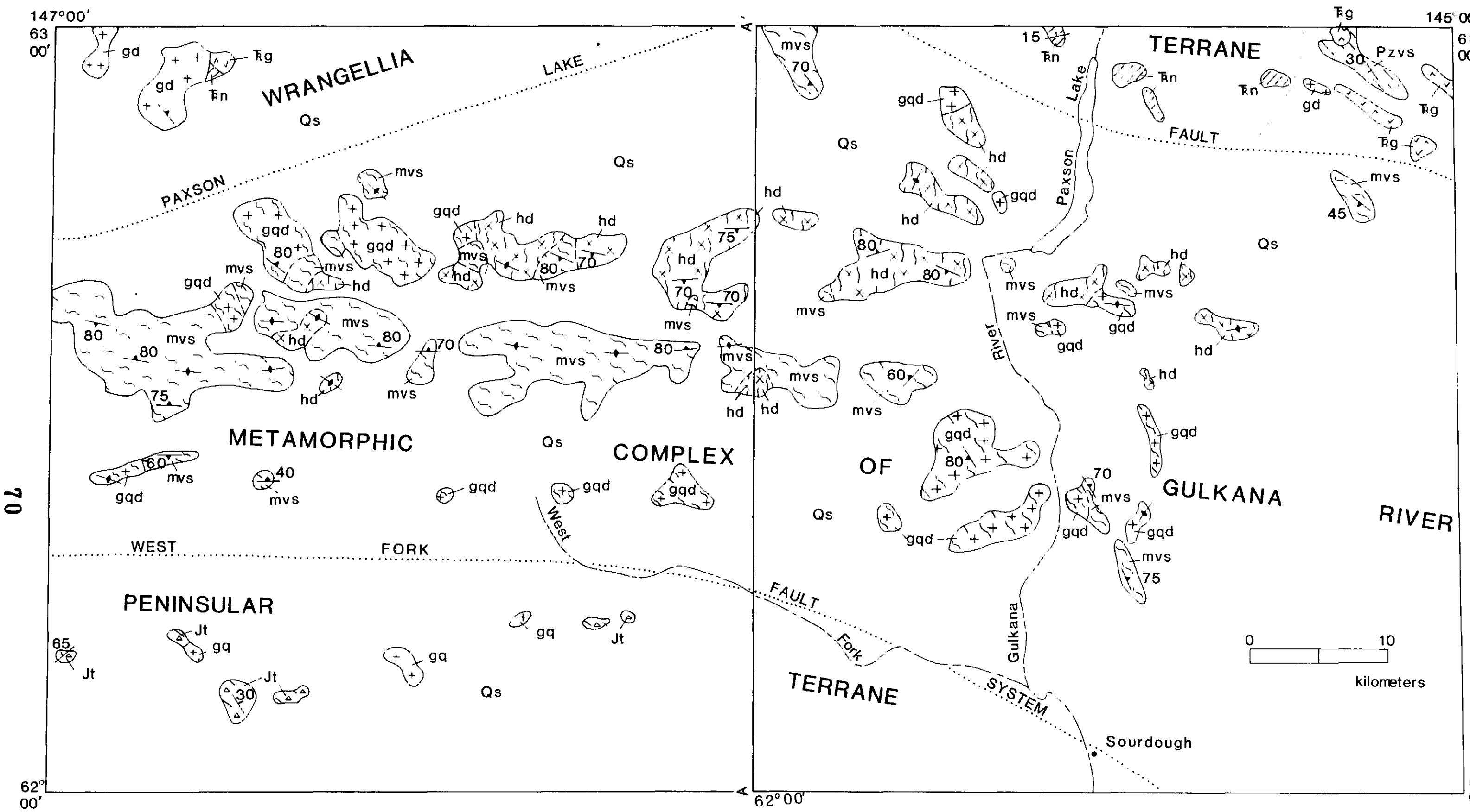
## SUMMARY OF GEOLOGY OF THE PENINSULAR TERRANE, METAMORPHIC COMPLEX OF GULKANA RIVER, AND WRANGELLIA TERRANE, NORTH-CENTRAL AND NORTHWESTERN GULKANA QUADRANGLE

**Warren J. Nokleberg,**
**W. Michael Wade,**
**Ian M. Lange,**
**and George Plafker**

Bedrock geologic mapping of the Gulkana C-3 through C-6 and D-3 through D-6 quadrangles was recently completed at a scale of 1:63,360 as part of the Alaska Mineral Resource Assessment Program and for the Trans-Alaska Crustal Transect (TACT) Program. Previous reconnaissance geologic mapping of bedrock in various parts of the study area was done at smaller scales by the U.S. Army Corps of Engineers (1960), Rose and Saunders (1965), and Ferrians (1971). Unpublished geologic mapping at a scale of 1:250,000 was done by T.E. Smith and T.K. Bundtzen (T.E. Smith, Alaska Div. of Geol. and Geophys. Surveys, written commun., 1982). From south to north, three major belts of bedrock occur in the study area: the Peninsular terrane (PT), the metamorphic complex of Gulkana River (GRC), and the Wrangellia terrane (WT) (fig. 38). Two major faults separate the three belts of bedrock: the West Fork fault system (WFFS) separates the PT and GRC; the Paxson Lake fault (PLF) separates the GRC and WT (fig. 38).

In the southern part of the study area, the PT consists, in part, of a sequence of volcanic conglomerate and sandstone, medium-bedded fine-grained dacite and hornblende andesite tuffs or flows, and lesser chert (fig. 38). The volcanic conglomerate and sandstone varies from thick bedded to massive and locally contains clasts up to 10 cm long of greenstone and volcanic porphyry in a matrix consisting of coarse sand-size greenstone, quartz, and sand- to pebble-size, angular to rounded, white feldspar grains. Pebble imbrication occurs locally in some layers. Submarine ashflows and waterlaid tuffs, locally silicified, occur in places within the volcanic conglomerates and sandstones. Some sparse chlorite-epidote hydrothermal alteration occurs in the matrix of volcanic rocks, and a few samples contain abundant metamorphic(?) granoblastic chlorite and epidote. The only structures recognized in this sequence are the low to moderate dips in bedding to the northeast or northwest (fig. 38). This unfossiliferous sequence is correlated with similar rocks of the Early Jurassic Talkeetna Formation that occurs on strike to the west in the Talkeetna Mountains quadrangle (Csejtey and others, 1978) and along the southern margin of the Copper River basin (Winkler and others, 1981). In the study area, Late Jurassic(?) hornblende granodiorite and biotite quartz monzonite apparently intrude the Talkeetna Formation in some places and are interpreted as part of the PT.

North of the PT, the GRC consists of three sequences (fig. 38): (1) metavolcanic and metasedimentary rocks that include an unfossiliferous sequence of chlorite schist derived from massive hornblende andesite and lesser clinopyroxene basalt, lesser amounts of weakly schistose hornblende andesite and clinopyroxene basalt, and sparse marble; (2) a younger sequence of schistose hornblende diorite and lesser gabbro; and (3) a still younger sequence of schistose granitic plutons, mainly biotite granodiorite and quartz monzonite, with lesser biotite-muscovite trondhjemite and quartz diorite. The metavolcanic rocks are generally massive with sparse outcrops of volcanic breccia and pillow andesite. These observations, and the interlayering of the metavolcanic and metasedimentary rocks, suggest a marine origin. Schistose granitic rocks are more prevalent in the southern part of the GRC, whereas the metavolcanic, metasedimentary, and mafic plutonic rocks are more prevalent in the northern part of the GRC (fig.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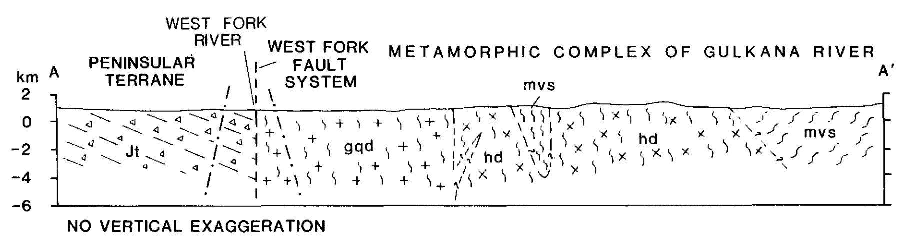
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METAMORPHIC COMPLEX OF GULKANA RIVER
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NO VERTICAL EXAGGERATION

EXPLANATION

WRANGELLIA TERRANE

| Unit | Description | Age |
|---|---|---|
| Qs | Surficial deposits | QUATERNARY |
| | Unconformity | |
| gd | Grandiorite and quartz diorite | CRETACEOUS(?) |
| | Intrusive contact | |
| TRg | Gabbro | LATE TRIASSIC |
| | Intrusive contact | |
| TRn | Nikolai Greenstone | LATE TRIASSIC |
| Pzvs | Volcanic and sedimentary rocks | LATE PALEOZOIC |
| | Fault | |

METAMORPHIC COMPLEX OF GULKANA RIVER

| Unit | Description | Age |
|---|---|---|
| gqd | Schistose granodiorite, quartz monzonite, and quartz diorite | PRE-LATE JURASSIC |
| | Intrusive contact | |
| hd | Schistose hornblende diorite and gabbro | PRE-LATE JURASSIC |
| | Intrusive contact | |
| mvs | Metavolcanic and metasedimentary rocks | PRE-LATE JURASSIC |
| | Fault | |

PENINSULAR TERRANE

| Unit | Description | Age |
|---|---|---|
| gq | Granodiorite and quartz monzonite | LATE JURASSIC(?) |
| | Intrusive contact(?) | |
| Jt | Talkeetna Formation | EARLY JURASSIC |

SYMBOLS

- Contact, dotted where concealed
- Fault, dashed where approximately located, dotted where concealed
- Strike and dip of bedding
  - 60 Inclined
  - Vertical
- Strike and dip of schistosity
  - 70 Inclined
  - Vertical
- Village
- River



**Figure 38.**--Simplified bedrock geologic map of the Gulkana C-3 to C-6 and D-3 to D-6 quadrangles, northern Copper River basin. Refer to text for description of map units and faults. Generalized section (A-A') shows structural style and relations among major units of the Peninsular terrane, Gulkana River complex, and Wrangellia terrane. Surficial geologic units omitted on section AA'. Heavy dash-dot contacts on section A-A' outline margins of possible buried mafic/ultramafic body along West Fork fault system.

38). A sequence of amphibolite, greenstone, and locally schistose granitic rocks, similar to that of the GRC, occurs on strike to the west in the Talkeetna Mountains quadrangle (Csjetey and others, 1978). However, when compared to the amphibole in the amphibolite unit to the west, the abundant hornblende in the GRC is interpreted to be primarily of igneous origin.

The GRC is weakly to strongly deformed and metamorphosed. The dominant major structure is an east-west striking, steeply dipping to vertical homocline in metavolcanic and metasedimentary rocks (fig. 38); the dominant minor structure is a locally intense schistosity or fluxion structure in all three units that strikes generally east-west and dips steeply to vertically, parallel to the regional strike of major units and bounding faults (fig. 38). Local extreme cataclasis occurs along the schistosity, with formation of mylonitic schist and local blastomylonite. Lower greenschist facies minerals, mainly chlorite, actinolite, epidote, albite, and white mica, occur along the schistosity. In more highly metamorphosed and deformed areas, hornblende, clinopyroxene, calcic plagioclase, and biotite in the metaigneous rocks are mostly to completely replaced by schistose metamorphic minerals, whereas in less deformed areas, hornblende, clinopyroxene, calcic plagioclase, and biotite are only partly replaced by weakly schistose metamorphic minerals. Schistose biotite-muscovite granodiorite and quartz diorite from two localities along the Richardson Highway yielded K/Ar biotite ages of 142 and 146 m.y. and white mica ages of 146 and 148 m.y. These K/Ar ages are interpreted as a Late Jurassic age of regional metamorphism and deformation and provide a minimum age for all three major bedrock sequences of the GRC.

The GRC may possibly correlate with the informally designated Haley Creek metamorphic complex (HC) in the northern Valdez quadrangle (Winkler and others, 1981; Plafker and others, 1985). Both have: (1) similar metavolcanic and metasedimentary protoliths, (2) similar suites of older schistose gabbro and diorite plutons and younger schistose granitic plutons, (3) a similar Late Jurassic K/Ar age of metamorphism and deformation, and (4) a similar geologic history, except for later thrusting of the HC. Major differences between the two complexes are more abundant metasedimentary rocks in the HC and locally higher grade, synplutonic metamorphism for the HC (Plafker and others, 1985). The HC also occurs as a klippe on top of the northern part of the Chugach terrane, but it is interpreted to be derived from the Peninsular-Wrangellia (composite) terrane to the north (Plafker and others, 1985).

North of the GRC, in the extreme northeastern part of the study area, the WT consists mainly of sparse exposures of equigranular and porphyritic andesite flows and lesser tuff, volcanic graywacke, argillite, and chert (fig. 38), which are correlated with similar late Paleozoic rocks in the Tangle subterrane of the WT in the Mount Hayes quadrangle to the north (Nokleberg and others, 1982, 1985). Possible east-west-striking folding is suggested by bedding that dips moderately to the north or south (fig. 38). Sparse large plutons of clinopyroxene gabbro also occur (fig. 38). The central and western WT consists mainly of massive, thick subaerial metabasalt flows of the Late Triassic Nikolai Greenstone, which forms the upper part of the Tangle subterrane. Major structures consist of sparse, low-angle, northward dips of crude layering in the Nikolai Greenstone (fig. 38). Sparse isolated plutons of locally schistose quartz diorite and biotite granodiorite of Cretaceous(?) age intrude the Nikolai Greenstone in the northwestern part of the study area (fig. 38). Throughout the WT, static, lower greenschist facies metamorphism is defined by pervasive, granoblastic, fine-grained chlorite, epidote, actinolite, and albite.

The WFFS and PLF, the two major faults in the study area, are everywhere covered by Quaternary surficial deposits, of mainly glacial or glaciolacustrine origin (fig. 38). The WFFS separates the Talkeetna Formation to the south from the GRC to the north, and strikes east-west along the western reaches of the West Fork of the Gulkana River in a broad valley that has a minimum width of 8 km (fig. 38). Geologic evidence for the WFFS consists mainly of the extensive, unmetamorphosed and gently dipping outcrops of the Early Jurassic Talkeetna Formation to the south juxtaposed against locally highly metamorphosed and deformed pre-Late Jurassic metamorphosed volcanic, sedimentary, and plutonic rocks of the GRC to the north. Geophysical evidence for the WFFS consists of the major east-west-striking West Fork magnetic high that occurs along the fault and that extends from the central part of the Gulkana quadrangle to the central part of the Talkeetna Mountains quadrangle (fig. 38) (Andreason and others, 1958; Campbell and Nokleberg, 1986). This magnetic high is interpreted as a near-vertical body of mafic or ultramafic rocks between the PT and GRC (Campbell and Nokleberg, 1986). The PLF separates the GRC to the south from the WT on the north, and it strikes mainly east-west along the northern part of the study area (fig. 38). Geologic evidence for the fault is: (1) the intensely deformed and steeply dipping GRC adjacent to the non-penetratively deformed and gently dipping WT, and (2) the presence of thick, massive hornblende andesites and two-mica granitic plutons in the GRC compared to few, if any, similar rocks in the WT. The PLF is most narrowly constrained on the southwest flank of Paxson Mountain to a colluvium-covered zone about 500 m wide. The PLF is correlated with a similar unnamed major fault to the west in the northeastern Talkeetna Mountains quadrangle that separates the WT to the north from a major unit of amphibolite, greenstone, and locally schistose granitic plutons to the south (Csejtey and others, 1978).

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## NATURE AND TIMING OF DEFORMATION ALONG THE CONTACT FAULT SYSTEM IN THE CORDOVA, BERING GLACIER, AND VALDEZ QUADRANGLES

**George Plafker, Warren J. Nokleberg John S. Lull, Sarah M. Roeske, and Gary R. Winkler**

Recent studies of the Contact fault system (CFS) for the Trans-Alaska Crustal Transect program and prior studies (Plafker and others, 1977; Winkler and Plafker, 1981) show that the CFS is a major early Tertiary suture that extends from the vicinity of Mount Saint Elias to Kodiak Island. Details of the structural fabric along the fault and in the units on either side of it are given by Nokleberg and others (1986); refraction data for the deep crustal structure across the CFS are presented by Fuis and others (1986).

In the Cordova quadrangle, the CFS consists of northward-dipping reverse faults (the Bagley, Gravina, and Landlock segments) that separate foliated flysch and greenstone of the Valdez Group of the Chugach terrane on the north from strongly deformed and weakly metamorphosed flysch and tholeiite of the Orca Group of the Prince William terrane on the south (fig. 39). Major splays of the CFS may bound successive accretionary wedges within the Orca Group. The Martin fault (Tysdal and others, 1976) diverges from the Bagley segment of the CFS in the Bering Glacier quadrangle east of the area shown in figure 39; it juxtaposes Eocene granitoid rocks and sedimentary rocks of the Orca Group on the north against intercalated volcanic and sedimentary rocks of the Orca Group on the south. The Rude River fault, which is largely concealed, bifurcates from the Bagley segment of the CFS near the head of Childs Glacier (loc. A, fig. 39) and extends westward along the Rude River. The fault juxtaposes dissimilar facies assemblages of the Orca Group and abruptly truncates structural trends in the southern block. Other important northeast-trending structures in the Orca Group include the Etches, Cordova, and Eyak faults and the Scott Glacier lineament; all of these structures may also bound accretionary slices.

The Bagley, Gravina, and Landlock segments of the CFS have differing characteristics that may reflect some post-Orca Group regional bending of an initially linear fault (fig. 39). The Bagley segment (Plafker and Lanphere, 1974) is remarkably linear and probably is nearly vertical. Minor structures, mainly slickensides, in rocks adjacent to the fault trace indicate that the youngest sense of movement was locally dextral strike-slip, although the original movement, as indicated by fold orientations, presumably was thrust or oblique thrust. At Miles Glacier (loc. A, fig. 39), the Bagley segment has been intruded by the Bagley pluton, an unfoliated pluton with similar geochemistry and K-Ar ages of 50.6 and 50.9 m.y. on opposite sides of the glacier, indicating that it cannot have undergone horizontal displacement of more than a few hundred meters since the early Eocene (Plafker and Lanphere, 1974).

The eastern part of the Gravina segment is sinuous and has a north dip as shallow as 30° (fig. 40); however, its dip in the western part from the Cordova Glacier to Port Fidalgo (locs. B, C, fig. 39) is very steep to vertical. The combination of high relief and varying dips results in an especially irregular map trace for this segment near Cordova Glacier. North of Sheep Bay, a pluton with K-Ar ages of 50.5 to 53.2 m.y. (Plafker and Lanphere, 1974) is terminated and highly sheared at the Gravina fault, indicating some post-early Eocene movement on this fault.

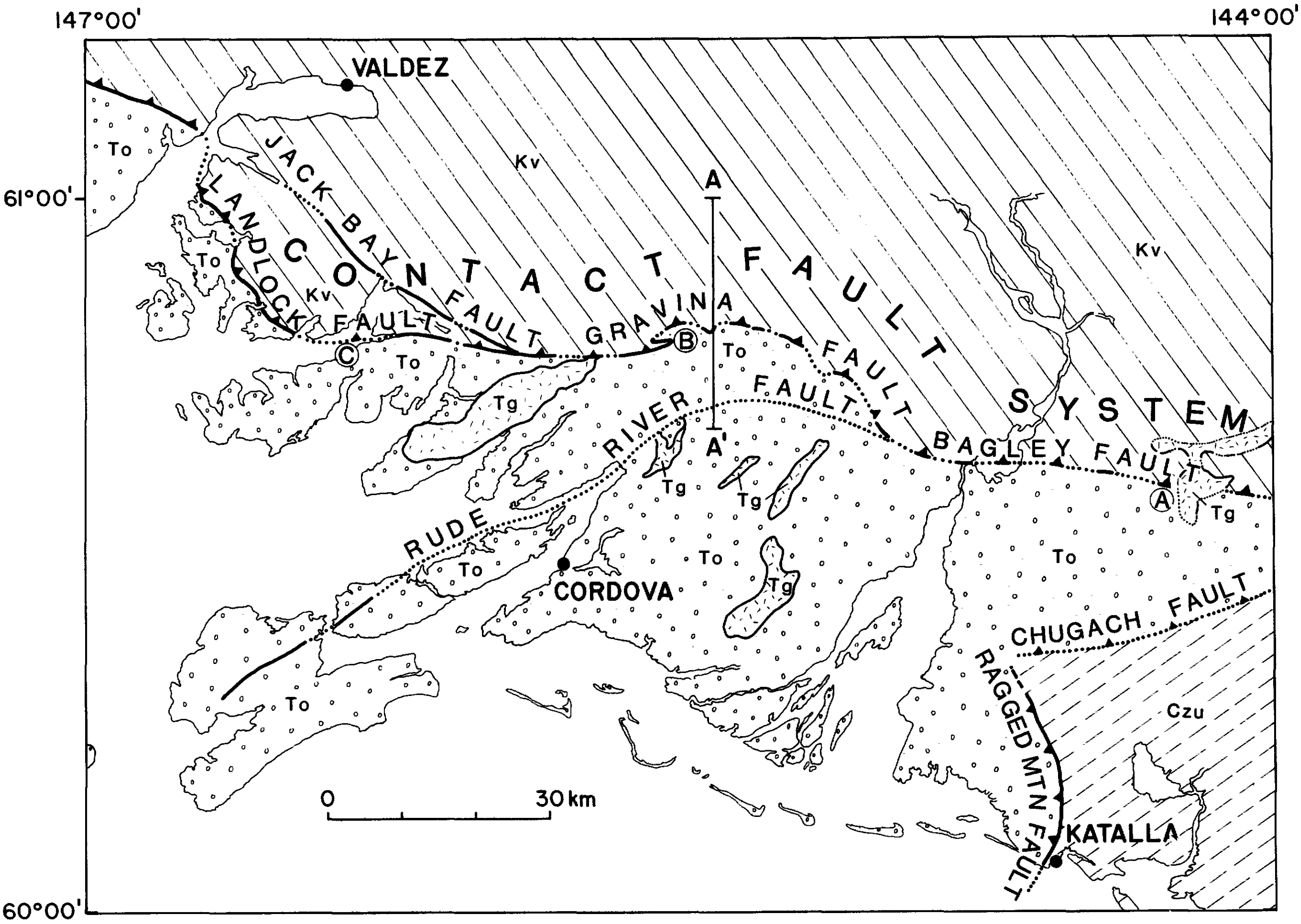


**Figure 39.**—Generalized geologic map showing the Contact fault system between the Chugach and Prince William terranes, and the Chugach and Ragged Mountain faults between the Prince William and Yakutat terranes. The Contact fault system consists of the Bagley, Gravina, and Landlock segments. Jack Bay, Rude River, and Martin faults are major splays of the Contact fault system. Localities on map (circled letters): A, Miles Glacier pluton; B, Cordova Glacier; C, Port Fidalgo. Revised from Winkler and Plafker (1981). Map symbols as on figure 40.

