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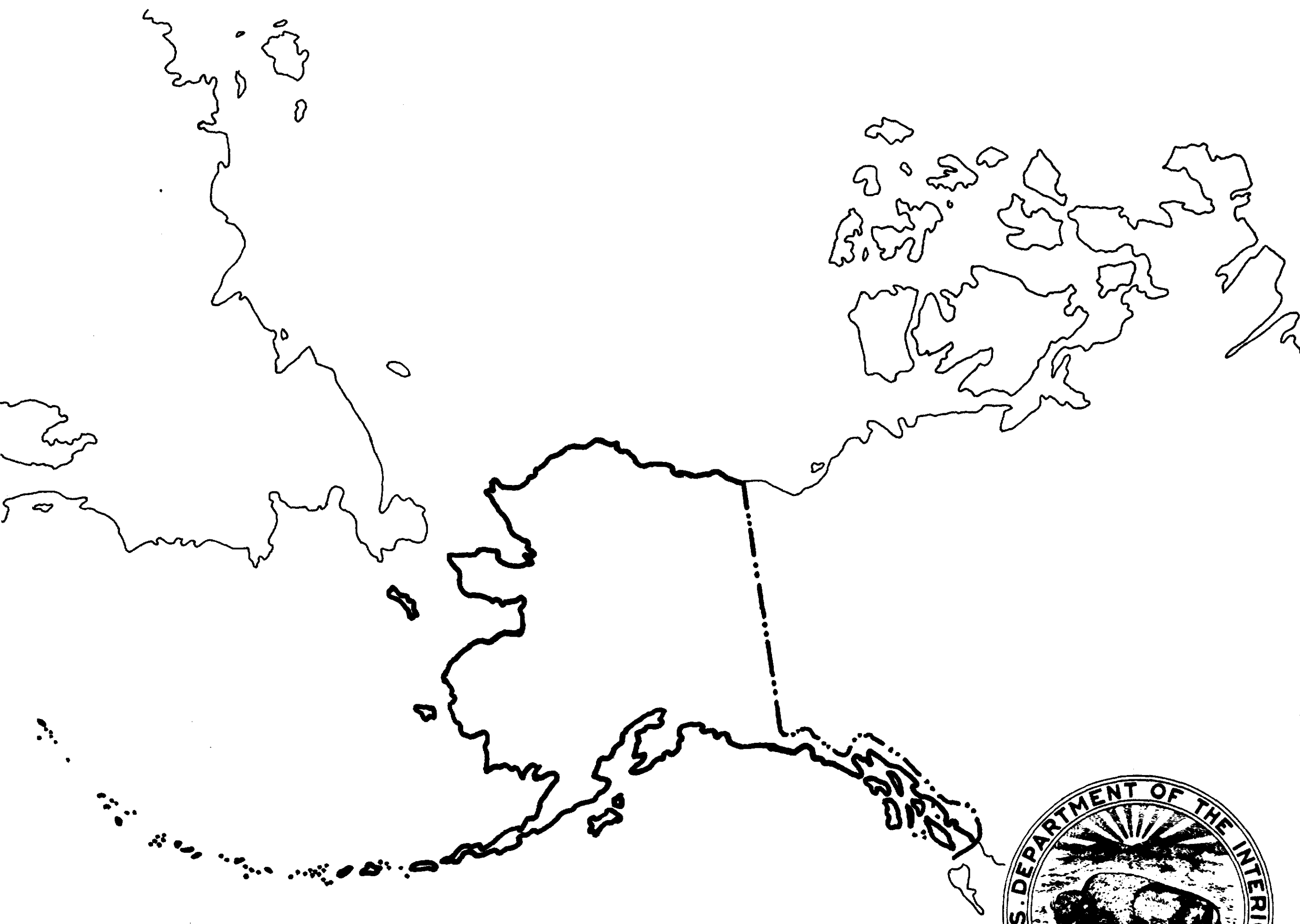
THE UNITED STATES GEOLOGICAL SURVEY IN ALASKA: ACCOMPLISHMENTS DURING 1982



U.S. Geological Survey Circular 939

The United States Geological Survey in Alaska: Accomplishments During 1982

Katherine M. Reed and Susan Bartsch-Winkler, Editors

U.S. GEOLOGICAL SURVEY CIRCULAR 939

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THE UNITED STATES GEOLOGICAL SURVEY IN ALASKA: ACCOMPLISHMENTS DURING 1982

Katherine M. Reed and Susan Bartsch-Winkler, Editors

ABSTRACT

This circular contains short topical and summary articles about the results of 1982 geologic studies on a wide range of subjects of economic and scientific interest. Included are lists of references cited for each article and a compilation of reports about Alaska written by members of the U.S. Geological Survey and published by the Geological Survey and other organizations.

INTRODUCTION

The U.S. Geological Survey investigates many aspects of the land and water in Alaska. Reports providing basic data and describing the results and significance of Geological Survey studies appear in both Geological Survey and other scientific publications. This circular contains brief discussions of the results of some Alaskan studies in 1982. The articles are arranged by geographic area, with a section for offshore topics. Figure 1 indicates the boundaries of these areas and the quadrangles included in each. Index maps, figures 2, 14, 22, 27, 49, 52, and 72, show the locations of the areas covered by the articles. The 1982 publications by Geological Survey scientists about Alaskan geology are listed at the back of the circular. Also included is a cross-index of subject matter in the text.

NORTHERN ALASKA

(Figure 2 shows study areas described.)

GRAVITY MEASUREMENTS SHOW LARGE SIZE OF RED DOG ZINC-LEAD-BARITE PROSPECT IN NORTHWESTERN ALASKA

By David F. Barnes and Robert L. Morin

Detailed gravity measurements over outcrops of the Red Dog zinc-lead-barite prospect in northwestern Alaska demonstrate the usefulness of gravimetry in preliminary evaluation of high-density-ore prospects. Measurements at the nearby Nimiuktuk barite deposit (Barnes, 1981; Barnes and others, 1982) had suggested that 1 1/2 days of surveying and gravimetry could provide a preliminary estimate of the tonnage and even a judgment of grade of that relatively small deposit. The Red Dog measurements were undertaken as a test of similar techniques on this larger, better known deposit, which has different and more complex mineralogy, is in more rugged topographic terrain, and which has already been outlined by drilling and sampling. The measurements were planned as a test of using gravimeter techniques to evaluate mineral deposits in remote areas where access, drilling, trenching, and large camps may be restricted. Cominco Alaska Inc., which is making a mining feasibility study of the prospect, provided lodging, transportation, and survey assistance; the hospitality and assistance of Gerry G. Booth, Richard Van Blaricom, Louis J. O'Connor, Joseph T. Plahuta, James B. Lincoln, Daniel P. Samson, and others of Cominco Alaska Inc. are gratefully acknowledged.

Tailleur (1970) first reported evidence of mineralization at the Red Dog prospect, and in 1975 and 1976 a contractor for the U.S. Bureau of Mines (Degenhart and others, 1978), after 18 man-days of study, recognized the deposit's potential size and grade and recommended drilling and further work. Later studies by Plahuta and others (1978) preceded exploratory drilling by Cominco in 1980 through 1982, which outlined 85 million tons of ore and led to a development agreement with the property owners (Collie, 1982). Although much is now known about the prospect after completion of more than 500 m of drilling, the 1982 gravimeter survey was designed to test an early-evaluation technique. This preliminary report relies primarily on data that were or could be obtained before the start of a drilling program, although some of the drilling results are mentioned here as corroboration or discussion. However, project planning benefited greatly from available knowledge of geologic mapping, test hole distribution, earlier geophysical surveys (mostly electromagnetic), and other products of preliminary exploration and study. Furthermore, the availability of an excellent grid of horizontal and vertical survey control and the assistance of Cominco surveyors in using and extending this grid enabled us to concentrate on gravity measurements without having to establish elevation control.

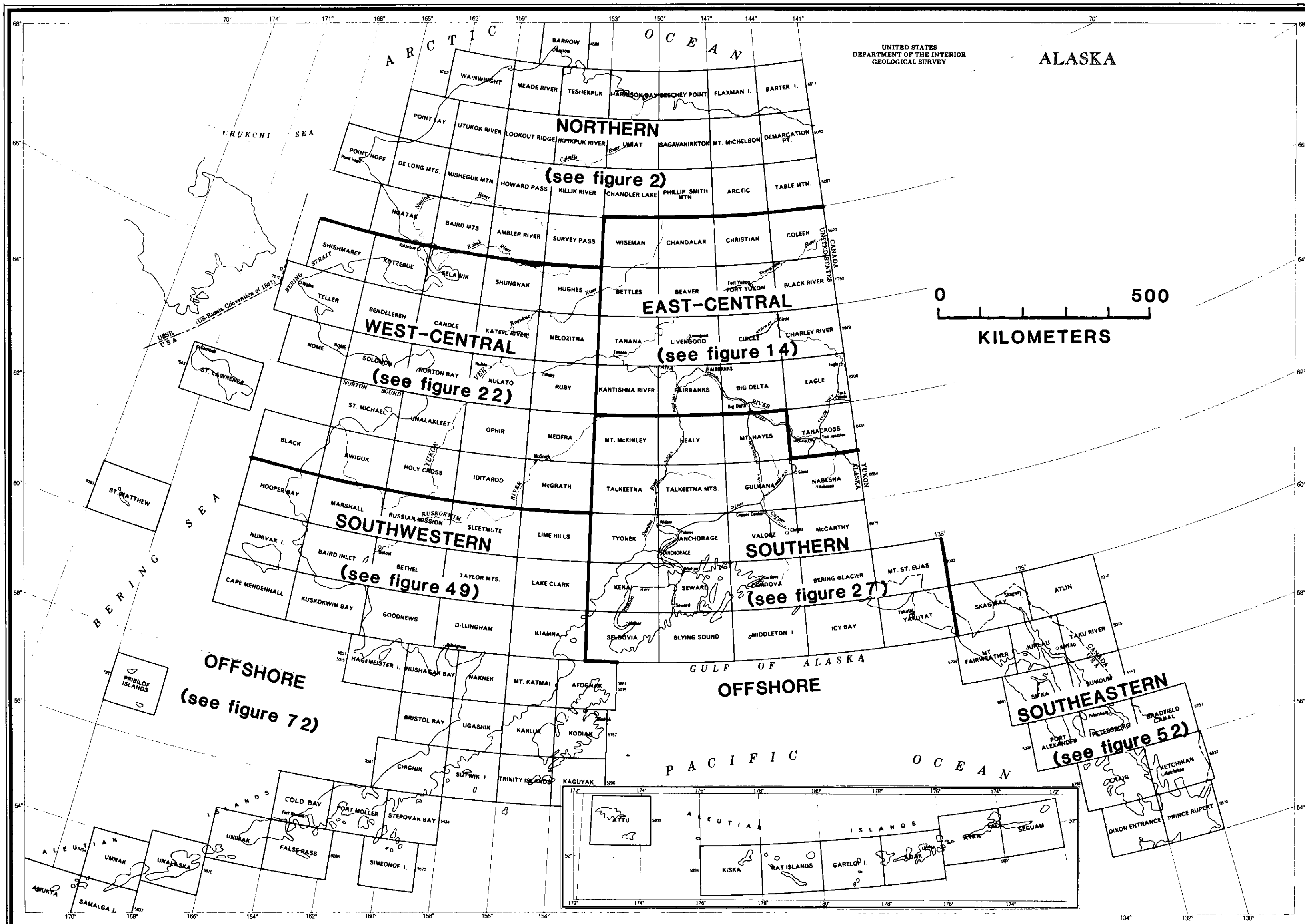

Figure 1.—Regions of Alaska used in this circular. Referenced figures indicate regional index maps that show areas discussed.

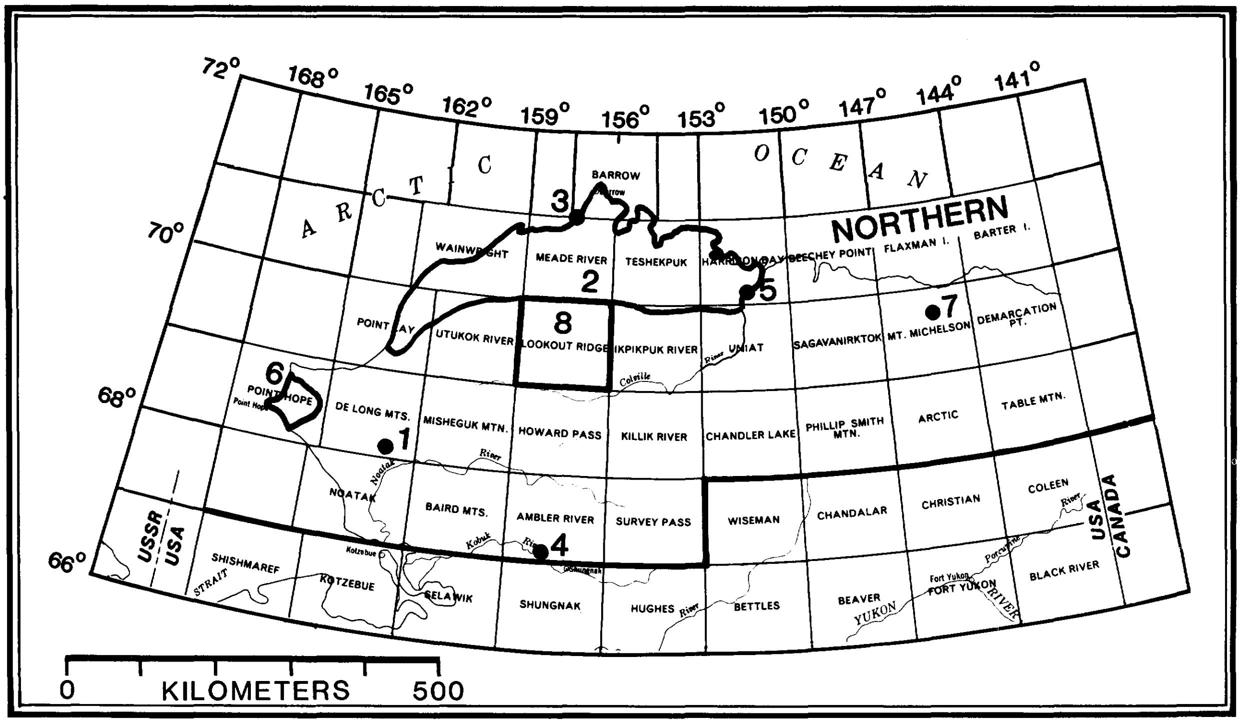

Figure 2.—Areas in northern Alaska discussed in this circular. A listing of authors and the inclusive pages of their articles follows. (1) Barnes and Morin, p. 1-5; (2) Brigham, p. 5-9; (3) Brouwers, Marincovich, and Hopkins, p. 9-12; (4) Hamilton and others, p. 12-15; (5) Marincovich, Brouwers, and Carter, p. 15-17: (6) Moore and others, p. 17-21; (7) Sohn, p. 21-23; and (8) Yeend, p. 23-25.

The Red Dog zinc-lead-barite deposit (lat. 68°04'N; long. 162°51'W; see fig. 3) consists of massive, stratiform sulfide mineralization and associated vein and breccia fillings in marine shales and cherts of the Lower Carboniferous Lisburne Group. Plahuta and others (1978) suggested a syngenetic origin by hydrothermal exhalative solutions involving a possible volcanic source. The nearly flat-lying deposit is cut by a deep creek channel that provides several exposures for outcrop study and sampling. Adjacent hilltops have summits 100 to 300 m above the creek channel, with intermediate slopes as steep as 20°. However, gravitational terrain effects neither mask the anomalies associated with the mineralization nor hamper their preliminary discussion before terrain corrections are complete.

The preliminary simple Bouguer gravity contours shown in figure 3 are based on some of nearly 250 measurements made by two men during six days of inclement weather in July 1982. To obtain regional control, most of the profiles extended significantly beyond the limits of the figure to total profile lengths of more than 2 km, and scattered measurements were made on nearby hilltops and other locations where positions and elevations relative to Cominco's grid were known. The grid consisted of staked points at intervals of 122 m (400 ft) both north-south and east-west. For gradient control, intermediate stations were established at intervals of 30.5 m (100 ft) or 61 m (200 ft) on a few east-west profiles across the deposit. The new survey control was established with an electronic survey transit containing a laser range-finder and built-in calculator, enabling two men to establish survey control in about half the time required for the gravity measurements. Both gravity and position surveys were made on foot, although helicopters provided transportation between the base camp and the deposit. Elevation datum was obtained from nearby vertical-angle bench marks, the gravity datum from the Alaskan gravity base station network (Barnes, 1968) updated to IGSN-71, and data reduction was based on a standard density of 2.67 g/cm^3, which was closely supported by 20 hand-specimen densities averaging 2.62 g/cm^3 for the country rock.

The 0.5-mGal contours show two gravity highs with amplitudes of about 2 and 4 mGal respectively. Terrain corrections for stations in the stream channel may tend to raise the gravity values measured between the highs and to eliminate the intermediate saddle, but the anomalies now seem fairly distinct. Planimetry of these contours and

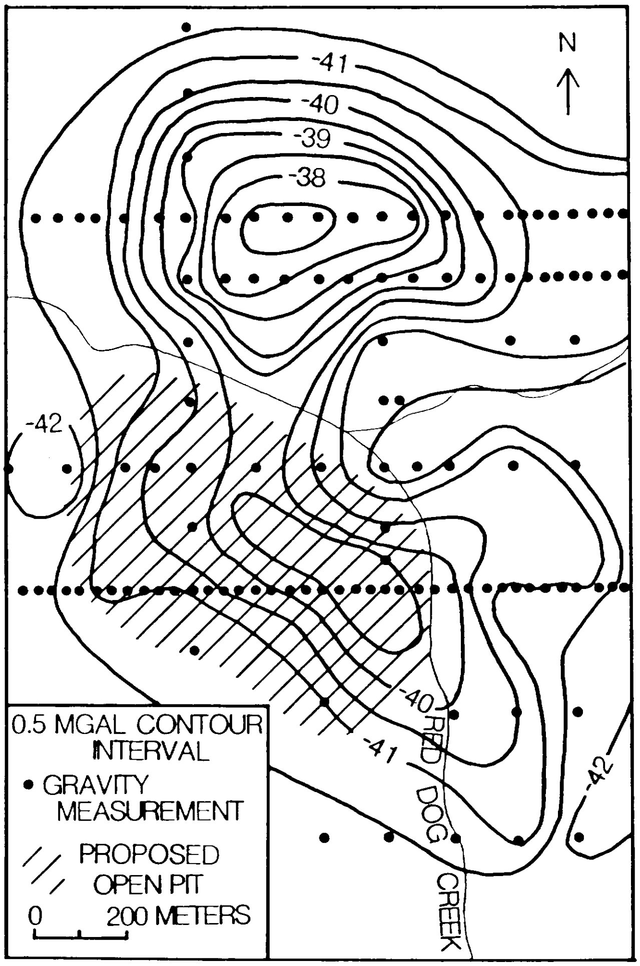

Figure 3.—Gravity contours at Red Dog prospect.

the use of Gauss' theorem (Nettleton, 1976) suggest that the anomaly south of the creek represents an additional mass of about 22 million tons and that north of the stream about 30 million tons. The actual tonnage depends on the ratio between the densities of the causative body and its country rock. There is sufficient difference in the mineralogy of outcrops south and north of the creek to suggest that the two anomalies should be considered separately.

Most of the hand specimens from north of the creek suggest that the predominant mineralization is barite, although deeper parts of the deposit may include other minerals. Ten barite hand specimens had densities ranging from 3.37 to 4.42 g/cm^3 with a mean of 4.07 g/cm^3. If this density is typical of the whole northern part of the deposit, its total tonnage probably exceeds 100 million tons of barite. Only a few holes have been drilled in the deposit, so this estimated tonnage cannot be verified.

Outcrops close to and south of the creek yielded many metalliferous specimens with high proportions of sphalerite, galena, and barite and a fairly wide range of densities. Eighteen hand specimens of this ore had densities ranging from 2.86 to 3.99 g/cm^3 with a mean of 3.81 g/cm^3. If such densities are typical of the southern deposit, it may contain more than 65 million tons of ore. Preliminary drilling reports (Collie, 1982) suggested an estimated tonnage of 85 millions tons with a grade of 17 percent zinc and 5 percent lead, but estimates of lower grade tonnages have not been published. The metalliferous grade of the hand specimens used for the density estimate is unknown. However, 18 rock and talus samples from this part of the deposit were assayed by Degenhart and others (1978), and the average mineralization of these samples was 11 percent zinc, 3 percent lead, and 34 percent barite, a mixture that would have a density close to that of the hand specimens. Thus, the ore from the drill holes is richer than that sampled on the surface. Furthermore, the data of Degenhart and others (1978) also suggest that the samples richest in zinc, lead, and silver contain less barite and thus have a lower density. Barite-free ore of the grade reported in the preliminary drilling would have a density of about 3.35 g/cm^3 and would suggest that the total tonnage of the southern deposit might exceed 100 million tons. Final quantitative evaluation of the gravimetric technique may require terrain corrections and better evaluation of density data, but the preliminary results are encouraging.

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Reviewed by Robert C. Jachens and Irvin L. Tailleur

MARINE STRATIGRAPHY AND AMINO ACID GEOCHRONOLOGY OF THE GUBIK FORMATION, WESTERN ARCTIC COASTAL PLAIN, ALASKA

By Julie K. Brigham

The Pliocene(?) and Pleistocene Gubik Formation consists of unconsolidated deposits mantling the Arctic Coastal Plain of northern Alaska (fig. 4) and includes marine, fluvial, eolian, and thaw lake facies (Black, 1964). Based on detailed stratigraphic study and amino acid geochronology between Point Barrow and Cape Beaufort, the beach, nearshore, and shallow-marine shelf deposits of this region have been subdivided into transgressive-regressive sequences representing at least six distinct high stands of sea level (table 1). The extent of epimerization of isoleucine (Ile) to alloisoleucine (aIle), as determined on enclosed fossil mollusks from the Gubik deposits, produced distinct ratios (aIle/Ile) that differentiate lithologically similar transgressive deposits. The results were derived from high performance liquid chromatography ion-exchange analysis of six genera: Hiatella, Mya, Macoma, Astarte, Siliqua, and Cyrtodaria. Only the results of analyses of H. artica, Mya truncata, and Mya arenaria are presented here (table 1).

The oldest unit recognized, of Pliocene(?) age, consists of a basal cobble gravel and silty clay enclosing mollusks that yield ratios of 0.82 in the free (naturally hydrolized) fraction and 0.216 in the total acid hydrolyzate (free plus peptide-bound amino acids) (Pliocene(?); table 1). Generally, the unit is less than 0.5 m thick and is laterally discontinuous where it overlies the Cretaceous bedrock of the Nanushuk Group (Molenaar, 1981) along Skull Cliff (fig. 4). Striated erratic cobbles and boulders, probably ice rafted, occur along the basal unconformity. No morphological evidence remains to indicate the spatial extent of this transgression inland across the coastal plain. Because it is the oldest unit recognized, this event may be correlative with the Beringian transgression (Hopkins, 1967, 1973) along the western coast of Alaska, which Hopkins now believes to have taken place about 2.5 m.y.B.P. (D. M. Hopkins, U.S. Geological Survey, written commun., 1983). The aIle/Ile ratios of samples collected from transgressive units at Nome (provided by D. M. Hopkins and analyzed by G. H. Miller, Univ. of Colorado, in 1976) are shown in table 1. The higher

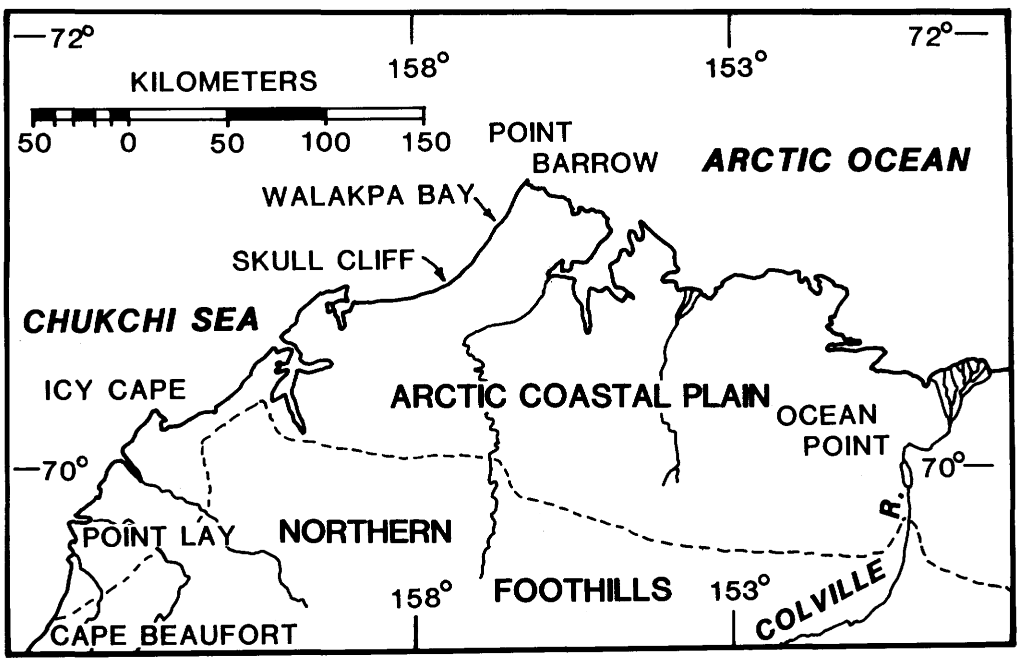

Figure 4.—Northern Alaskan coastal plain showing localities mentioned in the text. The dashed line indicates the boundary between physiographic subdivisions of northern Alaska.

Table 1.--Comparison of aIle/Ile ratios in molluscs representing marine transgressions of similar age in northern and western Alaska*

Horizon	Species**	aIle/Ile Free	aIle/Ile Total	Horizon	Species**	aIle/Ile Free	aIle/Ile Total
WESTERN ARCTIC COASTAL PLAIN				NOME, WESTERN ALASKA			
FLAXMAN FM.		No data					
PELUKIAN	Ha	ND***	0.014 ± .002	PELUKIAN	Ha	0.16 ± .001	0.040 ± .002
	Ma	ND	0.018 ± .002		Mt	0.28 ± .03	0.042 ± .005
MIDDLE PLEISTOCENE	Ha	0.40 ± .052	0.038 ± .007	KOTZEBUAN	Mt	0.46 ± .04	0.093 ± .001
	Mt-Ma	0.37 ± .085	0.038 ± .006				
EARLY PLEISTOCENE	Ha	0.51 ± .023	0.086 ± .009				
	Mt-Ma	0.37 ± .057	0.086 ± .013				
LATE PLIOCENE(?)	Ha	0.50 ± .072	0.15 ± .007	ANVILIAN	Ha	0.90 ± .02	0.47 ± .03
	Mt-Ma	0.52 ± .057	0.15 ± .023				
PLIOCENE(?)	Ha	0.82	0.22	BERINGIAN	Ha	1.04 ± .03	0.56 ± .07
	Ma	0.84	0.23		Mt	0.95 ± .03	0.56 ± .04

* Ratios represent the group mean ± 1 of peak height measurements. A single value indicates only 1 valve was analyzed.

** Ha = Hiatella arctica, Mt + Mya truncata, Mt-Ma = Mya truncata and Mya arenaria results combined

*** ND = Not detectable

absolute value of ratios from samples near Nome reflects the warmer effective diagenetic temperatures at that latitude.

Stratigraphic and paleontologic evidence, as well as amino acid data, indicate the existence of a younger high sea stand. In the western coastal plain, this transgression (late Pliocene, table 1) is probably represented by discontinuous shallow-marine deposits that crop out along Skull Cliff and either overlie the older Pliocene(?) sediments or rest directly on bedrock. In contrast, correlative marine deposits (tentative correlation based on faunal content and amino acid data) are continuously exposed at Ocean Point along the Colville River (Carter and Galloway, 1981). These sediments are believed to be about 2 m.y. old, based on identification of fossil sea otter remains from the Ocean Point exposures (Repenning, 1983). As with the older Pliocene(?) sediments, no correlative evidence has been found to indicate the southern extent of this transgression. The high sea stand may be correlative with the Anvilian transgression (Hopkins, 1967) as suggested by Carter and Galloway (1981), which Hopkins now believes to be about 2.2 m.y. old (D. M. Hopkins, written commun., 1983).

Shallow marine deposits of an early Pleistocene transgression (early Pleistocene, table 1) are extensively exposed along Skull Cliff and contain mollusks for which ratios are distinctly lower (i.e., are younger) than the two Pliocene(?) units. These deposits are locally folded and contorted (fig. 5) and are commonly truncated by an unconformity. Typically, they contain open-marine ostracode and molluscan taxa, including specimens of Natica janthostoma Deshayes, a gastropod presently limited to the northwest Pacific. Collections also include the ostracode Rabilimus paramirabilis (Brouwers, Marincovich, and Hopkins, this volume) and the gastropod Neptunea (Neptunea) lyrata leffingwelli, both of which are now extinct and known only in the late Pliocene and early Pleistocene marine strata of western Alaska. Ice-rafted erratic boulders are also common at the base of the unit. The inland extent of this transgression is not well defined, but shells of Hiatella arctica having similar alle/Ile ratios were collected from a map elevation of about 60 m above sea level (asl) from gravels exposed 25 km inland along the Kukpowruk River.

Shallow marine deposits of middle Pleistocene age unconformably overlie the older units (fig. 5) and are characterized by shells that have lower alle/Ile ratios. These beds are exposed in the upper half of the bluffs along Skull Cliff and are correlative with fossiliferous beach deposits found up to 30-33 m asl along the western edge of the coastal plain inland from Icy Cape and Point Lay. All ostracodes, foraminifers, and marine vertebrate and mollusk assemblages examined from these sediments are similar to the modern arctic faunas. Based on theoretical age estimates from the amino acid data, the absolute age of this transgression appears to be certainly greater than 200,000 yr and may be as old as 500,000 yr B.P. By inference, this event is probably correlative with the Kotzebuan transgression (Hopkins, 1973) which has been radiometrically dated to within that age range.

The last interglacial high sea stand, about 125,000 yr B.P. and correlative to the Pelukian transgression along western Alaska (Hopkins, 1973), is probably responsible for a prominent shoreline with gravel beach and lagoonal deposits up to 10-12 m asl. Due to the cold thermal history of the region (current mean annual temperature at Barrow, -12.6° C), isoleucine epimerization ratios are only slightly above those in modern samples (about 0.011). A radiocarbon analysis of wood from these deposits yielded a minimum age of greater than 36,000 yr B.P. (Beta-1766). Likewise, a date on marine shells of 31,200 $\pm$ 810/900 yr B.P. (DIC-2569) is considered to be a minimum estimate. (Compare with Carter and Robinson, 1980).

The youngest transgressive unit recognized on the coastal plain is the Flaxman Formation, believed to date from some latter part of ^{18}O stage 5, 90,000 or 105,000 yr B.P. (D. M. Hopkins, written commun., 1982). Flaxman sediments consist of marine silt and clay that locally contain large ice-rafted boulders. These deposits are present up to 7 m asl along the Beaufort coast but do not occur west of Barrow along the Chukchi Sea coast; hence no amino acid data are reported. An erosional bench at approximately 6 m asl, observed in Walakpa Bay, may have been formed contemporaneously with the Flaxman deposits. The lack of land area less than 8 m asl along the Chukchi Sea coast may have precluded the preservation of Flaxman deposits. Any sediments that may have been deposited offshore appear to have been eroded and reworked by the Holocene transgression (L. Phillips, U.S. Geological Survey, written commun., 1982).

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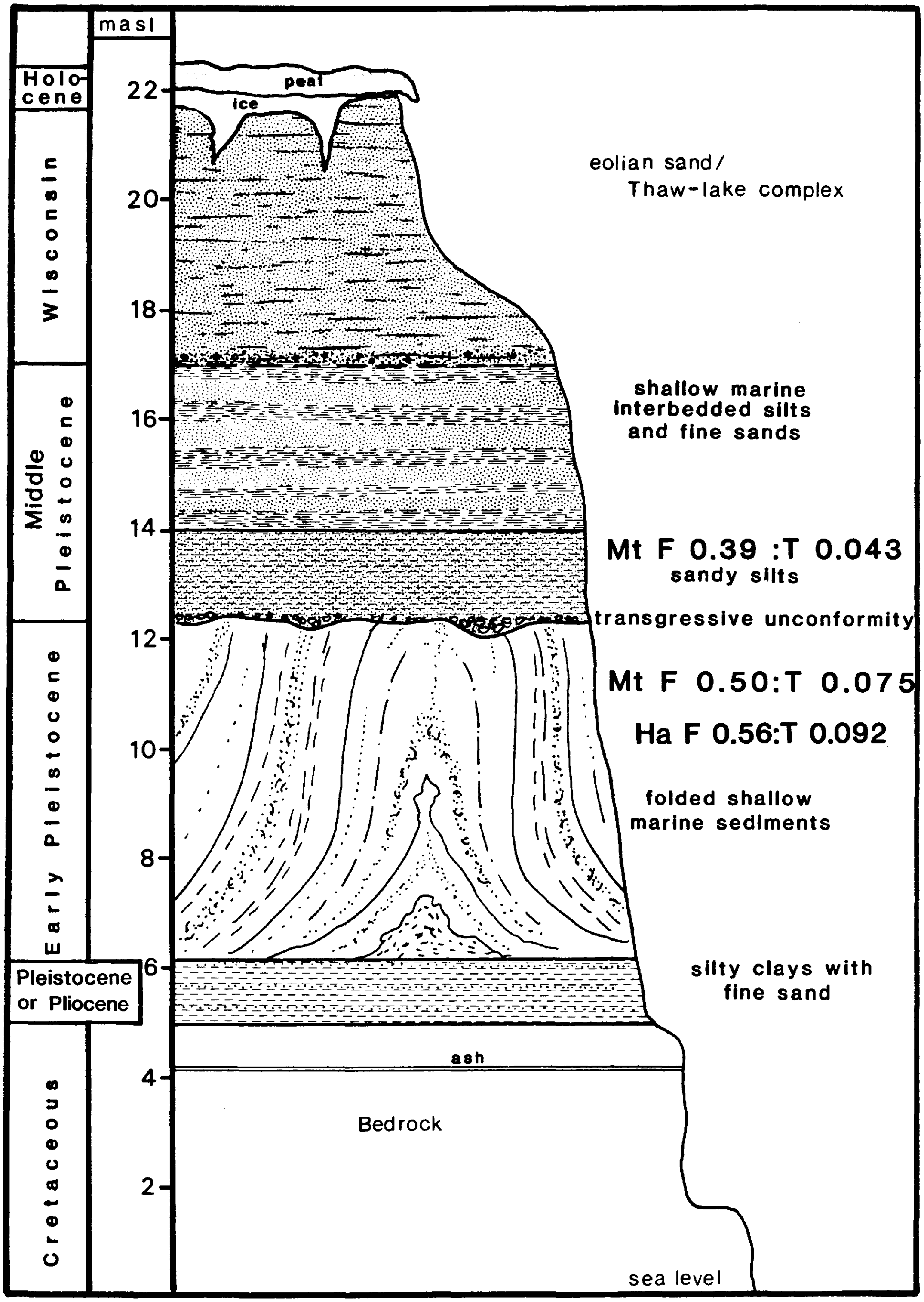

Figure 5.—Stratigraphy of the Gubik Formation at one section along Skull Cliff. (Compare with Williams, 1979). Isoleucine epimerization ratios in the Free (F) and Total (T) indicate a significant age difference between units. Ratios observed here are within the standard deviation of values listed on table 1 for sediments of this age. Mt, Mya truncata; Ha, Hiatella arctica.

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Reviewed by L. David Carter and David M. Hopkins

PALEOENVIRONMENTAL RECORD OF PLEISTOCENE TRANSGRESSIVE EVENTS PRESERVED AT SKULL CLIFF, NORTHERN ALASKA

By Elisabeth Brouwers, Louie Marincovich, Jr., and David M. Hopkins

Eustatic changes in sea level during the late Pliocene(?) and the Pleistocene are well recorded in the coastal areas and on the shallow continental shelf of northern Alaska. Records of transgressive events are preserved as terraces, wave-cut cliffs, beach ridges, and deposits of fossiliferous marine sediments. Hopkins (1967) proposed seven marine transgressions that are thought to be represented in coastal and offshore sediments of northern and western Alaska. He characterized each transgression on the basis of the position of sea level (altitude of shoreline), paleoclimate, radiometric dating, and faunal and floral content (particularly mollusks and pollen). Subsequent studies, especially those conducted in northern Alaska by J. K. Brigham, L. D. Carter, D. M. Hopkins, R. E. Nelson, K. A. McDougall, C. A. Repenning, P. A. Smith, J. R. Williams, and the present authors, have resulted in considerable modification and refinement of Hopkins' original scheme. Table 2 presents Hopkins' revised assessment of the chronology of late Cenozoic marine transgressions in western and northern Alaska.

In northern Alaska, younger Pleistocene deposits that preserve records of the Flandrian, Flaxman Formation, and Pelukian transgressions are fairly well known and their faunas described from coastal exposures and offshore boreholes (Hopkins and others, 1980). However, the older Pleistocene (Kotzebuan, Fishcreekan) and Pliocene (Anvilian, Beringian) transgressive events are not as well documented. These sediments are best exposed and preserved along the Chukchi Sea coast and at scattered localities inland on the Arctic Coastal Plain west of the Colville River (Hopkins, 1967; McCulloch, 1967).

Older Pleistocene marine sediments crop out in nearly continuous exposures from Peard Bay 80 km northeast to Point Barrow along the Chukchi Sea coast. The best exposed, most complete section is at Skull Cliff (fig. 6). The Gubik Formation here was divided into two units by Leffingwell (1919) and Meek (1923), the lower unit named the Skull Cliff unit and the upper unit named the Barrow unit by Black (1964). Reconnaissance studies by Williams (1979) and work by Brigham (1981a, 1981b, 1984) documented a more complex stratigraphic sequence; based on amino acid geochronology, five depositional sequences representing high sea-level stands of probable Pliocene and Pleistocene age are identified at Skull Cliff.

The oldest recognized unit of the Gubik Formation at Skull Cliff is a late Pliocene or Pleistocene cobble gravel and silty clay, which discontinuously overlies the Cretaceous bedrock. Amino acid racemization ratios suggest that this unit probably represents the Anvilian transgression. D. M. Hopkins (U.S. Geological Survey, written commun., 1983) believes that some Beringian sediments may also be present in Skull Cliff as 1- to 2-m thick remnants of a basal unit.

The oldest known Quaternary deposits exposed at Skull Cliff consist of highly folded beds of marine, fine-grained sand and silt containing numerous mollusks; these sediments formed the basal third of Black's (1964) Barrow unit. Amino acid racemization ratios suggest that the folded beds are early Pleistocene in age, correlative with the newly recognized Fishcreekan transgression (1.2 m.y.; L. D. Carter, unpub. data, 1982; Brigham, 1981a, 1981b; Brigham and Miller, 1982; Repenning, 1983). Molluscan assemblages in the deformed marine sands consist primarily of species that are endemic to the region today, as well as Neptunea (Neptunea) lyrata leffingwelli, an extinct gastropod, and Natica (Tectonatica) janthostoma, a gastropod now confined to the southwestern Bering Sea and the northwestern Pacific Ocean. Neither of these taxa occur in younger strata in northwestern Alaska. A small assemblage of ostracode species from the northeastern part of Skull Cliff similarly contains several extinct forms. Rabilimis paramirablis and Paracyprideis aff. P. pseudopunctillata have been documented in Beringian sediments and may range as young as Fishcreekan, but these species have not been collected from younger sediments.

Table 2.—Pliocene and Quaternary marine trangressions in Alaska as presently defined (Compiled by David M. Hopkins)

Name of transgression	Estimated age	Evidence for age	Changes from previous publications
FLANDRIAN	<18,000 yr	Radiocarbon dating of shelf cores; correlation with other areas	Kruzensternian Transgression of Hopkins (1967)
FLAXMAN FM.	90,000 to 105,000 yr	Correlation with oxygen-isotope curve in deep-sea cores; also beyond reach of radiocarbon dating in northern Alaska	Some deposits previously referred to non-existent mid-Wisconsinan Woronzofian Transgression of Hopkins (1967)
PELUKIAN	ca. 125,000 yr	Correlation with other areas and with oxygen-isotope curve in deep-sea cores	No change
KOTZEBUAN	ca. 240,000 yr	Correlation with oxygen-isotope curves; K-Ar dating on St. Paul Island, Pribilof Islands	Includes deposits of non-existent Einahnuhtan Transgression
FISHCREEKAN	1.2 m.y.	K-Ar dating on Pribilof Islands; amino acid geochronology	Newly named; previously confused with Anvilian deposits
ANVILIAN	2.2 m.y.	K-Ar dating on Pribilof Islands; stage of evolution of fossil sea otter at Ocean Point, Colville R.; amino acid geochronology	Previously thought to be 1.2 m.y. old on basis of K-Ar dating of Fishcreekan deposits on Pribilofs
BERINGIAN	ca. 2.5 m.y.	K-Ar dating on St. George Island, Pribilof Islands	Previously thought to be 2.2 m.y. on basis of K-Ar dating of Anvilian deposits on St. George Island

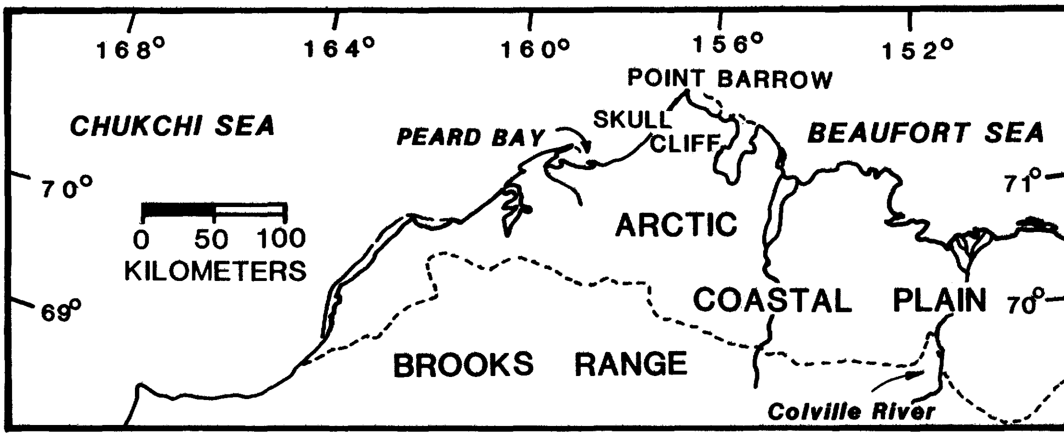

Figure 6.—Location of the Skull Cliff section along the western Arctic Coastal Plain.

Horizontal fossiliferous transgressive gravels unconformably overlie the folded sand-silt beds. Above this are flat-lying beds of sand and fine gravel with scattered mollusks, detrital coal beds, and chert pebbles (Williams, 1979). The gravel at the unconformity and the overlying sand and gravel formed the upper two-thirds of Black's Barrow unit. Amino acid racemization ratios suggest that these sediments are middle Pleistocene in age, correlative with the Kotzebuan transgression (about 240,000 yr, J. K. Brigham, INSTAAR, University of Colorado, oral commun., 1983). All the ostracode species recovered from this transgressive event are endemic to the region today. The ostracode assemblages from these Kotzebuan sediments contain a diverse group of species that appear to represent several progressively seaward facies of the inner neritic zone. The presence of a number of euryhaline and eurythermal species, including Rabilimis septentrionalis, Cytheromorpha spp., Heterocyprideis, Loxoconcha venepidermoidea, and Paracyprideis, indicates very shallow, nearshore conditions. Other species prefer a habitat with less fluctuating salinity conditions such as exist farther offshore; these species include Normanicythere leioderma, Robertsonites tuberculata, and Palmanella limicola.

A comparison of the Kotzebuan ostracode assemblage from Skull Cliff with the assemblage described from Anvilian sediments from the western Seward Peninsula (Hopkins and others, 1974) shows clear differences in the paleoenvironments. The Kotzebuan fauna is comprised of cold water, Arctic species that are endemic today to the Beaufort and northern Chukchi Seas. The Anvilian fauna is quite similar to faunas that occur in Pelukian deposits. Several Atlantic endemics are present, as well as several species that imply somewhat warmer water conditions than exist today.

The youngest transgressive event recorded along the Skull Cliff exposures is the Pelukian transgression, which has been dated as about 125,000 yr B.P. by amino acid geochronology and correlation with the oxygen isotope sequence. These sediments consist of interbedded sands and gravel with some peaty silt. The ostracode assemblages contain a number of endemic Atlantic species (Cytheropteron pyramidale, Eucytheridea macrolaminata, Finmarchinella angulata), as well as several species (Cythere lutea, Baffinicythere emarginata, and Eucytheridea n. sp.) that imply warmer water conditions than are present today.

In summary, mollusks and ostracodes have been recovered from sediments representing four transgressive events: Anvilian (late Pliocene?), Fishcreekan (early Pleistocene), Kotzebuan (middle Pleistocene), and Pelukian (late Pleistocene). Amino acid racemization analyses of selected molluscan species enables a relative chronology to be established for the sediments at Skull Cliff. Analysis of the mollusks and ostracodes enables a comparison to be made of the marine paleoclimate that existed during each transgression. The ostracodes and mollusks also provide a relative chronology based on the known sequence of climatostratigraphy as intepreted by Hopkins (table 2).

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Reviewed by David M. Hopkins and J. A. Wolfe

STRATIGRAPHY AND SEDIMENTOLOGY OF EPIGURUK BLUFF—A PRELIMINARY ACCOUNT

By Thomas D. Hamilton, Gail M. Ashley[1], Katherine M. Reed, and Douglas P. Van Etten

Epiguruk Bluff is situated on the Kobuk River 14 km west of Ambler (fig. 7). It varies in height from 12 m to 36 m, extends for about 3.5 km along the southwest side of the river, and truncates a surface dominated by sandy fluvial, eolian, and thaw-lake deposits that formed on the floor of the Kobuk River valley during late Quaternary time. Exposures along the face of the bluff exhibit a complex record of late Pleistocene alluviation, eolian activity, and soil formation, with associated development of sand wedges, ice wedges, thaw ponds, and other periglacial features. The bluff also is significant for its abundant organic record, including peat layers, grasses and woody shrubs in primary growth positions, and bones of Pleistocene animals. These remains occur in deposits of glacial, interglacial, and interstadial age, and, together with the host sediment, provide an unusally complete and detailed record of changing environments and biota during late Quaternary time.

Epiguruk Bluff was first studied by A. T. Fernald (1964), who measured a section near its south end. C. S. Schweger (1976) later mapped, dated, and studied the pollen content of deposits through a 900-m-long section at its northern end. In 1981 we measured a 2,600-m base line along the river and defined the principal stratigraphic units present. In 1982, we measured sections at 25-50-m intervals along the base line and drew a complete section of the bluff face (fig. 8). We described and sampled all major units and collected associated organic remains for identification and radiocarbon dating.

Buried soils exposed along the face of the bluff formed under diverse environments. Highly organic bog and flood-plain soils predominate near the present river level, with sandy "upland" tundra soils at higher positions. The paleosols characteristically are black to very dark gray, compact, and highly organic; they are more resistant to erosion than any other sediments in the bluff and are not as readily destroyed by channel cutting. These distinctive, generally continuous paleosols provide the stratigraphic framework within which the diverse sediments of the bluff can be placed (fig. 8). Two buried soils occur discontinuously through the central part of the bluff. The lower paleosol consists of vertical-sided, domelike masses of very compact muck and peat; these are regularly spaced at intervals of 15-25 m and appear to be thermokarst mounds (Péwé, 1982, p. 36-39). The upper paleosol, exposed laterally between about 700 m and 1,650 m (fig. 8), is less dense and contains abundant layers and lenses of wind-blown sand. Radiocarbon dates suggest that the upper paleosol began forming prior to 35,000 yr B.P. and was buried beneath accreting loess and alluvium about 24,000 yr B.P. The lower paleosol has not yet been dated.

Sediments in the bluff comprise four principal facies: channel, flood plain, eolian sand, and loess. Sediments of the channel facies are cross-bedded, well sorted, fine to medium sand. Contacts between cross-bed sets are sharp and erosional, commonly containing rip-up clasts and organic

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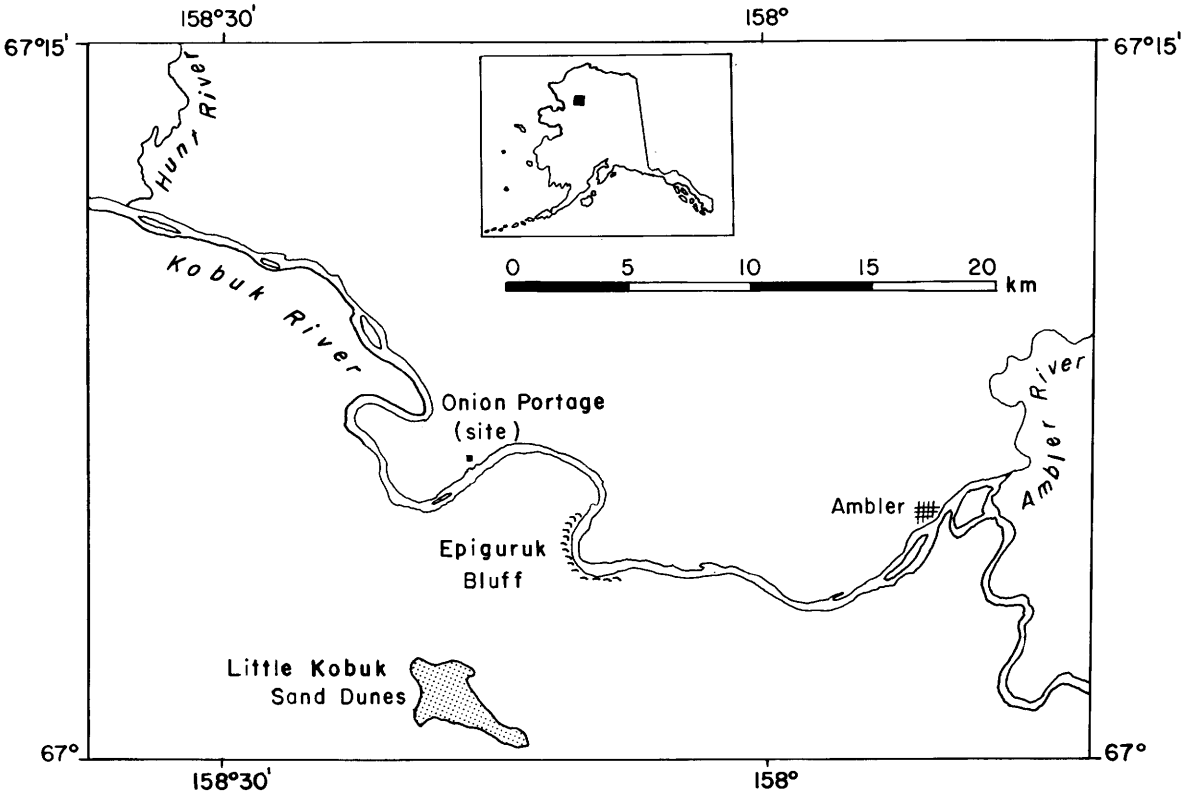

Figure 7.—Regional setting of Epiguruk Bluff.

detritus. Both small-scale crossbedding (<10 cm thick) from migrating ripples and large-scale crossbedding (20-50 cm thick) from migrating sand waves on bars indicate general paleocurrent direction to the west. The flood-plain facies consists of laterally persistent, horizontally bedded fine sand. The sedimentation pattern is rhythmic, alternating from ripple-cross-laminated, light gray, fine to medium sand beds 2-15 cm thick to drapes of dark gray, humic, micaceous, fine sand. These couplets, which probably represent flooding events, contain willows in growth positions, as well as small-scale dewatering and cryoturbation structures. The eolian sand facies consists of medium to fine sand with a few layers of coarse sand and minor (<5 percent) silt. Disruption of layering by plant rootlets is common. Large-scale (50-100 cm thick) bedding features consist of gently-dipping (10^{o}-14^{o}) sets that are laterally persistent for tens of meters. Parallel-laminated sedimentation packages 10-20 cm thick, with both normal and reverse grading, comprise each of the sets. The loess facies is composed of silty, very fine sand that dips gently northward and caps most of the section north of 1,150 m. The loess contains frost cracks and abundant in situ rootlets; it is generally structureless but has some faint laminations made visible by concentrations of mica or humus. The loess fines northward, indicating winnowing from the active sand dunes near the south end of the bluff (fig. 8).

Channel, point-bar, and flood-plain deposits that formed above the upper paleosol and its presumed correlatives indicate a major episode of alluviation of the Kobuk River. Sediments began to aggrade about 24,000 ^{14}C yr B.P., in phase with the onset of the last major glaciation of the Brooks Range (Hamilton, 1982), and had attained a height of 19 m above modern river level by about 17,000 yr ago. Textural and bedding characteristics indicate that the Kobuk River probably was wider and shallower than at present; it contained more sand bars and islands, probably had a braided channel pattern, and was bordered by an extensive vegetated flood plain that received both overbank and eolian sediment.

The alluvial and eolian deposits that formed during the height of the last glaciation later were incised by the Kobuk River and its tributaries. Downcutting probably began between about 18,000 and 16,000 yr ago. All the fluvial deposits at or near their maximum height (19 m) in the bluff are about 18,000-20,000 yr old, providing a maximum limiting age for downcutting. As glaciers waned, a decrease in sediment load would have caused a transition from a braided to a meandering channel pattern. Point-bar deposits dated at about 16,000 yr B.P. near the north end of the bluff indicate that

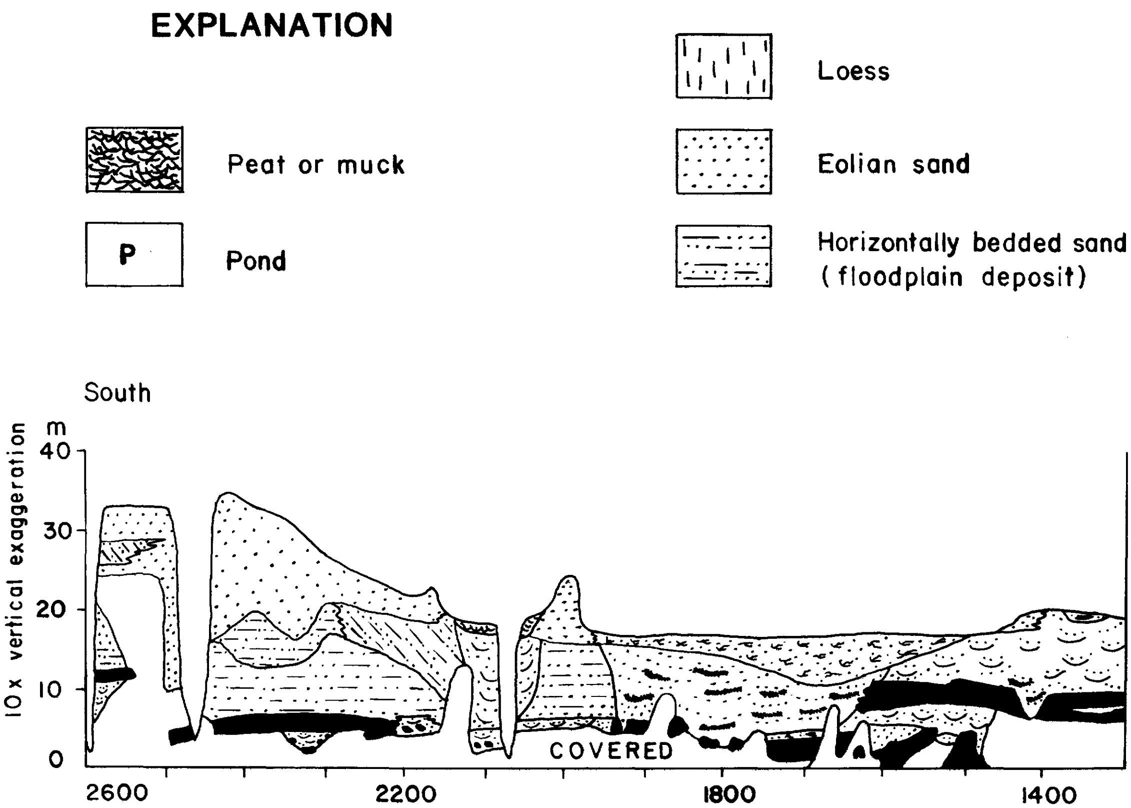

Figure 8.—Stratigraphic relations in Epiguruk Bluff.

the Kobuk River had assumed its modern meandering character by that time and was depositing sandy alluvium no more than 10 m above modern river level. [A similar history of alluviation and downcutting is documented for the nearby Koyukuk River valley (Hamilton, 1982).] Incision of deep channels into the upriver part of the bluff occurred about 10,500 yr B.P., after the Kobuk River had incised to a level near that of the present. By 8,800 yr B.P., thaw-lake development took place on finer grained deposits near the north end of the bluff, presumably in response to the warming climate.

Remains of large mammals recovered in situ from Epiguruk Bluff include Mammuthus primigenius, Mammuthus sp., and Equus sp. cf. E. (Asinus) lambei (C. A. Repenning, U.S. Geological Survey, written commun., 1982). The few small mammal remains recovered were all of modern aspect, and they were identified by C. A. Repenning (written commun., 1983) as Lemmus sibiricus, Spermophilus (Spermophilus) parryi, Dicrostonyx sp., and Microtus sp. Mollusks were found only beneath the lower paleosol and in thaw-lake sediments. These are provisionally identified as Lymnaea spp., Pisidium (lapponicum group), and Pisidium sp. The fauna is consistent with a cool, wet-tundra or riparian environment.

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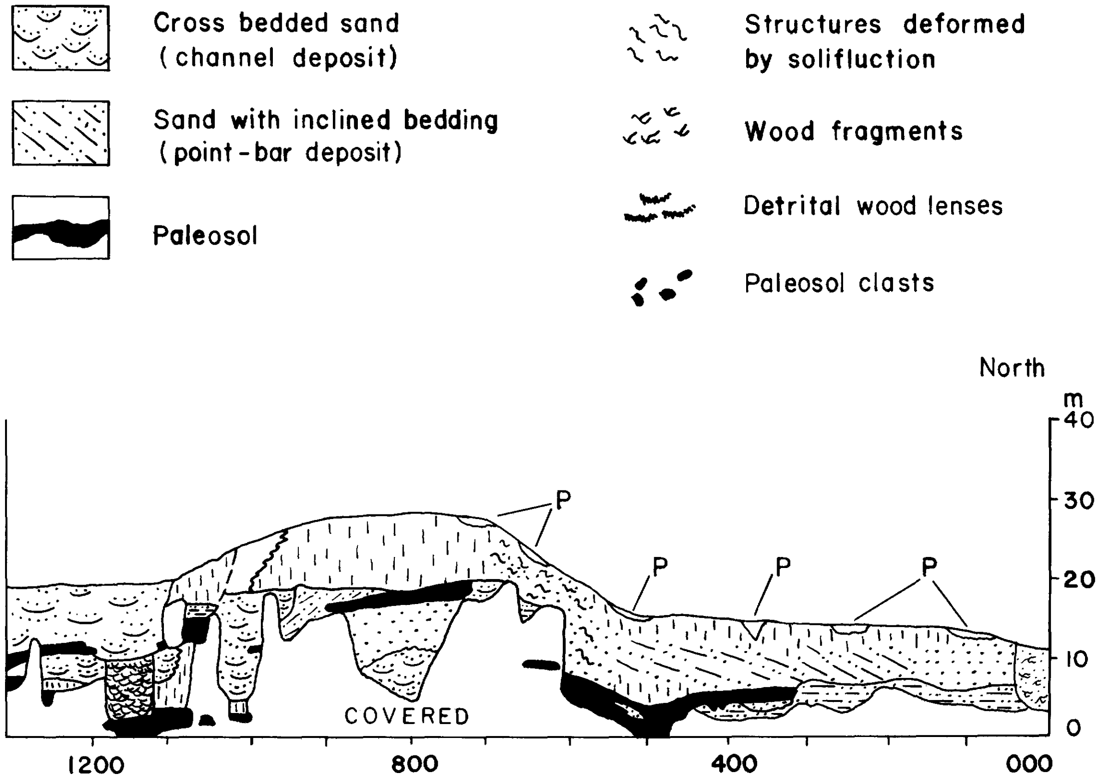

Figure 8.—Stratigraphic relations in Epiguruk Bluff (continued).

EARLY TERTIARY MARINE FOSSILS FROM OCEAN POINT, ARCTIC COASTAL PLAIN, AND THEIR RELATION TO ARCTIC OCEAN PALEOGEOGRAPHY

By Louie Marincovich, Jr., Elisabeth Brouwers, and L. David Carter

Marine invertebrates and palynomorphs at Ocean Point (lat. 70°05' N., long. 151°22' W.), along the lower Colville River, are the first major find of Paleogene fossils in northern Alaska and make up a unique faunal and floral assemblage. The fossils occur in a stratigraphic sequence at least 20 m thick that is composed mostly of moderately consolidated, gray sandy siltstone and unconsolidated tan and gray sand. These beds have been assigned to the Sentinel Hill Member of the Schrader Bluff Formation by Macbeth and Schmidt, (1973) and crop out for about 1.6 km along the west bank of the Colville River immediately north of Ocean Point. Well-preserved mollusks, brachiopods, ostracodes, benthic foraminifers, and dinoflagellates occur throughout the section. Palynomorphs are especially common in the thin carbonaceous laminations that are abundant in most parts of the section. The mollusks and ostracodes have been studied in the greatest detail, but uncertainties in identification at even the genus level have so far made determination of precise age and biogeographic affinities impossible. A fission-track date on zircon grains separated from a tephra layer in terrestrial strata beneath the marine deposits is 50.9 ± 7.7 m.y. (Carter and others, 1977).

At least 15 molluscan taxa, including two gastropods, occur in the Ocean Point fauna. These have been identified, some questionably, only to the genus level and, as yet, none to the species level. Most genera are known from both Mesozoic and Cenozoic faunas; however, some have been reported only from deposits of one or the other era. There is no evidence of reworking of the marine fossils. The bivalves *Integricardium*, *Oxytoma*, and *Yokoyamaina* have been previously found only in Mesozoic faunas, whereas the bivalve *Cyrtodaria* is known only in Paleocene to Holocene faunas. In addition, a large *Pecten* species has a primary shell structure indicative of post-Mesozoic pectinids (T. A. Waller, U. S. National Museum, oral commun., 1979). The large size of the Cretaceous to Holo- cene bivalve *Arctica* further suggests a Tertiary age, because Cretaceous species of *Arctica* are relatively small. With the exception of broadly defined genera, there are no molluscan taxa in common between the Ocean Point fauna and Paleo- gene faunas of southern Alaska (Burk, 1965; Kanno, 1971; Addicott and Plafker, 1971) or adjacent regions of the North Pacific.

Forty-two ostracode taxa occur at Ocean Point, though none has been identified to the species level and several genera are tentatively identified. Genera suggestive of a Cretaceous age, including Crassacythere and Brachycythere, occur with others characteristic of the Paleogene, such as Schuelerridea and Patellacythere. Some taxa, including "Paracyprideis", Eucytheridea", and "Roundstonia", appear to be ancestors of Neogene and Quaternary genera of northwestern Europe. A precise age determination based on ostracodes is not yet possible, but the closest affinities are with Eocene faunas of northwestern Europe (Keen, 1978), and the youngest possible age for the Ocean Point fauna is Oligocene.