The trace of the Landlock segment (Winkler and Plafker, 1975, 1981) is variable and has an average north dip of 45°. The relatively shallow dip of the Landlock segment is anomalous and apparently has caused this part of the fault system to be cut off within the Valdez Group of the upper plate by the younger, steeply dipping, and linear Jack Bay fault. The Jack Bay fault extends northwestward into the Valdez quadrangle to rejoin the CFS west of Valdez Arm where the trace of the CFS again becomes more nearly linear (Winkler and others, 1981).

Adjacent to the Gravina segment along cross section A-A' (fig. 39), deformation near the CFS is characterized by an accretionary fabric consisting of moderately appressed to tight, asymmetric, major and minor south-verging folds. Locally intense, penetrative schistosity is parallel to axial planes, and lineation is parallel to fold axes. Where they are not deformed around younger open folds, the axial planes, schistosity, and compositional layering have vertical to steep north dips, and fold axes and lineations plunge gently northeast or southwest.

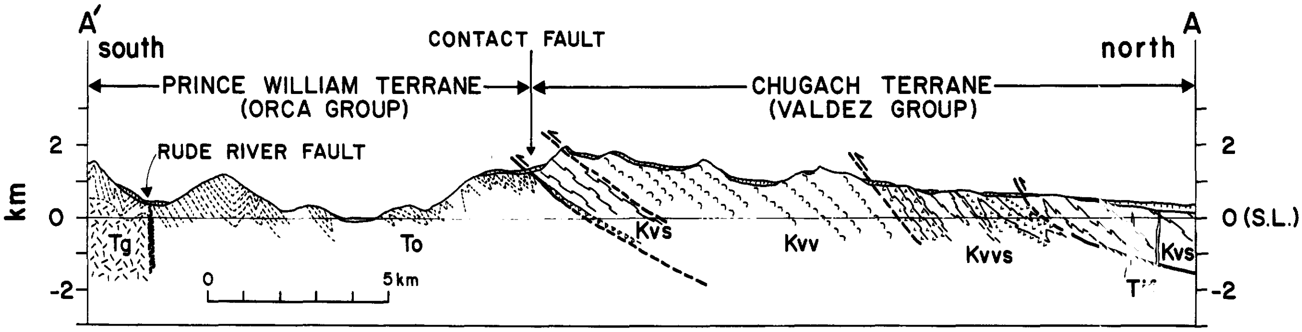


**Figure 40.**--Structure section across the Gravina segment of the Contact fault system along the Copper River Meridian (location shown on fig. 39). Trace of Gravina segment at the surface dips 30° north and is marked by a zone of sheared schistose argillite and greenschist about 40-80 m thick at the base of the upper plate. Tif, Dacitic dikes (Tertiary); Tg, Granodiorite (Tertiary); To, Orca Group (Tertiary) sedimentary and metasedimentary rocks; Kvs, Valdez Group (Late Cretaceous) metasedimentary rocks; Kvv, Valdez Group (Late Cretaceous) metabasalt; Kvvs, Valdez Group (Late Cretaceous) interbedded metasedimentary and metavolcanic rocks, undivided.

Parallellism of subhorizontal major and minor fold axes and lineations, where they are not affected by later folding, indicates oblique thrust convergence during accretion along the CFS near section A-A'. From the transect route eastward to the Copper River, structures in the Orca Group trend obliquely into, and are truncated by, both the CFS and Rude River faults (fig. 39).

The accretionary fabric associated with the CFS occurs in both the Valdez and Orca Groups along the transect. However, in the Valdez Group the accretionary fabric is superposed on an older fabric, and the older schistosity, compositional layering, and isoclinal folds are refolded into the younger south-verging, asymmetric folds. In contrast, in the Orca Group the younger accretionary fabric structures are formed in bedding and represent the first intense deformation of this group.

The older fabric in the Valdez Group along the transect consists of an intensely developed schistosity and lineation, parallel compositional layering, and sparse isoclinal folds. Axes of minor isoclinal folds generally parallel the intense lineation. Schistosity, compositional layering, and isoclinal fold axial planes generally strike east-west and dip moderately north. Lineations and isoclinal fold axes generally plunge horizontally to gently east-west. Lower greenschist facies metamorphic minerals are associated with the older fabric. This older fabric is interpreted as having formed during accretion of the Chugach terrane to the combined Peninsular and Wrangellia terranes along the Border Ranges fault system (Plafker and others, 1985).

Superposed on the older fabric in the Valdez Group, as well as on the accretionary fabric in both the Valdez and Orca Groups, are younger open folds and related structures that are associated with broad regional warping. This warping is best displayed in the arcuate trace of the CFS, particularly along the Gravina and Rude River segments (fig. 39). Planar structures in the Valdez and Orca Groups are similarly warped, especially south of the Rude River fault. The broad warp, concave to the south, defines part of the Alaskan orocline, the axis of which is marked by the transition from east-west to northeast-southwest trends (Carey, 1958). North-northeast strikes and near-vertical dips of axial planes of the open folds indicate mainly east-southeast to west-northwest subhorizontal compression. Limited counter-clockwise rotation of older structures suggests minor right-lateral movement in the zone adjacent to the CFS on the western limb of the bend.

Early Tertiary timing for accretion of the Orca Group and associated deformation along the CFS is suggested by the occurrence of local, sparse, parallel schistosity in some of the gabbro and granitic plutons in the Orca Group with K-Ar ages of 50 to 53 m.y. Most of the granitic plutons, however, are undeformed, suggesting syn- to post-accretion intrusion. During and subsequent to accretion of the Orca Group, the CFS and other major and minor accretionary structures were deformed into the broad oroclinal warp. The lack of comparable folds, metamorphism, and plutonism in middle and late Tertiary rocks of the Yakutat terrane to the southeast suggests that oroclinal warping occured prior to accretion of the Yakutat terrane in Neogene time (Plafker, 1983). Minor post-50-m.y. conjugate strike-slip shears that locally displace the Valdez and Orca Groups and the plutons reflect west- to northwest-directed compression that may be related to emplacement of the Yakutat terrane.

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# MAGNETOTELLURIC STUDY OF A COMPRESSED FLYSCH SYSTEM IN THE HEALY AND ADJACENT QUADRANGLES*

**William D. Stanley**

Two regional geophysical profiles using the magnetotelluric (MT) method were completed in summer 1985 across the central Alaska Range and Talkeetna thrust (Csetjey and others, 1982) in the Healy quadrangle. The MT method allows easy measurement of the electrical resistivity of the earth materials to depths of more than 20 km. The purpose of the survey was to study the vertical and horizontal extent of Mesozoic flysch units, composed mostly of black argillite (Béla Csejtey, Jr., USGS, written commun., 1985) that could serve as deep source rocks. The MT method was used because large electrical resistivity contrasts were expected between the Mesozoic shales that normally have low resistivity and other sedimentary and crystalline rocks that are more resistive.

Fifty soundings were completed along the two profiles (fig. 41). Most sounding sites were within 1 km of the Denali and Parks highways. Initial interpretation of the data reported herein used one-dimensional (horizontally layered), station-to-station Earth models which form geoelectrical cross-sections. The resulting sections can be analyzed for lithology and geologic structures. The survey region is geologically and geoelectrically complex, and interpretation of the data in terms of two- and three-dimensional models remains to be done. However, the preliminary interpretations show striking features related to mapped geology and indicate several unsuspected features.

Two preliminary geoelectrical sections are shown in figure 42 (A-A' and B-B'). The MT data were interpreted using four- and five-layer models; a typical MT sounding curve is shown in figure 43 along with the computer-generated model curve. Connections between individual sounding models in figure 42 were made arbitrarily and were constructed in some places with geological prejudice. For instance, between soundings 21 and 22 the contact between units of 1,500-5,000 ohm-m and 100-500 ohm-m could have any configuration based solely on one-dimensional interpretations; thus, a probable geological portrayal has been used. In future interpretations, additional constraints will be placed upon the overall model with the application of two-dimensional modelling methods.

The interpreted layer resistivities have been broken out into the ranges shown in figure 42. Patterns that suggest preliminary interpreted rock types were used: (1) Crystalline rocks are suggested by the highest resistivity ranges of 500-1,500 and 1,500-5,000 ohm-m, typical values for crystalline rocks observed elsewhere in the Cordillera; (2) shale is suggested by the lowest resistivity range, 1-20 ohm-m, values typical of Mesozoic shales in well logs and surface electrical soundings in other parts of the Cordillera; and (3) two patterns were used for the 100-500 ohm-m range, to distinguish units at the surface from those associated with deep crust, both of unspecified lithology.

It is easier to specify rock types for the highest and lowest resistivity ranges than for the 20-100 and 100-500 ohm-m ranges. For these intermediate ranges, the typical resistivities of several rock types overlap. Sandstones, volcanic rocks, and limestones all typically have resistivities in these ranges, whereas high-grade metamorphic rocks and plutonic rocks generally fall in the higher ranges, and shales and glacially derived Quaternary sediments fall in the lowest range due to their high clay content.

The sections depict several major tectonic and plutonic features:
(1) Soundings 21-36 (from Paxson west on the Denali Highway) are interpreted to be a 13-18 km-thick plate of the Wrangellia terrane (Jones and others, 1977); the plate thins to the northwest. The lower

* This study was funded by the Department of Energy, Morgantown Energy Technology Center, USGS-DOE Interagency Agreement DE-A121-83MC20422. The management of Denali National Park provided access to two measurement sites.

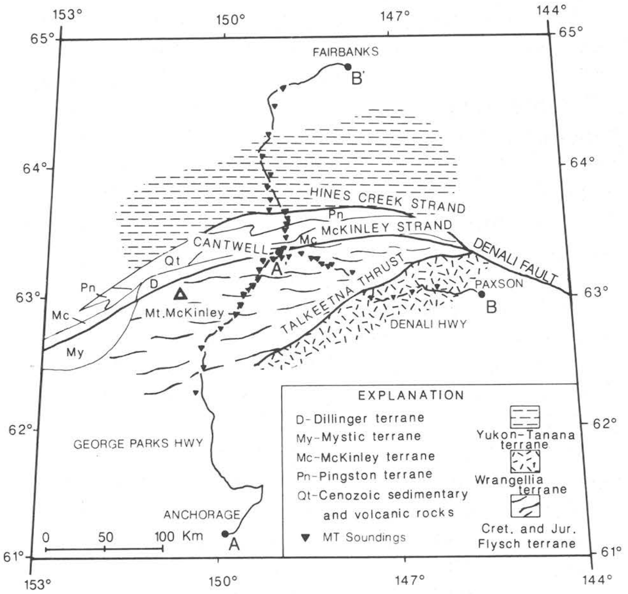


**Figure 41.**--Terrane map of central Alaska Range (modified from Jones and others, 1982) showing locations of magnetotelluric sounding locations (inverted triangles). The compressed flysch system between the Talkeetna thrust and the Denali fault was the focus of study. Several small terranes in the flysch basin that were identified by Jones and others (1982) were omitted for clarity.

surface of this 1,500-5,000-ohm-m plate may represent the base of the oceanic buildup complex (Nokleberg and others, 1985), which forms much of the Wrangellia terrane. The MT interpretation may provide evidence for thrusting of this plate over younger accreted components or, alternatively, the 100-500-ohm-m section beneath the more resistive plate may be lithospheric material associated with Wrangellia. (2) Resistive rocks at MT soundings 34-32, extending westward on the Denali Highway, may be recrystallized Mesozoic flysch rocks, part of the MacLaren metamorphic belt (Nokleberg and others, 1985) as suggested by D.L. Jones (Univ. of Calif., Berkeley, oral commun., 1985). Alternatively, much of this section may be made up of Cretaceous and Tertiary plutonic rocks, which crop out at numerous nearby locations. (3) Resistive rocks at soundings 4-8 southeast of Mt. McKinley on the George Parks Highway probably represent a Tertiary pluton identical to rocks mapped on either side of the profile in this area (Béla Csejtey, USGS, written commun., 1985). The resistive rocks at soundings 9, 7, and 39 are probably smaller separate masses of intrusive rock. (4) Low-resistivity rocks beneath soundings 6-27 south of Cantwell on the George Parks Highway, may be Jurassic and Cretaceous black argillite that forms massive cliffs to the east of the highway near the profile. South of these soundings, between sounding locations 8 and 6, large outcrops of black argillite occur; here, a thrust fault (Honolulu thrust) has been mapped (Béla Csejtey, unpubl. information, 1985) during a study of the Healy quadrangle. This sinuous thrust fault also occurs between soundings 32 and 41. The MT interpretation shows a corresponding section of low-resistivity rocks at this location (fig. 42). (5) Low resistivity rocks were interpreted from soundings 24-37, in a region north of the McKinley strand of the Denali fault. The rocks are within 3-5 km of the surface just north of the fault and extend to a depth of at least 30 km. Lithologic units that make up the low resistivity section may consist, in part, of Paleozoic argillite in the McKinley terrane (Jones and others, 1982). Mesozoic black argillite that occurs in massive exposures south of the Denali fault must also be considered a likely constituent of a major part of the conductive section. This interpretation implies that the Yukon-Tanana terrane is a thin-skinned thrust sheet overlying conductive sedimentary rocks for several tens of kilometers north of the McKinley strand of the Denali fault. (6) Resistive rocks were found north of the Denali fault at soundings 29-30, southwest of Fairbanks. Sounding 29 was made on schist and marble (correlatives of the now-abandoned Birch Creek Schist); the high resistivities are typical of such high-grade metamorphic rocks. Sounding 30 indicated a thinner section of high-resistivity rocks.

Though these results are only preliminary, they indicate that valuable information may be obtained through the use of

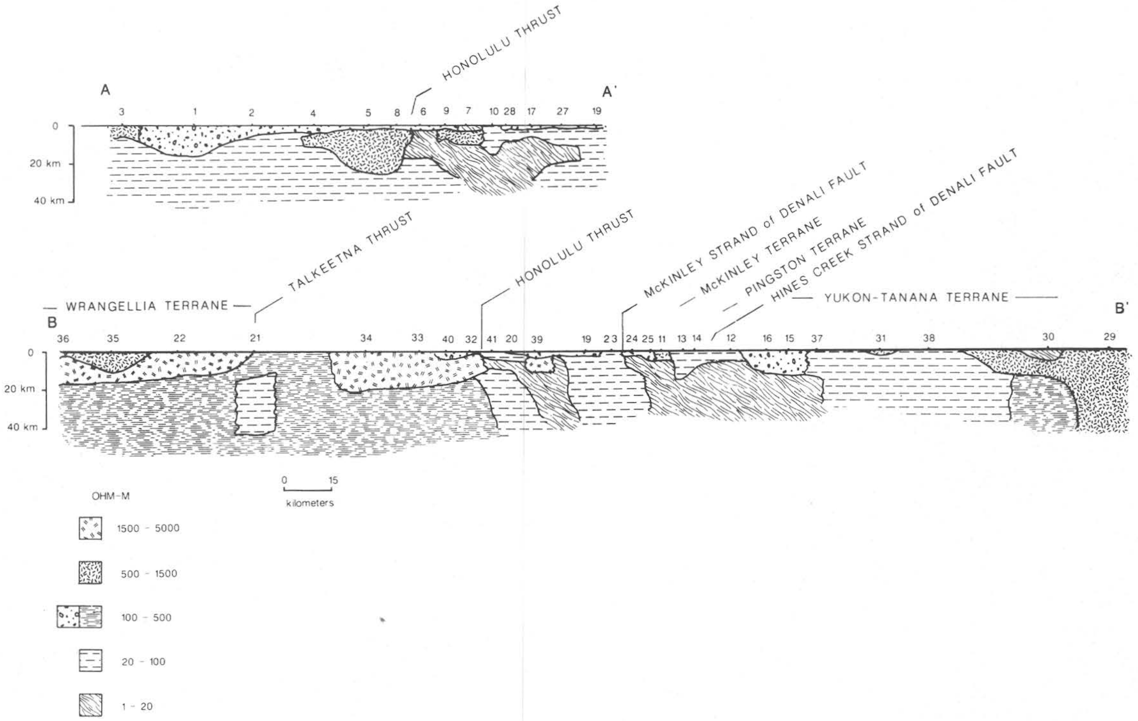


**Figure 42.**--Preliminary geoelectrical sections constructed from one-dimensional models for study area.

magnetotelluric methods in conjunction with geologic mapping in tectonic and mineral resource studies.

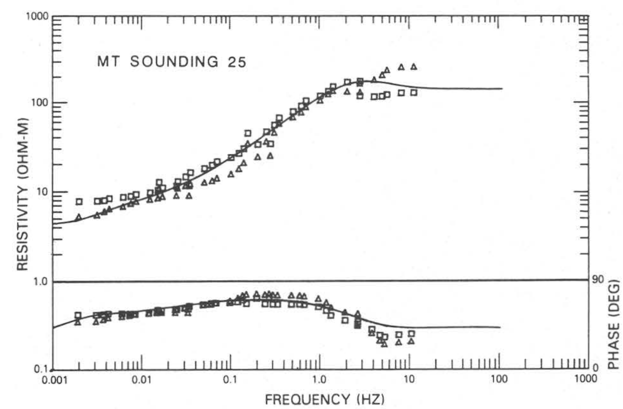


**Figure 43.**--Magnetotelluric sounding curve for sounding 25, showing resistivity in the upper part of the figure and phase in the lower part of the figure. Both parameters are modelled in interpretation of data. Triangles and squares are resistivity and phase measured along strike and across strike, respectively. Solid curve is a computed one-dimensional response for the model depicted in figure 42.

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Reviewers:
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## SEISMICITY IN SOUTHERN ALASKA, OCTOBER 1984-SEPTEMBER 1985

**Christopher D. Stephens, Kent A. Fogleman, John C. Lahr, and Robert A. Page**

Earthquake hypocenters determined using data from the USGS regional seismograph network in southern Alaska (fig. 44) for the period October 1984 through September 1985 are shown in figure 45. Details about the location procedures for the network, including magnitude determinations, completeness levels, and estimates of hypocenter errors are described in published catalogs (for example, Fogleman and others, 1986). Notable shocks or changes in the pattern of seismicity that occurred during this period are discussed below, but generally the distribution of seismicity is similar to that observed over the past few years (for example, Stephens and others, 1985). Principal features in the seismicity include a highly active, northwestward-dipping Wadati-Benioff zone below about 30 km depth along the Aleutian arc, which marks the descent of the subducted Pacific plate into the mantle beneath and north of Cook Inlet; a weakly active northeastward-dipping Wadati-Benioff zone south of the Wrangell volcanoes (Stephens and others, 1984); spatially clustered seismicity occurring at a relatively low rate within the overthrust North American plate; and broad patches of shallow (depths less than about 20 km) seismicity beneath the eastern part of the network, including the aftershock zone of the 1979 Mount Saint Elias earthquake (6.4 $m_b$, 7.1 $M_s$; fig. 44), a shallow thrust event that occurred north and east of Icy Bay (Stephens and others, 1980). Body-wave ($m_b$) and surface-wave ($M_s$) magnitudes are taken from the Preliminary Determination of Epicenters of the USGS National Earthquake Information Service (NEIS).