Benthic foraminifers from Ocean Point were thought to be of Late Cretaceous (Campanian) age by Macbeth and Schmidt (1973), based on presumed similarities with other northern Alaskan faunas of undoubted age. Their age interpretation is now thought to be in error, based on recent reexamination of the benthic foraminifers and the evidence provided by the several other fossil groups preserved at Ocean Point. Latest Cretaceous (Maestrichtian) pollen and spores from a probable lacustrine environment occur at Ocean Point, but only in clayey pebbles that clearly have been reworked into the sandy fossil beds (F. E. May, U.S. Geological Survey, written commun., 1978).

The molluscan fauna taxonomically most similar to that at Ocean Point occurs in the Cannonball Member of the Fort Union Formation of North and South Dakota. The Cannonball bivalve fauna consists of 30 species of late Paleocene (Thanetian) age, based on species similarities with European faunas in the London and Paris basins (Stanton, 1920; Cvancara, 1966). The Cannonball species in several bivalve genera recognized by Cvancara (1966), including Nucula, Nuculana, Arctica, Bicorbula, Panopea?, and possibly Corbicula, are very similar to the Ocean Point species. However, about 6,000 km separate the Ocean Point beds from the Cannonball outcrops in the Dakotas, so it is unlikely that many of the molluscan species are identical between the two faunas. Cvancara (1966) noted that the Cannonball bivalve fauna has closer affinities with Paleocene faunas of northern Europe than with the geographically nearer Paleocene faunas of the American Gulf Coast. The Ocean Point ostracode fauna also is similar at the genus level to northern European faunas (Keen, 1978), but it lacks close affinities to Gulf Coast faunas (Howe and Garrett, 1934; Howe and Chambers, 1935).

Ocean Point mollusks and microfossils are also similar to poorly known early Tertiary faunas of the Canadian Arctic Islands and West Greenland. In Canada, the Eureka Sound Formation is of Late Cretaceous to Oligocene(?) age and crops out extensively on Ellesmere, Axel Heiberg, and Banks Islands (between about lat. 70° and 80° N.)(Miall, 1981). This formation consists of marine and nonmarine deposits bearing a variety of plant and animal remains, including reptiles and primates, indicative of a warm climate (Miall, 1981; West and others, 1981). Marine beds in the Eureka Sound Formation have been poorly studied but do contain mollusks and microfossils (Miall, 1981; West and others, 1981) that suggest a shallow, relatively warm environment similar to that indicated by the roughly coeval Ocean Point fauna. Lower Tertiary marine strata of West Greenland contain temperate to warm faunas (Rosenkrantz, 1970; West and others, 1981) that are better known than those of the Canadian Arctic Islands and may be more closely compared with the Ocean Point fauna. The West Greenland Tertiary mollusk faunas are thought to be of early Paleocene (Danian) age. The fauna from the upper Danian Agatdal Formation contains taxa similar to those found in the Cannonball Member and at Ocean Point. Agatdal bivalve species in the genera Nucula, Nuculana, Corbula, Tellina, and Pecten, among others, are very similar to those in the Ocean Point beds and the Cannonball Member. Continuing study of the Ocean Point mollusks, ostracodes, and other fossil groups will determine the degree of similarity to faunas from northern Canada and West Greenland. The Ocean Point locality is the westernmost occurrence of this high-latitude early Tertiary fauna.

Our working hypothesis is that the Ocean Point fauna lived in an isolated part of the early Tertiary Arctic Ocean, virtually or completely cut off from the world ocean. Related faunas in northern Canada and West Greenland imply that this isolation may have affected a large portion of the Arctic Ocean basin. The time of onset and the duration of this early Tertiary isolation episode are not precisely known. Further study of the Ocean Point and related arctic faunas will add significant insight into the Cenozoic history of the Arctic Ocean and the plate tectonic events that affected marine connections with the North Atlantic and North Pacific Oceans.

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Reviewed by Charles D. Blome and Charles A. Repenning

PARAUTOCHTHONOUS MISSISSIPPIAN MARINE AND NONMARINE STRATA, LISBURNE PENINSULA, ALASKA

By Thomas E. Moore, Tor H. Nilsen, Arthur Grantz, and Irvin L. Tailleur

An unnamed Mississippian sequence of intercalated marine and nonmarine sandstone, siltstone, shale, coal, and small amounts of conglomerate, here informally called the Kapaloak sequence, crops out for a distance of more than 50 km in an arcuate pattern east of Point Hope on the Lisburne Peninsula (fig. 9). The unit is at least 600 m thick and is unconformably underlain by metamorphosed Ordovician and Silurian graptolitic marine shale, siltstone, and sandstone of the Iviagik Group (fig. 10) (Martin, 1970; Grantz and others, 1983). The Kapaloak sequence is successively overlain by an unnamed Upper Mississippian marine shale, limestone of the Lisburne Group, and Permian and Triassic chert and argillite of the Etivluk Group (Campbell, 1967; Mull and others, 1982). J. T. Dutro, Jr. (in Campbell, 1967) suggested that the Kapaloak sequence might be correlative with the Noatak Sandstone of the allochthonous sequence of the Endicott Group in the Brooks Range. Our recent work, however, suggests that the Mississippian clastic rocks of the Lisburne Peninsula are analogous to the autochthonous sequence of the Endicott Group, which crops out in the northern and southern Brooks Range and underlies the North Slope, and may instead be parautochthonous.

The Kapaloak sequence was referred to by Collier (1906) as his "lower (Carboniferous) formation" (p. 18) and by Campbell (1967) as his "undivided Mississippian sedimentary rocks" (p. 5). B. L. Mamay (in Tailleur, 1965) considered plant fossils from the Kapaloak sequence to be Early Mississippian; marine fossils from near its top were reported by Campbell (1967) to range from Early to Late Mississippian. Coal beds in the Kapaloak were measured and described at Kapaloak Creek near Cape Dyer by Tailleur (1965) and at Cape Thompson and near the Kukpuk River by Conwell and Triplehorn (1976).

The base of the Kapaloak sequence consists of 4 m of interbedded nonmarine granule to pebble conglomerate and coarse-grained sandstone which form discontinuous fining-upward conglomerate-sandstone couplets that were probably deposited by braided streams. The conglomerate clasts are 96 percent chert, 3 percent vein quartz, and 1 percent quartzite. The maximum clast size is 7 cm. The basal conglomerate at Cape Dyer contains clasts of the underlying shale. Sandstone from near the base of the Kapaloak contains an average of 75 percent quartz, 15 percent chert, 5 percent siltstone and argillite, and minor feldspar as framework grains.

The remainder of the Kapaloak sequence consists of about 60 percent nonmarine and 40 percent marine strata. The marine and nonmarine intervals are commonly separated by coal and carbonaceous shale layers as thick as 2 m that have detrital textures and are interpreted to represent interdistributary-bay deposits.

The marine intervals are as thick as 40 m and are characterized by lenticular, thin-bedded, very fine- to fine-grained sandstone and shale. The sandstone contains abundant oscillation ripple markings, shale drapes, rip-up clasts, and, near the top of the unit, hummocky cross-strata. Although no marine megafossils have been found, abundant

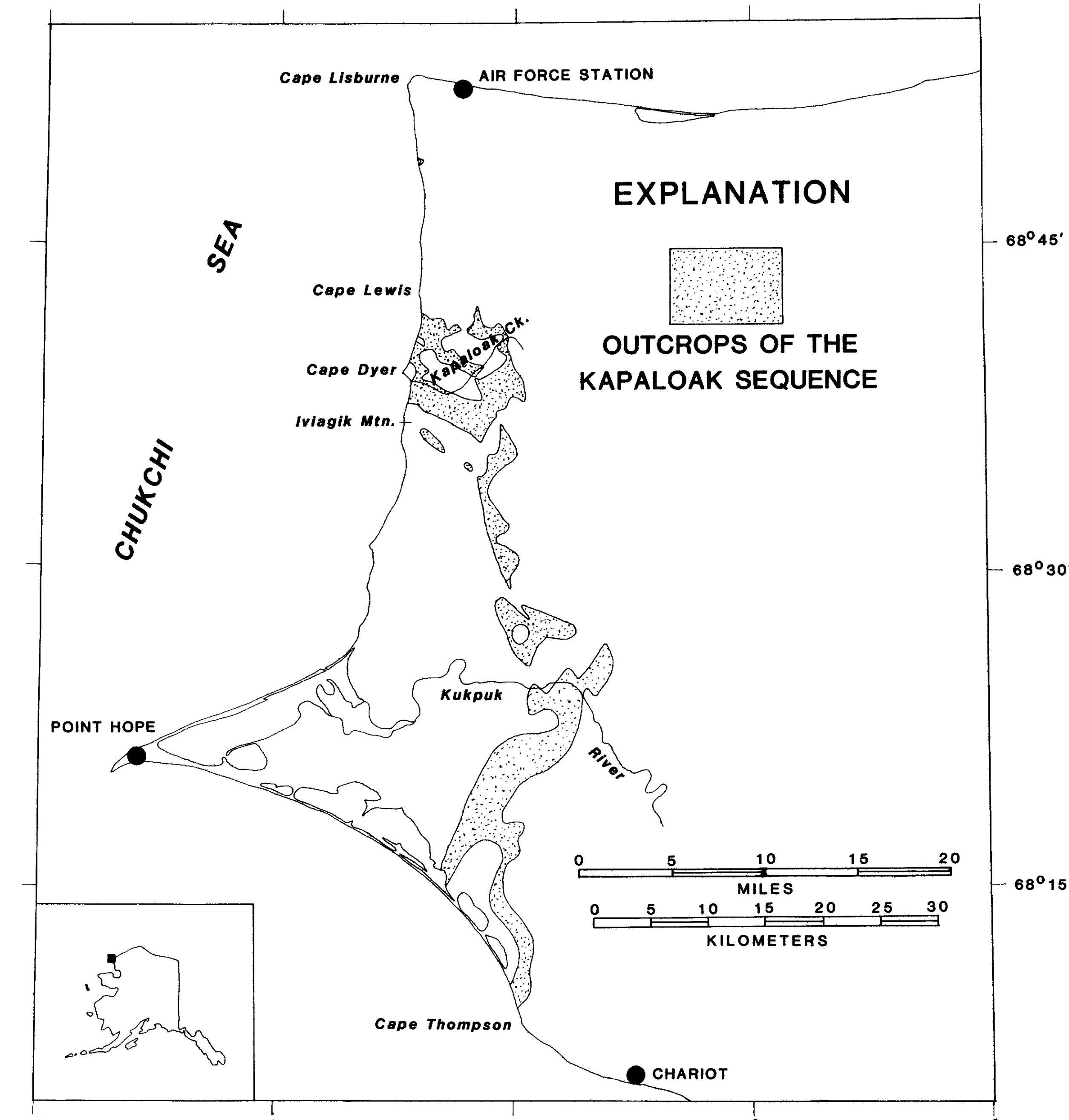

Figure 9.—Index map showing distribution of the Kapaloak sequence in the Lisburne Peninsula (modified from an unpublished geologic map by I. L. Tailleur).

and diverse ichnofossils, including Scolithus, are present. The trace fossils, sedimentary structures, and well-sorted character of sandstone suggest a marine origin for these intervals. The presence of mudcracks, debris-flow deposits, and channelized sandstone suggest that many of the marine intervals were deposited in tidal and nearshore marine environments.

The nonmarine intervals are as thick as 70 m and contain abundant plant fragments. They commonly form fining-upward cycles that consist, in ascending order, of channelized fine-grained sandstone with trough cross-strata, siltstone with current ripple markings, and mudstone with abundant root casts. These sequences are interpreted to represent the deposits of meandering streams. Thinner,

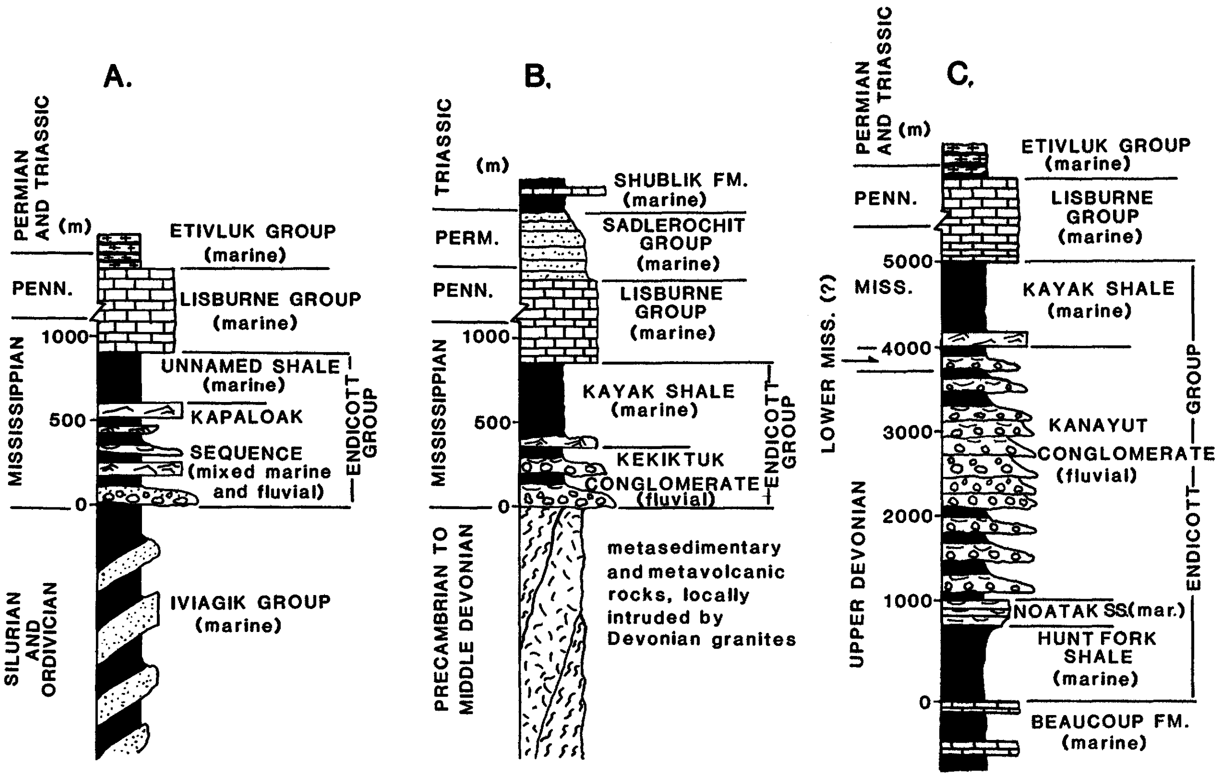

Figure 10.—Schematic columnar sections and inferred depositional environments of middle and upper Paleozoic strata in the Lisburne Peninsula and Brooks Range. A, the Lisburne Peninsula; B, the autochthonous sequence; and C, the allochthonous sequence (generalized) of the Brooks Range.

coarsening-upward nonchannelized cycles of mudstone to fine-grained sandstone with current ripple markings and in situ fossil trees are also common; they are interpreted to represent prograding levee deposits. Sandstone from the nonmarine intervals is commonly more poorly sorted than that in the marine units.

The Kapaloak sequence and the unnamed Upper Mississippian shale unit are here interpreted to be part of the Endicott Group which comprises the Upper Devonian to Upper Mississippian clastic rocks elsewhere in northern Alaska (Tailleur and others, 1967). In the Brooks Range, the Endicott Group has been divided into autochthonous and allochthonous parts (Mull and Tailleur, 1977; Nilsen, 1981). The autochthonous part consists, in ascending order, of the Lower Mississippian fluvial Kekiktuk Conglomerate and the tidal and marine Kayak Shale (Brosgé and others, 1962). The Kekiktuk Conglomerate, which is as thick as 310 m in the northeastern Brooks Range, rests unconformably on deformed Middle(?) Devonian and older rocks that consist principally of Ordovician and Silurian slate and (or) shale and argillite in the subsurface of the Barrow and Prudhoe Bay oilfield areas (Carter and Laufeld, 1975) and Precambrian to Middle Devonian metasedimentary, metavolcanic, and granitic rocks in the northeastern Brooks Range (fig. 10). The autochthonous sequence records fluvial deposition following a Late Devonian and Early Mississippian erosional event and later onlap of marine strata.

The allochthonous part of the Endicott Group consists of more than 4,000 m of conglomerate, sandstone, and shale that comprise, in ascending order, the marine Hunt Fork Shale and Noatak Sandstone, the nonmarine Kanayut Conglomerate, and the marine Kayak Shale (Bowsher and Dutro, 1957; Porter, 1966). The allochthonous part of the Endicott Group rests conformably on marine sandstone, shale, and carbonate rocks of Late Devonian age (Dutro and others, 1979) and is interpreted to represent the progradation and retrogradation of a fluvial-dominated delta during the Late Devonian and Early Mississippian (Nilsen, 1981; Nilsen and others, 1981). The allochthonous sequence of the Endicott Group and the overlying Lisburne Group are thought to have been thrust northward over the autochthonous sequence in the late Mesozoic (Brosgé and Tailleur, 1971, Mull and others, 1976; Roeder and Mull, 1978).

Like the autochthonous part of the Endicott Group, the upper Paleozoic rocks of the Lisburne Peninsula rest unconformably on metamorphosed Ordovician and Silurian clastic rocks and record an

Early Mississippian erosional event followed by fluvial deposition and a later marine trangression. Although the Kapaloak sequence is thicker and finer grained, the similar depositional facies and composition, together with the basal unconformity and Early Mississippian age strongly suggest that it is analogous to the fluvial Kekiktuk Conglomerate and overlying tidal to shallow-marine basal sandstone member of the Kayak Shale of the autochthonous part of the Endicott Group. The presence of the Etivluk Group in the stratigraphy of the Lisburne Peninsula, however, indicates that these Permian and Triassic rocks have an affinity with the allochthonous sequences of the western Brooks Range (C. F. Mayfield, U.S. Geological Survey, written commun., 1983). These conflicting lines of evidence suggest that the upper Paleozoic rocks of the Lisburne Peninsula should be considered as parautochthonous. The parautochthonous nature of the sequence may also be supported by geologic mapping that indicates eastward-directed thrusting during a later event (Campbell, 1961, 1967; Martin, 1970; Grantz and others, 1983).

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Reviewed by William P. Brosgé and Charles F. Mayfield

TETHYAN MARINE TRIASSIC OSTRACODES IN NORTHEASTERN ALASKA

By I. G. Sohn

Middle and Late Triassic ostracodes from the reference section of the Shublik Formation are similar on the generic level to taxa from the Tethyan Triassic of Europe and Asia. All but one specimen are from samples collected by R. L. Detterman in 1969 from the Shublik Formation at the reference section, 10.4 km (6.3 mi) N. 84° W. of the confluence of Fire Creek with the Sadlerochit River, lat. 69° 32'45" N., long. 145°12' W., Mt. Michelson quadrangle (Detterman and others, 1975, p. 14-16, fig. 6). Stage assignments for the units in this section are based on megafossils identified by N. J. Silberling (Detterman and others, 1975, p. 43).

In order to demonstrate the potential utility of ostracodes in biostratigraphy and paleoecology, illustrated specimens were selected from ostracodes in 64 previously disaggregated samples that were documented by collection numbers and by distance (measured in feet) from 105 to 475 ft above the base of the Shublik Formation. Table 3 lists the data for the collections from which specimens illustrated in figure 11 were taken.

Hyatobairdia sp. ex gr. H. arcuata Kristan-Tollmann, 1970, was originally described from the Rhaetian of the Austrian Alps. The illustrated specimen from Alaska is from a collection considered by Detterman and others (1975, fig. 6) to be Norian in age.

Ogmoconcha n. sp., from the Ladinian in the reference section, has the same taxodont hinge as the figured lower Carnian specimen from Hungary. A ventroposterior spine only on the right valve is also present on the Hungarian specimen. These common characters indicate that the Alaskan and Hungarian specimens are definitely congeneric.

Two specimens are here tentatively referred to Ogmoconcha Triebel, 1941. One specimen has an aggregate adductor muscle-attachment scar, and one valve has a taxodont hinge, both characteristic of a group of Triassic and Early Jurassic genera that belong to the Hungarella-Ogmoconcha complex.

New genus, n. sp. is anomalous in Triassic ostracode assemblages because it represents a typical Paleozoic beyrichiopsid element. Specimens of this species were recovered from four collections spanning a stratigraphic interval of 350 ft, beginning with the sample 125 ft above the base of the Shublik Formation. Because the matrices of the internal molds are similar to the sediments from which the specimens were extracted, these specimens may not be contaminants. Instead, they may be examples of Paleozoic ostracodes that survived the Permian-Triassic extinction (Sohn, 1965).

Preliminary processing of 52 samples from borehole cores in the National Petroleum Reserve area revealed ostracodes similar to those in the Shublik Formation. After further study, the Triassic ostracodes in northern Alaska may prove to be biostratigraphically useful.

Table 3.—Ostracode specimens from the Shublik Formation, northeastern Alaska

Stage	Collection number	Distance above base of formation [in meters (feet)*]	Figure 11
Ladinian	3-11	36.6-38.1 (120-125)	A-D, J
Ladinian	3-28	64.01-65.5 (210-215)	I
Norian	6-7	108.2-109.7 (355-360)	H,K
Norian?	6-10	112.8-114.3 (370-375)	L

* Original measurements were made in feet; table includes converted values.

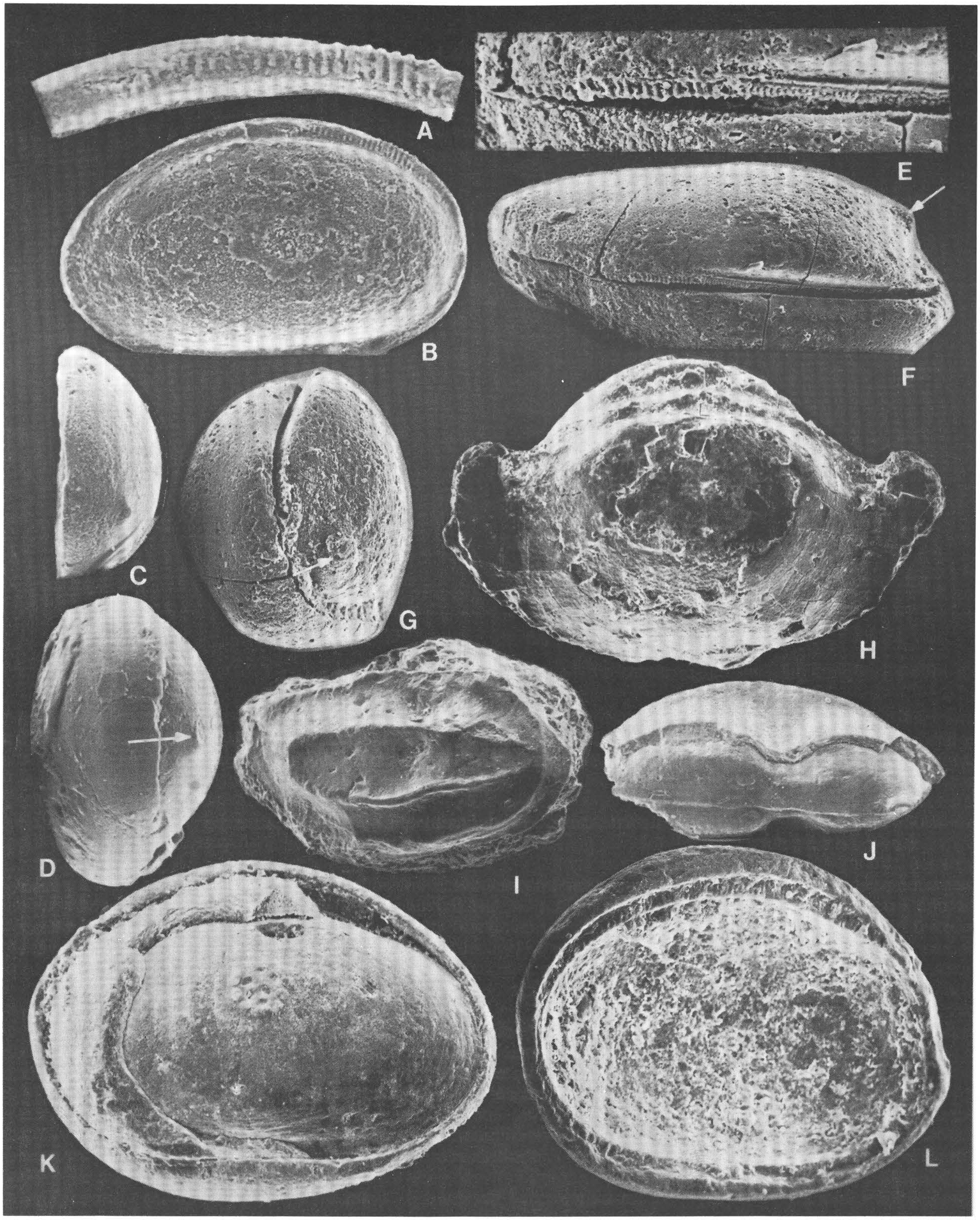
A
B
C
D
E
F
G
H
I
J
K
L

Figure 11.—Triassic ostracodes from northeastern Alaska. A-D: Ogmoconcha n. sp. A, Detail of posterior hinge of right valve, approx. X410; B, C, Inside and and posterior views of right valve, greatest length 0.42 mm, aapprox. X160; D, Posterior view of a carapace, ventroposterior spine on right valve shown by arrow, greatest length 0.45 mm, approx. X170. Ladinian, about 125 ft above base of the Shublik Formation. E-G: Ogmoconcha n. sp. E, Dorsal part of hinge (anterior to left), taxodont dentition similar to that shown in A, approx. X300; F, G, Dorsal and posterior views of carapace, ventroposterior spine onrightvalve shown by arrows, greatest length 0.68 mm, approx. X120. Lower Carnian marl from Csopak, Neszteri Valley, Hungary. Collected by Dr. J. Fulop, Hungarian Geological Society. H, Hyatobairdia sp. ex gr. H. arcuata Kristan-Tollmann, 1970. Left oblique view of carapace, greatest length 0.49 mm, approx. X180. Norian, about 355 ft above base of the Shublik Formation. I, J: New genus, n. sp.; I, Lateral view of internal mold of right valve on matrix, greatest length 0.9 mm, approx. X85. Ladinian, about 210 ft above base of the Shublik Formation; J, Dorsal view of internal mold of carapace (anterior to right), part of shell of left valve retained, greatest length 0.75 mm, approx. X90. Ladinian, 125 ft above base of the Shublik Formation. K, L: Gen. and sp. indet. ex gr. Ogmoconcha Triebel, 1941; K, Right view of carapace, part of the valve missing to expose mold of adductor muscle-attachment scar, greatest length 0.45 mm, approx. X190. Norian, same collection as specimen shown on H; L, Inside view of left valve with taxodont hingement, greatest length 0.45 mm, approx. X190. Norian?, 370 ft above base of the Shublik Formation.

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Reviewed by J. M. Berdan and J. T. Dutro, Jr.

SURFICIAL GEOLOGY OF LOOKOUT RIDGE QUADRANGLE

By Warren E. Yeend

The Lookout Ridge quadrangle is near the center of the National Petroleum Reserve in Alaska (NPRA). It spans the low hogbacks of the northern Arctic Foothills and the broad, featureless Arctic Coastal Plain physiographic provinces. The exploration for and possible ultimate development of hydrocarbons in the quadrangle require a knowledge of the distribution and engineering characteristics of the surficial deposits and bedrock in order to locate suitable foundations and construction materials and to minimize environmental degradation from drill pads, roads, airfields, and other man-induced changes.

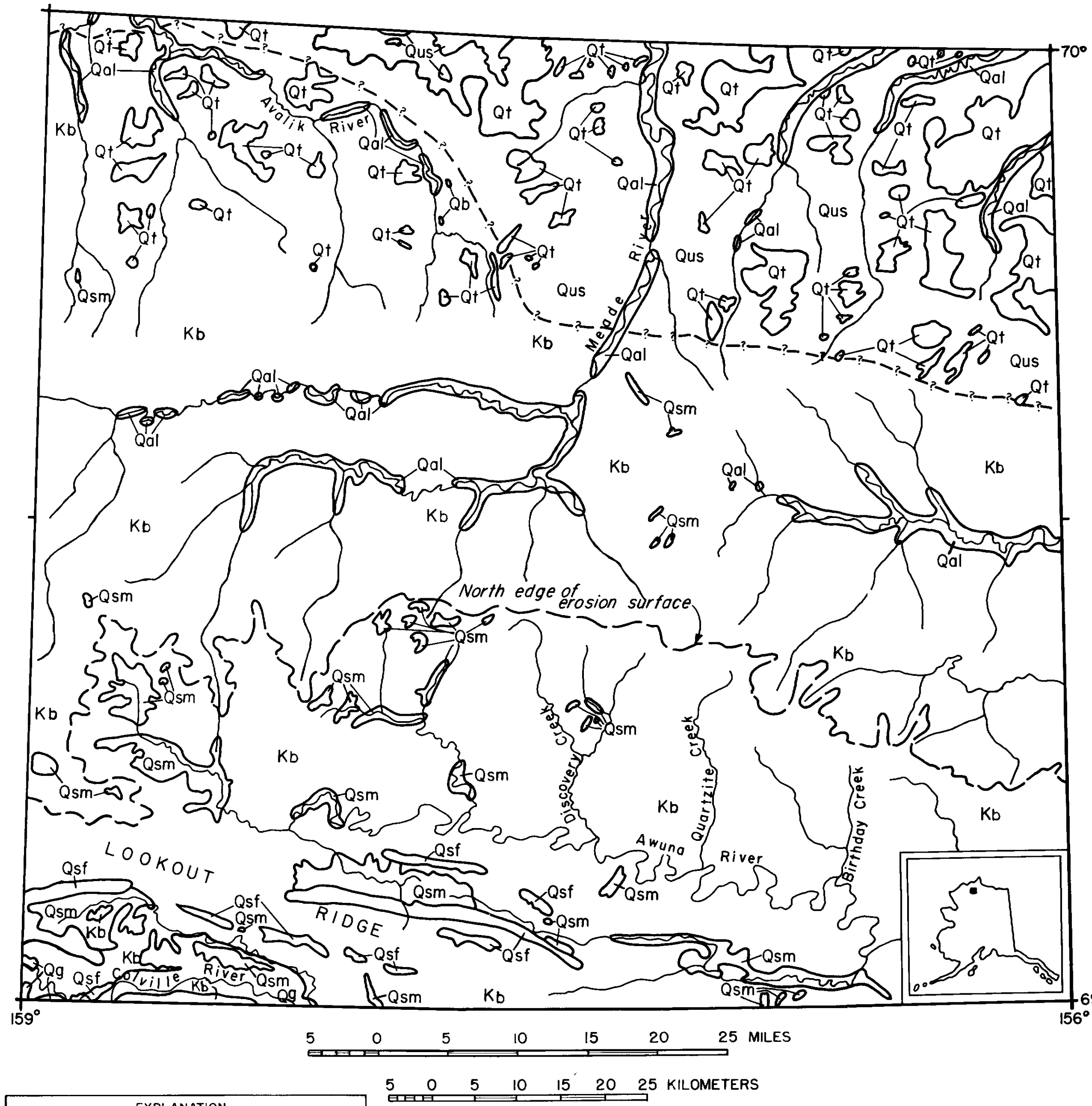
Approximately 70 percent of the quadrangle is underlain by a silty, clayey regolith (mapped as "bedrock, undivided") developed on truncated, east-trending, gentle folds in Cretaceous sandstone and shale (fig. 12). Locally, the regolith has been involved in mass movement that formed solifluction deposits on slopes within drainage depressions. Solifluction deposits cover nearly all the slopes bordering Lookout Ridge.

The lake-dotted depressions along some of the major drainageways in the southern one-third of the quadrangle are filled with organic-rich silt and muck that contain ice-wedge polygons. These ice-rich areas are subject to settling where permafrost is thawed by stripping or compaction of vegetation during construction activities.

Thaw-lake deposits mantle much of the silt and sand and cover part of the coastal plain in the northern one-quarter of the quadrangle. Numerous shallow lakes are remnants of larger Holocene(?) lakes (Williams and Yeend, 1979; Nelson, 1982).

Silt and sand mantle the northern part of the quadrangle, extending up to the level of the 400-ft contour. Much of this silt and sand is probably eolian, but some may be fluvial and marine. Well-washed and -sorted pea gravel occurs as scattered outcrops near the 300-ft contour in the headwaters of the Avalik River. This pea gravel may be an old beach deposit, although no marine fossils were found. Gravel for construction is scarce throughout all but the southwestern corner of the quadrangle along the Colville River. All other rivers and streams head in easily eroded Cretaceous sandstone and shale that break down quickly in the high-energy alluvial environment. The Meade River flood plain is characterized by sandy granule and small pebble gravel composed of friable rocks that would be generally unsuited for most construction uses.

An old (Tertiary?) south-sloping erosion surface is present in the southern one-third of the quadrangle (fig. 13). Well-rounded and broken pebbles and cobbles of yellowish brown, black, red, and gray radiolarian chert are present on the erosion surface



EXPLANATION

Unconsolidated deposits

QUATERNARY
- Qg — Floodplain and alluvial terrace deposits (gravel)
- Qsm — Silt and muck deposits
- Qsf — Solifluction mantle deposits
- Qt — Thaw lake deposits
- Qus — Upland silt
- Qb — Marine beach (?) deposits

Bedrock

CRETACEOUS
- Kb — Bedrock, undivided

——— — — —?— —?— Contact, dashed where approximately located, queried where extreme doubt as to position

Figure 12.—Map of surficial geology of the Lookout Ridge quadrangle.

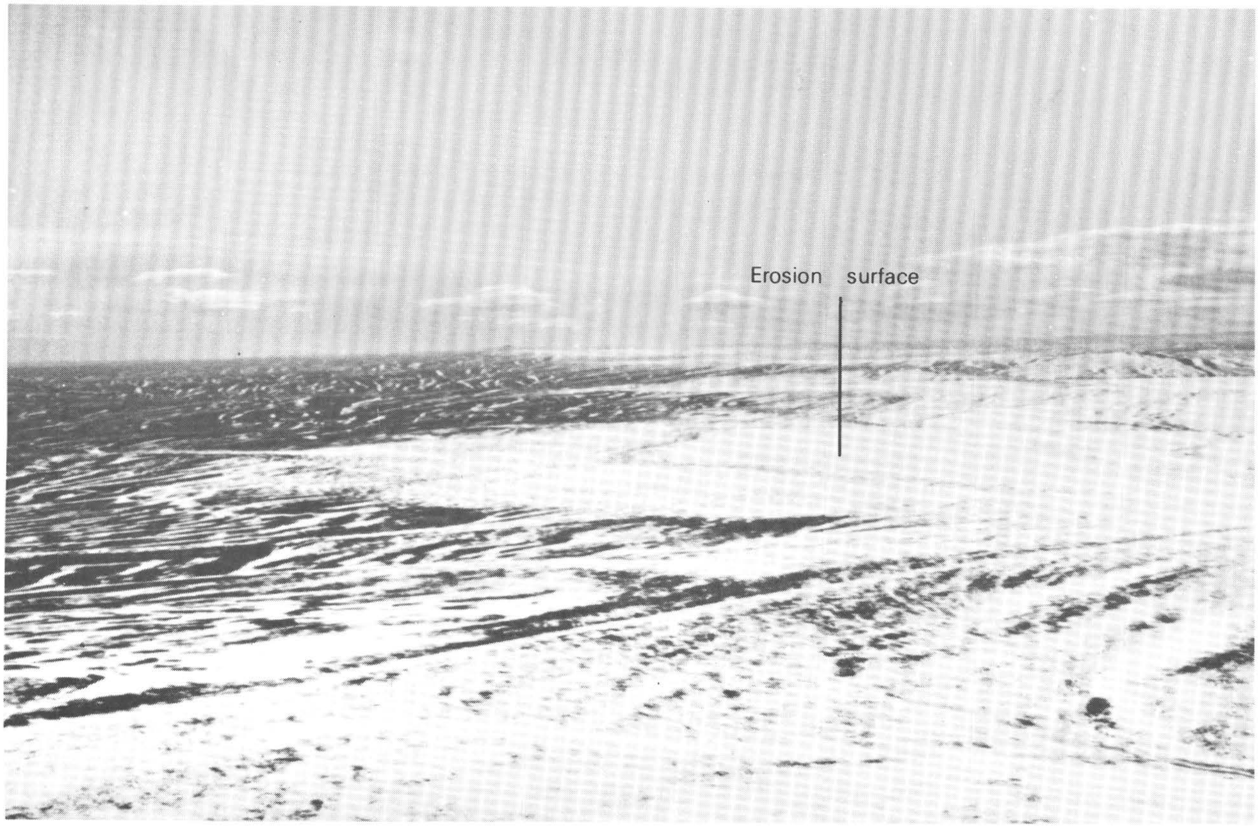


Figure 13.—Aerial view to the east of the snow-covered erosion surface cut across Cretaceous sandstone and shale. Streams that drain into the Arctic Ocean are working headward (to the right) and eroding this old surface.

bordering the entrenched meanders of the Awuna River. The fact that bedrock of this type does not occur locally suggests that the Brooks Range was the most likely source for the gravel. The erosion surface is now cut off from the Brooks Range drainages by the Colville River and Lookout Ridge.

Southerly tilting of the original north-sloping surface and uplift in the Lookout Ridge area, together with the headward erosion and development of the Colville River, are major events that have occurred subsequent to the development of the old surface. Discovery, Quartzite, and Birthday Creeks are consequent drainageways that have developed on this south-tilted surface. They flow into the Awuna River, a major drainageway that parallels the strike of the bedrock. The old erosion surface is also being destroyed by the headward erosion of streams that flow north to the coastal plain and eventually enter the Arctic Ocean (fig. 13). The Colville River, as it worked its way west, beheaded all the rivers that formerly flowed north from the Brooks Range across the Lookout Ridge quadrangle area to the Arctic Ocean. However, as the tributaries of the Meade and Ikpikpuk Rivers erode headward, they will eventually breach this old erosion surface and Lookout Ridge. They will then capture the drainage of the Colville and reestablish drainage in a more direct northerly route from the Brooks Range to the Arctic Ocean. The Utukok River to the west has already done this, shortening the Colville by beheading its headwaters.

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Reviewed by J. R. Williams and O. J. Ferrians, Jr.

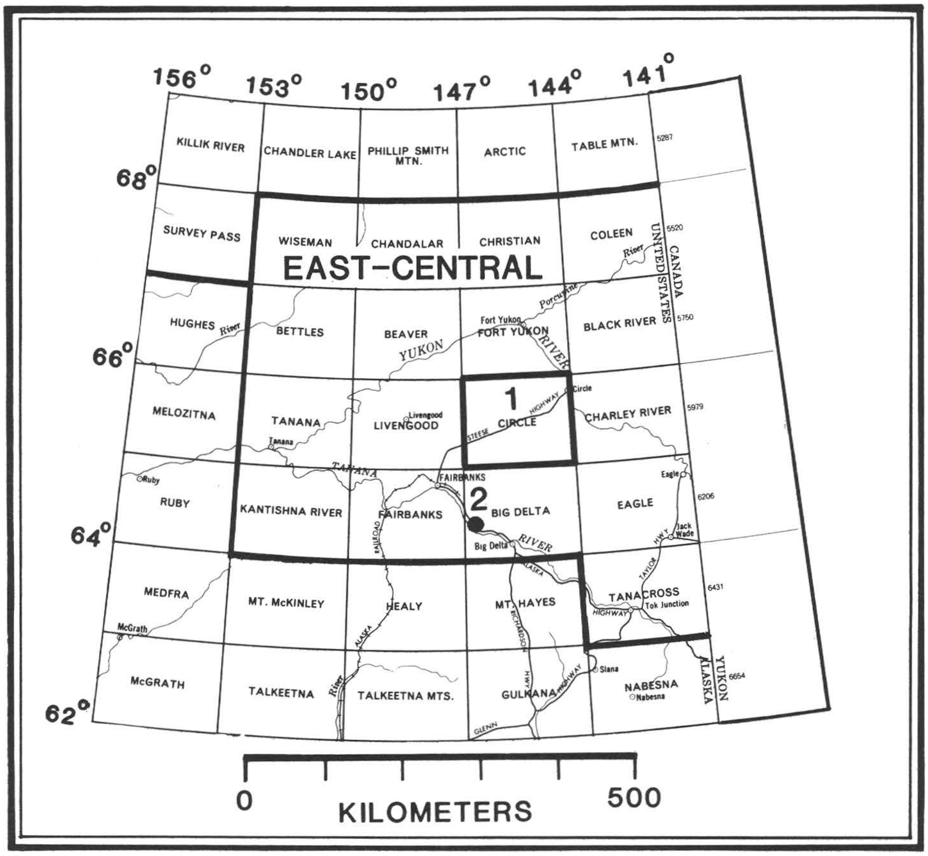


Figure 14.—Areas in east-central Alaska discussed in this circular. Authors and inclusive pages of their articles are: (1) Hamilton and Bischoff, p. 26–29; (2) Laird and Foster, p. 29–33.

EAST-CENTRAL ALASKA

(Figure 14 shows study areas described.)

URANIUM-SERIES DATING OF FOSSIL BONES FROM THE CANYON CREEK VERTEBRATE LOCALITY IN CENTRAL ALASKA

By Thomas D. Hamilton and James L. Bischoff

In a recent paper, Weber and others (1981) discussed the stratigraphy, fauna, and radiocarbon dating of late Pleistocene vertebrate fossils that were collected from a road cut near Canyon Creek, about 85 km southeast of Fairbanks (fig. 15). Bone fragments yielded an apparently finite radiocarbon date of 39,360 ± 1740 yr B.P., but tephra associated with the bones indicated that their true age might be greater than 56,600 radiocarbon years B.P. (Weber and others, 1981, p. 177). Four bone samples from the Canyon Creek locality have recently been dated by the uranium-series method in an attempt to resolve these contradictory age assessments.

The stratigraphic section at the bone locality, as described in Weber and others (1981), is capped by 13 m of eolian sandy silt that overlies colluvium and alluvial sand and gravel (fig. 16). Gravel at the base of the bluff (east of section shown in fig. 16) was deposited by the Tanana River when it flowed at a level about 15 m above present. The overlying bar deposits of the river (unit 3 in fig. 16) interfinger laterally with local alluvium (units 2 and 4) derived from Canyon Creek, which evidently was depositing a fan on the flood plain while the Tanana River was at its higher level. Units 5 and 6 are local alluvial and colluvial deposits that formed when the Tanana River was no longer depositing sediment at the site because of either downcutting or lateral migration across its flood plain. Unit 5 contains fragments of redeposited tephra. Unit 6 evidently formed by solifluction, and its upper surface later was subjected to ventifact formation and sand-wedge development during a period of exposure under thin and probably discontinuous vegetation. These processes may have occurred under a frost climate more rigorous than that of the present but, alternatively, could have been associated with with drier conditions and less winter snow cover (Carter, 1982). The sandy silt of unit 7 was winnowed from the flood plain of the Tanana River and transported to the site as loess. It was dated as late glacial (Pleistocene) and early Holocene (Weber and others,

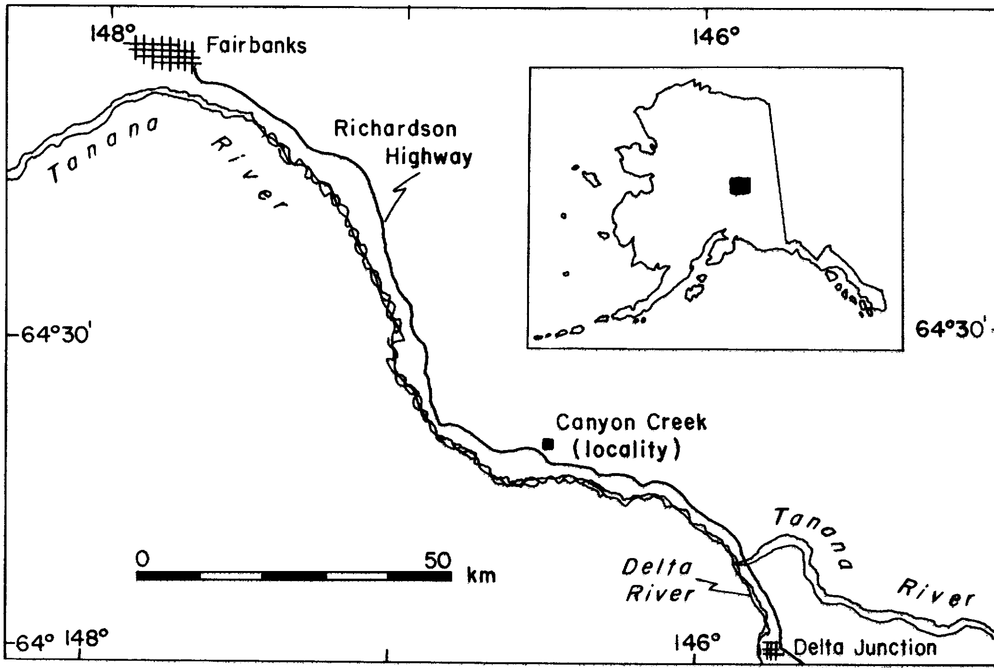
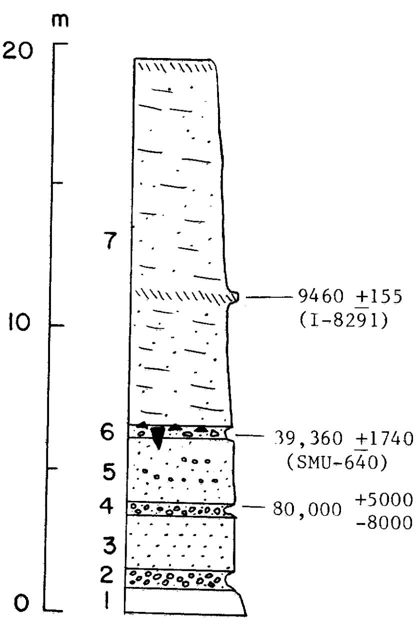


Figure 15.—Location of Canyon Creek site.

1981) on the basis of a radiocarbon date of 9,460 yr B.P. and analogies with the Tanana Valley pollen record of Ager (1975).

Uranium-series dating of bones has yielded reliable results in many cases, but in others contamination has been a serious problem (Bischoff and Rosenbauer, 1981). One way to judge the method's validity is by testing for internal concordancy between two independent decay schemes: $^{238}U - ^{230}Th$ and $^{235}U - ^{231}Pa$. Because ^{230}Th



UNIT	THICKNESS (m)	DESCRIPTION
7	12.8	LOESS-- Sandy silt. Bears Subarctic Brown Forest Soil profile at top and reddish gray paleosol at 8 m depth.
6	0.5	COLLUVIUM-- Angular rock fragments in oxidized sand matrix. Ventifacts at upper contact. Contains sand wedges and rodent burrows.
5	2.4	FLUVIAL SAND-- Poorly sorted. Contains fossil bones and redeposited tephra.
4	0.4	ARKOSIC GRAVEL-- Iron-stained. Fossil bones near upper contact.
3	1.8	FLUVIAL SAND --Interbedded with lenses of arkosic gravel near west end of bluff.
2	0.5	ARKOSIC GRAVEL--As in unit 4, but lacks fossil bones.
1	1.0	COVERED

Figure 16.—Stratigraphic section, Canyon Creek, at 90-m position above base line. Simplified from figure 3 and table 1 in Weber and others, 1981.

Table 4.—Bones from Canyon Creek site, Alaska, dated by uranium-series method

Sample no.	Strati-graphic unit	Identification	Comments
14-1	6	*Mammuthus* cranial fragment	Worn at edges, and probably redeposited
14-4	4	*Equus* humerus	Collected from upper unit 4 and (or) basal unit 5 prior to detailed mapping of the section.
14-6	4	*Equus* metacarpal	
14-10	4	*Equus* radius	

and ^{231}Pa have different decay rates and different chemical properties, any perturbation of the system should give discordant ages. Although ^{230}Th can yield ages back to about 350,000 yr, the concordancy test is limited to about 120,000 yr by the shorter half-life of ^{231}Pa. Successful application of the technique requires that the bone initially take up U but no Th or Pa and then become isolated from further isotopic exchange in a relatively short time. A $^{234}U/^{238}U$ ratio close to unity and absence of or low activity of ^{232}Th relative to ^{230}Th are additional criteria for a closed system and lack of contamination.

Four samples of bone from the Canyon Creek collection (table 4) were selected for dating. The samples were carefully cleaned and their outer surfaces scraped away. After acid dissolution, U and Th isotopes were isolated by ion exchange and solvent extraction, and activities of isotopes were determined by alpha spectrometry, generally following the methods of Ku (1966). Interferences caused by the phosphatic matrix were eliminated by fractional precipitations with $Fe(OH)_3$. Protactinium-231 was determined on a separate portion by analysis for the short-lived daughter ^{227}Th. Dates were calculated using equations given by Ku (1966) with rate constants published by Szabo (1980). All samples contained high concentrations of U, ranging from 100 to 250 ppm (table 5). Such high concentrations facilitated analysis of all pertinent isotopes and minimized the effects of small amounts of impurities. Thorium-232, a monitor of detrital contamination, was below the detection limit ([0.01 disintegrations per minute per gram) for all samples.

Sample 14-4, yielding a date of about 80,000 yr, satisfies all criteria for highest validity, including concordance between ^{230}Th and ^{231}Pa decay systems. The ^{230}Th date of sample 14-1 and the two replicate ^{230}Th dates on sample 14-6 are also close to 80,000 yr. The ^{231}Pa dates on these two samples are somewhat discordant at 56,000 yr, suggesting that Pa has migrated but that the ^{230}Th system has remained closed. The date on sample 14-1 suggests that it could have been redeposited from a stratigraphic level comparable to that of the other bones. Besides the discordance, there is no other isotopic evidence of contamination for samples 14-1 and 14-6, nor is there for sample 14-10, which yields a ^{230}Th date of only 36,000 yr and a ^{231}Pa date of 27,000 yr. This anomalously young date could be due to later uptake of U in an open system, or it may indicate that the bones comprise a mixed assemblage of younger and older specimens. Most of the bones deposited near the contact between units 4 and 5 probably are between about 72,000 and 85,000 yr old, but considerable redeposition may have taken place and the bones do not necessarily date their host sediments.

Units 4 and 5 were deposited at about the time that the Tanana River began downcutting into an extensive outwash train correlated by Blackwell (1965) with glacier advances of the Delta glaciation of Péwé (1953) in the Alaska Range. (See discussion in Weber and others, 1981.) If the dated bones are of the same age as their host sediments, they indicate that glaciation of Delta age may in part coincide with high-latitude cold phases and with episodes of ice-sheet growth that Dansgaard and others (1982) correlate with isotope stages 5b and 5d in the marine record. Moreover, our dates would

Table 5.—Uranium-series analyses of mammal bones from Canyon Creek, Alaska

Sample No.	U (parts per million)	Activity (in disintegrations per minute per gram)							Age x 10^3 yr	
		^{238}U	^{234}U	^{230}Th	^{232}Th	^{231}Pa	$^{234}U/^{238}U$	$^{230}Th/^{232}Th$	^{230}Th	^{231}Pa
14-1	109	81.7	105.9	53.47	<0.01	2.51	1.30	> 5,300	74 +3/-2	56 ± 5
14-4	250	186.5	258.5	136.6	<0.01	6.63	1.39	>13,600	80 +5/-8	73 +14/-10
14-6a	104	76.5	104.1	56.65	<0.01	2.56	1.36	> 5,600	81 ± 2	56 ± 3
14-6b	108	80.3	110.5	61.23	<0.01	NA	1.38	> 6,100	83 ± 4	
14-10	253	189.2	266.2	76.0	<0.01	3.66	1.41	> 7,600	36 +1/-2	27 ± 3

suggest that the Tanana Valley may have undergone a prolonged interstadial that possibly began as early as 80,000 yr ago and ended with the onset of glaciation of late Wisconsin age about 24,000 yr ago (Hamilton, 1982). If, alternatively, the bones represent a mixed assemblage, then both the ^{230}Th and ^{14}C dates of 36,000 and 39,360 yr B.P., respectively, could be valid and would indicate that the interstadial deposits at Canyon Creek represent a more restricted middle Wisconsin time span.

Our uranium-series dates should furnish a close approximation of the age of the tephra in unit 5, provided that neither bones nor tephra were redeposited from much older beds. The tephra fragments are abundant and unmixed with soil or sediment; they occur with no other clasts of unconsolidated material. We believe that they probably were eroded from a recent surface accumulation and that reworking of an older buried tephra layer is much less likely. The Canyon Creek tephra has recently been correlated with the Sheep Creek tephra of the Fairbanks area and with tephra in the Lost Chicken area of east-central Alaska and the Stewart River region of Yukon Territory (J. A. Westgate, Univ. of Toronto, oral commun., Jan. 24, 1983). It is older than 45,000 yr B.P. in the Yukon, according to Westgate, and older than 50,400 yr B.P. (USGS-1253) at the Lost Chicken locality (D. M. Hopkins, U.S. Geological Survey, written commun., Feb. 11, 1983). Stratigraphically, it appears to be younger than the last interglacial maximum. These age brackets are consistent with our date of about 80,000 yr B.P. for bones immediately underlying the tephra at the Canyon Creek locality and suggest that the sediments that enclose the bones cannot be much younger.

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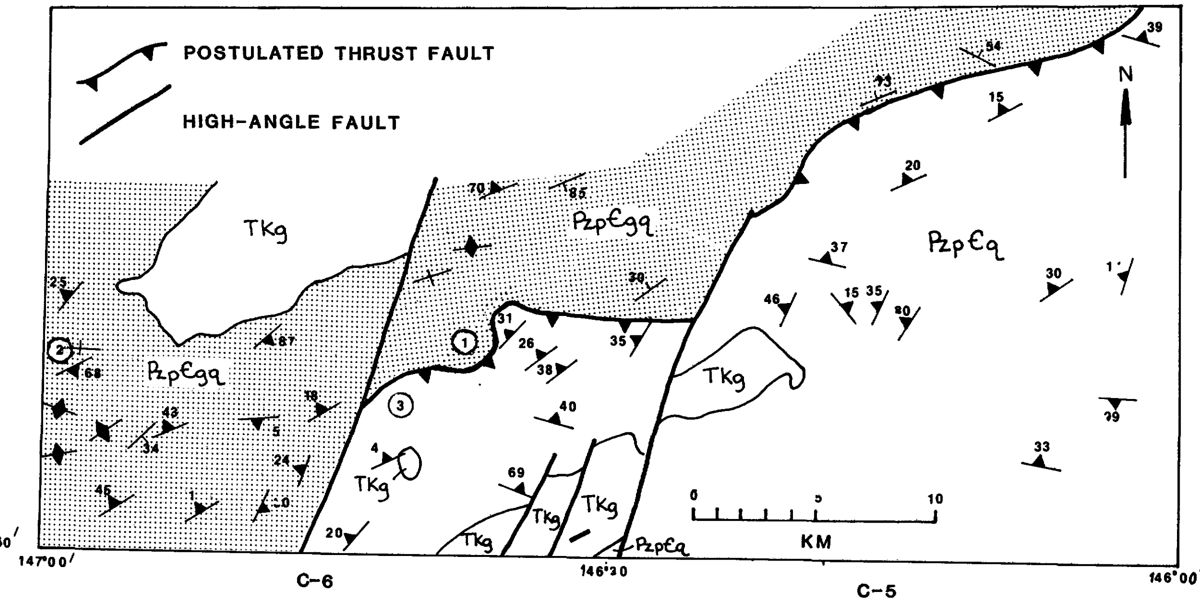
Reviewed by L. David Carter and David M. Hopkins.

DESCRIPTION AND INTERPRETATION OF A MYLONITIC FOLIATED QUARTZITE UNIT AND FELDSPATHIC QUARTZ WACKE (GRIT) UNIT IN THE CIRCLE QUADRANGLE, ALASKA

By Jo Laird and Helen L. Foster

The juxtaposition of a foliated quartzite unit, much of which is mylonitic, and a unit composed dominantly of weakly recrystallized, poorly sorted, feldspathic quartz wacke in the Circle C-5 and C-6 quadrangles (fig. 17) suggests that the contact is probably a fault. Mertie (1937) mapped the foliated quartzite and intercalated pelitic, mafic, and calc-silicate schists as the now-abandoned Birch Creek Schist, which he distinguished from a unit composed mostly of feldspathic quartz wacke, quartzite, slate, and argillite. The Birch Creek Schist was described as being more recrystallized than the latter unit.

Feldspathic quartz wacke, commonly called grit, is the dominant lithology of the less recrystallized unit in the Circle C-5 and C-6 quadrangles, and this group of rocks is referred to as the grit unit in this discussion. The grit is characteristically poorly sorted (fig. 18) and ranges from fine to coarse grained. Angular to subrounded quartz, minor feldspar, and lithic clasts occur with finer grained quartz, feldspar, white mica ± chlorite, opaques minerals, and accessory tourmaline, zircon, and epidote. Quartz clasts generally show undulatory extinction and, in places, strain lamellae. Both plagioclase (commonly twinned) and alkali feldspar (perthite, rarely microcline) clasts are generally somewhat sericitized. Lithic fragments include: polycrystalline quartz ± feldspar, granitic rock with granophyric texture, schist with crenulation cleavage, slate, and poorly sorted grit. Some pressure solution at the edges of the clasts and a weak foliation defined by sheet silicates indicates minor deformation. Little petrographic distinction



EXPLANATION

TKg	Granitic rocks (Paleocene and (or) Late Cretaceous)
PzpЄgq	Grit unit (Paleozoic and (or) Precambrian)
PzpЄq	Foliated quartzite unit (Paleozoic and (or) Precambrian)

Figure 17.—Geologic map of the southern Circle C-5 and C-6 quadrangles, Alaska (after Foster and others, 1983). Sample numbers are localities for photographed thin sections shown in figures 18, 19, and 20.

is observed between grits in this unit. (Compare figs. 18 and 19.)

The well-foliated quartzite unit adjoining the grit unit on the south and east is composed of quartz, plagioclase, white mica ± chlorite, perthitic alkali feldspar, biotite, carbonate, epidote, opaque minerals sphene, and accessory tourmaline (fig. 20). Porphyroclasts are common and include highly strained quartz, plagioclase, perthitic alkali feldspar, and polycrystalline quartz ± feldspar. Some plagioclase is twinned; both plagioclase and alkali feldspar may be somewhat sericitized. White mica, elongated porphyroclasts, and ribbon quartz define the foliation. The porphyroclasts, which generally show evidence of pressure solution, commonly are spindle-shaped and have "tails" composed of finer grained, polygonal quartz.

Farther southeast the foliated quartzite unit has a mineral assemblage composed of quartz + plagioclase + white mica ± chlorite, opaque minerals, biotite, and accessory tourmaline, zircon, and carbonate. Garnet enters the assemblage about 25 km south of the contact. Some samples contain strained quartz, plagioclase, and polycrystalline quartz ± plagioclase porphyroclasts (fig. 21), but others have no porphyroclasts. Quartz ribbon structures are common. Porphyroclasts have strain shadows of polygonal quartz strung out parallel to the foliation. Some porphyroclasts have pull-apart textures with fine-grained quartz filling the fractures. Quartz and feldspar grains are more polygonal and grain boundaries more sutured than just south of the contact.

The grit unit, which both crops out and occurs as rubble, commonly has a network of quartz veinlets. Deep, vegetation-covered saddles separate it, at the western edge of the C-5 and the eastern edge of the C-6 quadrangles, from the foliated quartzite unit, which crops out on topographic highs. Sheared and brecciated rocks of both units are seen near their contact at the eastern edge of the C-6 quadrangle, but the foliated quartzite unit shows significantly more deformation and recrystallization than the grit unit. (Compare figs. 18 and 20a.) The petrographic character of the quartzite is very much like that in recognized mylonite (thrust) zones (for example, fig. 9.27, Hobbs and others, 1976; Higgins, 1971). These observations, along with the very extensive shearing and the apparent restriction of the boxwork of quartz veinlets to the grit unit suggest that in the Circle quadrangle the contact between the grit unit and foliated quartzite unit is a fault and the foliated quartzite near the contact is a mylonite. The sinuous contact and shallow dip of foliation (fig. 17) reinforce a thrust fault interpretation.

The foliated quartzite with porphyroclasts farther southeast of the contact also has a mylonitic

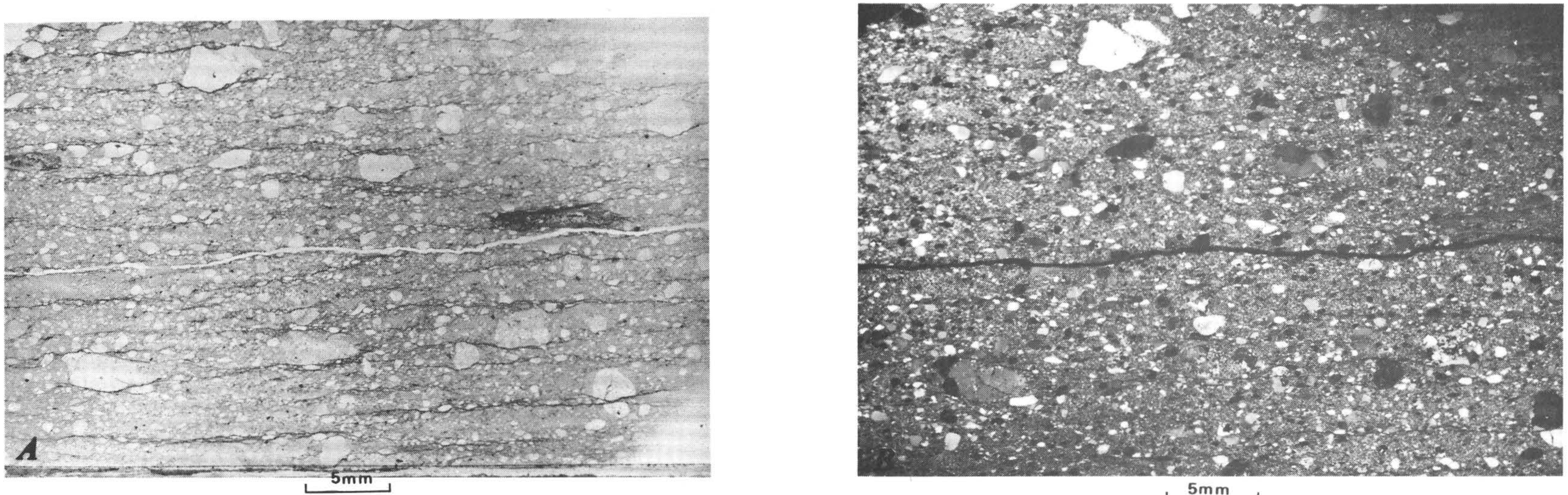


Figure 18.—Grit unit close to contact with the foliated quartzite unit in plane (a) and crossed (b) polarized light. Sample 79AWr359B, Circle C-6 quadrangle (fig. 17, no. 1). Estimated mode: quartz (60 percent), plagioclase (10 percent), alkali feldspar (10 percent), white mica (10 percent), Fe chlorite (8 percent), opaque minerals (2 percent), tourmaline (trace).