In September 1985, a magnitude 4.6 $m_b$ shock occurred at a depth of 185 km beneath Iliamna Lake near the downdip extent of the Aleutian Wadati-Benioff zone. A preliminary focal mechanism determined from regionally recorded P-wave polarities for this event has a steeply dipping tension axis and a nearly horizontal compression axis closely aligned with the local, north-northeast strike of the inclined seismic zone. This result is consistent with those of other studies of earthquake focal mechanisms from the southern Cook Inlet Wadati-Benioff zone (for example, Pulpan and Frohlich, 1985) where it has been suggested that north-south compression is an important component of the stress field within the subducted plate in this region.

Seismicity within the North American plate west of about longitude 148.5° W. (fig. 46) is concentrated along the volcanic arc and within distinct clusters, such as those beneath the northern Cook Inlet basin and north of the Castle Mountain fault. The most pronounced concentration of shallow epicenters is due to continuing aftershock activity from the August 1984 Sutton earthquake (fig. 44), which occurred on the Castle Mountain fault (Lahr and others, 1986). In July and August 1985, a tightly clustered swarm of about 40 earthquakes with magnitudes of 2.4 and smaller occurred near the intersection of the Caribou and Castle Mountain faults, about 15 km east of the Sutton aftershock zone. Earthquakes of this magnitude have been located by the network in this area in the past, but the relatively large number of smaller events located during the last year is probably due to systematic changes that include applying a lower magnitude threshold for processing events in this area and improved detection capabilities resulting from the installation of two seismographs near the Sutton

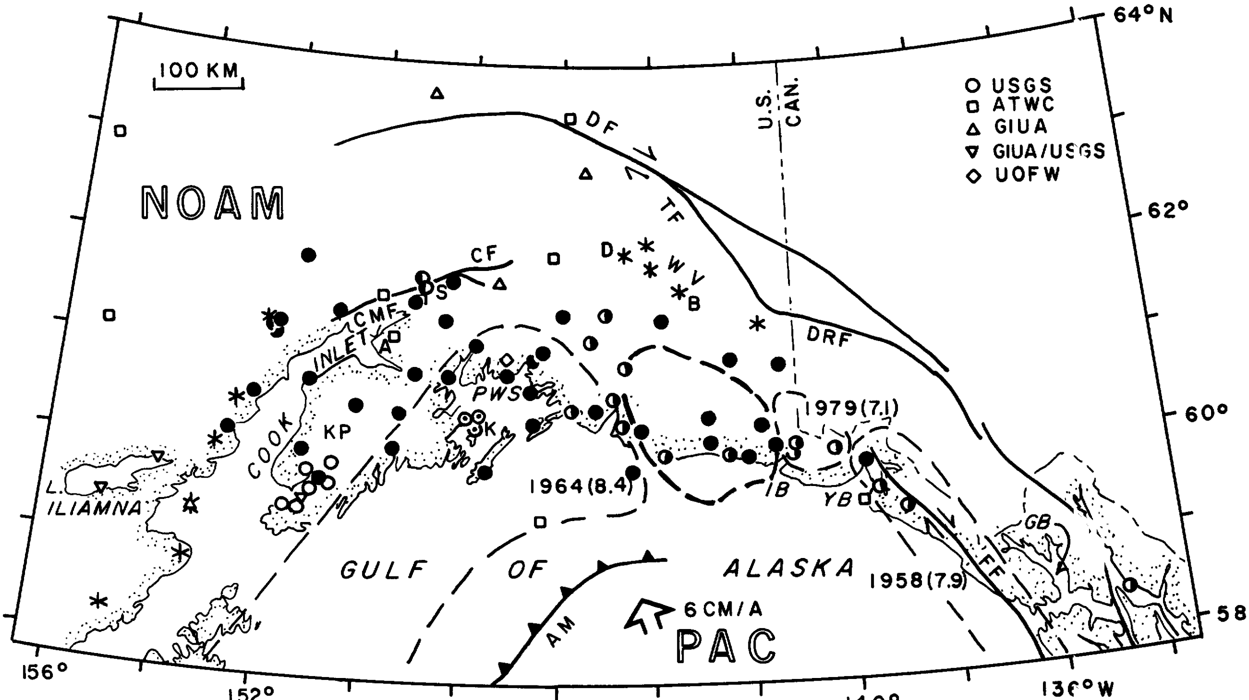


**Figure 44.**--Map of seismograph stations in southern Alaska during 1984-85. Symbol shape corresponds to operating institution as indicated: ATWC, Alaska Tsunami Warning Center; GIUA, Geophysical Institute, University of Alaska; UOFW, University of Washington. Symbol fill for USGS stations (circles) is as follows: solid, operated throughout time period; open, shut down in summer 1984; left fill, installed in fall 1984; right fill, shut down in summer 1985; center dot, installed in summer 1985. Asterisks indicate Quaternary volcanoes. Light dashed contours with corresponding years and magnitudes ($M_S$) in parentheses outline aftershock zones of large earthquakes since 1938 (after Sykes, 1971; Stephens and others, 1980); heavy dashed contour indicates inferred extent of Yakataga seismic gap. Heavy arrow indicates relative plate motion between Pacific (PAC) and North American (NOAM) plates (after Minster and others, 1974). Abbreviations are as follows: A, Anchorage; AM, Aleutian megathrust; B, Mt. Blackburn; CF, Caribou fault; CMF, Castle Mountain fault; D, Mt. Drum; DF, Denali fault; DRF, Duke River fault; FF, Fairweather fault; GB, Glacier Bay; IB, Icy Bay; K, Knight Island; KP, Kenai Peninsula; PWS, Prince William Sound; S, Sutton; TF, Totschunda fault; WV, Wrangell volcanoes; YB, Yakutat Bay.

aftershock zone (fig. 44). A notable lack of shallow activity beneath the southern Kenai Peninsula compared to preceding years is an apparent change due to the removal of six of the nine seismographs that had been operating in this area (fig. 44).

Beneath the eastern Kenai Peninsula, Prince William Sound, and the adjacent continental shelf--the area that ruptured in the 1964 earthquake (fig. 44)--the rate of activity is markedly lower than in the Wadati-Benioff zone to the west. This rate difference can be attributed in part to an increase in the magnitude threshold for event detection by the regional network with distance from the land-based network. However, the distribution of recent teleseismically recorded earthquakes, which constitute a more uniform data set, also indicate a low relative rate of activity in the offshore area. The nature of the events that are located

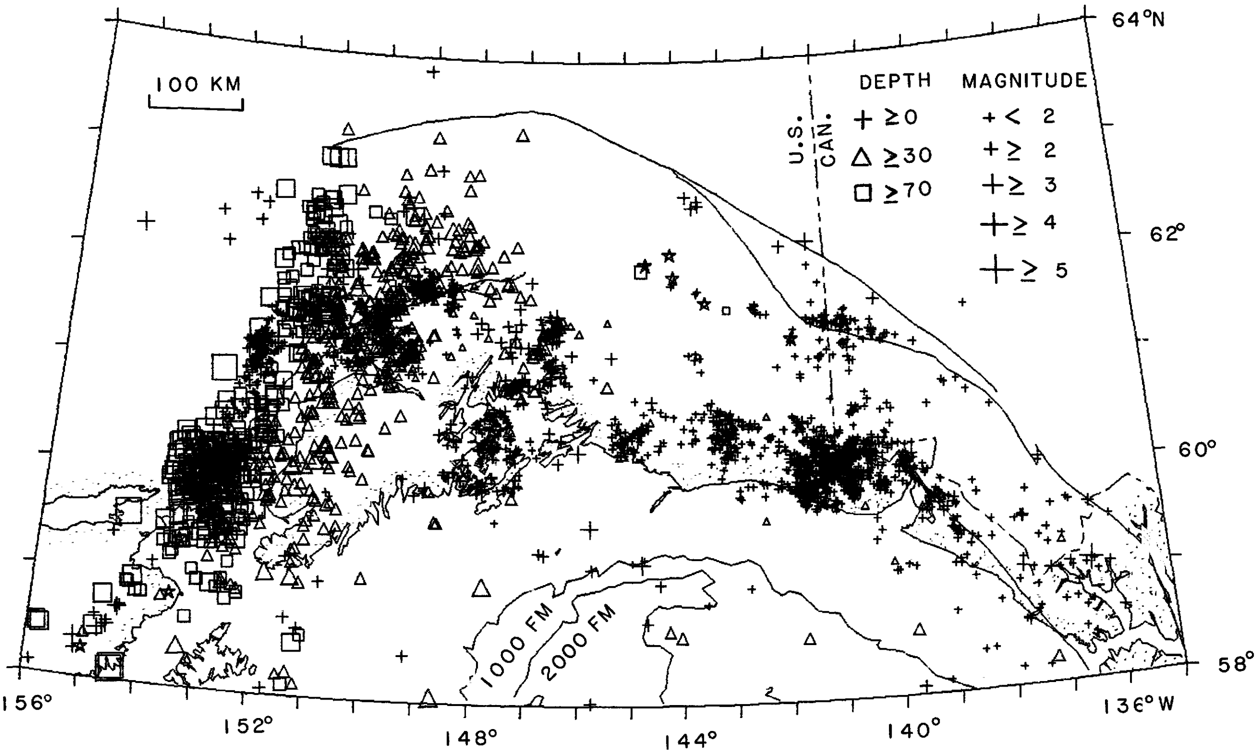


**Figure 45.**--Hypocenters of 3,842 earthquakes from the period October 1984-September 1985 located using data from the USGS regional seismograph network in southern Alaska. The magnitude level for completeness varies across the network due to uneven station spacing and to criteria that are imposed to selectively process additional small events in areas of interest. Stars indicate Quaternary volcanoes. See figure 44 for identification of map features.

offshore is uncertain because their depths are poorly constrained.

Beneath the Prince William Sound area there are several concentrations of seismicity. Page and others (1985) inferred that at least part of this activity is the result of normal faulting within the subducted Pacific plate below about 20 km depth. Other activity, such as the concentration near Valdez, clearly occurs within the upper 20 km of the North American plate. In August 1985, three seismographs were installed in the vicinity of Knight Island in western Prince William Sound to investigate the nature of a persistent cluster of shallow activity occurring beneath the island. Preliminary locations of well-controlled hypocenters in this area indicate that the depths of these events are concentrated between 15 and 27 km. However, additional work is necessary, such as incorporating results from the 1985 Trans-Alaska Crustal Transect (TACT) refraction survey across southern Prince William Sound (Fuis and Ambos, 1986) into the velocity model, to improve the locations of the Knight Island seismicity.

In January 1985, a shallow, magnitude 5.7 $m_b$ (5.1 $M_s$) shock occurred within the Mount Saint Elias aftershock zone near the U.S.-Canada border. This is the largest shock located by the network during this period. As is typical for shocks larger than about magnitude 5 $m_b$ that occur in the central part of the Mount Saint Elias aftershock zone (Stephens and others, 1985), the recent shock was followed by a relatively minor secondary aftershock sequence.

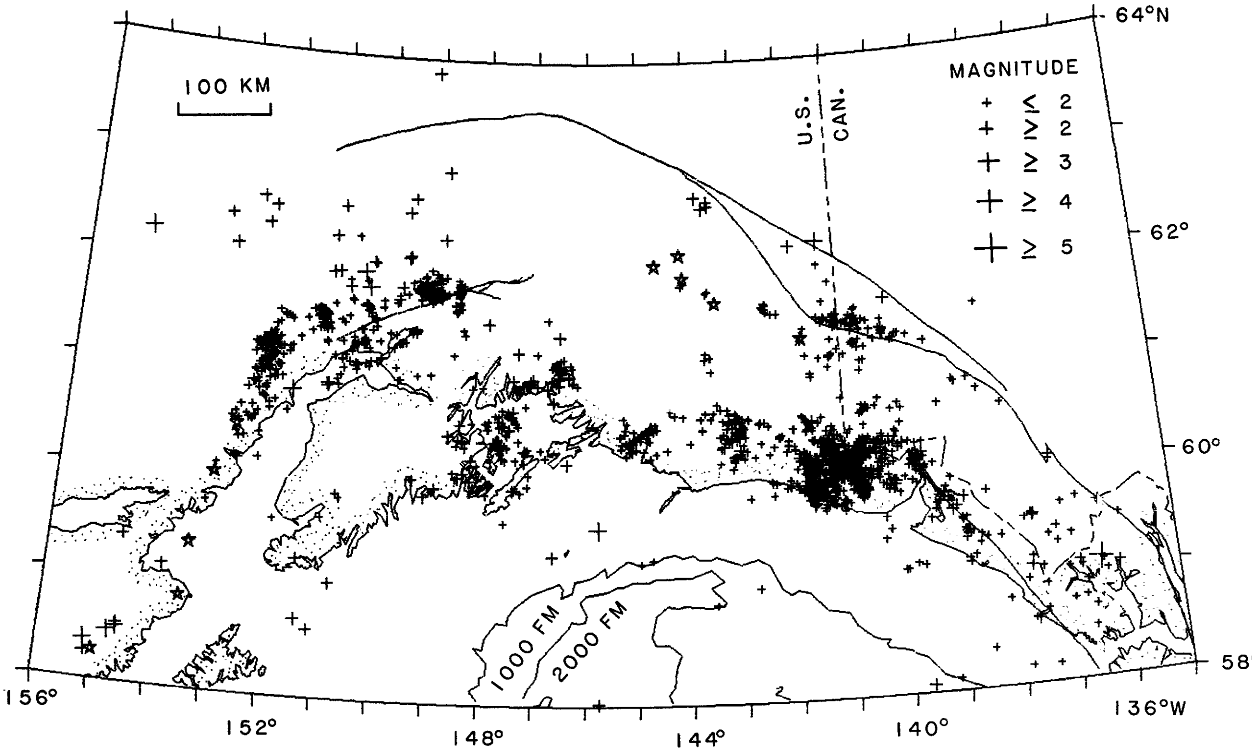


**Figure 46.**--Epicenters of earthquakes shown in figure 45 with depths of 20 km and shallower. Stars indicate Quaternary volcanoes. See figure 44 for identification of map features.

In September 1985, a shallow magnitude 5.3 $m_b$ shock occurred north of Glacier Bay and was felt at distances ranging up to 200 km from the epicenter (NEIS). Because this event was located well outside of the network, only four aftershocks, all smaller than coda-duration magnitude 2.5, were detected within 24 hours of the mainshock. Prior to this shock, the largest event in this area was a magnitude 6.0 shock in 1952. This area is characterized by both a high rate of relative uplift (40 mm/yr; Hicks and Shofnos, 1965) and persistent low-level seismicity, which Horner (1983) suggests are the result of a component of convergence across the Fairweather fault.

Relatively few events define the northeastward-dipping Wrangell Wadati-Benioff zone south of the Wrangell volcanoes (Stephens and others, 1984). However, two of the deepest earthquakes ever located in this zone by the network occurred in April 1985, one at a depth of 86 km beneath Mt. Drum, and the other at a depth of 80 km east of Mt. Blackburn.

The Yakataga seismic gap (fig. 44) is considered by McCann and others (1980) to be a likely site for a great (magnitude 7.8 or larger) earthquake within the next one or two decades. The spatial and temporal distributions of microearthquake activity within and around the gap have remained relatively stable since at least October 1979 (available data from earlier time periods is discontinuous because processing is not yet completed), so this estimate of timing cannot be refined on the basis of patterns of recent seismicity.

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Reviewers:
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## NEW RADIOCARBON DATES FROM THE MATANUSKA GLACIER BOG SECTION

**John R. Williams**

Radiocarbon dates from a newly exposed 6.6-m-thick bog section show that undeformed pond and peat deposits only 1,830 m from the terminus of the Matanuska Glacier have not been covered by ice in the past 13,000 years and that the underlying diamicton is more than 13,000 years old (fig. 47). The new exposure is similar to, but 30 m downvalley from, an exposure examined in 1954 (Williams and Ferrians, 1961).

The bog section is at the top of a 25-30-m-high bluff of banded, till-like diamicton on the north side of the Matanuska River valley. The section exposes 355 cm of black, uncompacted, undeformed peat overlying 305 cm of stratified, organic-rich, locally peaty, clayey to silty, and,

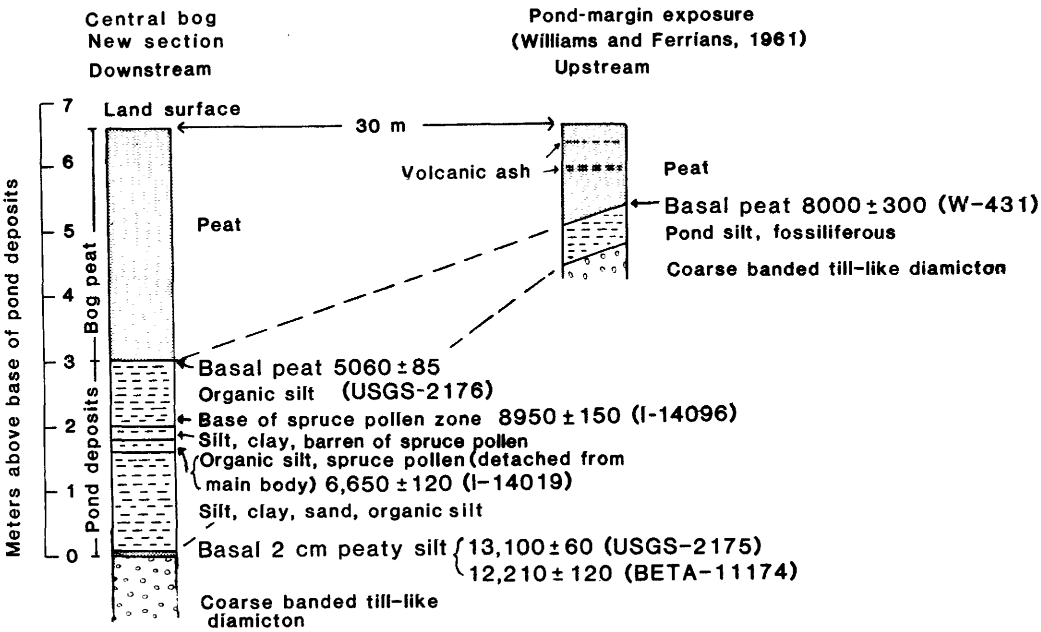


**Figure 47.**--Cross-section of part of Matanuska Glacier bog exposed at top of river bluffs 1.83 km west of glacier terminus and stratigraphic location of radiocarbon-dated samples.

in places, sandy pond deposits that contain shell horizons along former pond margins. The pond deposits overlie the till-like diamicton in the river bluff; the upper part of the latter sediments is winnowed and washed.

Four samples of the peaty and organic-rich pond deposits (fig. 47) were dated by radiocarbon methods. The sample from 200-220 cm, taken to date the zone in which the lower limit of spruce pollen was found (T.A. Ager, USGS, written commun., 1985), yielded a date of 8,950±150 yr B.P. A thin zone also rich in spruce pollen at 160-162 cm above the base of the section is below and detached from the main body of spruce-pollen-rich deposits. An anomalously young radiocarbon date of 6,650±120 yr B.P. indicates that this lower body of pond sediments was emplaced by sliding or by scour and redeposition of those sediments.

Dates from the base of the peat are 8,000±300 yr B.P. in the upstream exposure (fig. 47) and 5,060±85 yr B.P. in the center of the bog. The time span represented by the dates in that stratigraphic horizon may indicate that a change from pond to bog conditions took place 8,000 years ago near the edge of the depression and that open-water conditions persisted until 5,060 years ago in the center of the bog. This might be expected under conditions of declining surface- and ground-water levels as the nearby Matanuska River cut into valley-floor deposits.

The basal peaty pond silt, 0-2 cm above the diamicton, was dated at 12,210±120 yr B.P. and 13,100±60 yr B.P. (fig. 47). These basal dates limit the age of later glacial advances to within less than 1,830 m of the present glacier terminus and provide a minimum age for the drift underlying the old glacial valley floor.

The section is located in a former marginal channel of Wisconsin age that is apparently the innermost of a series of channels that extends 4-8 km downvalley from the glacier terminus. Beyond these channels and intervening moraines, no evidence for glacial halts or readvances exists for nearly 125 km west to the Elmendorf Moraine near Anchorage. Schmoll and others (1972) and Reger and Updike (1983, p. 202) cite radiocarbon dates that seem to indicate that the Elmendorf

Moraine was formed about 11,690-13,690 yr B.P. when the Knik Glacier, expanding 85 km beyond its present terminus, advanced into stagnant glacier ice that filled the Matanuska Valley. The new radiocarbon dates from the bog section near the present terminus of the Matanuska Glacier, together with similar age determinations from other areas in southern Alaska, make this interpretation questionable.