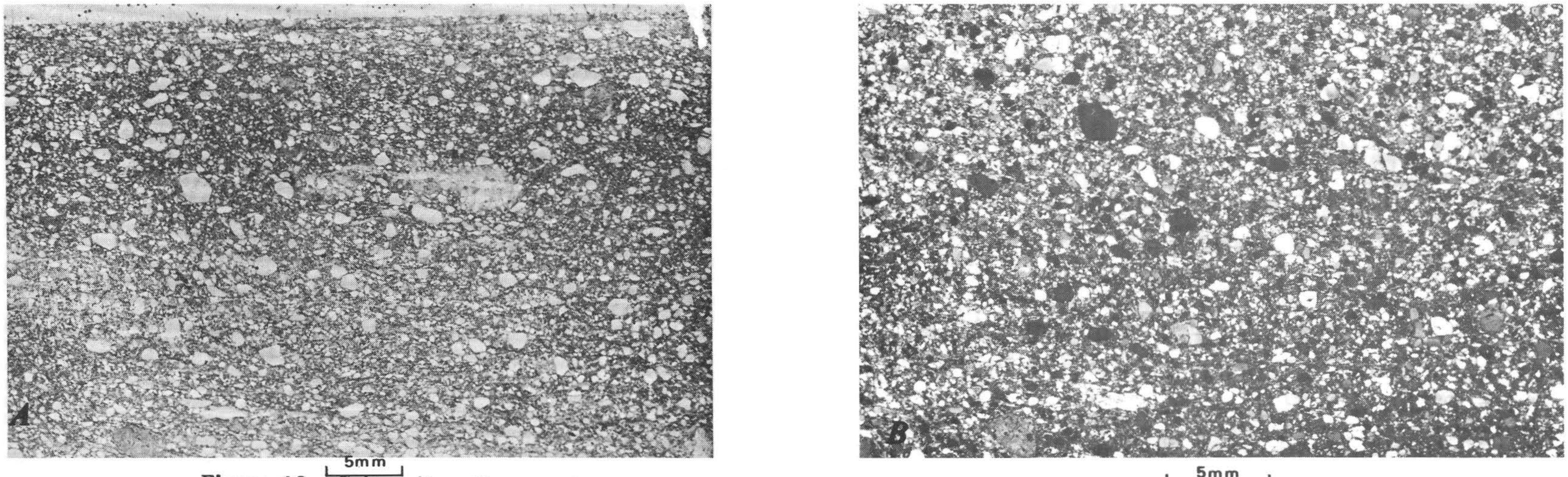


Figure 19.—Grit unit well away from contact with the foliated quartzite unit in plane (a) and crossed (b) polarized light. Sample 79AFr387B, Circle C-6 quadrangle (fig. 17, no. 2). Estimated mode: quartz (60 percent), alkali feldspar (15 percent), white mica (13 percent), plagioclase (10 percent), epidote (2 percent), tourmaline (trace).

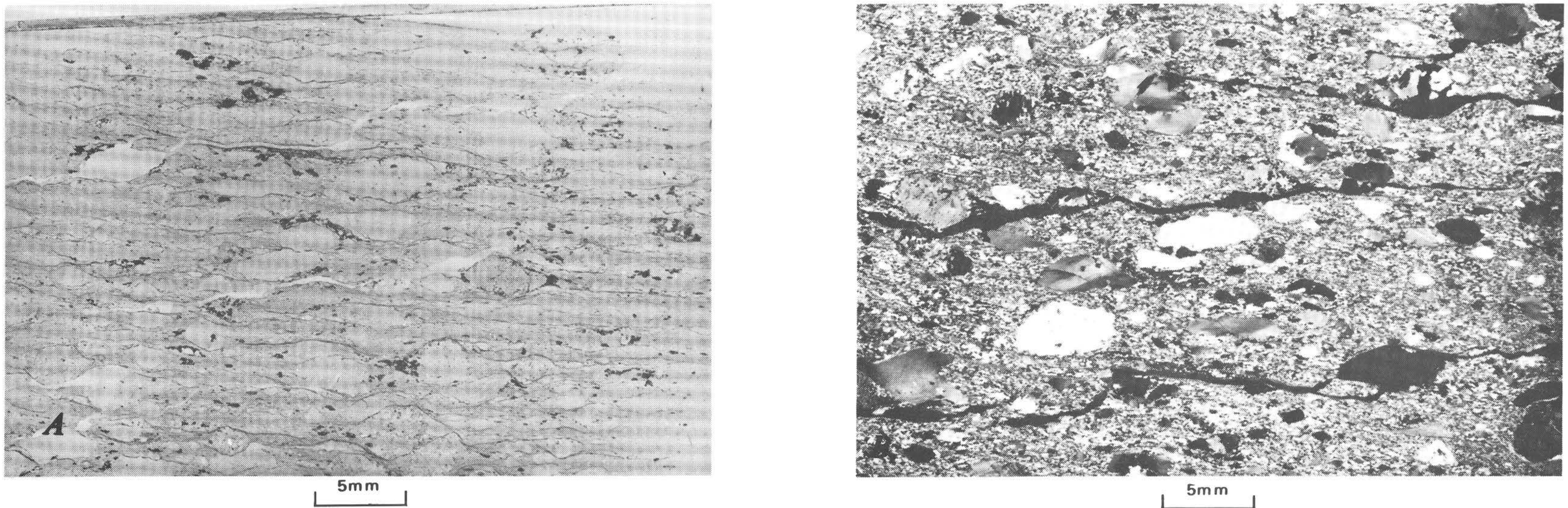


Figure 20.—Foliated quartzite unit close to contact with the grit unit in plane (a) and crossed (b) polarized light. Sample 79AWr356, Circle C-6 quadrangle (fig. 17, no. 3). Estimated mode: quartz (55 percent), plagioclase (20 percent), white mica (15 percent), opaque minerals (10 percent).

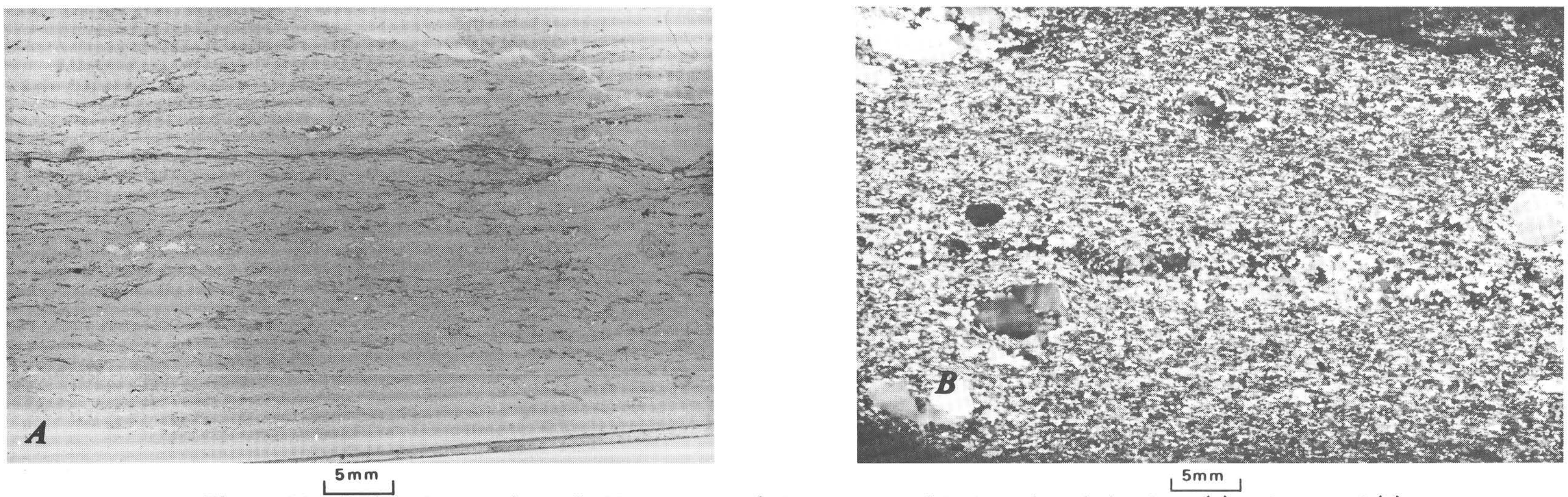


Figure 21.—Foliated quartzite unit 20 km south of the contact with the grit unit in plane (a) and crossed (b) polarized light. Sample 81AFr74D, Circle C-6 quadrangle (not shown on map). Estimated mode: quartz (60 percent), plagioclase (15 percent), opaque minerals (15 percent), white mica (10 percent).

texture, but it apparently has been more annealed because the quartz and feldspar grains are polygonal and boundaries are sutured. The metamorphic grade increases to the southeast. The foliated quartzite unit may lie along the sole of a thrust and have been subjected to greater depths of burial and higher temperatures of recrystallization with increasing distance south and east of the contact. However, mylonitic textures in quartzite may also be produced by plastic deformation during axial symmetric strain (Tullis and others, 1973) rather than by shear strain, and thus not all of the mylonitic textures in the foliated quartzite at a distance from the contact may be due to thrusting.

The possibility of a thrust relationship between the grit unit and the foliated quartzite unit has been recognized by others. Mertie (1937, p. 69) suggested that in the Livengood quadrangle (adjacent to the Circle quadrangle on the west), the grit unit is separated from the more highly metamorphosed rocks (Birch Creek Schist) by a thrust. He also suggested (1937, p. 54) that much of the deformation in the foliated quartzite unit might be due to thrusting from the southwest. Churkin and others (1982) postulate a major structural break between the grit (Beaver terrane) and the foliated quartzite (Yukon crystalline terrane) and suggest thrusting from the southeast.

Another hypothesis is that the grit unit and foliated quartzite unit form a continuous section without a major structural break (F. R. Weber, U.S. Geological Survey, oral commun., 1982). This interpretation is supported by the fact that the protoliths of the two units may have been similar. However, the bimodal appearance of the foliated quartzite may be formed by cataclasis of an equigranular sandstone (for example, fig. 9.27 in Hobbs and others, 1976) or a felsic igneous rock (Higgins, 1971), rather than from recrystallization of a poorly sorted sandstone. At the contact of the two units in the Circle C-5 quadrangle, there is a sharp textural discontinuity, and preliminary petrographic studies indicate that the grit unit there is somewhat richer in quartz and poorer in feldspar than the foliated quartzite unit. We presently prefer the interpretation that the grit and foliated quartzite units are tectonically juxtaposed, but other alternatives cannot be ruled out at the present stage of our investigation.

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Reviewed by R. M. Chapman and W. W. Patton, Jr.

WEST-CENTRAL ALASKA

(Figure 22 shows areas discussed in this section.)

GEOCHEMICAL SURVEY OF THE SOLOMON AND BENDELEBEN QUADRANGLES, SEWARD PENINSULA, ALASKA

By Harley D. King

A reconnaissance geochemical survey of the Solomon and Bendeleben quadrangles began in 1981 and continued in 1982; fieldwork is to be completed in 1983. Areas of previously reported mineral occurrences, as well as areas of known geochemical anomalies, were sampled as part of this survey to further define those areas and to provide reference information for interpretation of anomalies found elsewhere in the two quadrangles. Stream-sediment samples were collected at 1,030 sites, and heavy-mineral concentrates of stream sediments were collected at most of those sites. The <0.18-mm fraction of the stream sediments and the nonmagnetic fraction of the heavy-mineral concentrates were analyzed for 31 elements by semiquantitative spectrography. Stream-sediment samples were also analyzed for As, Au, Bi, Cd, Sb, and Zn by atomic absorption. Analytical data for the <0.18-mm stream-sediment samples collected from 592 sites in 1981 are available in King and others (1982).

Preliminary interpretation of the results indicates the presence of a number of geochemical anomalies in the area of this survey. Most of these reflect previously known mineralization or mineral occurrences. A notably anomalous area is briefly described below.

An area in the western half of the Bendeleben Mountains (fig. 23) is outlined chiefly by anomalous values of Cu, Pb, and Zn in <0.18-mm fractions of stream-sediment samples. (See histograms, fig. 24.) Detailed geochemical investigations by Asher (1970) and Bundtzen (1974) identified geochemical anomalies in the area of the Bendeleben Mountains west of the Pargon River. The present study extends the area of anomalous values of Cu, Pb, and Zn to the east of the Pargon River and anomalous

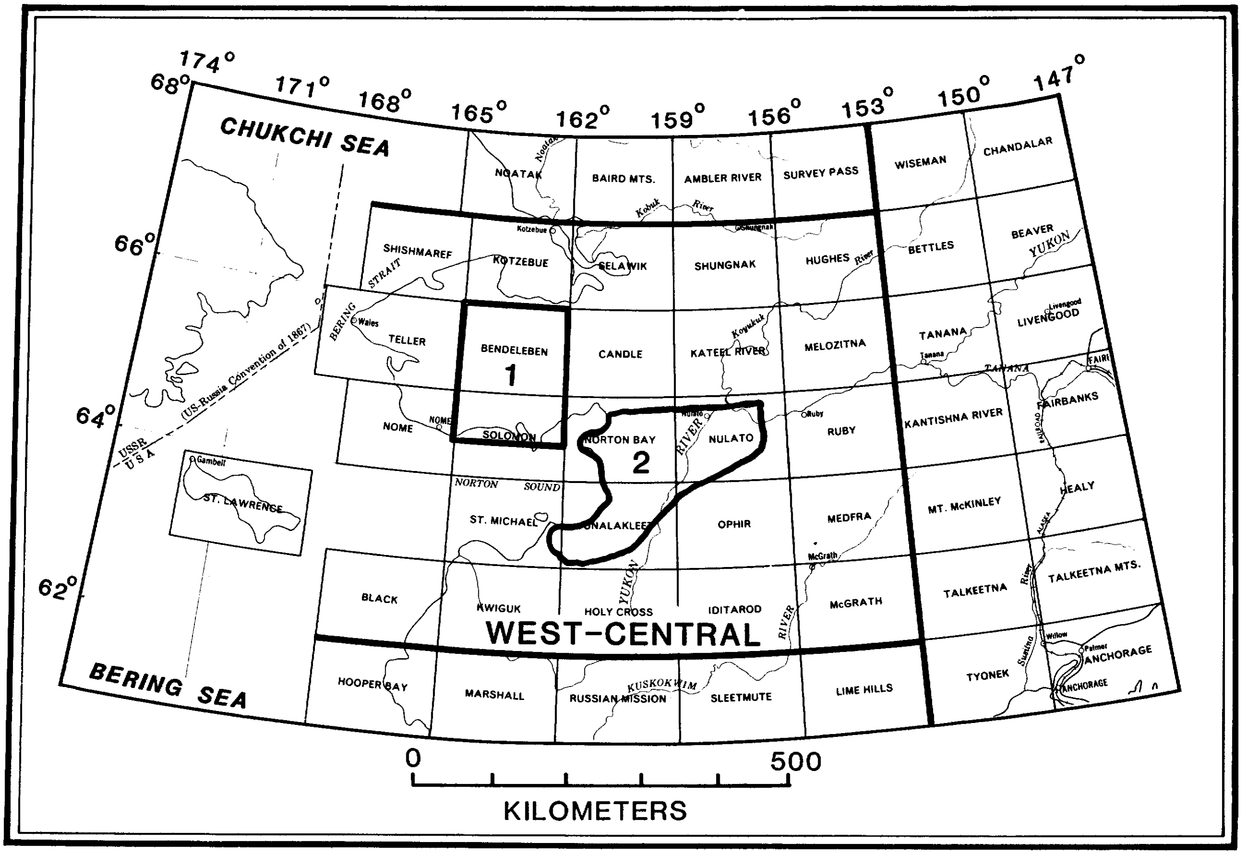

Figure 22.—Areas in west-central Alaska discussed in this circular. Authors and inclusive pages of their articles are: (1) King, p. 33-37; (2) Nilsen and Patton, p. 37-40.

values of Cu and Zn farther eastward to Boston Creek. The area is underlain chiefly by Precambrian metamorphic rocks that are intruded by Cretaceous granitic rocks and Cretaceous and Tertiary dikes (Sainsbury, 1974). This part of the Bendeleben Mountains has been uplifted less than adjacent parts (Hudson, 1977). Mineral lode deposits are known in this area, chiefly where higher analytical values were found. These areas are indicated on figure 23. Sainsbury (1974) reported lode deposits along altered northeast-trending dikes in the headwaters of Libby Creek and the Niukluk River, where old prospect pits disclosed notable amounts of galena, chalcopyrite, and sphalerite, as well as secondary copper minerals. Sainsbury also reported widely disseminated galena in the metamorphic rocks along the ridgeline east of the old prospect pits and suggested that the galena may represent the distal edge of a strata-bound lead-zinc-silver deposit. Hudson and others (1977) reported lode claims of undisclosed commodities on upper Boston and Ella Creeks, lode claims for copper in the upper Niukluk River area, and copper prospects near upper Nesbit Creek. The highly anomalous values in the analyses are apparently related to the known mineral occurrences. The significance of lower anomalous values of Cu, Pb, and Zn has not been determined, but they may be related to concealed deposits.

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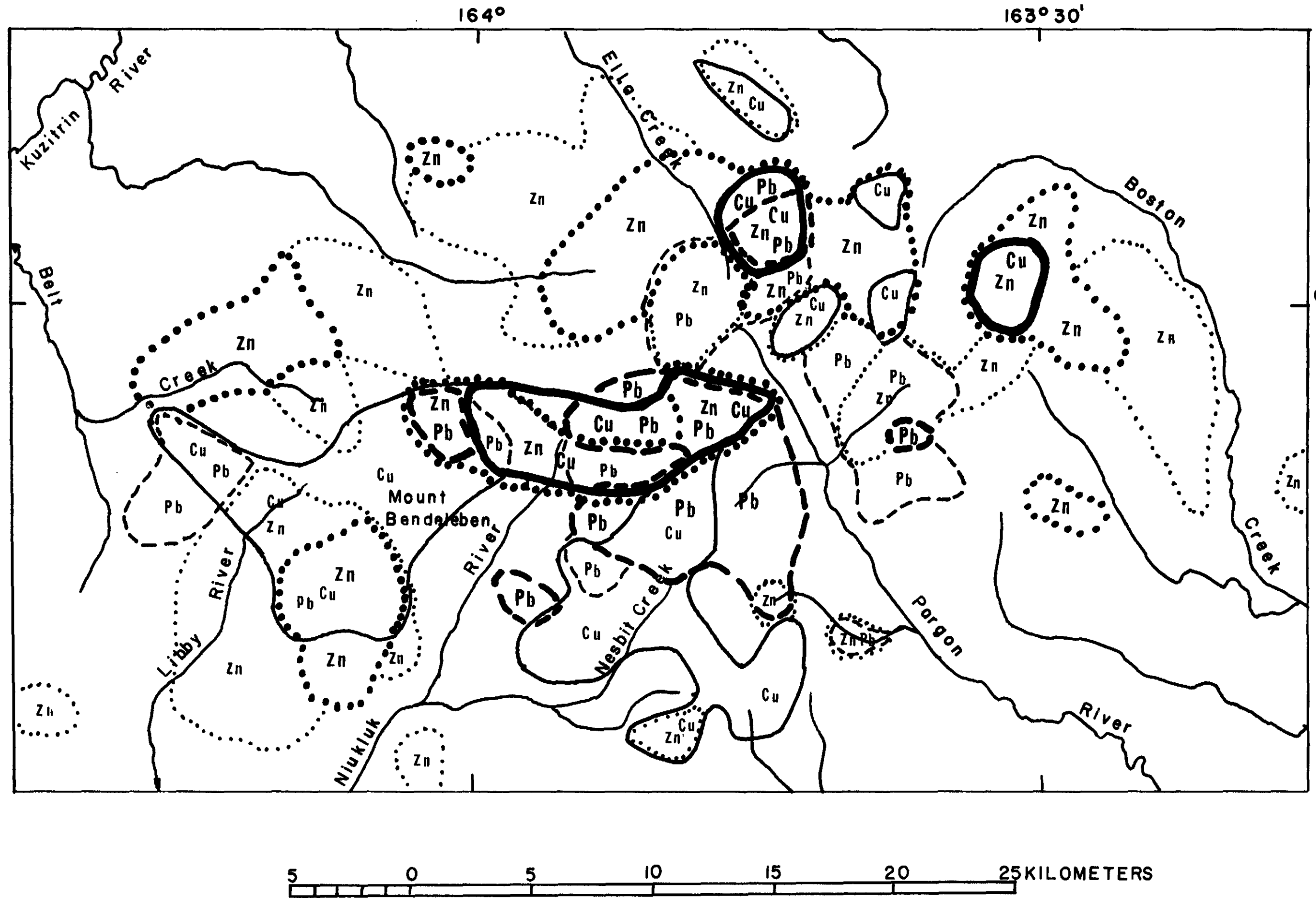
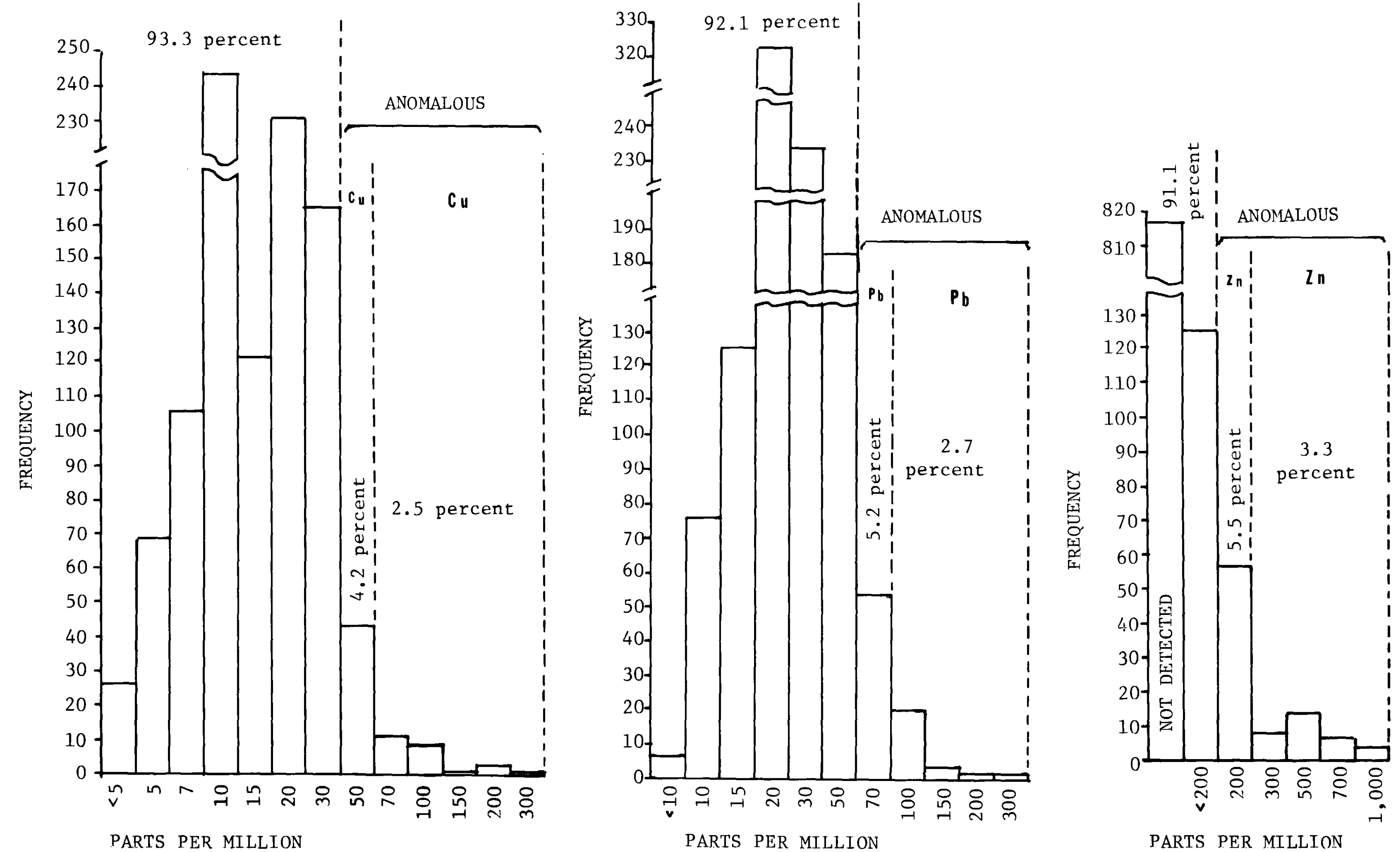


Figure 23.—Distribution of areas of anomalous copper, lead, and zinc in <0.18-mm stream-sediment samples of the base-metal anomaly in the western Bendeleben Mountains, Seward Peninsula, Alaska. Contours enclose areas as follows: heavy solid lines, ≥ 70 ppm Cu; light solid line, 50 ppm Cu; heavy dashed line, ≥ 100 ppm Pb; light dashed line, 70 ppm Pb; heavy dotted line, ≥ 300 ppm Zn; light dotted line, 200 ppm Zn.



Copper in <.18-mm stream-sediment samples

Lead in <.18-mm stream-sediment samples

Zinc in <.18-mm stream-sediment samples

Figure 24.—Histograms for Cu, Pb, and Zn in 1,030 <0.18-stream-sediment samples determined by semiquantitative spectrography, Solomon and Bendeleben quadrangles, Seward Peninsula, Alaska, showing symbols denoting anomalous concentrations, and percentage of total number of samples represented by each range.

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Reviewed by J. E. Kilburn and R. B. Tripp.

CRETACEOUS FLUVIAL TO DEEP-MARINE DEPOSITS OF THE CENTRAL YUKON-KOYUKUK BASIN, ALASKA

By Tor H. Nilsen and William W. Patton, Jr.

The wedge-shaped Cretaceous Yukon-Koyukuk basin occupies much of western Alaska (fig. 25). Older terranes bound it to the north in the Brooks Range, to the southeast in the Ruby geanticline, and to the west in the Seward Peninsula. The basin appears to extend southwestward beneath the Yukon-Koyukuk delta to the Bering Sea. It is transected between Unalakleet and Ruby by the northeast-trending Kaltag fault, which right-laterally offsets its margins 65-130 km (Patton and Hoare, 1968).

An ophiolite assemblage of Mississippian to Jurassic mafic volcanic and intrusive rocks crops out on the perimeter of the Yukon-Koyukuk basin and probably underlies it (Patton and others, 1977). The Cretaceous sedimentary rocks rest on widespread marine andesitic volcanic rocks dated as earliest Cretaceous (Neocomian) by K-Ar radiometric techniques and scattered marine mollusks

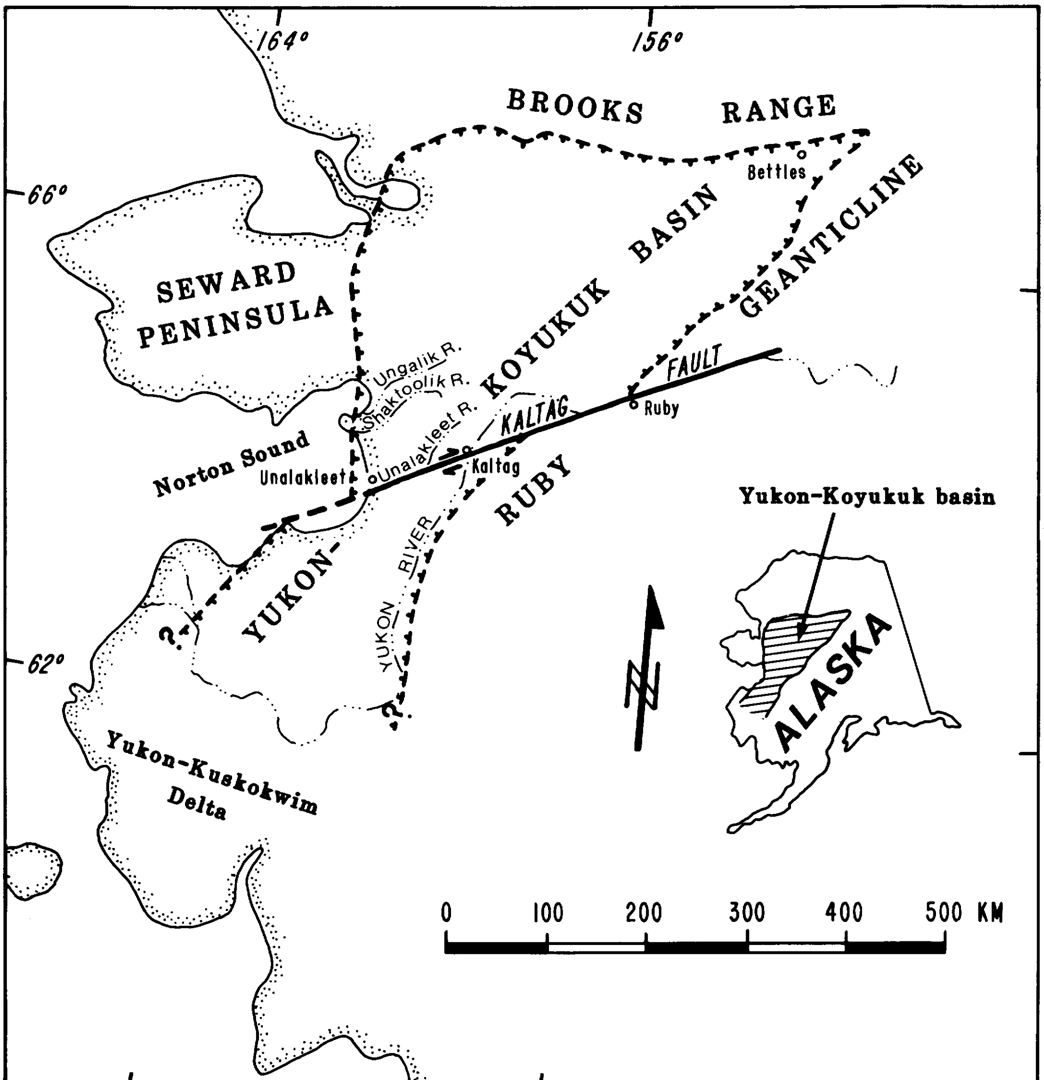

Figure 25.—Index map showing location of Yukon-Koyukuk basin and important geographic features.

(Patton, 1973). The Cretaceous sedimentary rocks are uncomformably overlain by Upper Cretaceous and lower Tertiary felsic volcanic rocks that have K-Ar dates of 85-58 m.y. and are locally interbedded with nonmarine sedimentary rocks (Patton, 1973).

The Cretaceous sedimentary strata underlie about half of the Yukon-Koyukuk basin (fig. 25). They contain scattered Middle Cretaceous (Albian and Cenomanian) marine fossils (Patton, 1973). The strata form two vaguely defined structural basins that are separated by a broad east-trending structural high underlain by Lower Cretaceous volcanic rocks. The Cretaceous sedimentary rocks have been extensively folded and faulted (Patton, 1973).

Patton (1973) recognized four major lithologic groups within the Middle Cretaceous deposits: (1) volcanogenic marine graywacke turbidites and mudstone of probable Albian age, more than 6,000 m thick, deposited in the two subbasins; (2) shallow-marine Albian to Cenomanian(?) calcareous graywacke that is locally conglomeratic, possibly as thick as 1,500 m, deposited along the western margin of the basin; (3) shallow-marine and nonmarine Albian to Cenomanian sandstone, siltstone, shale, and coal, at least 3,000 m thick, deposited in the southeastern part of the basin; and (4) nonmarine quartz conglomerate, at least 900 m thick, deposited along the northern and northeastern margins of the basin. In order to better understand the sedimentary history and paleogeography of the basin, we examined outcrops of the Cretaceous strata along the coast of Norton Sound north and south of Unalakleet, along the Shaktoolik and Unalakleet Rivers, and along the Yukon River between Ruby and Kaltag (fig. 25). As a result, we obtained sedimentologic information from the first three of Patton's lithologic groups that enables us to reconstruct the distribution of facies in the central part of the basin.

Nonmarine and shallow-marine facies are present along the western margin of the Yukon-Koyukuk basin north of Unalakleet (fig. 26). The westernmost or most proximal facies, exposed along the Ungalik River, consists of red-weathering, poorly sorted, subangular to well-rounded pebble to cobble conglomerate that contains clasts of volcanic rocks as long as 20 cm and scattered wood fragments. These east-transported conglomerates appear to have been deposited by streamflow and

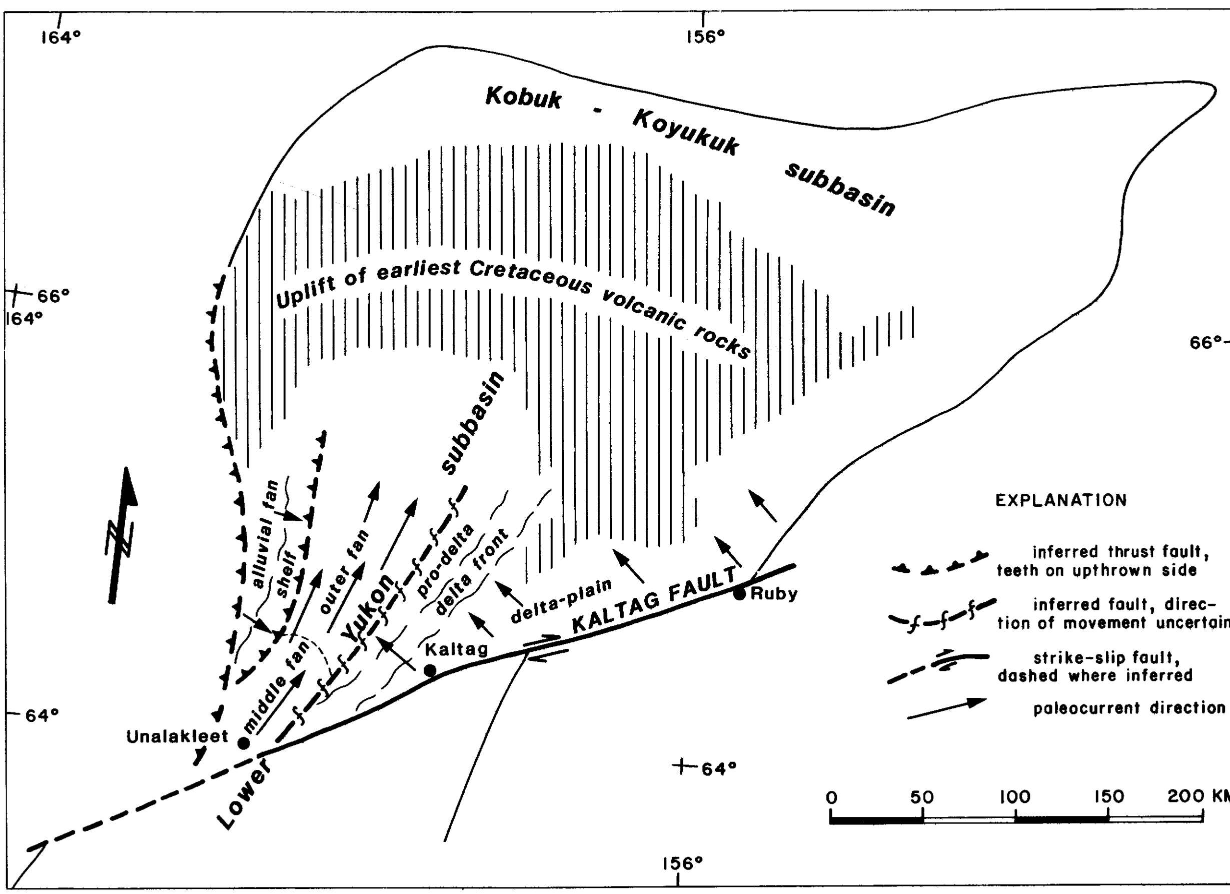

Figure 25.—Present distribution of depositional facies and transport directions for the central Yukon-Koyukuk basin.

debris-flow processes, probably on small alluvial fans adjacent to basin-margin highlands.

To the east and overlying the marginal alluvial-fan deposits are exposures of fan-delta and inner- to outer-shelf facies. The fan-delta sequences consist of intertonguing east-transported conglomerate and dominantly south-transported trough-cross-stratified calcareous sandstone. The conglomerate beds are commonly channeled, contain some debris-flow deposits with clasts as large as 50 cm, are largely quartzose, and are locally well sorted. The sandstone beds are pebbly, coarse to very coarse grained, have trough cross-strata with amplitudes as great as 1 m, and contain some oscillation ripple markings in fine-grained laminae. The conglomerate appears to have been transported by rivers into a high-energy shallow-marine setting, where it was reworked and redeposited by variably oriented currents.

Shelf deposits to the east consist of trough-cross-stratified, planar-stratified, and hummocky-cross-stratified fine- to medium-grained sandstone interbedded with mudstone. These deposits contain abundant indications of bioturbation, current and wave ripples, climbing ripples, and, locally, convolute laminations, contorted stratification, sole markings, and sand volcanoes probably indicative of storm redeposition. Paleocurrents from these deposits generally indicate sediment transport to the northeast, east, and southeast.

Along the Shaktoolik River (Patton and Bickel, 1956a), east of the nonmarine and shallow-marine deposits and structurally juxtaposed with them, are outer-fan and fan-fringe turbidites that were deposited by northeast-flowing turbidity currents. These turbidites form two northeast-trending belts: a western belt of graywacke and mudstone rich in carbonate detritus and an eastern belt of graywacke and mudstone rich in laumontitized volcanic detritus (Hoare and others, 1964). To the southwest, in coastal exposures north of Unalakleet and in inland exposures along the Unalakleet River, correlative turbidites consist of middle-fan channel and related overbank facies that were deposited by northeast-flowing turbidity currents.

The regional distribution of facies associations in the west-central part of the basin suggests that inner-fan deposits may be present farther to the southwest and basin-plain deposits to the northeast. However, the slope facies association, which should separate the shelf deposits to the east from the turbidite assemblage, is missing, probably as a result of post-depositional structural shortening of the basin.

Along the Yukon River between Ruby and Kaltag, the Middle Cretaceous sequence consists of a very thick assemblage of fluvio-deltaic and shallow-marine deposits (Patton and Bickel, 1956b). Coals are locally abundant in flood-plain and marginal-marine deposits, and megafossils are abundant in inner-shelf deposits. The western margin of the delta plain probably migrated considerably as a result of local progradation and retreat of different parts of the delta. Nonmarine debris-flow and braided-stream deposits are most common in the eastern part of this area, delta-plain meandering-stream deposits in the central, and delta-front deposits in the western part. Paleocurrents from the fluvio-deltaic deposits indicate that the direction of sediment transport was toward the north and west and that the source area was probably located to the southeast.

The depositional facies and paleocurrent patterns along the western and southeastern margins of the Yukon-Koyukuk basin suggest transport of sediment toward the center of the basin. The fluvio-deltaic and shelf deposits are juxtaposed with deep-sea fan facies deposited by northeast-flowing currents in the center of the basin, without preservation of intermediate slope deposits. This relation suggests that the Yukon-Koyukuk basin has been subjected to major post-Cenomanian structural shortening that was probably oriented northwest-southeast. Patton and Tailleur (1977) cite evidence for a strong east-west compressional event in the Bering Strait region in the Late Cretaceous or early Tertiary. They attributed the north-south folding and eastward-directed thrusting along the western margin of the Yukon-Koyukuk basin to this compression event. The crustal shortening of the basin indicated by our reconstruction of the Cretaceous sedimentation suggests that the effects of this compressional event may have extended eastward at least as far as the central part of the basin.

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SOUTHERN ALASKA

(Figure 27 shows study areas discussed in this section.)

EARLY PROTEROZOIC METAVOLCANIC ROCKS IN THE JARVIS CREEK GLACIER TECTONOSTRATIGRAPHIC TERRANE, MOUNT HAYES C-6 QUADRANGLE, EASTERN ALASKA RANGE, ALASKA

By John N. Aleinikoff and Warren J. Nokleberg

The Jarvis Creek Glacier tectonostratigraphic terrane occurs in a fault-bounded block immediately north of the Hines Creek fault in the Mount Hayes quadrangle (fig. 28). The terrane, along with other terranes in the quadrangle, occurs in an area formerly called the Yukon-Tanana terrane (Jones and others, 1981). The terrane extends across the entire northern part of the quadrangle and appears to extend at least 40 km westward into the Healy quadrangle (Sherwood and Craddock, 1979) and for at least 100 km eastward into the Tanacross and Nabesna quadrangles (Richter, 1976; Foster, 1970). In the Mount Hayes quadrangle, the terrane is structurally juxtaposed to the south against the Hayes Glacier terrane, which was locally formerly called the Pingston terrane by Jones and others (1981) and Aleinikoff and Nokleberg (1984). This terrane was renamed the Hayes Glacier terrane because it is significantly different from the Pingston terrane in the type area as described by Jones and others (1981).

In the western part of the Mount Hayes quadrangle (fig. 28), the terrane consists predominantly of varying proportions of muscovite-chlorite schist, calc-schist, meta-andesite to metarhyolite tuff and flows, metagraywacke, less quartzite or metachert, and sparse marble. The terrane is intensely deformed; it has abundant major and minor appressed to isoclinal folds that are generally north verging and have well-developed axial plane schistosity, as well as locally abundant major and minor faults. Axial planes of folds, schistosity, and faults are parallel to the Hines Creek fault and generally strike west-northwest and dip 35° to 40° south. The pervasive occurrence of metamorphic white mica, chlorite, and biotite indicates middle greenschist facies metamorphism. Locally abundant massive sulfide deposits occur in the terrane in the area west of Hayes Glacier at the locality for sample 81ANK 233A (fig. 28). These massive sulfide deposits are interlayered with the metavolcanic rocks and consist mainly of pyrite and pyrrhotite with sparse chalcopyrite and galena.

In order to better define and describe this terrane and associated massive sulfide deposits, we sampled a metarhyodacite (sample 81ANK 233A) from an area about 1.6 km north of the Hines Creek fault for zircon U-Pb geochronologic studies. The metarhyodacite contains relict phenocrysts of plagioclase, potassium feldspar, and quartz in a schistose fine-grained matrix of chlorite, white mica, feldspar, and quartz. Relict feldspar and quartz phenocrysts generally are intensely deformed into microaugen. An igneous parentage for the metarhyodacite is indicated by abundant complicated twinning in plagioclase, local preservation of normal and delicate oscillatory zoning in plagioclase, euhedral outlines of feldspar, and, rarely, resorbed embayments in quartz.

Zircons extracted from sample 81ANK 233A were divided by hand picking into two populations based on color: dark pinkish-red and gray. In transmitted light, the gray grains are opaque, whereas the red grains are pale pink. A careful analysis of zircon morphology was made in order to interpret age data from the sample. Previously, the only Early Proterozoic protolith ages in east-central Alaska have been either from Mississippian plutonic rocks with 2.2-b.y.-old inheritance (Aleinikoff and others, 1981) or from detrital zircons from Precambrian(?) or Paleozoic(?) metasedimentary rocks (Aleinikoff and others, 1984). Most of the zircons are subhedral to euhedral, although the red grains tend to be better preserved than the gray grains. Whereas some of the gray grains are fragmented and rounded, nearly all of the red ones have crystal faces. The red zircons appear to be igneous zircons that have not been recycled; they have none of the detrital characteristics of zircons obtained from quartzites in the area. Thus, the isotopic data from the red grains should determine the age of the protolith of the metarhyodacite. The gray zircons are metamict; previous experience with metamict zircons suggests that the Pb/U ages are more discordant and the data are more difficult to interpret.

One size fraction of gray zircon and three size fractions of red zircon were analyzed for U and Pb concentrations and Pb isotopic composition, following a method modified from Krogh (1973). The red size fractions have Pb/U ages ranging from about 1,370 to 1,730 m.y. and have $^{270}Pb/^{206}Pb$ ages tightly grouped between 1,978 and 2,002 m.y. They contain fairly low U concentrations, between about 350 and 500 ppm, and are only about 25 percent discordant, i.e., the $^{206}Pb/^{238}U$ age is about 25 percent lower than the $^{207}Pb/^{206}Pb$ age. In marked contrast, the gray size fraction contains about 2,000 ppm U and has Pb/U ages of 425 and 652 m.y. and

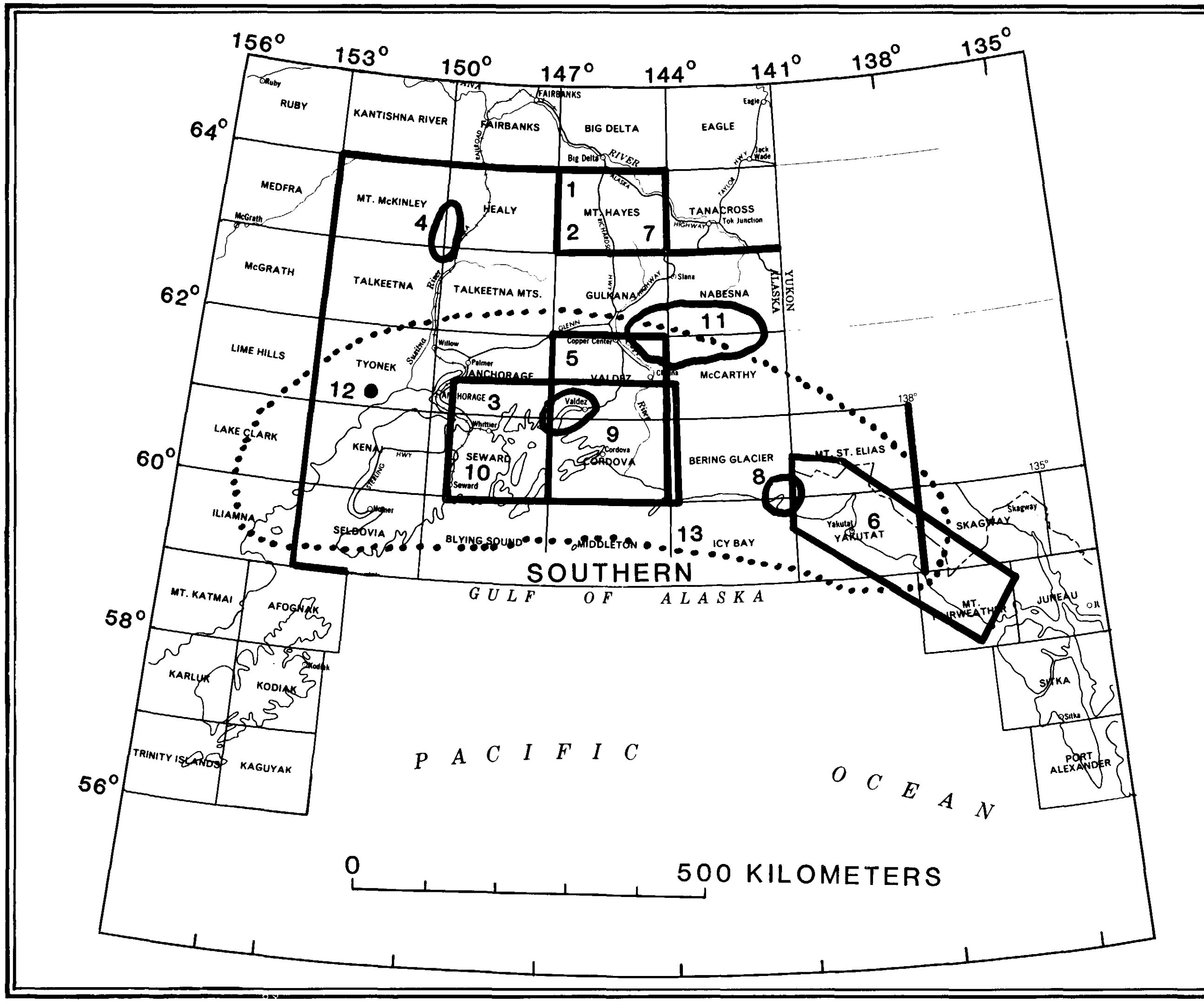

Figure 27.—Areas of southern Alaska discussed in this section. Authors and inclusive pages of their articles are: (1) Aleinikoff and Nokleberg, p. 40-44; (2) Campbell and Nokleberg, p. 44-47; (3) Goldfarb and others, p. 47-50; (4) Hillhouse and Csejtey, p. 50-52; (5) Miller, Dumoulin, and Nelson, p. 52-57; (6) Nilsen and others, p. 57-60; (7) Nokleberg, Lange, and Roback, p. 60-65; (8) Page and others; p. 65-67; (9) Pickthorn, p. 67-70; (10) Pickthorn and Nelson, p. 70-71; (11) Richter and others, p. 71-75; (12) Severson and Gough, p. 75-78; and (13) Stephens, Lahr, and Page, p. 78-82.

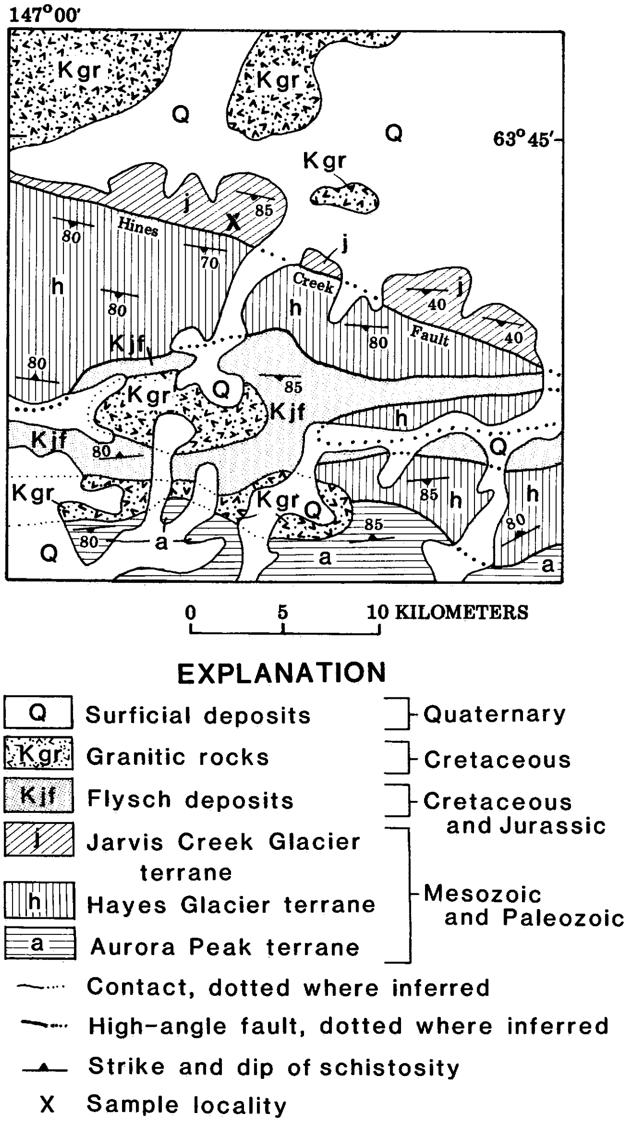

Figure 28.—Simplified geologic map of parts of the Mount Hayes C-5 and C-6 quadrangles, eastern Alaska Range, Alaska. X is location of sample 81ANK 233A.

a $^{207}Pb/^{206}Pb$ age of 1,545 m.y. This fraction is about 75 percent discordant, undoubtedly due to Pb loss caused by a combination of its high U concentration and Phanerozoic thermal events. The high U content has caused metamictization which is responsible for the opacity of the gray grains.

Isotopic data are plotted on a concordia diagram (fig. 29) and are listed in table 6. Of the red grains, the (+100)RED fraction is the least discordant and has the lowest U concentration and the youngest $^{207}Pb/^{206}Pb$ age, so that a best-fit line through the three points has an upper intercept age of about 2,000 m.y. but a geologically impossible lower intercept age of -304 ± 69 m.y. This array can be

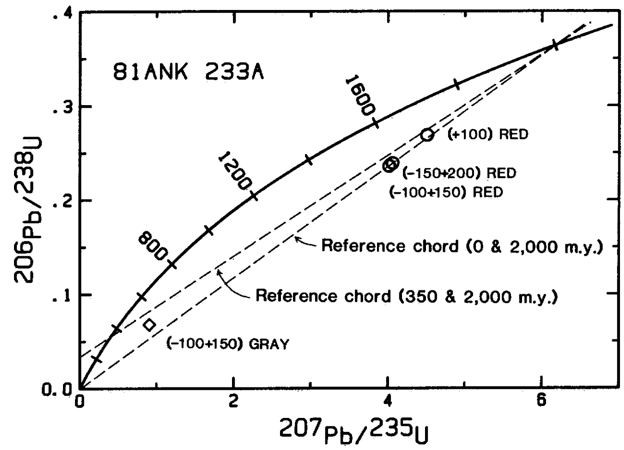

Figure 29.—Concordia plot of four data points from sample 81ANK 233A. Reference chord between 0 and 2,000 m.y. passes through the two finer size fractions, suggesting an extrusive age of 2,000 m.y., with an amount of discordance being due to modern dilatancy lead loss. The (+100)RED and (-100+150) GRAY data points plot within a triangle formed by the previously mentioned reference chord, a reference chord between 350 and 2,000 m.y., and a segment of the concordia curve. This relation suggests both Mississippian and modern Pb loss.

caused by two possibilities: (1) the (+100)RED fraction lost slightly more radiogenic Pb than the two finer fractions sometime in the Phanerozoic (probably about 350 m.y. ago, as this is a well-documented plutonic-metamorphic event in this region) (Aleinikoff and others, 1981; 1984); or (2) the finer grain sizes contain slightly more inherited radiogenic Pb than the (+100)RED fraction. The first possibility implies that the protolith of the metarhyodacite had a minimum age of about 2,000 m.y., whereas the second possibility implies a maximum age of 1,980 m.y. This problem cannot be resolved without additional geochronologic work on other samples. However, we favor the former possibility, which suggests that the present coordinates of the data points for the (-100+150) RED and (-150+200)RED fractions are due primarily to modern dilatancy Pb loss, and thus the $^{207}Pb/^{206}Pb$ age of about 2,000 m.y. for these fractions is interpreted as the extrusive age of the rhyodacite protolith. This is shown graphically in figure 29 where the two finer grained fractions plot on the reference chord between 0 and 2,000 m.y. The (+100)RED and (-100+150)GRAY fractions probably have suffered at least two episodes of Pb loss, perhaps at about 350 m.y. and the present (fig. 29).

Previous geochronologic studies show that this region was affected by several Phanerozoic plutonic events, including at least one major regional metamorphism at about 350 m.y., and at least two

Table 6.—U/Pb isotopic data for zircons from sample 81ANK 233A (location: 63°42'51" N.; 146°45'22" W)

Sample	Concentrations (ppm) U	Concentrations (ppm) Pb	Atomic percent ^{204}Pb	Atomic percent ^{206}Pb	Atomic percent ^{207}Pb	Atomic percent ^{208}Pb	Ages (m.y.) $^{206}Pb/^{238}U$	Ages (m.y.) $^{207}Pb/^{235}U$	Ages (m.y.) $^{207}Pb/^{206}Pb$
(+100)RED	354	113	0.10	74.4	10.4	15.1	1,536	1,732	1,978
(-100+150)RED	492	137	0.11	74.9	10.7	14.3	1,367	1,636	2,002
(-150+200)RED	422	112	0.04	77.8	10.1	12.0	1,383	1,645	1,999
(-100+150)GRAY	2012	156	0.04	76.2	7.8	15.9	424	652	1,545

major Mesozoic and Cenozoic tectonic deformations. Therefore, the zircons in sample 81ANK 233A may have undergone several episodes of minor Pb loss, in addition to modern dilatancy Pb loss, so that the exact age of the protolith is obscured. However, based on both zircon morphology and the tight grouping of the $^{207}Pb/^{206}Pb$ ages of the three red fractions, we conclude that the age of the protolith is about 2,000 m.y.

The Early Proterozoic age of about 2,000 m.y. for this metarhyodacite is unusual and makes it one of the two oldest rocks yet known in Alaska. Other Early Proterozoic ages in Alaska and western Canada include: (1) the Kilbuck terrane of southwestern Alaska with an age of 2,050 m.y. (Turner and others, 1983); (2) a detrital zircon fraction from a quartzite with an approximate age of 1,900 m.y., possibly derived from rocks similar to sample 81ANK 233A, in the north-central part of the terrane in the Mount Hayes quadrangle (J. N. Aleinikoff, unpub. data); (3) an approximate 1,960-m.y. age for a gneiss dome forming part of the basement rocks of the Shuswap metamorphic complex, British Columbia (Wanless and Reesor, 1975); (4) an approximately 1,970-m.y. event affecting granitic rocks in the Bear-Slave province boundary, Northwest Territories (Frith and others, 1977); (5) a 2,000-m.y. event affecting Pb isotopes in Ni sulfides in the Churchill-Superior province boundary in Manitoba (Cumming and others, 1982); and (6) an initial rifting event with maximum age of 2,010 m.y. in the Wopmay Orogen, Northwest Territories (Bowring and Van Schmus, 1982). The 2,000-m.y. age for the metarhyodacite, and presumably for the associated more mafic and siliceous metavolcanic rocks in this part of the terrane indicates a period of Andean-type arc volcanism of Early Proterozoic age. This period of volcanism constitutes an important part of the geologic history for this part of the terrane and may be extremely useful in reconstruction of terranes in Alaska and western Canada.

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Reviewed by Cynthia Dusel-Bacon and Marvin A. Lanphere.

MAGNETIC PROFILE ACROSS ACCRETED TERRANES, MOUNT HAYES QUADRANGLE, EASTERN ALASKA RANGE, ALASKA

By David L. Campbell and Warren J. Nokleberg

The Mount Hayes quadrangle in the eastern Alaska Range contains a series of accreted tectonostratigraphic terranes (Jones and others, 1981; Nokleberg and others, 1982). South of the Denali fault, these terranes are juxtaposed along several thrust faults that trend subparallel to the Denali fault. These thrusts include the Broxon Gulch thrust, which forms the boundary between the Wrangellia and Maclaren terranes, and the Eureka Creek thrust, which separates two subterranes of Wrangellia, the Slana River subterrane on the north and the Tangle subterrane on the south (fig. 30a).

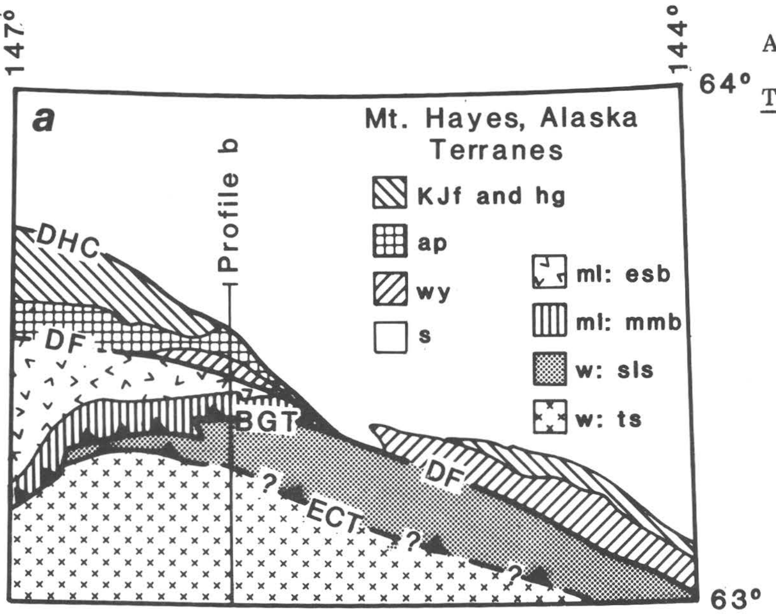
The Maclaren terrane is made up of penetratively deformed and regionally metamorphosed plutonic rocks, and schist, phyllite, argillite and metagraywacke. The northern part of the Maclaren terrane consists of the East Susitna batholith, predominantly metamorphosed diorite and granodiorite with lesser quartz monzonite. The southern part consists of the Maclaren Glacier metamorphic belt, where metasedimentary rocks increase in metamorphic grade northward from lower greenschist facies to amphibolite facies metamorphism.

The Wrangellia terrane (Jones and others, 1981) is predominantly made up of Paleozoic island-arc rocks which are disconformably overlain by the Triassic Nikolai Greenstone and locally by younger rocks. The Slana River subterrane, on the north, consists of andesite flows, breccia, epiclastic rocks, argillite, and limestone of Pennsylvanian to Permian age (Richter and others, 1977). The Triassic Nikolai Greenstone consists of massive, subaerial, amygdaloidal basalt flows. Compared with the Slana River subterrane, the Tangle subterrane on the south contains thinner sequences of upper Paleozoic rocks and thicker sequences of overlying Nikolai Greenstone. The upper Paleozoic rocks of the Tangle subterrane are predominantly aquagene tuff with minor andesite. Extensive gabbro dikes, sills, and stratiform cumulate mafic and ultramafic rock intrude the sedimentary and volcanic rocks of the Tangle subterrane.

A model of magnetic structures (fig. 30b) was made along flight line C41 of the aeromagnetic map of the Mount Hayes quadrangle (State of Alaska, 1974). Figure lb was made using a program for the Hewlett-Packard model 85 desktop computer (Campbell, 1983) and is one of a series being prepared for the AMRAP assessment of the quadrangle. The fields are calculated on a datum (dashed line) which drapes topography at 0.3048 km, the nominal position of the survey aircraft. All bodies are presumed magnetized in the direction of the Earth's present-day field, of 57,000 nT magnitude, 76° inclination, and 28° easterly declination. The effective magnetic susceptibility for each body is listed in table 7. For several bodies, particularly the Nikolai Greenstone and ultramafic rocks, the effective magnetic susceptibility includes both an induced ("true" susceptibility) component and an additional viscous Natural Remanent Magnetization (NRM) component that is parallel to the Earth's present field and of magnitude comparable to the true component. J. W. Hillhouse (U.S. Geological Survey, oral commun., Dec. 1982) has measured such NRM components on samples of the Nikolai Greenstone from the Mount Hayes and other nearby quadrangles.

In the calculations, the bodies shown were assumed to extend unchanged and perpendicular to the plane of the section for varying strike-length distances Y1 and Y2 (table 7) and then to be vertically cut off (the "2 1/2-dimensional geometry" of Shuey and Pasquale, 1973). The strike lengths were chosen to match strike extents of corresponding anomalies on the aeromagnetic map.

Contacts shown on this section are thought to be correct to ± 20,, at least near the surface. Errors in attitude modelling can be caused by using wrong strike lengths, by differences in shape from the idealized prism supposed by the program, by nonuniform susceptibilities of source bodies, and by



ABBREVIATIONS USED IN THIS FIGURE:

Terranes and subterranes
- w = Wrangellia terrane
- ts = Tangle subterrane
- sls = Slana River subterrane
- ml = Maclaren terrane
- mmb = Maclaren Glacier metamorphic belt
- esb = East Susitna batholith
- wy = Windy terrane
- ap = Aurora Peak terrane—formerly the McKinley terrane of Jones and others (1981) in this quadrangle
- KJf = Deformed upper Mesozoic flysch
- hg = Hayes Glacier terrane—formerly the Pingston terrane of Jones and others (1981) in this quadrangle
- s contains many subterranes not shown here

Faults
- DF = Denali fault
- DHC = Hines Creek strand of Denali fault
- BGT = Broxon Gulch thrust
- ECT = Eureka Creek thrust

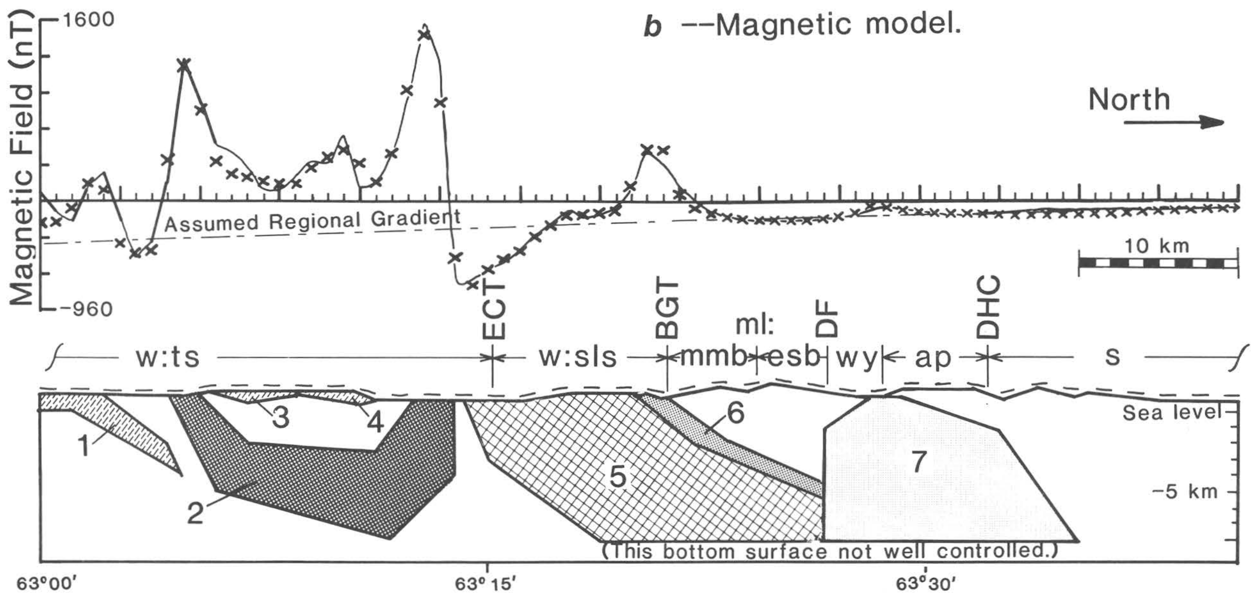


Figure 30.—(a) Sketch map of Mt. Hayes quadrangle showing boundaries of selected terranes and location of the profile shown in figure 30b. (b) Model of magnetic bodies along the profile. The cross section in the lower part of the figure shows the asssumed magnetic structures. In the upper part of the figure, a plot comparing the observed magnetic field (solid curve) and calculated magnetic field (line of x's) that results from these structures is shown. On the cross section, darker shading indicates rock units having higher susceptibilities, and the unshaded areas represent non-magnetic or very low susceptibility rocks. No vertical exaggeration.