The two oldest radiocarbon dates for undeformed deposits so close to the present terminus of the Matanuska glacier are consistent with those of other recent studies in southern Alaska that document glacier recession from extended late Pleistocene positions by 13,000 yr B.P. and only minor readvances since that time. For example, Denton (1974) noted that the base of muskeg overlying till 3.2 km beyond the terminus of Russell Glacier at the head of White River in the northwest St. Elias Mountains is dated at 11,270±200 yr B.P. (Y-2,306), and Bartsch-Winkler and others (1983, p. 813) showed that extrapolation of dated postglacial sediments in a test hole at the head of Turnagain Arm near Anchorage to those in a nearby 301-m well indicates that estuarine and shallow-water marine sedimentation 8 km downvalley from the Holocene moraine of Portage Glacier has been taking place since deglaciation at least 14,000 years ago. Similar dates on deglaciation are reported from (1) Hidden Lake about 21 km from Skilak Glacier in western Kenai Peninsula, which was deglaciated at least 13,710 years ago, according to Ager (1983); (2) the lower Copper and Katalla Valleys, where Sirkin and Tuthill (1969, 1971, 1984) found the ages of some basal bog sections close to modern glaciers to range from 10,000 to more than 14,000 years; and (3) Muir Inlet of Glacier Bay, where Molnia and others (1984) found that glaciers retreated to their present position before 11,000 years ago. In addition, Miller (1973, p. C-17) reported that Mendenhall Glacier near Juneau has remained within 5 km of its present position during the past 10,000 years and that the elevated postglacial marine deposits along its valley walls were probably laid down 10,000-13,000 years ago when the glacier was upvalley from its present terminus.

The notable exception to this pattern of small advances is the postulated advance of Knik Glacier to a point 85 km beyond its present position sometime between about 11,690 and 13,690 yr B.P., based on the assumption that the dated part of the Bootlegger Cove Formation lies below the Elmendorf Moraine (Schmoll and others, 1972; Reger and Updike, 1983; Schmoll and Yehle, 1986). Additional data are needed to delineate glacier advances in the middle and lower Matanuska Valley and to clarify the stratigraphic relations between the Bootlegger Cove Formation and the Elmendorf Moraine.

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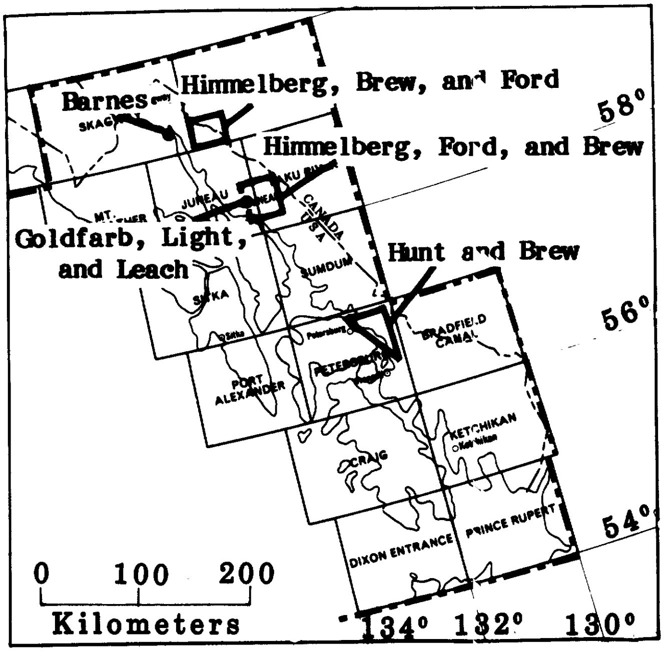


**Figure 48.**--Map showing areas in southeastern Alaska discussed in this circular.

# SOUTHEASTERN

(Figure 48 shows study areas described.)

## GRAVITY DATA INDICATE LARGE MASS AND DEPTH OF THE GABBRO BODY AT HAINES

**David F. Barnes**

Interpretation of the 50-mGal gravity high measured at Haines suggests that the gabbroic intrusion that apparently causes the anomaly is very massive and may extend through a significant part of the earth's 40-km crust. The intrusion is one in a belt of several zoned ultramafic bodies that have been mapped along most of the length of southeastern Alaska (Taylor, 1967; Brew and Morrell, 1983). Reconnaissance gravity measurements indicate that most of these bodies cause small gravity highs ranging from 2- to 25-mGal magnitude; the only other measurement of an ultramafic anomaly greater than 35 mGal in southeastern Alaska was made at Duke

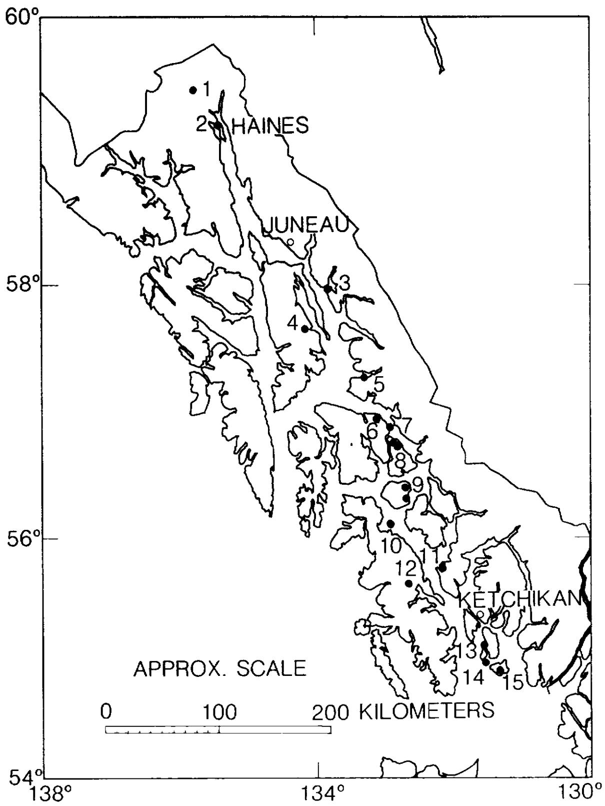


**Figure 49.**--Locations of ultramafic bodies in southeastern Alaska. Numbers reference names and anomaly magnitudes in table 6. Adapted from maps of Taylor (1967), Souther and others (1979), Hutchinson and others (1979), and Brew and Morrell (1983).

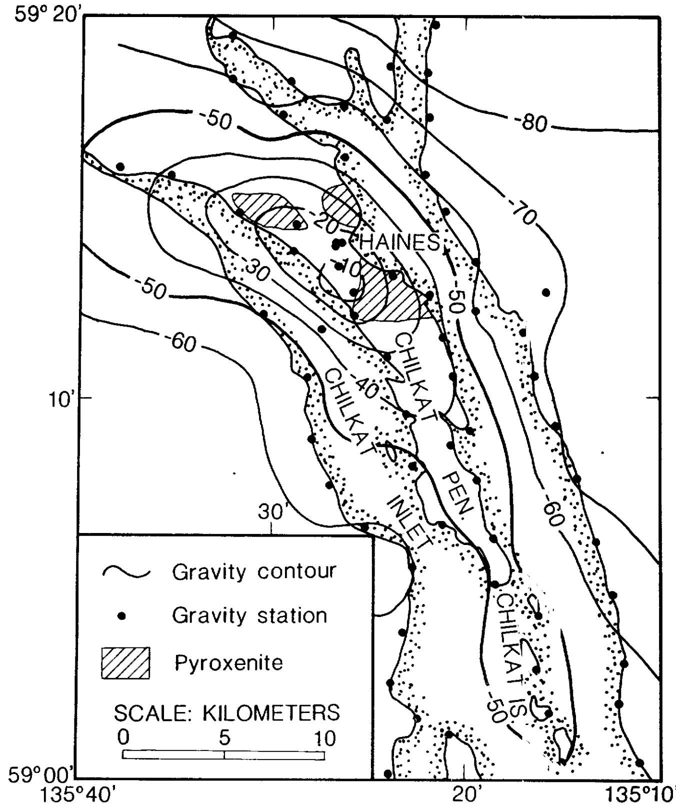


**Figure 50.**--Complete Bouguer gravity contours and pyroxenite outcrops near Haines.

Island, 550 km to the southeast at the other end of the archipelago. Southeastern Alaska gravity data consist primarily of shoreline measurements at intervals of 2-3 km, and many stations are probably closer to the margins than to the centers of ultramafic outcrops. Thus, maximum amplitudes associated with each body may not have been measured. Nevertheless, gravity anomalies at Haines and Duke Island are larger than those at any of the other ultramafic outcrops.

Figure 49 shows the locations of the larger and better known zoned ultramafic bodies in southeastern Alaska. Table 6 lists the amplitudes of measured gravity anomalies at each of these ultramafic bodies and notes locations of the measurements relative to the gabbroic outcrops. Table 6 also shows the relative magnitude of the Haines anomaly and indicates the limited sampling available for many of the other ultramafic bodies. All these ultramafic bodies crop out near the crest and western edge of a steep gravity gradient represented on the Alaska state gravity map (Barnes, 1977) as a group of closely spaced and nearly parallel contours, which are interpreted as representing crustal thickening beneath the high mountains along the Alaska-Canada border. Only the Haines and Duke Island ultramafic bodies cause gravity anomalies large enough to be detected on that State map (Barnes, 1977). Much of the data in table 6 was obtained from more recent detailed surveys

**Table 6.**--Amplitudes of gravity anomalies at zoned ultramafic bodies in southeastern Alaska

| Map Location | Ultramafic body | Gravity Anomaly (mGal) | Comments |
|---|---|---|---|
| 1 | Klukwan | -- | No data |
| 2 | Haines | 50 | Good control |
| 3 | Snettisham | 14 | North edge |
| 4 | Mt Davidson | -- | No data? |
| 5 | Windham Bay | 5 | On shoreline |
| 6 | Kane Peak | 20 | Near center |
| 7 | Sukoi Islands | 13 | On shoreline |
| 8 | Point Frederick | 5 | On shoreline |
| 9 | Deep Bay and Meter Bight | 5 | On shoreline |
| 10 | Blaske Islands | 10 | East edge |
| 11 | Union Bay | *28 | Near center |
| 12 | Salt Chuck | *26 | Near center |
| 13 | Annette Island | 8 | On shoreline |
| 14 | Percy Islands | 13 | East edge? |
| 15 | Duke Island | 35 | On shoreline |

* Unpublished data from J.F. Wynn (USGS, written commun., Dec. 9, 1985)

(Barnes, 1984; and unpublished surveys). Taylor (1967) summarized the petrology of most of these bodies in southeastern Alaska and compared them to similar ultramafic bodies occurring in an outcrop belt in the Ural Mountains, but his discussion did not mention the Haines body, which was shown only as a spot on his map.

Figure 50 compares the Haines gravity contours with the pyroxenite outcrops mapped by Robertson (1956); a gravity map of a broader area around the anomaly was published by Barnes and others (1978). The gravity data clearly suggest that the three outcrops shown are part of a single intrusion, which is partly covered by thin layers of alluvium and metabasalt. The 50-mGal anomaly is approximately elliptical in shape and suggests that the body causing the anomaly has a width or minimum diameter of about 5-6 km. Measurements on a few samples collected during the fieldwork suggest that the pyroxenite has a density of about 3.35 g/cm$^3$ and that the mean density for most bedrock in southeastern Alaska is about 2.75 g/cm$^3$ (Barnes and others, 1978). Thus, the density contrast between the pyroxenite and adjacent rocks is probably close to 0.6 g/cm$^3$. If the 50-mGal anomaly were caused by a slab of unlimited lateral extent and this density, its thickness would be about 2 km. The small elliptical shape of the anomaly suggests, however, that a more localized model is required for an adequate estimate of the body's depth. A two-dimensional analysis of a northeast-trending profile through the center of the anomaly by the method of Talwani and others (1959) suggests that the anomaly amplitude and shape can be approximated by a rectangular body with a width of 5 km, a depth of 7 km, and infinite NW-SE extent. Using the 2 1/2-dimensional geometry of Rasmussen and Pedersen (1979) and limiting the extent of the rectangular body to distances of 5 km northwest and southeast of the profile, the most probable cause of the 50-mGal gravity high might be a body with a density contrast of 0.4 g/cm$^3$, a width of 5 km, a depth of about 10 km, and a length of 10 km. Another approximation can be obtained by using Nettleton's (1976) curves for the gravitational attraction of vertical cylinders. In this analysis, with this density contrast and a radius of 3 km, a thickness of 20 km is needed to cause a 50-mGal anomaly. Calculations using a formula from Donald Plouff (USGS, written commun., Sept. 9, 1971) indicate that the gravity gradient over the edge of such a cylinder would be close to the 12-mGal/km maximum gravity gradient measured between one station on an island near the center of Chilkat Inlet and another on the northeast shore of the Inlet. Despite the variation in these models, most suggest that the pyroxenite body has sufficient depth to penetrate a significant part of the thickness of the earth's crust, which an unreversed seismic refraction line (Hales and Asada, 1966) measured as about 40 km beneath the highway north of Haines. Furthermore, it is unlikely that a 0.6-g/$cm^3$ density contrast could persist through such great depths, and accordingly, the models should perhaps be deeper.

The pyroxenite ultramafic body at Haines is the only one of southeastern Alaska's zoned ultramafic bodies with sufficient gravity coverage to satisfactorily define the size and shape of its anomaly. Most of the other mafic intrusions produce gravity anomalies with lower magnitudes and of smaller areal extent, but with additional data, calculations based on three-dimensional models might show that some or all of the ultramafic intrusions have comparable depths.

The anomaly at Haines extends southeastward over the Chilkat Peninsula and Chilkat Islands, where it has an amplitude of about 10 mGal. The metabasalt that crops out along this trend also has a high density (3 samples exceeded 3.0 g/$cm^3$) and is probably thick enough to explain this southeastward extension of the anomaly. Similar metabasalts have also been mapped in the mountains northwest of Haines, but there are no gravity measurements to show a possible northwestern extension of the anomaly. However, the continuity of these rocks suggests that a small gravity high may extend northwestward as far as another large ultramafic body near Klukwan.

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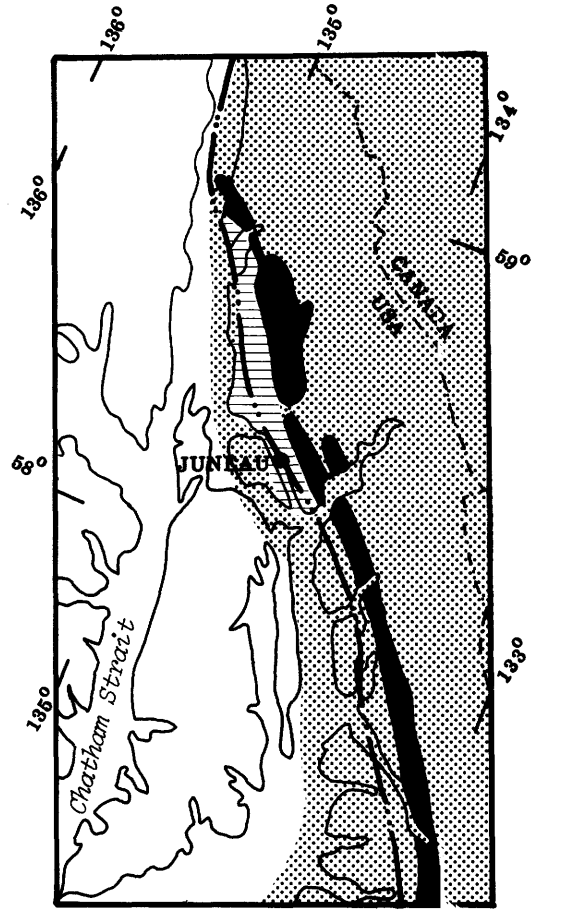
## NATURE OF THE ORE FLUIDS AT THE ALASKA-JUNEAU GOLD DEPOSIT

**Richard J. Goldfarb, Thomas D. Light, and David L. Leach**

The Juneau gold belt extends for almost 200 km along the western flank of the Coast Mountains in southeastern Alaska (fig. 51). The belt lies in the western metamorphic zone of the northern Coast plutonic-metamorphic complex of Brew and Ford (1984); (Brew and Morrell, 1983), which is of latest Cretaceous and Tertiary age. The Alaska-Juneau gold deposit (AJ), located near the center of the gold belt, yielded 3.5 million oz of gold (Berg, 1984), ranking it as the largest gold deposit in the State. The AJ consists of a dense network of vertically oriented, lens-shaped auriferous quartz veins that range in width from a few centimeters to as much as one meter. Quartz is the dominant gangue mineral, and ankerite is especially common along the vein margins; pyrrhotite, galena, sphalerite, and arsenopyrite are the most abundant sulfide minerals. The four main ore bodies at the AJ are hosted by slate and quartzite (fig. 52), which are metamorphosed to amphibolite facies grade (Ford and Brew, 1973). An obvious spatial relation exists between the auriferous vein systems and small diorite sills hosted in the slate. Ore shoots in the slate are preferentially oriented parallel to the long dimension of these igneous bodies (Twenhofel, 1952) and crosscut compositional layering and cleavage. Intense hydrothermal alteration of




**EXPLANATION**

Northern Coast plutonic-metamorphic complex of Brew and Ford (1984)

Coast plutonic complex sill

Juneau gold belt

Coast Range megalineament

0 25 50
kilometers

**Figure 51.**--Location of the Juneau gold belt (after Brew and Ford, 1984).

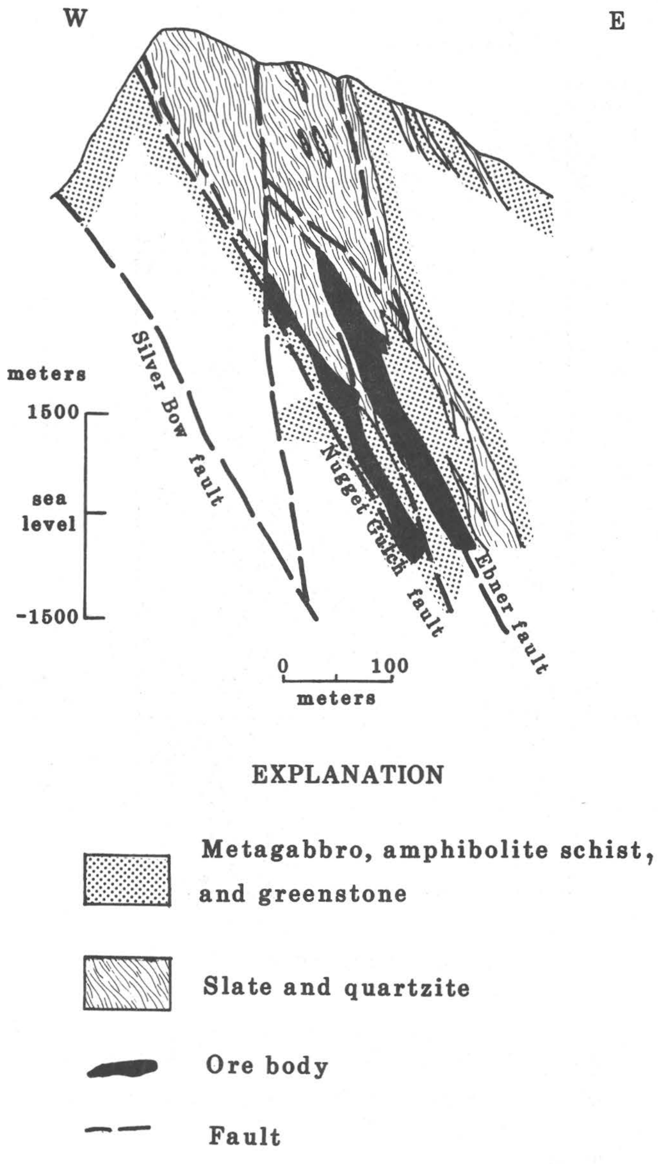



**Figure 52.**--Cross-section showing simplified geology of the Alaska-Juneau ore body (after Wernecke, 1932).

the host rocks has produced abundant biotite, ankerite, and albite adjacent to the quartz veins.

A reconnaissance fluid-inclusion study was undertaken to determine the nature of the ore fluids and the environment of ore deposition at the AJ deposit. Fluid-inclusion data can provide valuable P-T-X information, which can be used to place constraints on conditions of ore deposition. Fluid inclusions in quartz samples from the auriferous veins are as large as 50 microns in maximum dimension. Inclusions occur singly, in isolated three-dimensional groups, in random three-dimensional distributions in individual quartz crystals, or as planar arrays extending in all directions from the ore minerals. Two types of fluid inclusions in the quartz crystals have been identified through microthermometry methods.

Type 1 fluid inclusions appear as 2-phase liquid $CO_2$ + liquid $H_2O$ at room temperature and can be characterized by the $H_2O$-$CO_2$-$N_2$-NaCl system. Generally, a $CO_2$ vapor phase forms upon cooling. The fluids range in composition from $H_2O$-dominant inclusions containing a few mole percent $CO_2$ to inclusions that are $CO_2$ dominant with no visible $H_2O$. Gas hydrate melting occurs between +7.8° C and +10.2° C. Melting of $CO_2$ ranges from the $CO_2$ triple point (-56.6° C) to -61.8° C, indicative of relatively pure $CO_2$ to $CO_2$ containing as much as 50 mole percent $N_2$.

Type 1 inclusions commonly occur as sets of inclusions in groups or planes with relatively constant gas/liquid ratios. However, there are wide ranges in gas/liquid ratios between sets of type 1 inclusions, demonstrating that the $CO_2$ content of the ore-forming fluids varied significantly. In some samples, sets of inclusions show highly variable gas/liquid ratios, which may represent recrystallization of the inclusion walls (necking down); however, some sets of inclusions with variable ratios clearly result from heterogeneous trapping of immiscible fluids. Homogenization temperatures for $CO_2$ vapor to $CO_2$ liquid range from -13° C to +17° C, corresponding to densities of 0.98 to 0.85 g/cm$^3$, respectively. Homogenization of the $CO_2$ fluid with the $H_2O$ liquid occurs at temperatures between 230 and 280° C.

Type 2 fluid inclusions are water-dominant with no detectable gases. Ice-melting temperatures for 2-phase inclusions range from -1.2° C to -6.5° C, corresponding to 2-10 weight percent NaCl equivalent salinity. Type 2 inclusions homogenize in the range of 145° C to

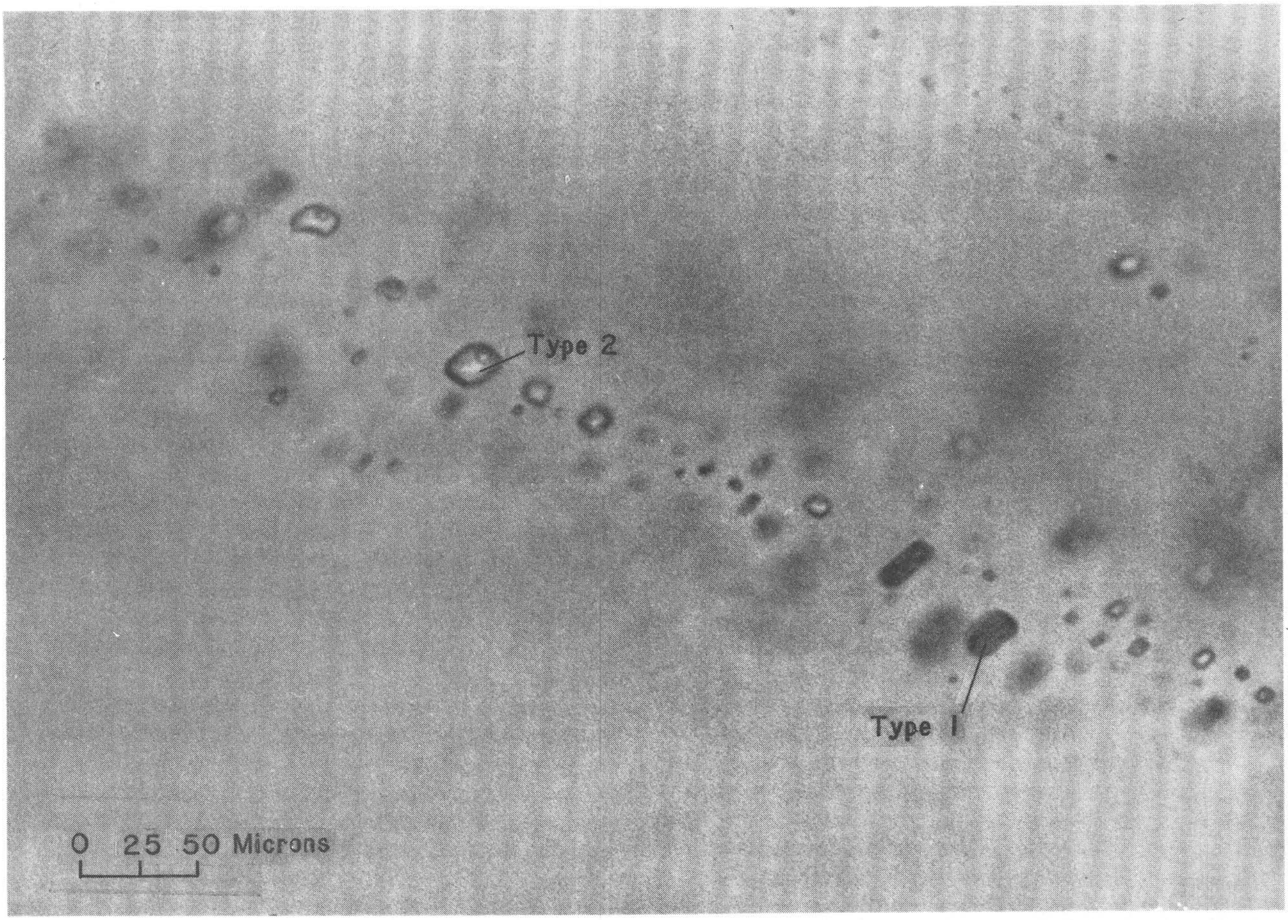



**Figure 53.**--Photomicrograph of typical planar array of type 1 and type 2 fluid inclusions from gold-bearing quartz.

190° C. Several groups of one-phase inclusions froze between -35° C and -40° C, but ice melting was not observed; these may represent a later entrapment of a second generation of aqueous fluids.

In many samples, type 1 and type 2 inclusions were interpreted to be coevally trapped, suggesting that they represent immiscible fluids derived from a boiling system (fig. 53). Both pressure and temperature of fluid trapping can be estimated from the intersection of isochores in P-T space for the pure $CO_2$ and pure $H_2O$ fluids. Isochores determined from measured homogenization temperatures of selected pairs of $H_2O$- and $CO_2$-dominant inclusions, intersect at about 230° C and 1.5 Kb. This P-T condition is a minimum estimate, since both $N_2$ contamination of the $CO_2$ and the presence of NaCl in the $H_2O$-dominant inclusions change the slope of the isochores and shift the intersection to higher P-T space.

A second minimum pressure estimate results from use of the quartz decrepitation data of Leroy (1979). Just prior to homogenization, inclusions with diameters of 15 microns decrepitated at temperatures between 250° C and 255° C, indicative of maximum internal pressures or minimum trapping pressures of about 1.2 Kb. This agrees closely with the value determined from the crossing isochore method.

The estimate of a minimum trapping pressure of 1.5 Kb corresponds to a minimum depth of mineralization of 5.5 km, where fluid pressure equals lithostatic pressure. If fluid pressure contained a significant component of hydrostatic load, depth of mineralization would be greater. Restriction of the AJ ore bodies to slate suggests that locally the hydrothermal fluids were channeled through shear cleavage in the slate units. Boiling of hydrothermal fluids occurred where fluid pressure exceeded confining pressure, possibly during uplift of the northern Coast plutonic-metamorphic complex.

In metamorphic belts where the geothermal gradient exceeds 12° C/km, uplift results in thermal expansion of pore waters and waters in microfractures, eventually resulting in hydraulic fracturing (Norris and Henley, 1976). Fluid-inclusion data from this study are consistent with a geothermal gradient of about 40° C/km. Our findings support a model of deep-seated hydrothermal fluids that were focused along structural zones in the western part of the northern Coast plutonic-metamorphic complex. Ore deposition took place at temperatures greater than 230° C, pressure exceeding 1.5 Kb, and a depth of at least 5.5 km, and was accompanied by intense alteration and hydrofracturing of the host rocks. Loss of volatiles during boiling produced dramatic shifts in $f_{O_2}$, $f_{S_2}$, and pH and resulted in deposition of the ores.

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## CHEMICAL COMPOSITION OF OLIVINE AND ORTHOPYROXENE IN PERIDOTITE OF THE COAST PLUTONIC-METAMORPHIC COMPLEX NEAR SKAGWAY

**Glen R. Himmelberg,**
**David A. Brew,**
**and Arthur B. Ford**

Several peridotite bodies occur discontinuously along a northwest-trending belt approximately 30 km long and 8 km wide within the informally designated Coast plutonic-metamorphic complex of Brew and Ford (1984) in the Atlin A-8 quadrangle (Himmelberg and others, 1985). The peridotite was emplaced into hornblende-plagioclase gneiss (and amphibolite). It is intruded by hornblende-biotite tonalite in which the peridotite bodies occur widely as rafts ranging in maximum dimension from a few meters to several kilometers. The protolith peridotite is mostly harzburgite (olivine, 70-85 percent; orthopyroxene, 15-30 percent; chromite, 1-3 percent) with minor dunite and poikilitic wehrlite. The peridotite has been partly to completely recrystallized by the hornblende-biotite tonalite; the degree of recrystallization is related to the size of the peridotite raft enclosed in the tonalite and, for the larger peridotite bodies, proximity to the contact with tonalite. Recrystallized mineral associations include:

(1) forsterite-talc-tremolite-chlorite-magnetite
(2) forsterite-tremolite-enstatite-chlorite-magnetite±talc
(3) forsterite-tremolite-enstatite-chlorite-green spinel-magnetite± plagioclase

These mineral associations are indicative of the hornblende hornfels facies (Evans, 1977). Using the P-T-$H_2O$ phase relations given by Evans (1977), temperatures of approximately 500° C and 600° C are indicated for mineral associations (1) and (2), respectively, for an arbitrary solid and fluid pressure of 1 kbar. For higher pressures, the temperatures would be slightly higher.

Chemical compositions recalculated to fit the structural formulas of olivine and orthopyroxene in the peridotite are given in tables 7 and 8. Samples 84GH48H, 84GH37A, and 84GH37C are completely recrystallized to a granoblastic texture. Sample 84GH48H is composed of metamorphic mineral association (1) and has olivine of composition $FO_{90.6}$ [100 Mg/(Mg+Fe+Mn)]. Samples 84GH37A and 84GH37C are composed of metamorphic mineral association (2) and the olivines are of compositions $FO_{95.6}$ and $FO_{96.2}$; the coexisting orthopyroxene has compositions of $EN_{94.8}$ [100 Mg/(Mg+-Fe+Mn)] and $EN_{95.1}$, respectively. Olivine compositions generally show no systematic change with metamorphic grade, but are a function of the Mg/Fe ratio of the rock, the oxygen fugacity, and the nature and numbers of coexisting phases (Evans, 1977). The difference in FO values in the recrystallized harzburgite is probably largely related to magnetite abundance. Sample 84GH48H, which has olivine with the lower FO value, has less magnetite than the other two samples. In addition, olivine in the sample coexists with abundant talc, which has the highest Mg/Fe ratio of all the minerals in ultrabasic rocks (Evans, 1977). Magnetite is rare in primary harzburgite that has not been recrystallized. Varying magnetite abundances could reflect either different oxygen fugacities during recrystallization or different degrees of serpentinization (which produced magnetite) prior to metamorphic recrystallization.

Metamorphic orthopyroxenes are very low in $Al_2O_3$ and CaO, which is characteristic of orthopyroxene in the hornblende-hornfels facies (Evans, 1977). According to the experimental data of Gasparik (1984), the $Al_2O_3$ content of the orthopyroxenes indicates a temperature of equilibration of about 400° C or less.

Samples 84GH33A and 84GH35D are harzburgite and dunite that have primary anhedral granular textures and gently curved grain boundary segments. The harzburgite is about 60 percent serpentinized. Olivine is locally recrystallized to granoblastic polygonal texture with 120° grain-boundary intersections, and minor metamorphic development of tremolite and talc postdates the serpentinization. The dunite contains minor serpentine veins that make up 2 percent or less of the total volume, with no evidence of recrystallization. Olivine in the harzburgite and dunite has compositions of $FO_{90.7}$ and $FO_{90.1}$, respectively. Orthopyroxene in the harzburgite has a composition of $EN_{90.7}$. $Al_2O_3$ and CaO content of the orthopyroxene are very low, as in the recrystallized orthopyroxene, suggesting that there has been some metamorphic re-equilibration. However, because the Mg/Fe ratio of olivine and orthopyroxene are not necessarily affected by metamorphism and the analyzed grains show no textural evidence of recrystallization, we interpret the FO and EN values for the olivine and orthopyroxene as approximating primary compositions.

Our current understanding of the nature of the peridotite bodies is summarized as follows: (1) original peridotite was dominantly harzburgite with a restricted olivine/orthopyroxene ratio of about 75/25 and with accessory chromite; (2) original texture was medium to coarse grained anhedral granular; (3) presumed primary olivine and orthopyroxene compositions were approximately $FO_{90}$ and $EN_{91}$, respectively; (4) the contact with the

Table 7.-- Chemical compositions and structural formulas of olivine in peridotite in the Atlin A-8 quadrangle, southeastern Alaska (-, not detected; n.d., not determined)

| Constituents | 84GH 48H | 84GH 37A | 84GH 37C | 84GH 33A | 84GH 35D |
|---|---|---|---|---|---|
| Composition, in weight percent | | | | | |
| $SiO_2$ | 41.0 | 41.0 | 41.5 | 40.9 | 40.5 |
| FeO* | 8.85 | 4.27 | 3.66 | 8.93 | 9.55 |
| MnO | 0.20 | 0.11 | 0.10 | 0.14 | 0.16 |
| MgO | 49.0 | 53.3 | 52.8 | 49.8 | 49.4 |
| CaO | - | .02 | .06 | - | - |
| NiO | .13 | n.d. | .14 | n.d. | n.d. |
| $Cr_2O_3$ | .08 | n.d. | - | n.d. | n.d. |
| Sum | 99.23 | 99.08 | 98.22 | 99.83 | 99.61 |
| Formula per 3 cations | | | | | |
| Si | 1.009 | .997 | 1.009 | .999 | .993 |
| Fe | .183 | .086 | .074 | .183 | .196 |
| Mn | .004 | .002 | .002 | .003 | .003 |
| Mg | 1.801 | 1.915 | 1.911 | 1.814 | 1.808 |
| Ca | - | - | - | .001 | - |
| Ni | .003 | - | .003 | - | - |
| Cr | - | - | - | - | - |
| 100 Mg/(Mg+Fe+Mn) | 90.6 | 95.6 | 96.2 | 90.7 | 90.1 |

*Total iron calculated as FeO

hornblende-plagioclase gneiss, where observed, is sharp, with no evidence of thermal metamorphism and no apophyses of peridotite into the gneiss (probably a fault contact); and (5) the peridotite has been metamorphosed to hornblende-hornfels facies by intrusion of hornblende-biotite tonalite.

The premetamorphic characteristics of the peridotite suggest to us that it is an alpine-type harzburgite of refractory mantle material that was tectonically emplaced into the crust (Coleman, 1977; Dick, 1977; Himmelberg and Loney, 1980). Alpine-type harzburgites are commonly part of ophiolite suites that include gabbro and mafic volcanic rock (Coleman, 1977). Although such mafic rock types have not been recognized to be associated with the harzburgite of the Atlin A-8 quadrangle, the hornblende-plagioclase gneiss and amphibolite in the vicinity may be their metamorphosed equivalents. The absence may also be attributed to fault emplacement of the harzburgite.

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**Table 8.**--Chemical compositions and structural formula of orthopyroxene in peridotite in the Atlin A-8 quadrangle, southeastern Alaska (-, not detected; n.d., not determined)

| Constituents | 84GH 37A | 84GH 37C | 84GH 33A |
|---|---|---|---|
| | Composition, in weight percent | | |
| $SiO_2$ | 57.0 | 57.8 | 56.5 |
| $TiO_2$ | - | 0.04 | - |
| $Al_2O_3$ | 0.70 | .63 | 0.21 |
| $Cr_2O_3$ | n.d. | 3.22 | n.d. |
| FeO* | 3.55 | .13 | 6.33 |
| MnO | .13 | 36.8 | .17 |
| MgO | 37.6 | .06 | 35.4 |
| CaO | .06 | - | .09 |
| $Na_2O$ | - | - | - |
| Sum | 99.09 | 98.68 | 98.70 |
| | Formula per 4 cations | | |
| Si | 1.948 | 1.989 | 1.966 |
| $Al^{IV}$ | .028 | .011 | .009 |
| $Al^{VI}$ | - | .014 | - |
| Ti | - | .001 | - |
| Cr | - | - | - |
| Fe | .101 | .093 | .184 |
| Mn | .004 | .004 | .005 |
| Mg | 1.916 | 1.886 | 1.833 |
| Ca | .002 | .002 | .003 |
| Na | - | - | - |
| 100 Mg/ (Mg+Fe+Mn) | 94.8 | 95.1 | 90.7 |
| Ca | .1 | .1 | .1 |
| Mg | 94.9 | 95.2 | 90.7 |
| Fe | 5.0 | 4.7 | 9.1 |

*Total iron calculated as FeO


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Reviewers:
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# THE OCCURRENCE AND CHEMICAL COMPOSITION OF CHLORITOID IN THE METAMORPHIC ROCKS OF THE COAST PLUTONIC-METAMORPHIC COMPLEX NEAR JUNEAU

**Glen R. Himmelberg, Arthur B. Ford, and David A. Brew**

A regional metamorphic belt containing mineral assemblages that reflect conditions of metamorphism ranging from prehnite-pumpellyite to upper amphibolite facies is exposed along the western margin of the informally designated Coast plutonic-metamorphic complex of Brew and Ford (1984) in southeastern Alaska. The metamorphic belt consists dominantly of intermixed pelitic and semipelitic metasedimentary rocks and mafic metavolcanic and intrusive rocks. Impure calcareous metasedimentary rocks, quartzite, and quartz dioritic and granodioritic orthogneiss are also present.

The first appearances of the Barrovian index minerals biotite, garnet, staurolite, kyanite, and sillimanite were documented by Forbes (1959) on Blackerby Ridge near Juneau (fig. 54). These index minerals are common in the pelitic rocks of the Coast plutonic-metamorphic complex and were later mapped by Ford and Brew (1973, 1977) and Brew and Ford (1977) from Taku Inlet to about 10 km northwest of Mendenhall Glacier (fig. 54). Chloritoid has not previously been recorded as occurring in this area.

Chloritoid, a relatively uncommon constituent in pelitic rocks of low to medium metamorphic grade, has been found in the Juneau area in only three samples from two closely spaced localities near the Mendenhall Glacier Visitors Center (fig. 54). The chloritoid occurs in two mineral associations:

1. Quartz-white mica-chlorite-plagioclase-chloritoid-opaque oxide (sample 68Bd145A)
2. White mica-chlorite-plagioclase-chloritoid-biotite-amphibole-opaque oxide±quartz±carbonate minerals (samples 68Bd159A, B)

In the first association, chloritoid generally occurs as lath-shaped idioblastic grains that commonly show simple or lamellar twinning. In the second association, it occurs as bow-tie structures or as bundles of radiating to spherulitic twinned crystals. The pleochroism of the chloritoid in all the samples is bluish gray to straw yellow. Most grains of chloritoid contain abundant minute inclusions of an opaque material.

The chemical compositions and structural formulas of chloritoid in the three samples from near Juneau are given in table 9. The formula of chloritoid is $(FeMg)Al_2SiO_5(OH)_2$. The amount of $H_2O$ in the chloritoid was calculated according to the method of Zen (1981) by assuming a fully hydrated stoichiometric formula. The chloritoid in the three samples, although from different mineral associations, has a uniform composition, with the cation ratio Fe/(Fe+Mg+Mn) ranging only from 0.85 to 0.88.

La Tour and others (1980) and Deer and others (1982) have reviewed chloritoid occurrences and stability. Chloritoid generally occurs in the chlorite zone of the greenschist facies but in some places is stable to the lower amphibolite facies. Ganguly (1968, 1969) and Ganguly and Newton (1968) showed that chloritoid commonly gives way to staurolite-bearing assemblages at higher grades. The absence of chloritoid in more than 75 percent of pelitic rocks of appropriate grade implies a strong dependency of chloritoid occurrence on bulk composition of the rock (Winkler, 1979). Hoscheck (1967) compared bulk compositions of chloritoid-bearing and chloritoid-free pelitic rocks and showed that chloritoid is favored in rocks with bulk compositions that have relatively high $Al_2O_3$ contents, high FeO/(FeO+MgO) ratios, and low abundances of CaO and alkalis.