Table 7.—Strike lengths used in computing the model. Y1 = eastward extent of body, and Y2 = westward extent of body, perpendicular to plane of the figure.

Body no.	Susceptibility (cgs units x 10^{-3})	Y1 (km)	Y2 (km)
1	5.0	9	9
2	7.0	10	30
3	3.5	0.3	12
4	3.5	0.3	0.3
5	1.5	12	15
6	5.0	10	6
7	0.5	6	6

survey aircraft not being in the assumed position. Details of the bodies at depth are not well controlled by the modelling process, and their bottoms (particularly for bodies 5, 6, and 7) could well be half as deep or twice as deep as shown.

Despite the above uncertainties, several features are apparent in the magnetic model. First, the thick high-susceptibility unit (body 2) making up part of the Amphitheatre Syncline is known from sparse outcrops along its northern flank to be cumulate mafic and ultramafic rocks. The aeromagnetic map of the quadrangle shows that body 2 is far more extensive than was indicated by geologic mapping alone. Bodies 1, 3, and 4 appear from outcrops to be the Nikolai Greenstone. The horizontal attitude of bodies 3 and 4 in the aeromagnetic data indicates the disconformity known to exist between the Nikolai Greenstone and underlying rocks. The north-dipping tongue of body 1, not well constrained by the modelling, suggests either a fault between it and body 2, or else that there are separate units of the Nikolai Greenstone above and below the cumulative mafic and ultramafic unit. The former interpretation is preferred because Nokleberg and others (1982) have mapped a major thrust fault between bodies 1 and 2.

The position of the Eureka Creek thrust shown on the figure is not precisely controlled geologically at the place where it cuts the section. Assuming that the Eureka Creek thrust forms the southern boundary of body 5, figure lb suggests that its true position may be 1-2 km south of the marked location. Body 5 then would represent most or all of the Slana River subterrane, a moderately magnetic sliver of island-arc volcanic and sedimentary rocks. The sole of the thrust fault is assumed to be 8 km below sea level, but this depth is not known with certainty.

The Broxon Gulch thrust appears in this model to be dipping about 30° north and to be associated with a high-susceptibility sheet (body 6). This sheet lies entirely in the Slana River subterrane south of the thrust fault near the profile and also occurs a few kilometers southeast of the profile in the Slana River subterrane but farther from the mapped position of the thrust. Thus, body 6 may represent one of several fault-bounded units of cumulative mafic and ultramafic rocks in the area that dip moderately north (Nokleberg and others, 1982).

Body 7 is a weakly magnetized zone that straddles an unnamed fault separating the Windy terrane from the Aurora Peak terrane. The shape and structure of this body or complex of bodies are not well determined magnetically due to its low susceptibility. The northern and southern edges, which may represent the Hines Creek strand of the Denali fault and the Denali fault, respectively, may be substantially different from that shown in figure 30 but still provide a fit to the observed magnetic field.

General features on this magnetic model are similar to features on the cross section by Nokleberg and others (1981) based solely on geologic mapping. Magnetic modelling, as shown by this example, may provide a relatively inexpensive tool to test the accuracy of such cross sections.

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Reviewed by John Cady and Carol Finn.

A STATISTICAL INTERPRETATION OF GEOCHEMICAL DATA FROM CHUGACH NATIONAL FOREST

By Richard J. Goldfarb, Peter F. Folger, Suzanne M. Smaglik, and Richard B. Tripp

A geochemical reconnaissance survey of the Chugach National Forest (fig. 31), southcentral Alaska, was conducted between 1980 and 1982 to identify regions with favorable mineral resource potential. Sediment and heavy-mineral, panned concentrate samples were collected at 2,200 sites on stream channels and active medial and lateral moraines. Each sample of sediment and heavy-mineral concentrate (nonmagnetic fraction) was analyzed semiquantitatively for 31 elements using an optical emission spectrograph.

R-mode factor analysis with varimax rotation was used to define the geochemical associations in both the sediment and concentrate data bases; this analysis places similarly behaving experimental variables (elements) into groups termed "factors." Specific types of ore deposits may contain an anomalous geochemical signature having a distinct suite of trace elements. Therefore, certain factors might be used to define the deposit types in the study area.

Table 8 shows the factor loadings for the four factors determined to be significant (eigenvalues greater than one) in the sediment data base. Factor loadings depict the influence of each factor on a variable and may be interpreted similarly to correlation coefficients. Related to the loadings are the factor scores, which will be discussed below. The scores for each sample on each factor measure the "effect" of that factor on each individual sample.

Factor 1 loadings (table 8) were interpreted as defining sediments derived from submarine tholeiitic basalt in the Chugach National Forest. Samples with high scores for factor 1 loadings cover most of Knight Island, where copper deposits are known to occur in ophiolite sequences; chalcopyrite was observed in most of the corresponding heavy-mineral concentrates. Knight Island has the highest geochemical favorability in the study area for submarine exhalative deposits in volcanic host rocks.

High scores onto factor 1 also characterize the Nuchek Creek region on Hinchinbrook Island, Solomon Gulch, Brown Basin, Wortmanns Glacier, Heney Glacier, and Allen Glacier. (See fig. 31.) All of these localities have interbedded volcanic rocks in the dominant flysch units; concentrate samples from these locations also contain chalcopyrite. They are thus considered target areas for possible volcanic rock-hosted massive sulfide mineralization. Anomalous scores (above the 95th percentile) for factor 1 are also associated with Glacier Island, Ragged Mountain, part of the Resurrection Peninsula, and an area east of Ptarmigan Lake. The lack of chalcopyrite in concentrate samples from these areas implies that the volcanic rocks are not associated with any significant Cu mineralization.

Factors 2 and 3 (table 8) represent lithogeochemical associations. High scores onto factor 2 define known intrusive bodies and possibly undiscovered plutons from Van Cleve Lake north to the Wernicke River, the Cleave Creek Glacier region, Marshall Pass, and to the south of Wortmanns Glacier. Anomalous scores for the B-Ba association of factor 3 characterize the Mount Alyeska region, much of the Kenai Peninsula, and Montague Island and may represent the more clay-rich sediments in the Orca and Valdez Groups.

Factor 4 (table 8) is associated with sediments in part derived from flysch-hosted massive sulfide bodies. Factor scores show LaTouche Island having the highest favorability for these ore deposits. The Ellamar region including Bligh Island, southeast Knight Island, and a belt east of Orca Inlet, are characterized by signatures similar to that of LaTouche Island. Weaker signatures appear at Thumb Cove on Resurrection Peninsula, southern Chenega Island, and a few locations near Whale Bay.

A strong association with factor 4 is also observed for Montague Island. Anomalous scores from the northern part of the island predominantly reflect Mn- and Zn-enriched sediments. The southern part of Montague Island is characterized by anomalous Zn in sediments and the presence of sphalerite in concentrates. Approximately half the concentrates from the island contain chalcopyrite, which is unusual in that this is the only place in eastern Prince William Sound where abundant chalcopyrite is found in flysch lacking any spatial association with volcanic rocks. These data suggest a possible relation to the Prince William Sound massive sulfide deposits.

R-mode analyses run on the concentrate data resulted in the significant loadings listed in table 9. Factor 1 shows anomalous scores for samples scattered throughout the western part of the national forest, which is underlain by greenstone, flysch and an intrusive body. These samples may be relatively enriched in Ca-Mg silicates.

Factor 2 defines areas of flysch from which the greatest concentrations of sulfides (pyrite, chalcopyrite, galena, arsenopyrite, minor sphalerite) are seen in heavy-mineral concentrates. This sulfide

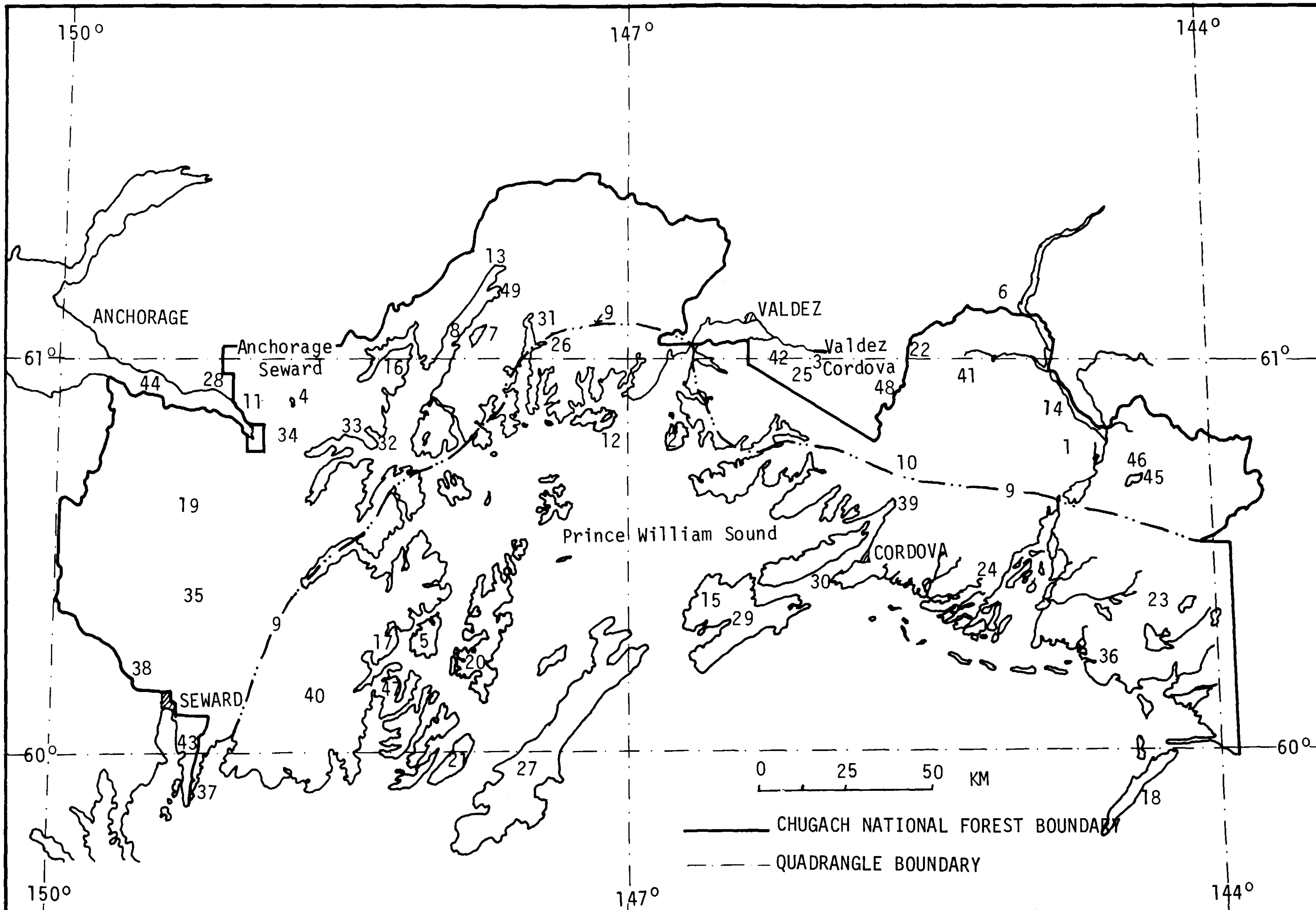

Figure 31.—Map of Chugach National Forest. Numbers refer to locations in the accompanying list.

Location Names (to accompany figure 31)

1. Allen Glacier
2. Bligh Island
3. Brown Basin
4. Carmen Lake
5. Chenega Island
6. Cleave Creek Glacier
7. Coghill Lake
8. College Fiord
9. Contact Fault
10. Cordova Glacier
11. Crow Pass
12. Glacier Island
13. Harvard Glacier
14. Heney Glacier
15. Hinchinbrook Island
16. Hobo Bay
17. Jackpot Point
18. Kayak Island
19. Kenai Peninsula
20. Knight Island
21. LaTouche Island
22. Marshall Pass
23. Martin River Glacier
24. McKinley Peak
25. Meteorite Mountain
26. Miner's Lake
27. Montague Island
28. Mount Alyeska
29. Nuchek Creek
30. Orca Inlet
31. Pedro Glacier
32. Pigot Bay
33. Pigot Glacier
34. Portage-Bear Valley area
35. Ptarmigan Lake
36. Ragged Mountain
37. Resurrection Peninsula
38. Resurrection River
39. Rude River
40. Sargent Icefield
41. Schwan Glacier
42. Solomon Gulch
43. Thumb Cove
44. Turnagain Arm
45. Van Cleve Lake
46. Wernicke River
47. Whale Bay
48. Wortmanns Glacier
49. Yale Glacier

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signature is especially strong in the Crow Pass-Carmen Lake-Pigot Glacier trend, Harvard and Yale Glacier areas in College Fiord, in a belt on the Kenai Peninsula along the western side of the Contact fault, and to the east of Miner's Lake. All these areas are adjacent to areas with favorable Au signatures (R. J. Goldfarb, unpub. data), but they lack samples with anomalous Au values. In the eastern half of the Chugach National Forest, this factor shows favorable geochemical signatures for base metal sulfides near Brown Basin and Meteorite Mountain, Cordova Glacier, an unnamed glacier at the head of the Rude River, and Cleave Creek Glacier.

Table 8.—Factor loadings for the first four factors after varimax rotation of the stream sediment data. Total variance explained by the four factors equalls 66 percent. Loadings less than 0.40 have been omitted.

Factors	1	2	3	4
Fe	0.72	--	--	--
Mg	.80	--	--	--
Ca	.50	--	-0.58	--
Ti	.72	--	--	--
Mn	--	--	--	0.65
B	--	--	.81	--
Ba	--	0.55	.66	--
Be	--	.51	--	.53
Co	.59	--	--	.63
Cr	.82	--	--	--
Cu	.53	--	--	.56
La	--	.66	--	--
Ni	.77	--	--	--
Pb	--	.53	--	.48
Sc	.74	--	--	--
Sr	--	.73	--	--
V	.81	--	--	--
Y	--	.64	--	--
Zn	--	--	--	.66
Zr	--	.75	--	--
Percent of total data variance explained by factor	35.6	14.1	8.9	7.7

Favorable lode and/or placer gold locations are indicated by high factor 4 scores which are located in the Valdez Group on the Kenai Peninsula from Turnagain Arm on the north to the Resurrection River near Seward. Other favorable areas for Au in the Valdez Group are the Portage-Bear Valley area, Pigot Bay north to Hobo Bay, a region southwest of Coghill Lake, the region between Miner's Lake and Pedro Glacier, Cleave Creek Glacier, and lower Schwan Glacier. In the Orca Group, Au signatures characterize McKinley Peak, Jackpot Point, and scattered localities southeast of the Sargent Icefield.

Anomalous factor 5 scores are associated with the Yakataga Formation and the Poul Creek Formation on Kayak Island and on the adjacent mainland south of the Martin River Glacier. Many of the elements with high factor 5 loadings might reflect relatively high concentrations inherent to shaley layers in the two formations, but there is also evidence for the presence of Zn-rich massive sulfide deposits. The concentrates with anomalous scores are rich in sphalerite and barite. Plafker (1974) noted intercalated basaltic fragmental rocks and minor pillow basalts in the Poul Creek Formation, providing evidence of a possible volcanogenic source for the Zn. However, the economic significance of this geochemical anomaly is uncertain at the present time.

Scores for factors 3, 6 and 7 are more difficult to interpret. They appear to be, for the most part,

Table 9.—Factor loadings for the first seven factors after varimax rotation of the heavy-mineral concentrate data. Total variance explained by the seven factors equals 65 percent. Loadings less than 0.25 have been omitted.

Factors	1	2	3	4	5	6	7
Fe	.36	.82	--	--	--	--	--
Mg	.85	--	--	--	--	--	--
Ca	.66	--	--	--	--	-.29	--
Mn	.82	--	--	--	--	--	--
Ag	--	.27	--	.87	--	--	--
As	--	.44	--	.41	--	--	.38
Au	--	--	--	.91	--	--	--
Ba	--	.29	.63	--	.31	--	-.29
Bi	--	--	--	.34	.34	--	.45
Co	--	.88	--	--	--	--	--
Cr	.67	--	--	--	--	--	--
Cu	--	.74	--	--	--	--	--
La	.42	--	.53	--	--	.40	--
Mo	--	--	--	--	.69	--	--
Nb	--	--	.61	--	--	-.26	--
Ni	--	.86	--	--	--	--	--
Pb	--	.63	--	.32	--	--	--
Sb	--	--	--	.25	.43	--	--
Sc	.68	--	--	--	--	.39	--
Sn	--	--	--	--	.25	.35	.59
Sr	--	--	.57	--	.29	--	-.36
V	.84	--	--	--	--	--	--
W	--	--	--	--	--	--	.66
Y	--	--	.78	--	--	--	.26
Zn	--	.28	--	--	.73	--	--
Zr	--	--	.77	--	--	--	--
Th	--	--	--	--	--	.74	--
Percent of total data variance explained by factor	17.9	14.6	11.2	7.6	5.5	4.3	4.1

related to varying amounts of rare earth and other late-stage magmatic element enrichments associated with intrusive bodies.

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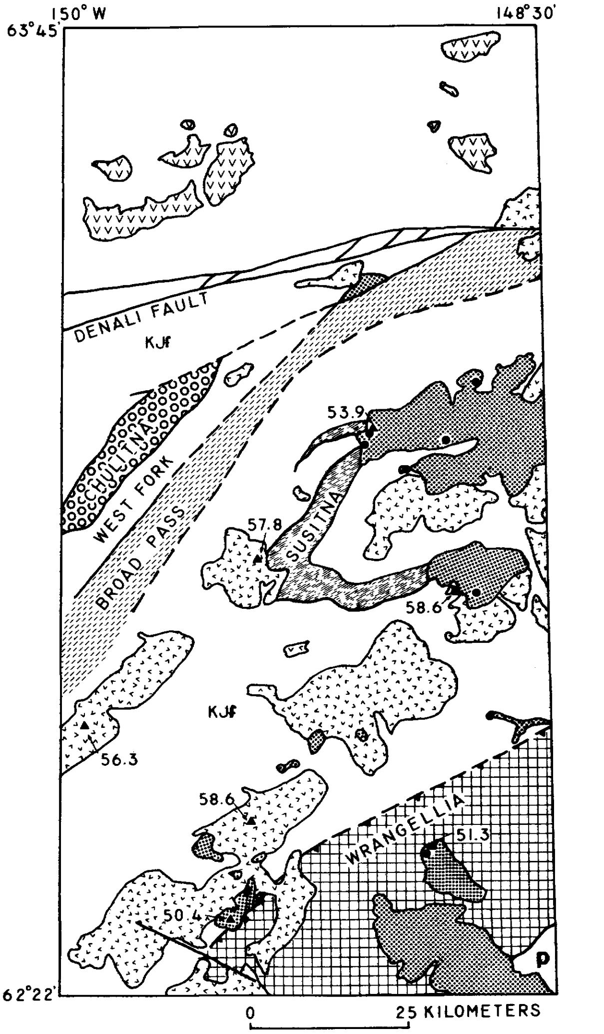
PALEOMAGNETISM OF EARLY TERTIARY VOLCANIC ROCKS IN THE NORTHERN TALKEETNA MOUNTAINS

By J. W. Hillhouse, C. S. Grommé, and Béla Csejtey, Jr.

Paleomagnetic studies have demonstrated that large parts of southern Alaska, such as Wrangellia and the Peninsular terrane, have histories of movement that differ from the Mesozoic motion of the North American plate. The net motions of the



TERTIARY VOLCANIC ROCKS

VOLCANIC ROCKS (PALEOCENE) OF THE CANTWELL FORMATION

PALEOCENE GRANITIC ROCKS

Figure 32.—Geologic map showing selected tectonostratigraphic terranes south of the Denali fault, the Peninsular terrane, the deformed upper Mesozoic flysch terrane (KJf), paleomagnetic sample sites (dots), and potassium-argon dated localities (triangles). Ages given in m.y. Map modified from Csejtey and others (1978) and Jones and others (1983).

Alaskan terranes are generally northward relative to the craton as indicated by the comparatively low paleolatitudes of the terranes (Stone and others, 1982; Hillhouse, 1977). According to some (Csejtey and others, 1982; Jones and others, 1982), geologic evidence suggests that by the beginning of the Tertiary, Wrangellia, as part of the Talkeetna superterrane, had accreted to North America and has moved little, if any, relative to the craton since that time. However, on the basis of paleomagnetic results from lower Tertiary sedimentary rocks of the Peninsular terrane, Stone and others (1982) speculate that Wrangellia and the Peninsular terrane, which were amalgamated in the Jurassic to form the Talkeetna superterrane, had moved northward relative to North America throughout the Tertiary at approximately 6 cm/yr. At this rate, the total relative displacement for the past 50 m.y. equals 3,000 km. This study of early Tertiary volcanic rocks mapped by Csejtey and others (1978) in the northern Talkeetna Mountains adds to the paleomagnetic data base and tests the large Tertiary displacement suggested by Stone and others (1982).

Our sample localities (fig. 32) are in andesite flows and dacites of early Eocene age, which have potassium-argon ages of 50.4 ± 2.0 m.y. (whole rock), 51.3 ± 2.5 m.y. (whole rock), and 53.9 ± 1.6 m.y. (sanidine). The volcanic suite is associated with small plutons, principally of biotite-granodiorite composition, which range in age from 58.6 to 56.3 m.y. The Tertiary volcanic rocks overlie Upper Triassic pillow basalt of the Susitna terrane (Jones and others, 1981), Lower Permian volcanic rocks of Wrangellia, and Lower Cretaceous flysch. In contrast to the isoclinally folded flysch, which entrains klippen of older rock sequences, the Eocene volcanic rocks are mildly deformed. The plutons and volcanic rocks effectively knit together the structural collage of terranes that lie between the McKinley strand of the Denali fault and the Talkeetna River.

Preliminary paleomagnetic results from 25 sites (8 specimens per site) in the Eocene volcanic rocks give a paleomagnetic pole at 69.8°N, 176.7°E, α_{95} = 12.7°. Field directions of reversed polarity were measured at 22 sites, and 3 sites gave directions of normal polarity. The normal and reversed groups give antipodal mean directions, indicating that the results are free of secondary components. Evidence that the magnetization was acquired before folding, a minimum criterion for reliability of the results, is provided by a successful fold test. That is, the dispersion of magnetic directions significantly decreased when the directions from both limbs of a syncline were corrected for tilt of the bedding.

Paleomagnetic studies (Diehl and others, 1980) of igneous rocks in Wyoming and Montana indicate that the early Eocene paleomagnetic pole of North America was located at 81.7°N, 171.2°E (α_{95} = 4.4°). From this reference pole, we calculate that the Eocene paleolatitide of the Talkeetna locality was 69°N, slightly lower than the mean paleolatitude (75°N) that was obtained from the Talkeetna volcanic rocks. The difference, which implies southward drift of the Talkeetna Mountains, is not significant in the context of the combined 95-percent confidence limits. By taking the square root of the sum of the squares of the confidence limits, we obtain an overall confidence limit (13°) that incorporates the uncertainties in the cratonic reference pole and the Talkeetna observations. Although the mean of the Talkeetna results strongly suggests that relative northward drift of Wrangellia had ceased by 50 m.y. ago, the overall confidence limit permits up to 7° (800 km) of northward drift relative to the craton. A similar conclusion was obtained from a paleomagnetic study (Hillhouse and Grommé, 1982) of Paleocene volcanic rocks in the Cantwell Formation located north of the Denali fault. The paleomagnetism of lower Tertiary volcanic rocks of the central Alaska Range and the northern Talkeetna Mountains is consistent with plate-tectonic models (Csejtey and others, 1982) in which northward drift of Wrangellia relative to the craton was completed before 50 m.y. ago. Thus, our results are not consistent with the high rate (6 cm/year) of Tertiary northward drift that was proposed by Stone and others (1982) for Wrangellia and the Peninsular terrane.

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Reviewed by J. Hagstrum and D. L. Jones.

A TRANSECT OF METAMORPHIC ROCKS ALONG THE COPPER RIVER, CORDOVA AND VALDEZ QUADRANGLES, ALASKA

By Marti L. Miller, J. A. Dumoulin, and Steven W. Nelson

The lower Tertiary Orca Group is juxtaposed against the Upper Cretaceous Valdez Group along the Contact fault system (Winkler and Plafker, 1974, 1981; Plafker and others, 1977)(fig. 33). In both groups, turbidites are the dominant rock type, with lesser mafic volcanic rocks (table 10). The Valdez Group, on the north, has traditionally been considered to be of higher metamorphic grade than the Orca Group (Moffit, 1954; Tysdal and Case, 1979; Winkler and Plafker, 1981; Winkler and others, 1981). In 1982, we made a transect across the regional strike of the rocks and the contact between the two groups. The transect area follows the Copper River for 85 km from the Cordova quadrangle north into the Valdez quadrangle and extends for about 25 km on either side of the river (fig. 33). We planned, by systematic sampling of the area, to examine the metamorphic differences between the Orca and Valdez Groups. We found, however, that a strong thermal metamorphic event has overprinted and obscured regional metamorphic relations. We believe intrusion of Tertiary granite (fig. 33) to be responsible for this metamorphism. (Figures 33 and 34 and tables follow this article.)

Thin sections from 112 localities were studied, including those from 22 stations occupied by George Plafker in 1971, and Plafker, Travis Hudson, and R. G. Tysdal in 1974. The metamorphic variation seen in 190 thin sections was divided into five categories based on mineral assemblage changes indicative of increasing metamorphic grade (table 11). Mineral assemblages in graywackes and, to a lesser extent, mafic volcanic rocks were used. The five categories as expressed in graywackes are: (A) authigenic minerals, (B) fine-grained white mica, (C) minor and very fine-grained biotite, (D) abundant biotite, and (E) andalusite, cordierite, or garnet. Mineral assemblages in the mafic volcanic rocks distinguish fewer categories: (B) actinolite, (C) hornblende/actinolite, and (D, E) hornblende plus plagioclase. The distribution of these metamorphic mineral categories (fig. 33) shows a general increase to higher grade to the east-northeast. This trend is primarily marked by the change from category C to category D (in graywacke, abundant biotite), and the relation of categories continues across the Contact fault.

Three lines of evidence suggest contact metamorphism is responsible for the metamorphic grade increase in this area. First, granitic plutons intrude both the Orca and Valdez Groups, and near granite contacts, the intruded strata show local injection migmatite and schistose to gneissic segregations of felsic and mafic minerals. Distant from the dynamic effects of intrusion, both Orca and Valdez sequences that contain category D assemblages retain primary sedimentary features such as cross-beds and burrows. Second, the minerals characteristic of categories D and E in the graywackes indicate low-pressure, high-temperature metamorphism (Williams, Turner, and Gilbert, 1954; Turner, 1981). Third, features indicative of thermal metamorphism occur in category E rocks of both the Orca and Valdez Groups. These features include totally recrystallized granoblastic textures and porphyroblastic andalusite or cordierite growing across the penetrative fabric of the rocks. The areal extent of the thermally metamorphosed rocks is many times that of the exposed granite. However, numerous small stocks, dikes, and stoping features throughout the area from the Martin fault to north of the Wernicke Glacier suggest that the area lies over the upper part of a large, partially-exposed, granitic complex.

In a regional overview, Hudson and Plafker (1982) suggested that the eastern Chugach Mountains form what they informally termed the Chugach metamorphic complex which stretches for 200 km from the Copper River to at least the Canadian border. Our transect overlaps some of the western part of the Chugach metamorphic complex (fig. 34). Hudson and Plafker mapped amphibolite facies schist and gneiss at the core of the complex and suggested that this high-grade central zone grades outward into an intermediate-zone of epidote amphibolite to amphibolite facies schist, and then to the more typical "***subgreenschist to greenschist facies slate, phyllite, and semischist" (Hudson and Plafker, 1982, p. 1280) of the Valdez Group. They further stated that the Chugach metamorphic complex (1) occurs only within the Valdez Groups, (2) is apparently truncated by the Contact fault system, and (3) "***is the result of a regional progressive metamorphism***" which is "***strongly thermal in nature" (p. 1288).

Category D rocks are apparently equivalent to the "intermediate-zone" of Hudson and Plafker. However, unlike the "intermediate-zone", category D assemblages are not restricted to the Valdez Group, but cross the Contact fault and also occur in the Orca Group. We believe that in our study area evidence for regional metamorphism has been obscured by thermal metamorphism associated with Tertiary plutonism. A study of all metamorphic rocks from the Chugach National Forest is under way in an attempt to document metamorphic mineral assemblage and fabric changes from the lower grade rocks in the west to the higher grade metamorphic complex in the east.

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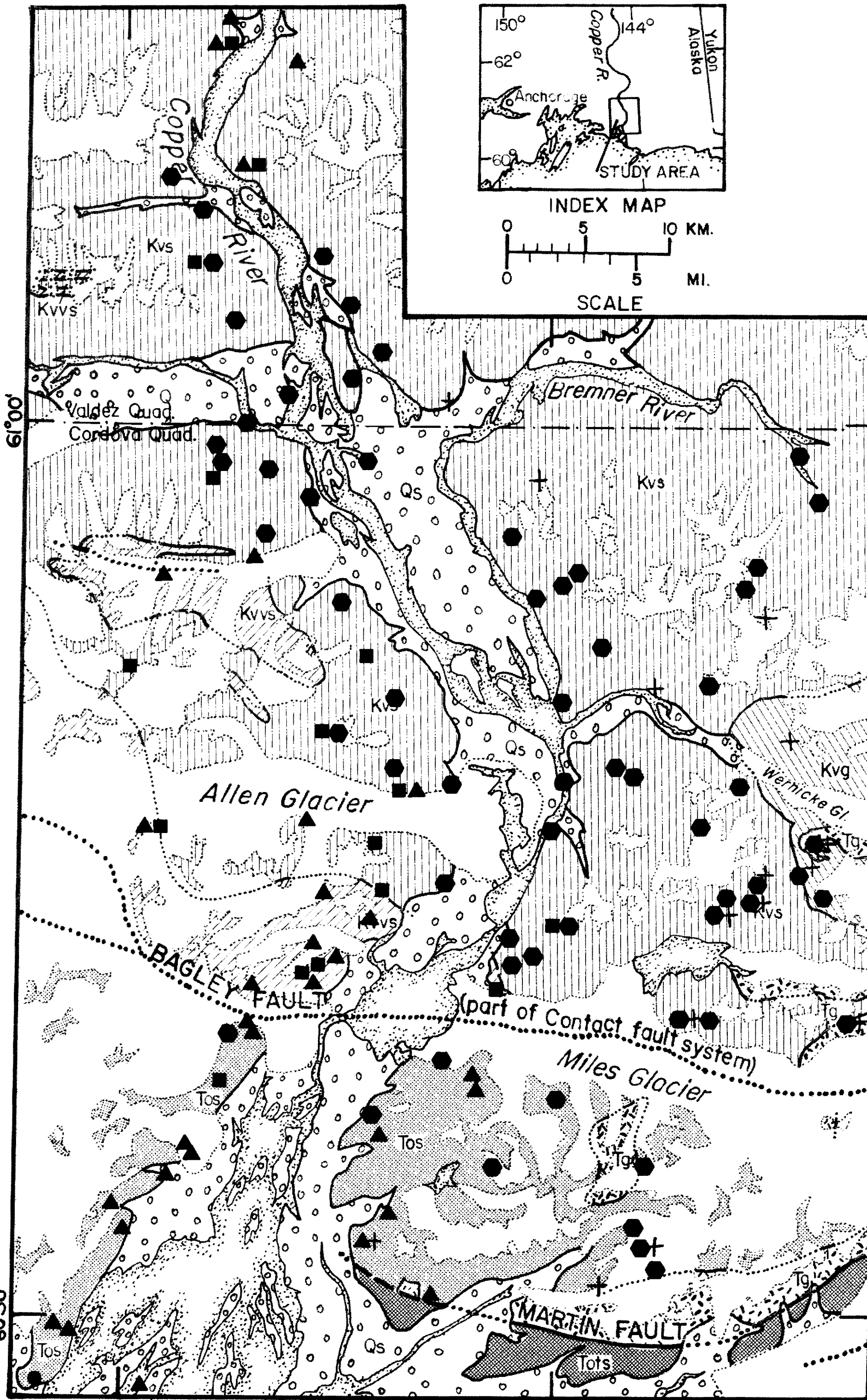

Figure 33.—Geologic map of the Copper River transect area. Symbols represent metamorphic categories A through E. An expanded description of map units is given in table 10.