Chloritoid in the Juneau area occurs in the biotite zone of the greenschist facies just below the garnet isograd. The local occurrence of chloritoid in the greenschist facies and the absence of chloritoid in the higher grade rocks prior to the appearance of staurolite suggests that the rare occurrences of chloritoid in the Coast plutonic-metamorphic complex in

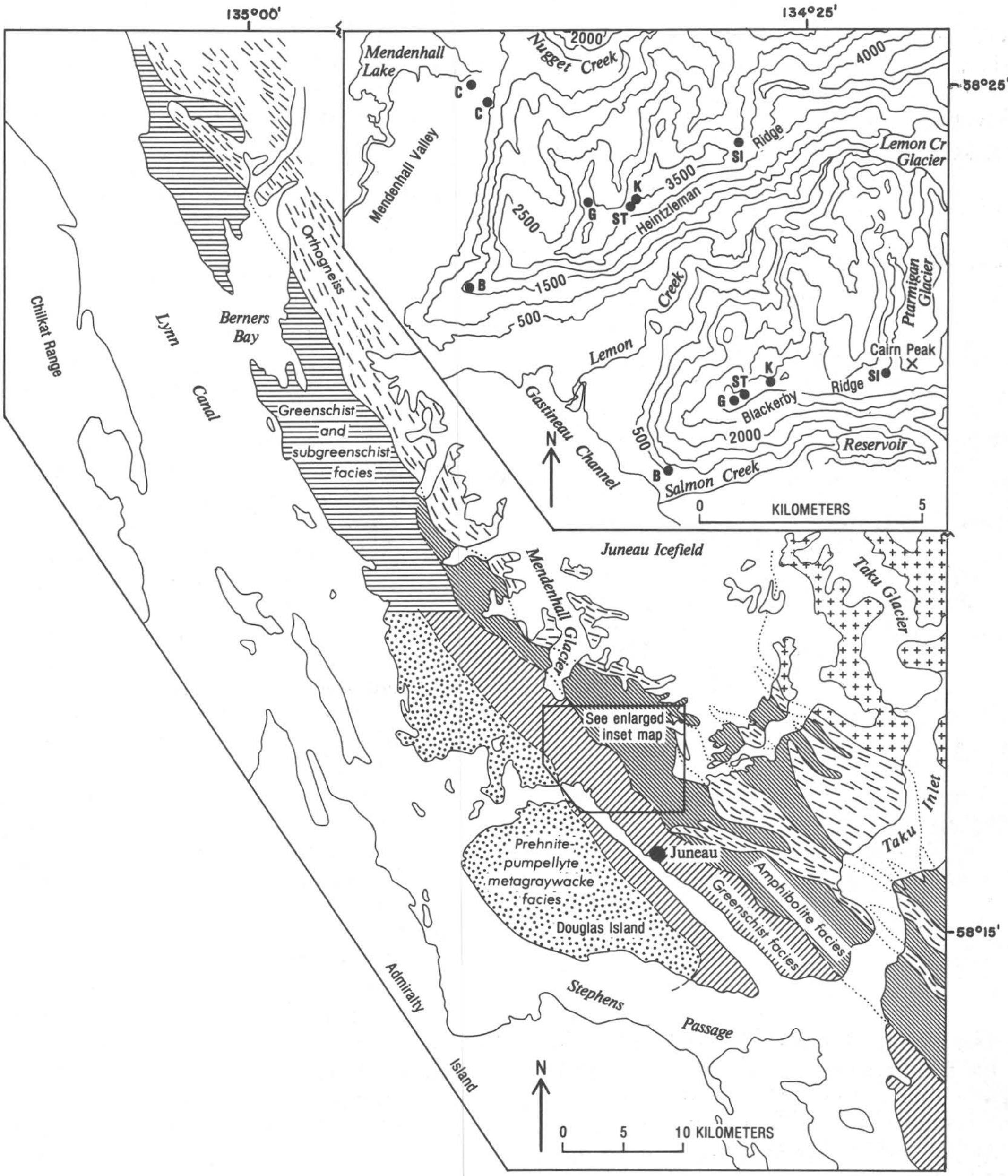


**Figure 54.**--Sketch map of the Juneau area, showing distribution of metamorphic facies and relation to plutonic units of the west margin of Coast plutonic-metamorphic complex. Dashed-line pattern shows approximate foliation trends in orthogneiss plutons. Crosses denote little-foliated post-tectonic granitic plutons. Inset topographic map (500-ft contours) shows first occurrences of biotite (B), garnet (G), staurolite (ST), kyanite (K), and sillimanite (SI), and occurrences of chloritoid (C) in northeastward transects on Heintzleman and Blackerby Ridges. After Himmelberg and others (1984).

**Table 9.**--Chemical compositions and structural formulas of chloritoid in schists near Juneau, Alaska (-, not detected)

| Constituent | 68Bd 145A | 68Bd 159A | 68Bd 159B |
|---|---|---|---|
| Composition, in weight percent | | | |
| $SiO_2$ | 23.8 | 23.5 | 23.8 |
| $TiO_2$ | - | 0.05 | 0.03 |
| $Al_2O_3$ | 39.2 | 39.9 | 40.3 |
| FeO* | 25.8 | 25.3 | 24.9 |
| MnO | 0.28 | .38 | .78 |
| MgO | 1.91 | 2.17 | 1.99 |
| CaO | - | - | .05 |
| Subsum | 90.99 | 91.30 | 91.85 |
| $H_2O$** | 7.09 | 7.13 | 7.18 |
| Sum | 98.08 | 98.43 | 99.03 |
| Formula per 4 cations | | | |
| Si | 1.00 | .99 | .99 |
| Ti | - | .002 | .001 |
| Al | 1.96 | 1.97 | 1.98 |
| Fe | .91 | .89 | .87 |
| Mn | .01 | .01 | .03 |
| Mg | .12 | .14 | .12 |
| Ca | - | - | .002 |
| Fe/(Fe+Mg+Mn) | .88 | .86 | .85 |

*Total iron calculated as FeO
**Calculated by method of Zen (1981) assuming fully hydrated stoichiometric formula

the Juneau area are controlled by bulk composition. Garnet-biotite geothermometry for the lower garnet zone (Himmelberg and others, 1984) indicates that the chloritoid formed at a temperature less than 495° C.

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## GEOMETRIC STRUCTURAL ANALYSIS OF PART OF THE WESTERN COAST PLUTONIC-METAMORPHIC COMPLEX EAST OF WRANGELL

**Susan J. Hunt**
**and David A. Brew**

A geometric structural analysis was performed on measurements of joints, foliations, lineations, and axial planes from 168 stations in the western part of the informally designated Coast plutonic-metamorphic complex of Brew and Ford (1984) east of Wrangell (fig. 55). The main goal of this analysis was to provide insight into the structural effects related to metamorphic and plutonic events. According to Brew and Ford (1983, 1984, 1985) and Monger and others (1982), these events are related to deformation and intrusion in a linear belt that involves only one or two terranes and one overlap assemblage. These same events are recorded in several different accreted terranes, according to Berg and others (1978). Another goal for this study was to test whether the Coast Range Megalineament (Brew and Ford, 1978), a prominent lineament in the study area, is a major structural boundary. No geometrical structural analyses have been published from this or any other part of the Coast plutonic-metamorphic complex in southeastern Alaska. Loney (1965), however, made a structural analysis of a large area on Admiralty Island, 100 km to the west-northwest, that has an uncertain relation to this study area.

The area studied is approximately 20 km x 35 km, located between the Stikine River and Eastern Passage (fig. 56). It is underlain by two types of plutonic bodies. The first type is a part of a regionally mapped tonalite sill of latest Cretaceous and (or) early Tertiary age [about 65-70 Ma (See time scale of Palmer, 1984); Gehrels and others, 1984], and the second consists of several equidimensional garnet-biotite granodiorite bodies of Late Cretaceous age (about 90 Ma; M.A. Lanphere, USGS, written commun., 1986). Also present are schist, metavolcanic, and minor marble country rocks of Paleozoic(?) to Mesozoic(?) age and of inferred Permian and Triassic age (Brew and others, 1984). Metamorphic grade ranges from greenschist to amphibolite facies, generally increasing eastward. The rocks are strongly foliated and locally contain isoclinal folds. Near the center of the mapped area is a zone of rhyolite dikes and sills of middle Tertiary age (Brew and others, 1984), which generally parallel compositional layering.

The study area was subdivided into 25 domains and subdomains of similar structural orientation that included parts of the plutonic bodies (A-N, fig. 56). These domains and subdomains were analyzed separately. The data for each domain were plotted on the lower hemisphere of a Schmidt equal-area net. The structural analysis used 199 foliation, 27 lineation, 13 fold, 281 joint, and 22 dike orientation measurements. Most domains included a single intrusive-metamorphic contact of fairly constant orientation. Details concerning procedures and data from individual domains are given in Hunt (1986); only composite diagrams are considered in this discussion.

Southwest of the Coast Range Megalineament, foliation is well developed in the metamorphic rocks and to a lesser degree in the plutonic rocks. The contoured plot (fig. 57A) shows a strong northwest-striking orientation (mostly between 348° and

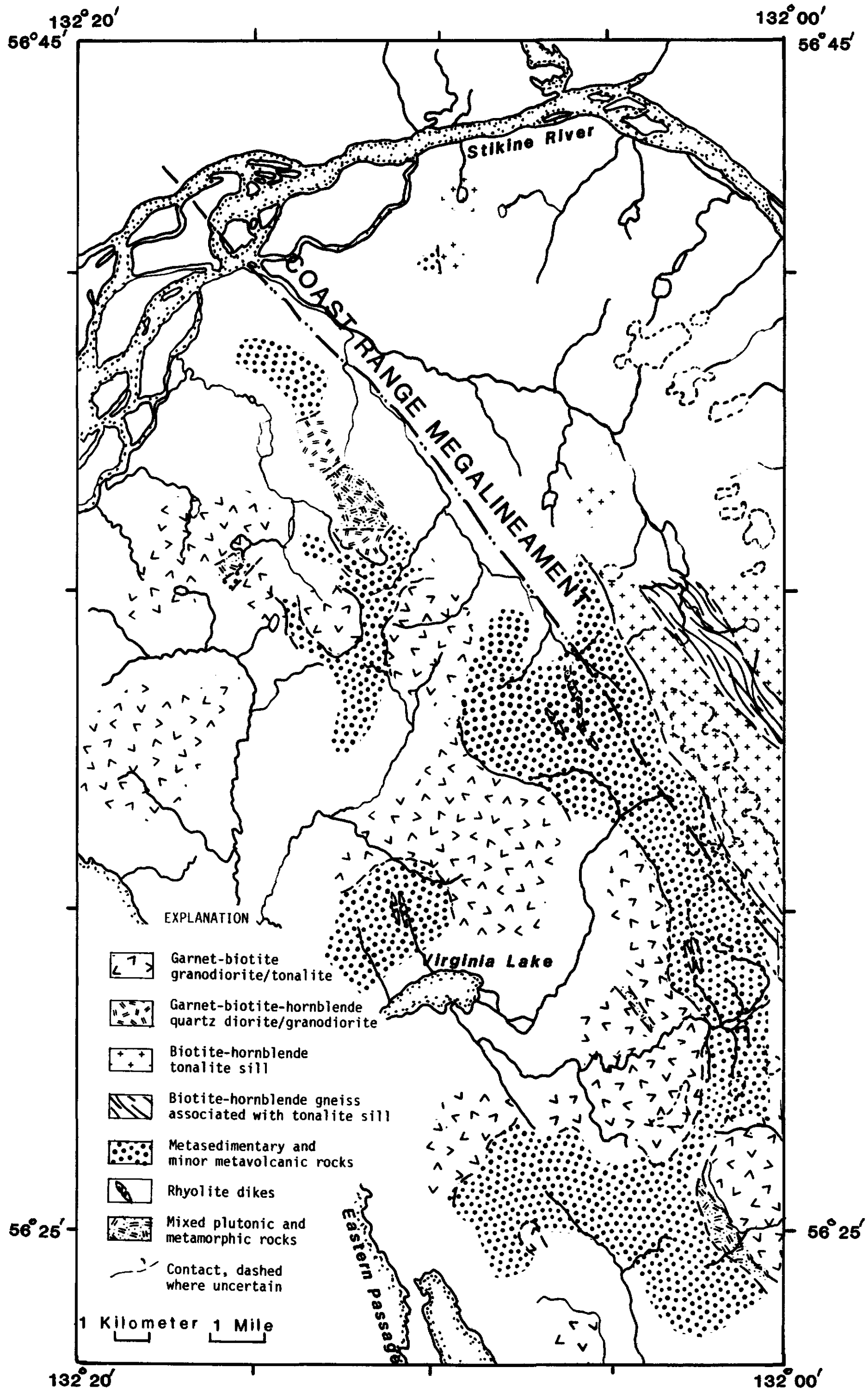


**Figure 55.**--Generalized geologic map of the east-central part of the Petersburg quadrangle.

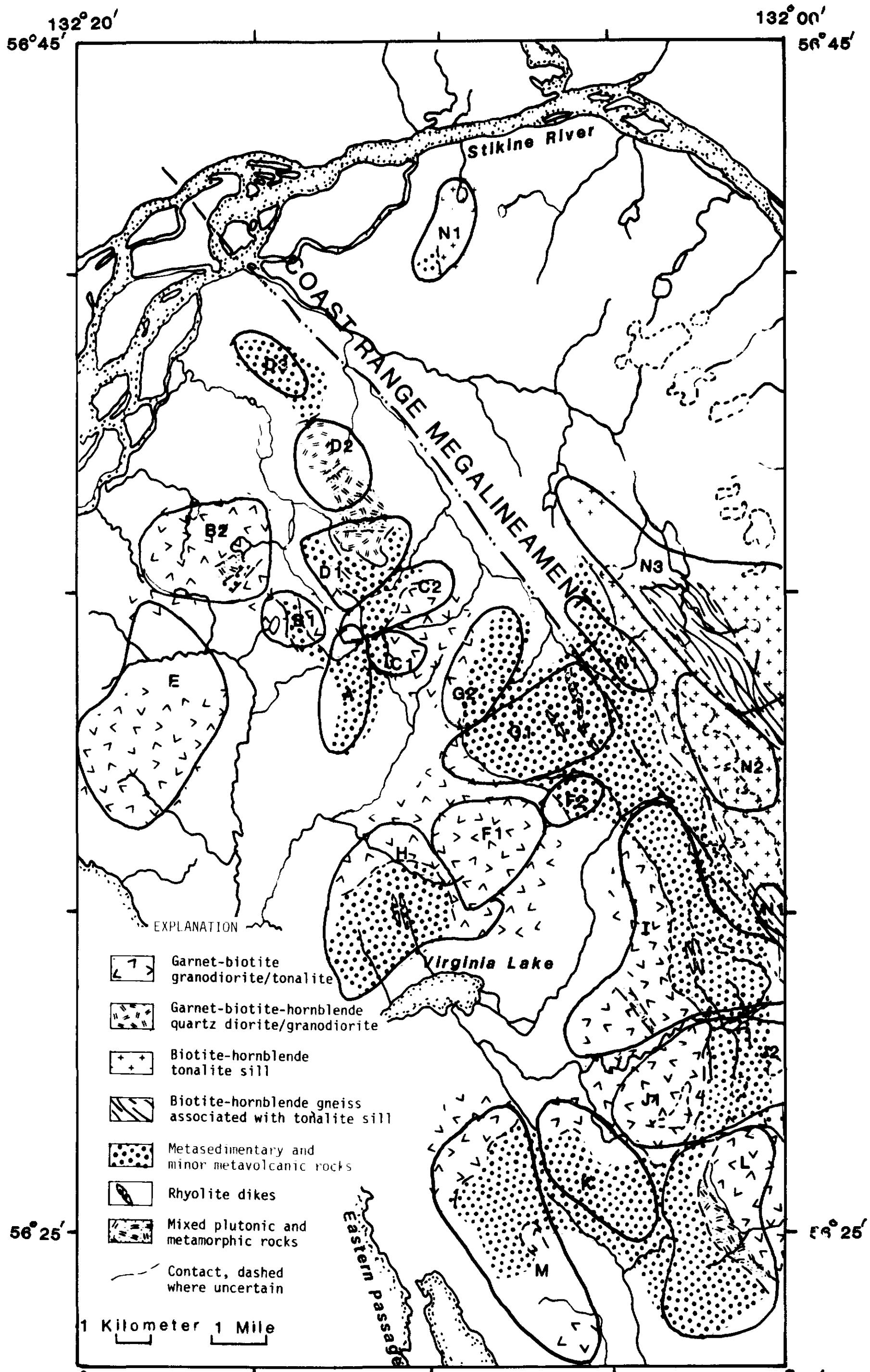


**Figure 56.**--Generalized geology and boundaries of structural domains (heavy lines) which are described in the text.

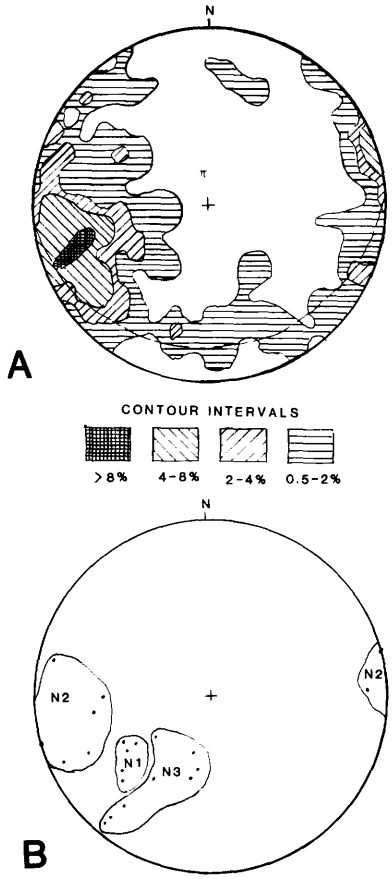


**Figure 57.**--Equal-area plots of poles to foliation (A) 179 measurements from the domains southwest of Coast Range Megalineament; (B) 20 measurements taken northeast of the Coast Range Megalineament in domains N1, N2, and N3.

338°) and dips between 58° and 84° NE. A significant number of poles to foliation define a shallowly south-dipping girdle, suggesting deformation of the foliation about a later-formed, steeply north-plunging Pi axis. Plots from individual domains (Hunt, 1986) indicate that this girdle reflects the local influence of the 90-Ma intrusions. In general, the foliation in both metamorphic and plutonic rocks strikes parallel to contacts of the somewhat equant intrusions with surrounding country rock (fig. 57A); gradual changes in orientation of both contacts and foliation indicate that the intrusion of the granodiorite bodies has deformed the foliation in the country rocks.

Foliation in the metamorphic rocks generally dips steeply (50°-90°) away from intrusive bodies. These relations indicate that pre-existing metamorphic foliation has been deformed by the intrusion of the 90-Ma plutons. Douglass and others (1985) suggest that this metamorphic foliation post-dates Cenomanian (early Late Cretaceous) fossils and pre-dates the intrusion of 100-110-Ma mafic-ultramafic bodies.

Plots of foliations northeast of the megalineament are shown in figure 57B. The strike in domain N (fig. 56) seems to be slightly more northerly and the dip shallower than those shown in figure 57A, but the differences are not significant. Domain N has three subdomains, although their significance is not clear (fig. 57B). Domain N1, which contains foliation measurements close to the schist-tonalite "sill" contact, shows a tight distribution and 45° dip (shallower than in the schists to the west); strikes of this foliation approximately parallel the contact. Domains N2 and N3 differ in that the strike changes from NS to NW and the dip shallows slightly from N2 to N3.

In summary: a regional 340°-striking foliation dominates the study area; it generally parallels compositional layering and metamorphic contacts, but it is influenced locally by the later intrusion of 90-Ma plutons. The same regional orientation is also apparent in the tonalite "sill" and adjacent metamorphic rocks on the northeast side of the Coast Range Megalineament.

Lineations in the area consist of aligned minerals and mineral streaks on foliation surfaces. Twenty-seven measurements are plotted in figure 58A; they show a wide variation in orientation. There are not enough values in any one domain to justify detailed analysis, but domains on the northeast side of the large central pluton (domains G, J, N, and the northern part of I on fig. 56) have generally NE-plunging lineations. Those on the south side of the pluton (domains H, M, and K) have lineations with SE plunges. Here again, domain N northeast of the megalineament does not appear to be different from the area to the southwest.

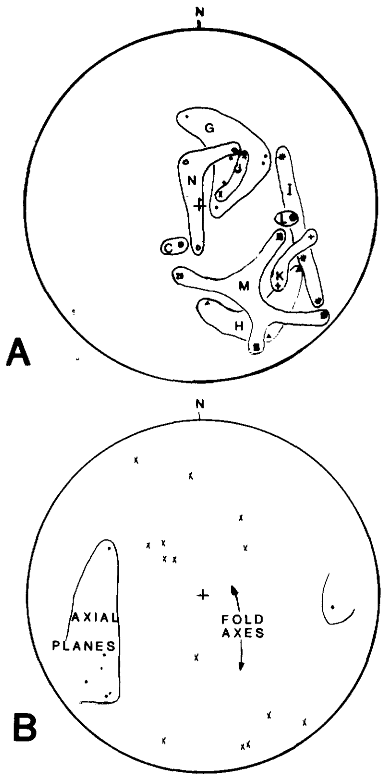


**Figure 58.**--Equal-area plots: (A) 27 lineations (heavy lines group points by domain); (B) 14 fold axes and 7 poles to axial planes.

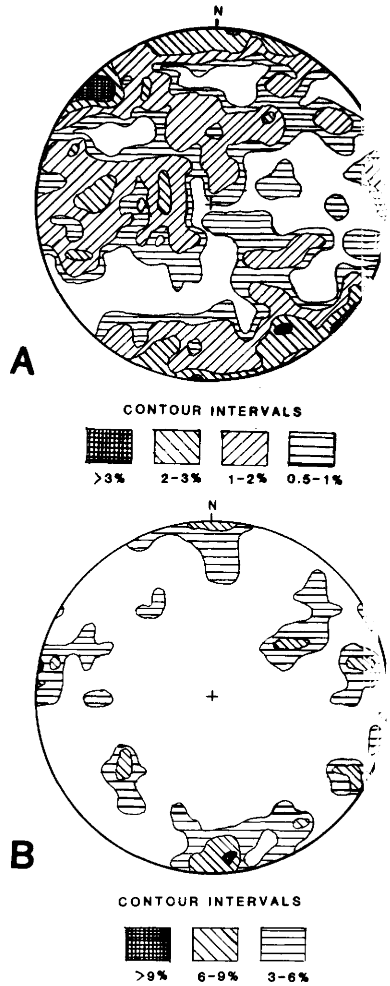


**Figure 59.**--Contoured equal-area plots of poles to joints: (A) 249 measurements from the area southwest of the Coast Range Megalineament; (B) 32 measurements from the area northeast of the Coast Range Megalineament. Lowest contour interval includes all poles.

Isoclinal folds are not abundant in the map area, but they were noted within the metamorphic sequence, most commonly close to an intrusive contact. Orientations of fold axes and associated axial planes are plotted in figure 58B. Except for one fold axis in each of domains C, H, and I, all the plotted points are from domains J and L. Axial planes are subparallel to the foliation.

Joints measured display a wide variation in orientation on both the southwest side of the Coast Range Megalineament (249 measurements, fig. 59A) and on its northeast side (32 measurements, fig. 59B). Several maxima can be discerned among the broad scatter of points in figure 59A: a strong NE-striking vertical set, a nearly EW-striking vertical set, and a broad NS-striking, east-dipping set. When separated into plutonic and metamorphic groups (Hunt, 1986), the joints on the southwest side of the megalineament show obvious differences. Joint patterns in the metamorphic rocks there indicate a strong NE orientation, with possibly two maxima--one at 040° and one at 070°--and the joints are mostly steeply dipping. Joint patterns in the plutons include a prominent EW-trending vertical set and a N-S set that dips moderately to the east. Apparently, joints in the plutonic rocks formed in response to a different stress pattern from that which produced the joints in the metamorphic rocks. Joint measurements northeast of the megalineament (fig. 59B) comprise a small data set. The orthogonal (NS/EW) sets are present, as is a set subparallel to foliation, and the 040°-070° set seen in the metamorphic rocks on the southwest side is conspicuously absent. Thus, the tonalite "sill" joints are similar to those in the granodiorite-tonalite-quartz diorite bodies on the opposite side of the megalineament.

This study indicates that a regional metamorphic and deformational event imparted a NW-striking foliation to all the metasedimentary country rocks. The foliation in rocks south of the Coast Range Megalineament was subsequently deformed by numerous 90-Ma plutons. The foliation was warped into parallel alignment with the intrusive contact and both the foliations and lineations tend to dip and plunge, respectively, away from the plutons. Douglass and others (1985) describe superposition of metamorphic effects associated with the 65-70-Ma tonalite sill on both a metamorphism associated with the 90-Ma granodiorite bodies and on a pre-110-Ma metamorphism. The last is the same as the one that is here concluded to precede the intrusion of the 90-Ma plutons. This complicated metamorphic history may have produced the minor structural contrasts on either side of the Coast Range Megalineament indicated by our data. The results of this structural analysis indicate, however, that in the study area, the megalineament is not a major structural discontinuity, confirming the conclusions of Brew and Ford (1978, 1981).

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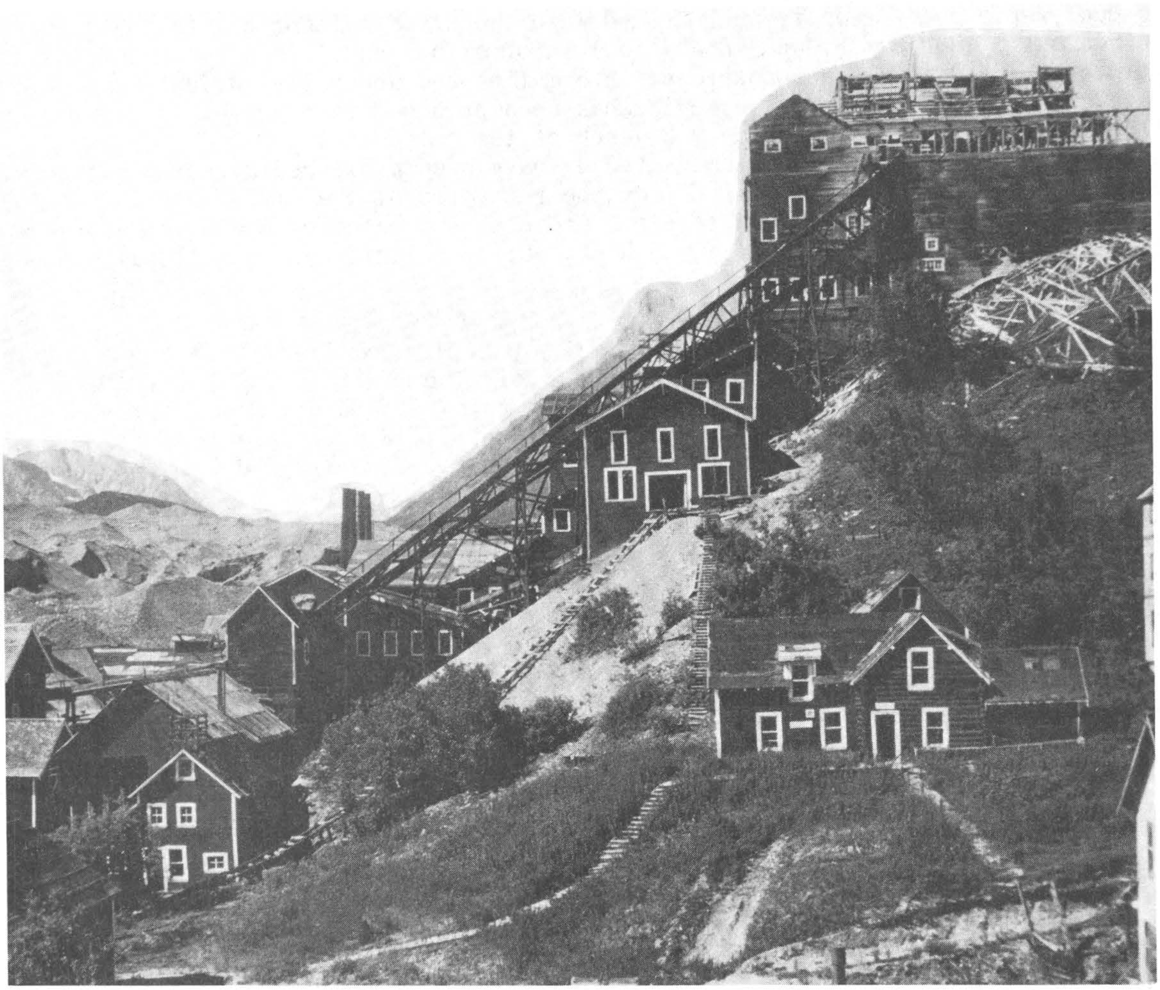

**BIBLIOGRAPHY I--Data releases and folio reports prepared for the Alaska Mineral Resource Assessment Program and the Regional Alaska Mineral Resource Assessment Program listed alphabetically by quadrangle**

**compiled by Gary R. Winkler**

The following bibliography of reports prepared for the Alaska Mineral Resource Assessment Program (AMRAP) is current as of January 1, 1986. It is composed of two lists. The first list consists all reports prepared for the regional program (or RAMRAP) and categorizes the reports within large regions of Alaska, but it includes several reports that are statewide in scope. The nominal map scale of these reports ranges from 1:600,000 to 1:1,000,000. The second list is of publications that describe varied aspects of the geology and mineral resources of the 49 1°x3° quadrangles (mostly at 1:250,000 scale) for which studies are complete or under way in the program. The citations are arranged alphabetically by quadrangle (or by pairs of quadrangles if two adjacent quadrangles were studied together).

The quadrangle-based reports are divided into two broad categories: (1) data releases that include basic geochemical and geophysical information, and (2) folio components that provide interpretation of basic geoscience information and generally include maps. For many quadrangle studies, a summary circular has been produced that serves as a guide to the folios and includes a short summary of each of the components.

Hundreds of additional reports and articles have been published by scientists participating in the program, either derived from their studies of the 1:250,000-scale quadrangles or developed separately through research projects supported by AMRAP. These journal articles, maps, or abstracts are not cited here, but are significant additional products of the program.

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Regional reports:

**STATEWIDE**


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## CHIGNIK AND SUTWIK ISLAND QUADRANGLES

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LeCompte, J.R., 1981, Landsat features map of the Ketchikan and Prince Rupert quadrangles, Alaska: U.S. Geological Survey Open-File Report 81-1139, 2 sheets, scale 1:250,000.

Steele, W.C., and Albert, N.R.D., 1978, Map showing interpretation of Landsat imagery of the Ketchikan and Prince Rupert quadrangles, Alaska: U.S. Geological Survey Open-File Report 78-73-K, 2 sheets, scale 1:250,000.

Geochronology:

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Other:

Berg, H.C., and Cruz, E.L., 1982, Map and table describing fossil collections and related samples in the Ketchikan and Prince Rupert quadrangles, southeastern Alaska: U.S. Geological Survey Open-File Report 82-1088, 1 sheet, scale 1:250,000.

Circular:

Berg, H.C., 1982, The Alaska Mineral Resources Assessment Program--Guide to information about the geology and mineral resources of the Ketchikan and Prince Rupert quadrangles, southeastern Alaska: U.S. Geological Survey Circular 855, 24 p.

**KILLIK RIVER AND CHANDLER LAKE QUADRANGLES**

DATA RELEASES:

Stream-sediment geochemistry:

Barton, H.N., Odland, S.K., O'Leary, R.M., and Day, G.W., 1982, Geochemical data for the Killik River and Chandler Lake quadrangles, Alaska: U.S. Geological Survey Open-File Report 82-1026, 53 p.

Sutley, S.J., Duttweiler, K.A., and Hopkins, R.T., 1984, Analytical results and sample locality maps of stream-sediment and panned-concentrate samples from the Killik River 1°x3° quadrangle: U.S. Geological Survey Open-File Report 84-406, 18 p., 1 sheet, scale 1:250,000.

Sutley, S.J., Duttweiler, K.A., and Hopkins, R.T., 1984, Analytical results and sample locality maps of stream-sediment and panned-concentrate samples from the Chandler Lake 1°x3° quadrangle: U.S. Geological Survey Open-File Report 84-412, 30 p., 1 sheet, scale 1:250,000.

Aeromagnetic map:

U.S. Geological Survey, 1983, Aeromagnetic map of the Killik River and Chandler Lake 1°x3° quadrangle, Alaska: U.S. Geological Survey Open-File Report 83-607, 1 sheet, scale 1:250,000.

Other:

Brosgé, W.P., Reiser, H.N., Dutro, J.T., Jr., and Nilsen, T.H., 1979, Geologic map of Devonian rocks in parts of the Chandler Lake and Killik River quadrangles, Alaska: U.S. Geological Survey Open-File Report 79-1224, 1 sheet, scale 1:200,000.

Kelley, J.S., 1984, Geologic map of a portion of the Chandler Lake B-1 quadrangle, Alaska: U.S. Geological Survey Open-File Report 84-77, 3 sheets, scale 1:63,360.

Kelley, J.S., 1984, Geologic map and sections of a portion of the Chandler Lake A-1 and A-2 quadrangles, Alaska: U.S. Geological Survey Open-File Report 84-555, 1 sheet, scale 1:63,360.

Kelley, J.S., and Peterson, D.M., 1983, Preliminary bibliography of geologic literature on the Killik River and Chandler Lake quadrangles and parts of neighboring quadrangles, north-central Brooks Range, Alaska: U.S. Geological Survey Open-File Report 83-232, 30 p.

FOLIO COMPONENTS:

Surficial geology:

Hamilton, T.D., 1979, Surficial geologic map of the Chandler Lake quadrangle, Alaska: U.S. Geological Survey Miscellaneous Field Studies Map MF-1121, 1 sheet, scale 1:250,000.

Hamilton, T.D., 1982, Quaternary stratigraphic sections with radiocarbon dates, Killik River quadrangle, Alaska: U.S. Geological Survey Open-File Report 82-606, 33 p., 1 sheet, scale 1:250,000.

Hamilton, T.D., 1980, Quaternary stratigraphic sections with radiocarbon dates, Chandler Lake quadrangle, Alaska: U.S. Geological Survey Open-File Report 80-790, 33 p., 1 sheet, scale 1:250,000.

Hamilton, T.D., 1980, Surficial geologic map of the Killik River quadrangle: U.S. Geological Survey Miscellaneous Field Studies Map MF-1234, 1 sheet, scale 1:250,000.

## LAKE CLARK QUADRANGLE

DATA RELEASES:

Stream-sediment geochemistry:

Crim, W.D., Cooley, E.F., Hampton, J.O., McDanal, S.K., and O'Leary, R.M., 1979, Spectrographic and chemical analyses of geochemical samples collected during 1978 from the Lake Clark quadrangle, Alaska: U.S. Geological Survey Open-File Report 79-871, 48 p., 1 sheet, scale 1:250,000.

King, H.D., Crim, W.D., Cooley, E.F., McDanal, S.K., O'Leary, R.M., and Cohen, D.K., 1978, Spectrographic and chemical analyses of geochemical samples from the Lake Clark quadrangle, Alaska: U.S. Geological Survey Open-File Report 78-788, 58 p., 1 sheet, scale 1:250,000.

Aeromagnetic map:

U.S. Geological Survey, 1978, Aeromagnetic map of Lake Clark quadrangle, Alaska: U.S. Geological Survey Open-File Report 78-1086, 1 sheet, scale 1:250,000.

FOLIO COMPONENTS:

Geology:

Nelson, W.H., Carlson, Christine, and Case, J.E., 1983, Geologic map of the Lake Clark quadrangle, Alaska: U.S. Geological Survey Miscellaneous Field Studies Map MF-1114-A, 2 sheets, scale 1:250,000.

Geophysics:

Case, J.E., and Nelson, W.H., 1985, Geologic interpretation of aeromagnetic map of the Lake Clark quadrangle, Alaska: U.S. Geological Survey Miscellaneous Field Studies Map MF-1114-G, 2 sheets, scale 1:250,000.

Geochemistry:

King, H.D., Cooley, E.F., McDanal, S.K., and Crim, W.D., 1985, Distribution and abundance of tin and tungsten in nonmagnetic heavy-mineral concentrate samples, Lake Clark quadrangle, Alaska: U.S. Geological Survey Miscellaneous Field Studies Map MF-1114-H, 1 sheet, scale 1:250,000.

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King, H.D., Cooley, E.F., O'Leary, R.M., Tripp, R.B., and Crim, W.D., 1985, Distribution and abundance of gold, tin, and tungsten in two geochemical sampling media, Lake Clark quadrangle, Alaska: U.S. Geological Survey Miscellaneous Field Studies Map MF-1114-D, 1 sheet, scale 1:250,000.

King, H.D., O'Leary, R.M., Cooley, E.F., and Crim, W.D., 1985, Distribution and abundance of arsenic, copper, lead, molybdenum, silver, and zinc in two geochemical sampling media, Lake Clark quadrangle, Alaska: U.S. Geological Survey Miscellaneous Field Studies Map MF-1114-E, 1 sheet, scale 1:250,000.

Mineral Occurrences and Resources Assessment:

Nelson, W.H., King, H.D., Case, J.E., Tripp, R.B., Crim, W.D., and Cooley, E.F., 1985, Mineral resource map of the Lake Clark quadrangle, Alaska: U.S. Geological Survey Miscellaneous Field Studies Map MF-1114-B, 2 sheets, scale 1:250,000.

Telegeology:

Steele, W.C., 1985, Map showing interpretation of Landsat imagery of the Lake Clark quadrangle, Alaska: U.S. Geological Survey Miscellaneous Field Studies Map MF-1114-F, 1 sheet, scale 1:250,000.

## McCARTHY QUADRANGLE

DATA RELEASES;

Stream-sediment geochemistry:

O'Leary, R.M., McDanal, S.K., Day, G.W., McDougal, C.M., and Robinson, Keith, 1976, Spectrographic and chemical analyses of geochemical samples from the McCarthy quadrangle, Alaska: U.S. Geological Survey Open-File Report 76-824, 806 p.

Aeromagnetic map:

U.S. Geological Survey, 1976, Aeromagnetic map of the McCarthy 1:250,000-scale (1°x3°) quadrangle, Alaska: U.S. Geological Survey Open-File Report 76-170, 1 sheet, scale 1:250,000.

FOLIO COMPONENTS:

Geology:

MacKevett, E.M., Jr., 1976, Geologic map of the McCarthy quadrangle, Alaska: U.S. Geological Survey Miscellaneous Field Studies Map MF-773-A, 1 sheet, scale 1:250,000.

Geophysics:

Case, J.E., and MacKevett, E.M., Jr., 1976, Aeromagnetic map and geologic interpretation of an aeromagnetic map of the McCarthy quadrangle, Alaska: U.S. Geological Survey Miscellaneous Field Studies Map MF-773-D, 2 sheets, scale 1:250,000.

Geochemistry:

Robinson, Keith, McDougal, C.M., and Billings, Theodore, 1976(1977), Distribution and abundance of copper in stream sediments and moraine debris, McCarthy quadrangle, Alaska: U.S. Geological Survey Miscellaneous Field Studies Map MF-773-F, 1 sheet, scale 1:250,000.

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Robinson, Keith, McDougal, C.M., McDanal, S.K., and Billings, Theodore, 1976(1977), Distribution and abundance of silver in bedrock, mineralized, vein, and altered rock samples, McCarthy quadrangle, Alaska: U.S. Geological Survey Miscellaneous Field Studies Map MF-773-K, 1 sheet, scale 1:250,000.

Robinson, Keith, McDougal, C.M., McDanal, S.K., and Billings, Theodore, 1976(1977), Distribution and abundance of gold in bedrock, mineralized, vein, and altered rock samples, McCarthy quadrangle, Alaska: U.S. Geological Survey Miscellaneous Field Studies Map MF-773-L, 1 sheet, scale 1:250,000.

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Mineral Occurrences:

MacKevett, E.M., Jr., 1976, Mineral deposits and occurrences in the McCarthy quadrangle, Alaska: U.S. Geological Survey Miscellaneous Field Studies Map MF-773-B, 2 sheets, scale 1:250,000.

Resources Assessment:
Singer, D.A., and MacKevett, E.M., Jr., 1976(1977), Mineral resources map of the McCarthy quadrangle, Alaska: U.S. Geological Survey Miscellaneous Field Studies Map MF-773-C, 1 sheet, scale 1:250,000.
Telegeology:
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Other:
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Circular:
MacKevett, E.M., Jr., Albert, N.R.D., Barnes, D.F., Case, J.E., Robinson, Keith, and Singer, D.A., 1977, The Alaska Mineral Resource Assessment Program--Background information to accompany folio of geologic and mineral resource maps of the McCarthy quadrangle, Alaska: U.S. Geological Survey Circular 739, 23 p.
Geologic Map (colored):
MacKevett, E.M., Jr., 1978, Geologic map of the McCarthy quadrangle, Alaska: U.S. Geological Survey Miscellaneous Investigations Series Map I-1032, 1 sheet, scale 1:250,000.

## MEDFRA QUADRANGLE

DATA RELEASES:

Stream-sediment geochemistry:
King, H.D., Cooley, E.F., Gruzensky, A.L., and Spiesman, D.L., Jr., 1983, Distribution and abundance of gold and silver in nonmagnetic and moderately magnetic heavy-mineral-concentrate and minus-80-mesh stream-sediment samples and silver in ash of aquatic-bryophyte samples, Medfra quadrangle, Alaska: U.S. Geological Survey Open-File Report 80-811-H, 3 sheets, scale 1:250,000.
King, H.D., Cooley, E.F., O'Leary, R.M., Tripp, R.B., McDanal, S.K., and Spiesman, D.L., 1979, Spectrographic and atomic-absorption analyses of geochemical samples from the Medfra quadrangle, Alaska: U.S. Geological Survey Open-File Report 79-959, 44 p., 1 sheet, scale 1:250,000.
King, H.D., Cooley, E.F., and Spiesman, D.L., Jr., 1983, Distribution and abundance of arsenic and bismuth in nonmagnetic and moderately magnetic heavy-mineral concentrate samples and arsenic in minus-80-mesh stream-sediment and ash of aquatic-bryophyte samples, Medfra quadrangle, Alaska: U.S. Geological Survey Open-File Report 80-811-K, 3 sheets, scale 1:250,000.
King, H.D., O'Leary, R.M., Risoli, D.A., and Galland, D.W., 1983, Distribution and abundance of antimony and mercury in minus-80-mesh stream-sediment and antimony in non-magnetic and moderately magnetic heavy-mineral-concentrate samples, Medfra quadrangle, Alaska: U.S. Geological Survey Open-File Report 80-811-L, 2 sheets, scale 1:250,000.
King, H.D., Risoli, D.A., Cooley, E.F., O'Leary, R.M., Speckman, W.A., Spiesman, D.L., Jr., and Galland, D.W., 1980, Final results and statistical summary of analysis of geochemical samples from the Medfra quadrangle, Alaska: U.S. Geological Survey Open-File Report 80-811-F, 136 p., 1 sheet, scale 1:250,000.
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King, H.D., Tripp, R.B., O'Leary, R.M., and Cooley, E.F., 1983, Distribution

and abundance of copper, lead, and zinc in nonmagnetic and moderately magnetic heavy-mineral-concentrate, minus-80-mesh stream-sediment, and ash of aquatic-bryophyte samples, Medfra quadrangle, Alaska: U.S. Geological Survey Open-File Report 80-811-I, 6 sheets, scale 1:250,000.

Aeromagnetic map:

U.S. Geological Survey, 1979, Aeromagnetic map of the Medfra 1°x3° quadrangle, Alaska: U.S. Geological Survey Open-File Report 79-380, 1 sheet, scale 1:250,000.

Other:

Miller, T.P., Moll, E.J., and Patton, W.W., Jr., 1980, Uranium- and thorium-rich volcanic rocks of the Sischu Creek area, Medfra quadrangle, Alaska: U.S. Geological Survey Open-File Report 80-803, 9 p.

FOLIO COMPONENTS:

Geology:

Patton, W.W., Jr., Moll, E.J., Dutro, J.T., Jr., Silberman, M.L., and Chapman, R.M., 1980, Preliminary geologic map of Medfra quadrangle, Alaska: U.S. Geological Survey Open-File Report 80-811-A, 1 sheet, scale 1:250,000.

Geophysics:

Patton, W.W., Jr., Cady, J.W., and Moll, E.J., 1982, Aeromagnetic interpretation of the Medfra quadrangle, Alaska: U.S. Geological Survey Open-File Report 80-811-E, 15 p., 2 sheets, scale 1:250,000.

Mineral Occurrences:

Schwab, C.E., Patton, W.W., Jr., and Moll, E.J., 1981, Mineral occurrence map of the Medfra quadrangle, Alaska: U.S. Geological Survey Open-File Report 80-811-B, 2 sheets, scale 1:250,000.

Resources Assessment:

Patton, W.W., Jr., and Moll, E.J., 1983, Mineral resource assessment map of the Medfra quadrangle, Alaska: U.S. Geological Survey Open-File Report 80-811-G, 3 sheets, scale 1:250,000.

Telegeology:

LeCompte, J.R., 1981, Maps showing interpretation of Landsat imagery of the Medfra quadrangle, Alaska: U.S. Geological Survey Open-File Report 80-811-D, 2 sheets, scale 1:250,000.

Geochronology:

Moll, E.J., Silberman, M.L., and Patton, W.W., Jr., 1981, Chemistry, mineralogy, and K-Ar ages of igneous and metamorphic rocks of the Medfra quadrangle, Alaska: U.S. Geological Survey Open-File Report 80-811-C, 2 sheets, scale 1:250,000.

Circular:

Patton, W.W., Jr., Moll, E.J., and King, H.D., 1984, The Alaska Mineral Resource Assessment Program--Guide to information contained in the folio of geologic and mineral resource maps of the Medfra quadrangle, Alaska: U.S. Geological Survey Circular 928, 11 p.

## MOUNT HAYES QUADRANGLE

DATA RELEASES:

Stream-sediment geochemistry:

O'Leary, R.M., Risoli, D.A., Curtin, G.C., and McDanal, S.K., 1981, Spectrographic and chemical analyses of stream-sediment and glacial-debris samples from Mount Hayes quadrangle, Alaska: U.S. Geological Survey Open-File Report 81-226, 58 p., 1 sheet, scale 1:250,000.

Panned-concentrate geochemistry:

O'Leary, R.M., Risoli, D.A., Curtin, G.C., Tripp, R.B., McDougal, D.M., and Huston, D.L., 1982, Final analytical results of stream-sediment, glacial-debris, and non-magnetic heavy-mineral concentrate samples from the Mount Hayes quadrangle, Alaska: U.S. Geological Survey Open-File Report 82-325, 1 sheet, scale 1:250,000.

Bedrock geochemistry:

Zehner, R.E., O'Leary, R.M., Day, G.W., Sutley, S.J., Nokleberg, W.J., and Koch, R.D., 1985, Emission spectrographic and atomic absorption analyses of rock, mineral deposit, prospect, and occurrence samples, Mount Hayes quadrangle, eastern Alaska Range, Alaska: U.S. Geological Survey Open-File Report 84-774, 141 P., 1 sheet, scale 1:250,000.

Aeromagnetic map:

Alaska Division of Geological and Geophysical Surveys, 1973 (revised, 1977), Aeromagnetic map, Mount Hayes quadrangle, Alaska: Alaska Division of Geological and Geophysical Surveys Open-File Report AOF-10, 1 sheet, scale 1:250,000.

Other:

Zehner, R.E., Cobb, E.H., Nokleberg, W.J., and Albert, N.R.D., 1980, Geologic bibliography of the Mount Hayes quadrangle, Alaska: U.S. Geological Survey Open-File Report 80-513, 29 p.

FOLIO COMPONENTS:

Geology:

Nokleberg, W.J., Albert, N.R.D., Bond, G.C., Herzon, P.L., Miyaoka, R.T., Nelson, W.H., Richter, D.H., Smith, T.E., Stout, J.H., Yeend, Warren, and Zehner, R.E., 1982, Geologic map of the southern part of the Mount Hayes quadrangle, Alaska: U.S. Geological Survey Open-File Report 82-52, 1 sheet, scale 1:250,000.

## MOUNT KATMAI AND NAKNEK QUADRANGLES

DATA RELEASES:

Aeromagnetic map:

Andreasen, G.E., and others, 1963, Aeromagnetic map of part of the Naknek quadrangle, Alaska: U.S. Geological Survey Geophysical Investigations Map GP-353, 1 sheet, scale 1:125,000.

## NABESNA QUADRANGLE

DATA RELEASES:

Aeromagnetic map:

Alaska Division of Geological and Geophysical Surveys, 1973, Aeromagnetic map, Nabesna quadrangle, Alaska: Alaska Division of Geological and Geophysical Surveys Open-File Report AOF-13, 1 sheet, scale 1:250,000.

FOLIO COMPONENTS:

Geology:

Richter, D.H., 1975, Geologic map of the Nabesna quadrangle, Alaska: U.S. Geological Survey Miscellaneous Field Studies Map MF-655-A, 1 sheet, scale 1:250,000.

Geophysics:

Barnes, D.R.and Morin, R.L., 1975, Gravity map of the Nabesna quadrangle, Alaska: U.S. Geological Survey Miscellaneous Field Studies Map MF-655-I, 1 sheet, scale 1:250,000.

Griscom, Andrew, 1975, Aeromagnetic map and aeromagnetic interpretation of the Nabesna quadrangle, Alaska: U.S. Geological Survey Miscellaneous Field Studies Map MF-655-H, 1 sheet, scale 1:250,000.

Geochemistry:

Marsh, S.P., 1975, Geochemical and generalized geologic map showing distribution and abundance of copper in the Nabesna quadrangle, Alaska: U.S. Geological Survey Miscellaneous Field Studies Map MF-655-B, 1 sheet, scale 1:250,000.

Marsh, S.P., 1975, Geochemical and generalized geologic map showing distribution and abundance of lead in the Nabesna quadrangle, Alaska: U.S. Geological Survey Miscellaneous Field Studies Map MF-655-C, 1 sheet, scale 1:250,000.

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Marsh, S.P., 1975, Geochemical and generalized geologic map showing distribution and abundance of chromium in the Nabesna quadrangle, Alaska: U.S. Geological Survey Miscellaneous Field Studies Map MF-655-E, 1 sheet, scale 1:250,000.

Marsh, S.P., 1975, Geochemical and generalized geologic map showing distribution and abundance of cobalt in the Nabesna quadrangle, Alaska: U.S. Geological Survey Miscellaneous Field Studies Map MF-655-F, 1 sheet, scale 1:250,000.

Marsh, S.P., 1975, Perspective diagrams showing geochemical abundance of silver, lanthanum, molybdenum, nickel, vanadium, yttrium, and zinc in the Nabesna quadrangle, Alaska: U.S. Geological Survey Miscellaneous Field Studies Map MF-655-G, 2 sheets, scale 1:250,000.

Mineral Occurrences:

Richter, D.H., Singer, D.A., and Cox, D.P., 1975, Mineral resources map of the Nabesna quadrangle, Alaska: U.S. Geological Survey Miscellaneous Field Studies Map MF-655-K, 1 sheet, scale 1:250,000.

Telegeology:

Albert, N.R.D., 1975, Interpretation of earth resources technology satellite imagery of the Nabesna quadrangle, Alaska: U.S. Geological Survey Miscellaneous Field Studies Map MF-655-J, 2 sheets, scale 1:250,000.

Other:

U.S. Geological Survey, 1975, Status of land in the Nabesna quadrangle, Alaska: U.S. Geological Survey Miscellaneous Field Studies Map MF-655-L, 1 sheet, scale 1:250,000.

Circular

Richter, D.H., Albert, N.R.D., Barnes, D.F., Griscom, Andrew, Marsh, S.P., and Singer, D.A., 1975, The Alaska Mineral Resource Assessment Program--Background information to accompany folio of geologic and mineral resource maps of the Nabesna quadrangle, Alaska: U.S. Geological Survey Circular 718, 11 p.

Geologic Map (colored)

Richter, D.H., 1976, Geologic map of the Nabesna quadrangle, Alaska: U.S. Geological Survey Miscellaneous Investigations Series Map I-932, scale 1:250,000.

## PETERSBURG QUADRANGLE

DATA RELEASES:

Stream-sediment geochemistry:

Cathrall, J.B., Day, G.W., Hoffman, J.D., and McDanal, S.K., 1983, A listing and statistical summary of analytical results for pebbles, stream sediments, and heavy-mineral concentrates from stream sediments, Petersburg area, southeast Alaska: U.S. Geological Survey Open-File Report 83-420-A, 281 p., 1 sheet, scale 1:250,000.

Bedrock geochemistry:

Karl, S.M., Koch, R.D., Hoffman, J.S., Day, G.W., Sutley, S.J., and McDanal, S.K., 1985, Trace element data for rock samples from the Petersburg, and parts of the Port Alexander and Sumdum quadrangles, southeastern Alaska: U.S. Geological Survey Open-File Report 85-146, 698 p., 1 sheet, scale 1:250,000.

Aeromagnetic map:

U.S. Geological Survey, 1979, Aeromagnetic map of Petersburg area, Alaska: U.S. Geological Survey Open-File Report 79-832, 1 sheet, scale 1:250,000.

Other:

Burrell, P.D., 1984, Map and table describing the Admirality-Revillagigedo intrusive belt plutons in the Petersburg 1:250,000 quadrangle, southeastern Alaska: U.S. Geological Survey Open-File Report 84-171, 6 p., 1 sheet, scale 1:250,000.

Burrell, P.D., Cobb, E.H., and Brew, D.A., 1982, Geologic bibliography of the Petersburg project area, Alaska: U.S. Geological Survey Open-File Report 82-483, 30 p.

U.S. Geological Survey, 1978, Aeroradioactivity of Kosciusko Island, Alaska: U.S. Geological Survey Open file Report 79-831, 1 sheet, scale 1:63,360.

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Geology:

Brew, D.A., Ovenshine, A.T., Karl, S.M., and Hunt, S.J., 1984, Preliminary reconnaissance geologic map of the Petersburg and parts of the Port Alexander and Sumdum 1:250,000 quadrangles, southeastern Alaska: U.S. Geological Survey Open-File Report 84-405, 43 p., 2 sheets, scale 1:250,000.

Geochemistry:

Cathrall, J.B., Day, G.W., Hoffman, J.D., and McDanal, S.K., 1983, Distribution and abundance of copper, determined by spectrographic analysis, in the minus-80-mesh fraction of stream sediments, Petersburg area, southeast Alaska: U.S. Geological Survey Open-File Report 83-420-B, 1 sheet, scale 1:250,000.

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Report 83-420-C, 1 sheet, scale 1:250,000.

Cathrall, J.B., Day, G.W., Hoffman, J.D., and McDanal, S.K., 1983, Distribution and abundance of lead, determined by spectrographic analysis, in the minus-80-mesh fraction of stream sediments, Petersburg area, southeast Alaska: U.S. Geological Survey Open-File Report 83-420-D, 1 sheet, scale 1:250,000.

Cathrall, J.B., Day, G.W., Hoffman, J.D., and McDanal, S.K., 1983, Distribution and abundance of lead, determined by spectrographic analysis, in the nonmagnetic fraction of heavy-mineral concentrates from stream sediments, Petersburg area, southeast Alaska: U.S. Geological Survey Open-File Report 83-420-E, 1 sheet, scale 1:250,000.

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## TALKEETNA QUADRANGLE

DATA RELEASES:

Geochemistry:

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by Warren J. Nokleberg, W. Michael Wade, Ian M. Lange, George Plafker and David L. Campbell

Nature and timing of deformation along the Contact Fault system in the Cordova, Bering Glacier, and Valdez quadrangles,
by George Plafker, Warren J. Nokleberg, John S. Lull, Sarah M. Roeske, and Gary R. Winkler

STRUCTURAL GEOLOGY/TECTONICS (continued)



VOLCANOLOGY

# AUTHORS




## ACKNOWLEDGMENTS

James R. LeCompte, Menlo Park, California, edited the articles for conformity with USGS standards of stratigraphic nomenclature.

Kathleen Krafft, Reston, Virginia, carefully read the final copy for errors and editorial conformity.

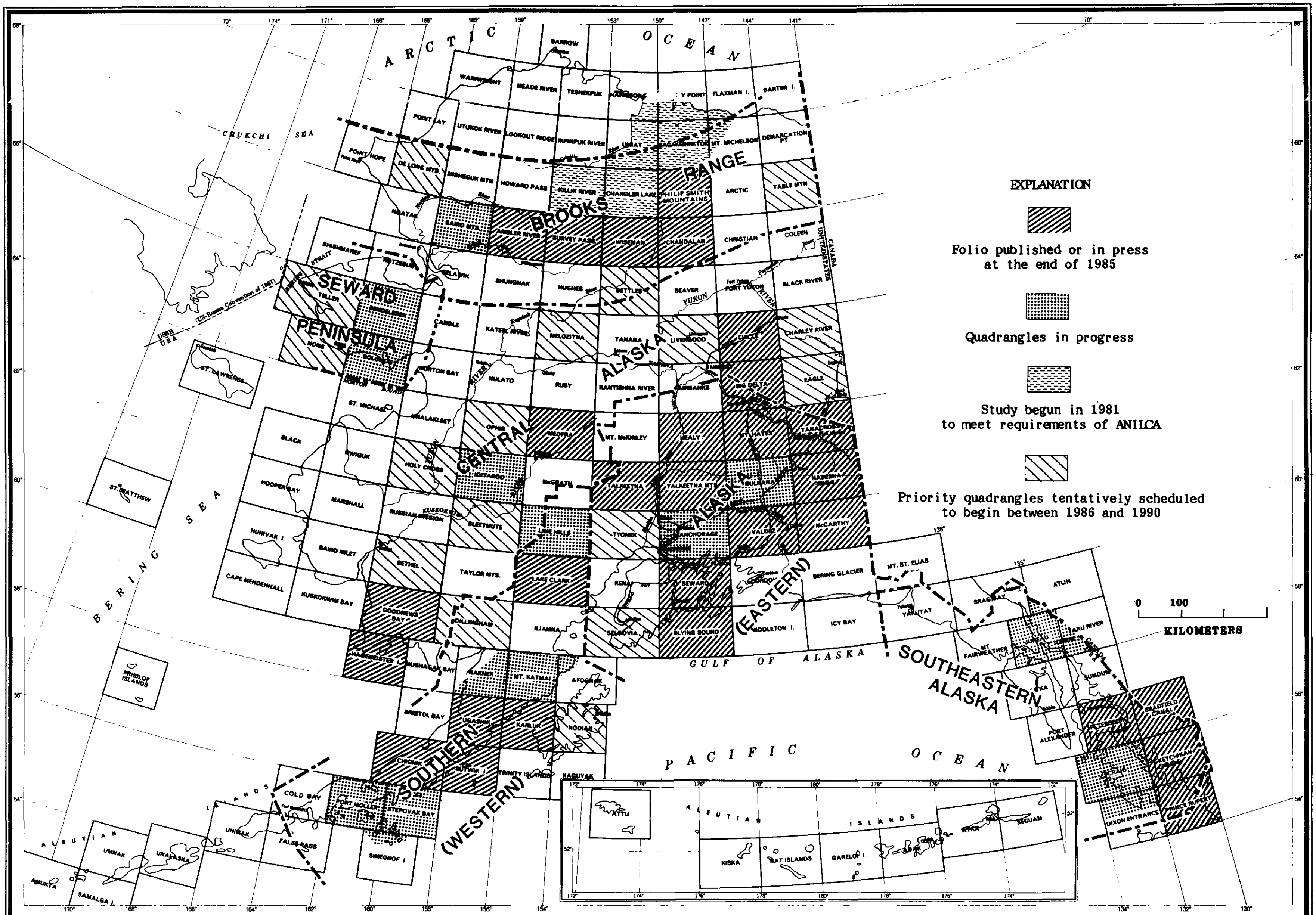
Audrey G. Schmidt, Reston, executed the cover design.

Virginia Shepard and Jane Marks, Anchorage, Alaska, managed telecommunications between Anchorage, Menlo Park, Denver, Colorado, and Reston.





*U.S. GOVERNMENT PRINTING OFFICE: 1986- 491-405:20121


ARCTIC OCEAN
CHUKCHI SEA
BERING SEA
BERING STRAIT
GULF OF ALASKA
PACIFIC OCEAN
ALEUTIAN ISLANDS
BROOKS RANGE
SEWARD PENINSULA
CENTRAL ALASKA
(EASTERN)
SOUTHERN (WESTERN)
SOUTHEASTERN ALASKA
UNITED STATES
CANADA
USSR
USA
(US-Russia Convention of 1867)
YUKON RIVER
KUSKOKWIM RIVER
BARROW
WAINWRIGHT
MEADE RIVER
TESHEKPUK
HARRISON BAY
BEECHEY POINT
FLAXMAN I.
BARTER I.
POINT LAY
UTUKOK RIVER
LOOKOUT RIDGE
IKPIKPUK RIVER
UMIAT
SAGAVANIRKTOK
MT. MICHELSON
DEMARCATION PT
POINT HOPE
DE LONG MTS.
MISHEGUK MTN
HOWARD PASS
KILLIK RIVER
CHANDLER LAKE
PHILIP SMITH MOUNTAINS
ARCTIC
TABLE MTN.
NOATAK
BAIRD MTS.
AMBLER RIVER
SURVEY PASS
WISEMAN
CHANDALAR
CHRISTIAN
COLEEN
SHISHMAREF
KOTZEBUE
SELAWIK
SHUNGNAK
HUGHES
BETTLES
BEAVER
FORT YUKON
BLACK RIVER
TELLER
BENDELEBEN
CANDLE
KATEEL RIVER
MELOZITNA
TANANA
LIVENGOOD
CIRCLE
CHARLEY RIVER
NOME
SOLOMON
NORTON BAY
NULATO
RUBY
KANTISHNA RIVER
FAIRBANKS
BIG DELTA
EAGLE
ST. LAWRENCE
ST. MICHAEL
UNALAKLEET
OPHIR
MEDFRA
MT. McKINLEY
HEALY
MT. HAYES
TANACROSS
BLACK
KWIGUK
HOLY CROSS
IDITAROD
McGRATH
TALKEETNA
TALKEETNA MTS.
GULKANA
NABESNA
ST. MATTHEW
HOOPER BAY
MARSHALL
RUSSIAN MISSION
SLEETMUTE
LIME HILLS
TYONEK
ANCHORAGE
VALDEZ
McCARTHY
NUNIVAK I.
BAIRD INLET
BETHEL
TAYLOR MTS.
LAKE CLARK
KENAI
SEWARD
CORDOVA
BERING GLACIER
MT. ST. ELIAS
CAPE MENDENHALL
KUSKOKWIM BAY
GOODNEWS BAY
DILLINGHAM
ILIAMNA
SELDOVIA
BLYING SOUND
MIDDLETON I.
ICY BAY
YAKUTAT
SKAGWAY
ATLIN
MT FAIRWEATHER
JUNEAU
TAKU RIVER
PRIBILOF ISLANDS
HAGEMEISTER I.
NUSHAGAK BAY
NAKNEK
MT. KATMAI
AFOGNAK
SITKA
SUMDUM
BRISTOL BAY
UGASHIK
KARLUK
KODIAK
PETERSBURG
BRADFIELD CANAL
PORT ALEXANDER
CHIGNIK
SUTWIK I.
TRINITY ISLANDS
KAGUYAK
CRAIG
KETCHIKAN
COLD BAY
PORT MOLLER
STEPOVAK BAY
DIXON ENTRANCE
PRINCE RUPERT
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AMUKTA
SAMALGA I.
ATTU
KISKA
RAT ISLANDS
GARELOI I.
ADAK
ATKA
SEGUAM
EXPLANATION
Folio published or in press at the end of 1985
Quadrangles in progress
Study begun in 1981 to meet requirements of ANILCA
Priority quadrangles tentatively scheduled to begin between 1986 and 1990
0 100
KILOMETERS

GPO 836762