EXPLANATION

GEOLOGIC UNITS

Qs	Surficial deposits (Holocene)
Tg	Granitoid rocks (Eocene)

ORCA GROUP (Eocene? and Paleocene)
—Divided into:

Tos	Sedimentary rocks
Tots	Tuffaceous sedimentary rocks

VALDEZ GROUP (Upper Cretaceous)
—Divided into:

Kvs	Sedimentary rocks
Kvvs	Volcanic and interbedded volcanic and sedimentary rocks
Kvg	Paragneiss

SYMBOL	METAMORPHIC CATEGORIES
●	A–authigenic
▲	B–fine-grained biotite
■	C–very fine-grained biotite
⬢	D–abundant biotite
+	E–andalusite, cordierite, or garnet

MAP SYMBOLS

Fault—
dashed where inferred
dotted where concealed

Contact—
dashed where inferred
dotted where concealed

Table 10.—Description of map units (to accompany figure 33)

Qs Undifferentiated surficial deposits (Holocene)

Tg Granitic to granodioritic intrusive rocks (Eocene)

Orca Group (Eocene? and Paleocene)—Divided into:

Tos Sandstone, siltstone, and mudstone in thin- to thick-bedded turbidites. Regional metamorphic grade ranges from zeolite to prehnite-pumpellyite facies with local metamorphism to the chlorite zone of the greenschist facies

Tots Tuffaceous sedimentary rocks—Abundant tuffaceous sedimentary rocks and volcaniclastic sandstone. Regional metamorphic grade is the same as in unit Tos

Valdez Group (Upper Cretaceous)—Divided into:

Kvs Sedimentary rocks, deformed turbidites with a range in regional metamorphic grade from zeolite to greenschist facies

Kvvs Volcanic rocks and interbedded volcanic and sedimentary rocks—Regional metamorphic grade is same as in unit Kv

Kvg Paragneiss—Largely metasedimentary rocks with well developed gneissic foliaton. Metamorphosed to hornblende hornfels or amphibolite facies

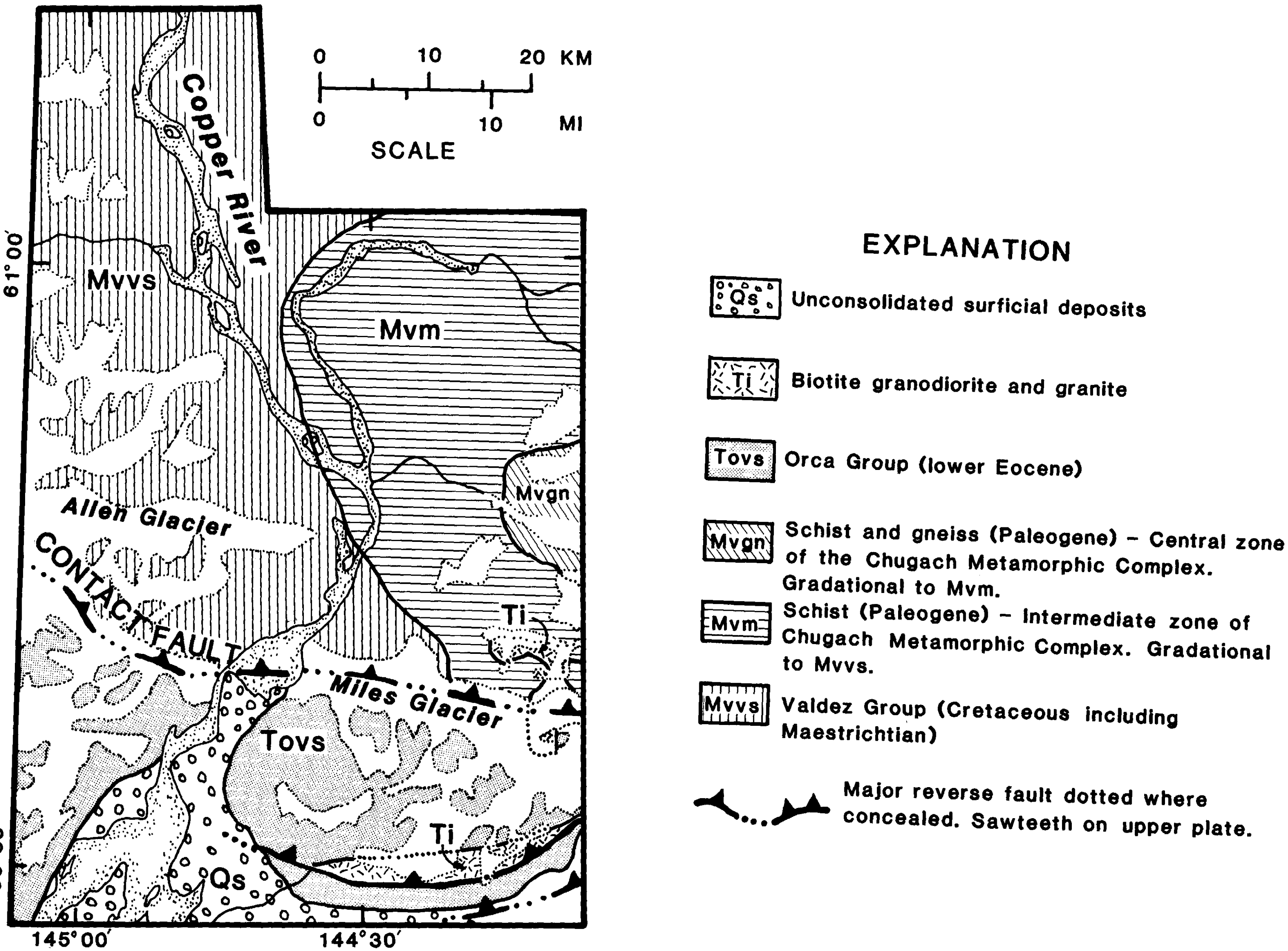

Figure 34.—Map showing the metamorphic complex of Hudson and Plafker (1982, fig. 2) and simplified explanation of map units. The map boundaries are the same as those in figure 33.

Table 11.—Description of metamorphic categories of the Copper River transect study

Category	Rock Type	Minerals Present	Approximate Facies Equivalent*
A	graywacke	authigenic matrix with phyllosilicates	Laumontite-phyllosilicate
B	graywacke	fine-grained white mica + chlorite ± epidote	Lower greenschist
	mafic volcanic rocks	actinolite + albite + sphene	
C	graywacke	fine-grained white mica ± chlorite ± epidote + minor very fine-grained biotite	Greenschist or Albite-epidote amphibolite
	mafic volcanic rocks	hornblende/actinolite + chlorite + epidote + albite	
D	graywacke	biotite > muscovite ± chlorite ± epidote ± sphene	Upper greenschist or Hornblende hornfels
	mafic volcanic rocks	hornblende + plagioclase ± biotite ± garnet	
E	graywacke	biotite > muscovite ± andalusite ± cordierite ± garnet	Amphibolite or Hornblende hornfels
	mafic volcanic rocks	same as Zone D	

* From Williams, Turner, and Gilbert (1954); Galloway (1974); Winkler (1976); and Turner (1981)

SEDIMENTOLOGY OF FLYSCH OF THE UPPER MESOZOIC YAKUTAT GROUP, MALASPINA DISTRICT, ALASKA

By Tor H. Nilsen, George Plafker, Dorothy E. Atwood, and Edwin R. Hill

The sedimentology of the Yakutat Group was studied at the northwestern margin of its outcrop area near the Samovar Hills (fig. 35). It is exposed along the southern flank of the Saint Elias Mountains between Icy Bay and Lituya Bay and is also known to underlie parts of the contiguous coastal lowland as well as the continental shelf and slope. The Yakutat Group is part of a small tectonostratigraphic terrane, the Yakutat block, that is presently moving northwestward with the Pacific plate along the Fairweather transform fault (Plafker and others, 1978). The Yakutat block is obliquely underthrusting the Chugach terrane (Plafker and others, 1977; Nilsen and Zuffa, 1982), a belt of upper Mesozoic deep-sea strata that is exposed for 2,000 km along the southern margin of Alaska from the Sanak Islands to the Sitka area, along the west-trending Chugach-Saint Elias thrust fault system. The Yakutat Group has also previously been considered to be part of the Chugach terrane by Berg and others (1972).

The Yakutat Group consists of an Upper Jurassic(?) and Lower Cretaceous melange facies and an Upper Cretaceous flysch facies (Plafker, 1967; Plafker and others, 1977). The melange facies

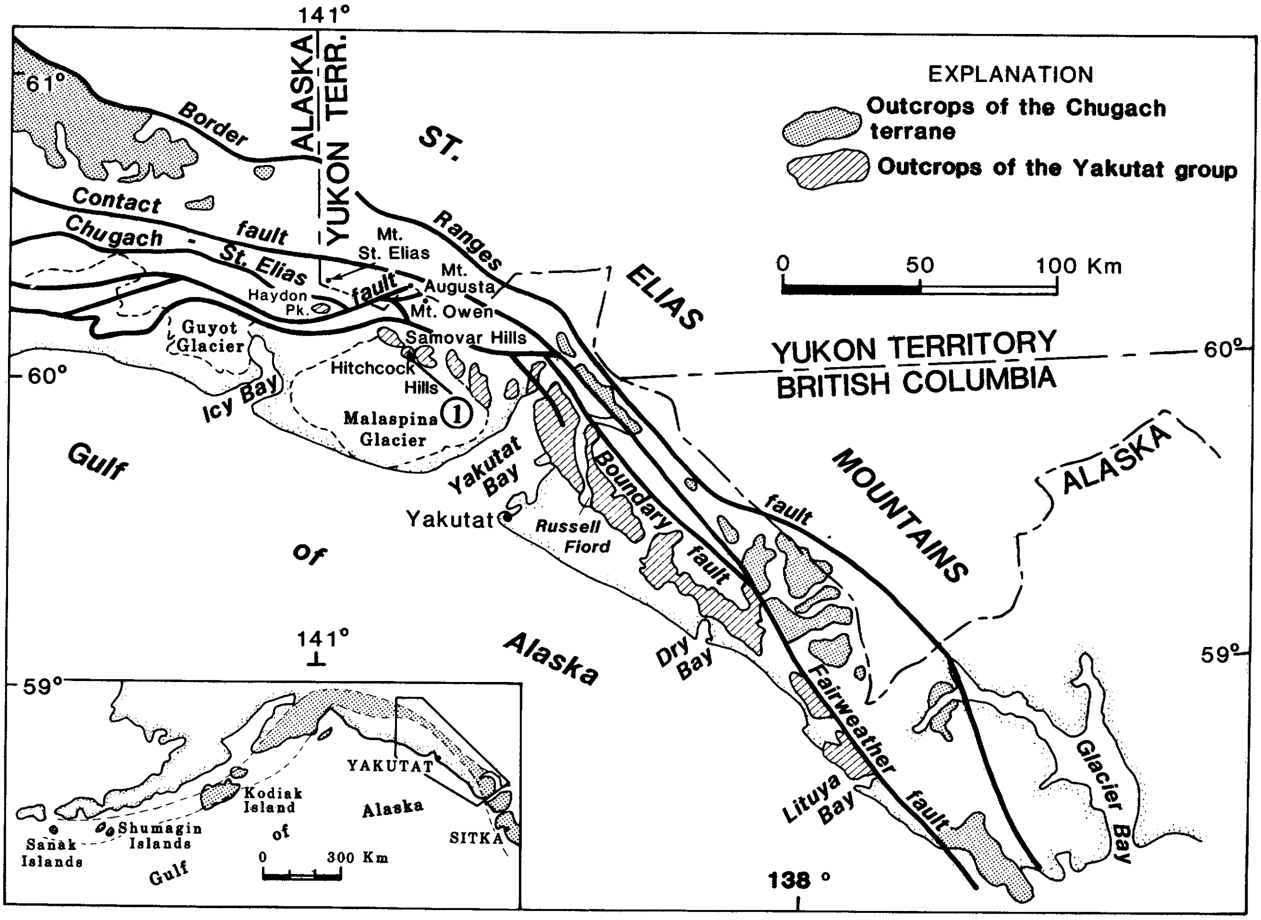

Figure 35.—Index map of part of southeastern Alaska showing location of outcrops of the Yakutat Group and principal geologic features. Circled number 1 indicates location of measured section shown in figure 36. Geology modified from Beikman (1980).

is generally exposed along the northeastern margin of the outcrop belt and the flysch facies along the southwestern margin. However, extensive faulting, particularly along the northern margins of the outcrop area, has locally resulted in a complex interleaving of the two facies in structurally bound blocks and slivers to the extent that the two facies have not everywhere been differentiated in reconnaissance-scale geologic mapping of the region (Plafker and Miller, 1957).

The Yakutat Group in the Malaspina District is largely a turbidite sequence consisting of dense, hard, poorly sorted, brown-weathering, gray sandstone alternating with thin-bedded black argillite. Thick beds of pebble and cobble conglomerate are abundant near Yakutat Bay, and conglomeratic debris-flow deposits are abundant in presumably equivalent strata north of the Chugach-Saint Elias fault near Haydon Peak, Mount Augusta, and Mount Owen. Blocks of mafic volcanic rocks and green tuff are present in the melange facies of the Yakutat Group, but are not known in the flysch facies. Although the northwest-trending Yakutat Group south of the Chugach-Saint Elias fault is only slightly metamorphosed to zeolite facies, presumably equivalent west-trending strata exposed in thrust slices north of the Chugach-Saint Elias fault are metamorphosed to prehnite-pumpellyite and lower greenchist facies. The Yakutat Group internally is highly deformed and typically forms fault-bounded slices characterized by steeply dipping, locally overturned, tightly folded, and highly sheared outcrops.

Although age-diagnostic fossils have not been found in the Malaspina District, the flysch facies of the Yakutat Group is considered to be mainly Late Cretaceous, probably Campanian, based on one Inoceramus schmidti collected from the Russell Fiord area east of Yakutat Bay (Plafker and others, 1977) and a sparse foraminiferal fauna from wells drilled for oil near Yakutat. The basement upon which the Yakutat Group was deposited, although not known, is thought to be oceanic crust, based on relations elsewhere along the Gulf of Alaska

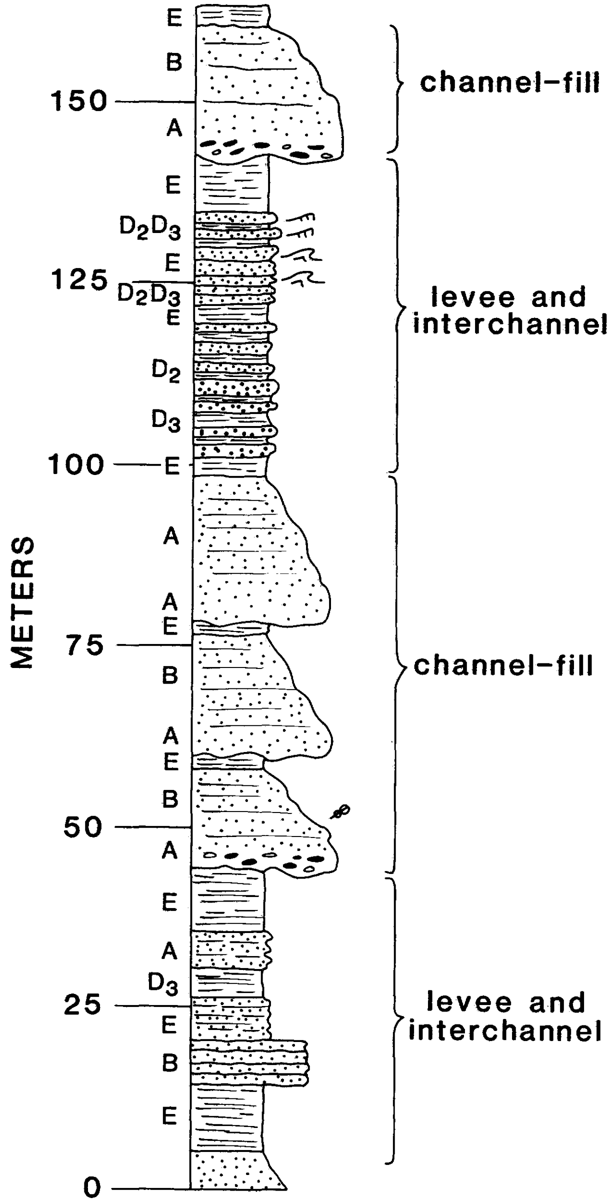

Figure 36.—Representative measured section of the Yakutat Group from the Hitchcock Hills. Letters refer to Mutti and Ricci Lucchi (1972, 1975) facies.

(Plafker and others, 1977). The sequence was deformed and accreted prior to emplacement of widespread discordant early Eocene granitic plutons. The Yakutat Group in the Samovar Hills is overlain with 90-degree angular unconformity by lower and middle Eocene shallow-marine and nonmarine coal-bearing strata (Plafker and Miller, 1957).

We measured four sections that total 1,050 m in the flysch facies of the Yakutat Group in 1982. The sections consist mostly of middle-fan channel, levee, and interchannel facies associations (fig. 36). The channel-fill deposits consist mostly of facies A and B sandstones of Mutti and Ricci Lucchi (1972, 1975) in the lower parts of the fill and B_1 and E in the upper parts.

The levee facies association consists chiefly of interbedded facies B_1 and E sandstone turbidites and shale. Some facies C and D beds are also included. Bedding thickness for the levee turbidites averages about 10 cm. The levee deposits locally develop into thick bundles of wavy-bedded facies B_1 sandstones, and small-scale synsedimentary slumps are locally present. The individual beds that comprise the levee deposits are laterally discontinuous and commonly change in both thickness and grain size within the scale of the outcrop.

The interchannel facies association consists of thinly interbedded facies E, D_2, and D_3 sandstone turbidites and shale. Bedding thickness for the interchannel turbidites averages about 3 cm.

Paleocurrents obtained from 35 measurements of flute casts and other sole markings near the Samovar Hills indicate dominant transport toward the south, with a vector mean and standard deviation of 183° ± 30°. Plafker previously obtained one northwestward-directed and a few southward-directed paleocurrents from the Yakutat Group in the Yakutat Bay area.

Sandstone from the flysch facies of the Yakutat Group contrasts markedly with presumably correlative rocks elsewhere in the Chugach terrane. More quartzofeldspathic and less rock-fragment detritus (Winkler and Plafker, 1981) and a higher percentage of plutonic rock fragments (Zuffa and others, 1980) characterize the Yakutat Group.

A suggested tectonic interpretation of the Yakutat Group is that it originally formed an extension of the Chugach terrane along the continental margin southeast of Chatham Strait and has subsequently been displaced to its present position as part of the Yakutat block by offset along the Queen Charlotte-Fairweather fault system (Plafker, 1983). This reconstruction does not appear to be supported by sedimentologic data from the Yakutat Group in the Samovar Hills area because: (1) the petrography of the Yakutat Group turbidite sandstones does not match that of comparable middle-fan deposits of the Sitka Graywacke or the rest of the Chugach terrane, but is similar to the inner-fan deposits of the Sitka Graywacke (John Decker, Alaska Division of Geological and Geophysical Surveys, oral commun., April 18, 1983); and (2) the relatively fine-grained middle-fan deposits of the Yakutat Group in the Samovar Hills area do not match the coarsely conglomeratic inner-fan deposits of the Sitka Graywacke on Baranof and southern Chicagof Island. However, we cannot be certain that the Yakutat Group turbidites are in fact coeval with the undated Sitka Graywacke. Nevertheless, it is clear that sedimentologic data of the type reported here can provide important constraints on palinspastic reconstructions of the Yakutat Group and the displacement history of the Yakutat block.

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Reviewed by D. L. Jones and G. W. Moore.

PRELIMINARY ACCRETIONARY TERRANE MODEL FOR METALLOGENESIS OF THE WRANGELLIA TERRANE, SOUTHERN MOUNT HAYES QUADRANGLE, EASTERN ALASKA RANGE, ALASKA

By Warren J. Nokleberg, Ian M. Lange, and Robert C. Roback

A major portion of the southern part of the Mount Hayes quadrangle consists of the Wrangellia terrane (fig. 37). Recent studies of this terrane show a long and complicated stratigraphic, structural, and tectonic history (Nokleberg and others, 1981a, 1982, 1983); a lengthy, complex metallogenetic history (Lange and others, 1981; Nokleberg and others, 1981b); and a relationship between metallogenesis and specific events in the origin, migration, and accretion of Wrangellia. Previous studies of the metallogenesis of Wrangellia in southeastern Alaska have been made by Berg (1979).

In the southern Mount Hayes quadrangle, Wrangellia is subdivided into a northern (Slana River) subterrane and a southern (Tangle) subterrane (fig. 37) (Nokleberg and others, 1981a; 1982). The Slana River subterrane consists of: (1) a thick sequence of upper Paleozoic island-arc rocks consisting of submarine andesite flows, breccia, epiclastic rocks, volcanic graywacke, argillite, and marble of the Pennsylvanian Tetelna Volcanics, and the overlying Pennsylvanian and Permian Slana Spur Formation; (2) altered andesite and dacite porphyries, dikes, and sills of Permian age; (3) argillite and limestone of the Permian Eagle Creek Formation; (4) masssive subaerial, amygdaloidal basalt flows of the disconformably overlying Triassic Nikolai Greenstone; (5) cumulate mafic and ultramafic sills of probable Triassic age that may be comagmatic with the Nikolai Greenstone; (6) disconformably overlying thin Upper Triassic limestone (too thin to be shown on fig. 37); (7) unconformably overlying Upper Jurassic and Lower Cretaceous rocks of the Gravina-Nutzotin belt that consist of volcanic flows and flysch composed of volcanic graywacke and argillite; and (8) isolated plutons of probable Early Cretaceous age that may be comagmatic with the volcanic rocks of the Gravina-Nutzotin belt. The Slana River subterrane is bounded to the north by the Broxson Gulch thrust and to the south by the Eureka Creek thrust (fig. xx); it is intensely faulted, particularly near these thrusts. The Slana River subterrane exhibits upper greenschist facies metamorphism with a weak cleavage or schistosity. Quartz veins and epidote-chlorite alteration are common, particularly in volcanic rocks and greenstone, and are interpreted as a late-stage phase of regional metamorphism. A whole-rock K-Ar isochron for the Nikolai Greenstone in the McCarthy quadrangle indicates a middle Cretaceous age of metamorphism (Silberman and others, 1981). The Slana River subterrane in the Mount Hayes quadrangle is the most common variation of Wrangellia in the eastern Alaska Range; similar stratigraphic sections occur in the Nabesna and McCarthy quadrangles (Richter, 1976; MacKevett, 1978).

The Tangle subterrane differs from the Slana River subterrane in having: (1) a thinner sequence of upper Paleozoic island-arc rocks; (2) a thicker sequence of the Nikolai Greenstone with a thick basal unit of pillow basalt; (3) no Gravina-Nutzotin belt rocks; and (4) fewer faults. The Tangle subterrane is fault bounded. The nearest known occurrence of Wrangellia that is similar to the Tangle subterrane is on Vancouver Island (Muller and others, 1974).

Six principal types of mineral deposits and occurrences found in this part of Wrangellia (table 12) (Lange and others, 1981; Nokleberg and others, 1981b) are:

(1) Small areas of sulfides up to a few meters wide, with anomalous concentrations of Cu, Pb, Zn, Ag, and Au, occur in fracture zones and are disseminated in hydrothermally altered volcanic rocks of the Tetelna Volcanics and Slana Spur Formation. Common sulfide minerals are chalcopyrite, bornite, sphalerite, and pyrite.

(2) Pods and lenses, up to a few meters thick, of massive sulfides (primarily chalcopyrite and pyrite) that contain anomalous concentrations of Cu, Ag, and Au occur in skarn and hydrothermally altered volcanic rocks of the Slana Spur Formation adjacent to shallow Permian andesite and dacite porphyries.

(3) Locally disseminated sulfides, with anomalous concentrations of Cu, Pb Zn, and Ag, occur in shallow Permian andesite and dacite porphyries; many of these porphyries are propylitically altered, contain calcite and chalcopyrite veinlets, and are laced with disseminated chalcopyrite and pyrite.

(4) Disseminated grains and lenses of Ni-bearing chromite occur in layers in cumulate mafic and ultramafic rocks in extensive and thick sills of probable Triassic age that intrude the Nikolai Greenstone and older rocks of both subterranes.

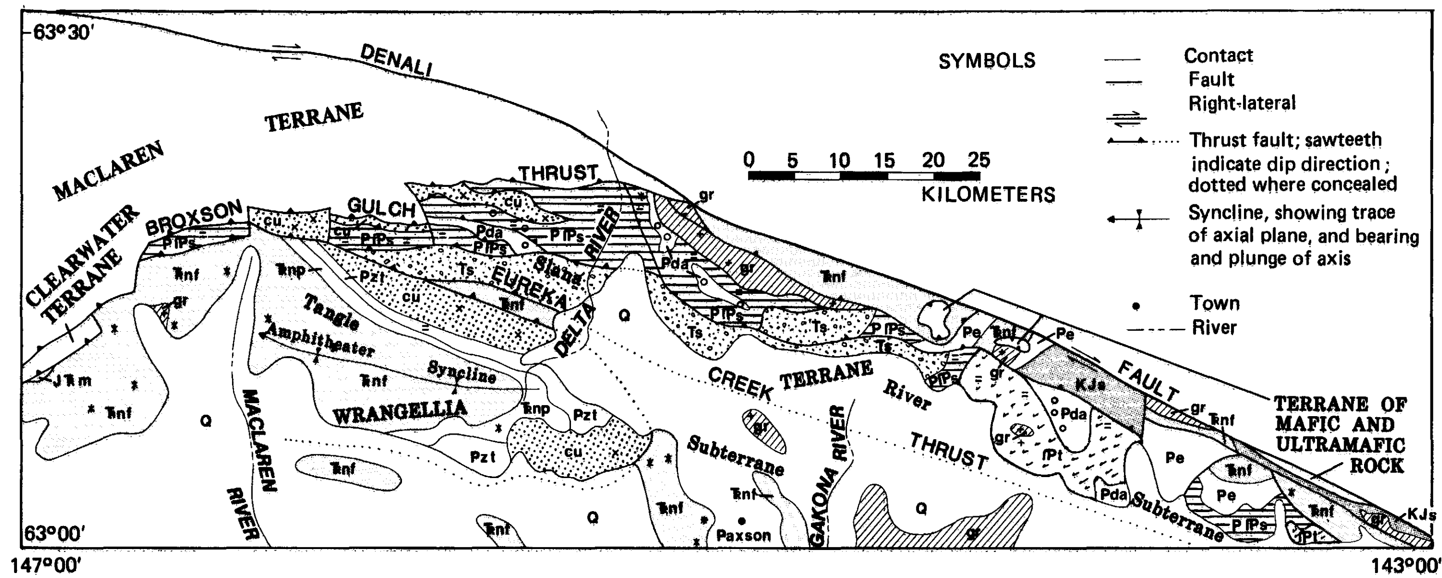
(5) Disseminated sulfides with anomalous concentrations of Cu, Ag, and Au occur in Early Cretaceous granodiorite and quartz diorite that intrude Wrangellia. These granitic rocks commonly contain very fine-grained K-feldspar, quartz, secondary sericite, and pyrite, with minor chalcopyrite in fractures and isolated grains. Sulfide-bearing skarn locally occurs adjacent to the granitic rocks.

(6) Sulfides with anomalous values of Cu, Ag, and Au occur in quartz veins and in associated epidote-chlorite-actinolite alterations in the Nikolai Greenstone and older volcanic rocks. The common sulfide minerals in the quartz veins and altered areas are chalcopyrite, bornite, and pyrite. Local weathering of Cu-sulfide has formed sparse malachite and azurite.

The first three types of deposit in the volcanic and associated rocks of late Paleozoic age in the Slana River subterrane appear to be quite similar to those of the modern Fiji island arc (Lange and others, 1981).

We have formulated a preliminary accretionary terrane model (table 12) that relates the origin of the principal mineral deposits and occurrences to the origin, migration, and accretion of Wrangellia. Initially, subduction formed an island arc during the Pennsylvanian and Permian, with submarine deposition of volcanic and sedimentary rocks of the Tetelna Volcanics and Slana Spur Formation and intrusion of shallow porphyries. During the initial building of the island arc, disseminated and vein sulfide deposits formed in volcanic flows and associated volcaniclastic rocks. Later, during the culmination of the island-arc, sulfide deposits formed in skarn and volcanic rocks adjacent to shallow porphyries. Subsequent rifting during the Middle and (or) Late Triassic resulted in submarine and subaerial extrusion of the Nikolai Greenstone and emplacement of associated cumulate mafic and ultramafic rocks that contain disseminated grains and lenses of chromite. In the Late Jurassic and Early Cretaceous, subduction and formation of an island arc along the leading edge of Wrangellia occurred during migration of Wrangellia toward North America. This resulted in deposition of the volcanic and sedimentary rocks of the Gravina-Nutzotin belt, in nearly contemporaneous intrusion of plutonic rocks containing disseminated and vein base-metal sulfides, and in local skarn formation adjacent to granitic rocks. Finally, Wrangellia was accreted onto the western margin of North America during middle or Late Cretaceous time, resulting in regional greenschist facies metamorphism and deformation that culminated in formation of late stage quartz sweat veins and altered areas with Cu-, Ag-, and Au-bearing sulfides. During accretion, Wrangellia was compressed and telescoped, thereby causing the accretion of the Tangle subterrane onto the Slana River subterrane.

This preliminary model for the metallogenesis of Wrangellia is an important predictive tool for mineral resource assessment and for discovery of new mineral deposits and occurrences in Wrangellia here and elsewhere in western North America. Additional petrologic, chemical, and isotopic studies in progress will result in refinement of the model and make it more useful for mineral resource assessment and exploration.



EXPLANATION

MINERAL DEPOSITS AND OCCURENCES

- ✱ Chalcopyrite, bornite, and pyrite in quartz veins and in altered areas in volcanic rocks
- \+ Chalcopyrite and pyrite in fractures and as disseminated grains in altered granodiorite and quartz diorite, and in skarn next to granitic rocks
- × Disseminated grains and lenses of chromite in cumulate mafic and ultramafic rocks
- ○ Chalcopyrite and pyrite in massive sulfides in skarn and volcanic rocks adjacent to porphyry, chalcopyrite and pyrite as disseminated grains and in veins in porphyries
- = Chalcopyrite, bornite, spalerite, and pyrite as disseminated grains in fractures and in hydrothermally altered andesite and dacite

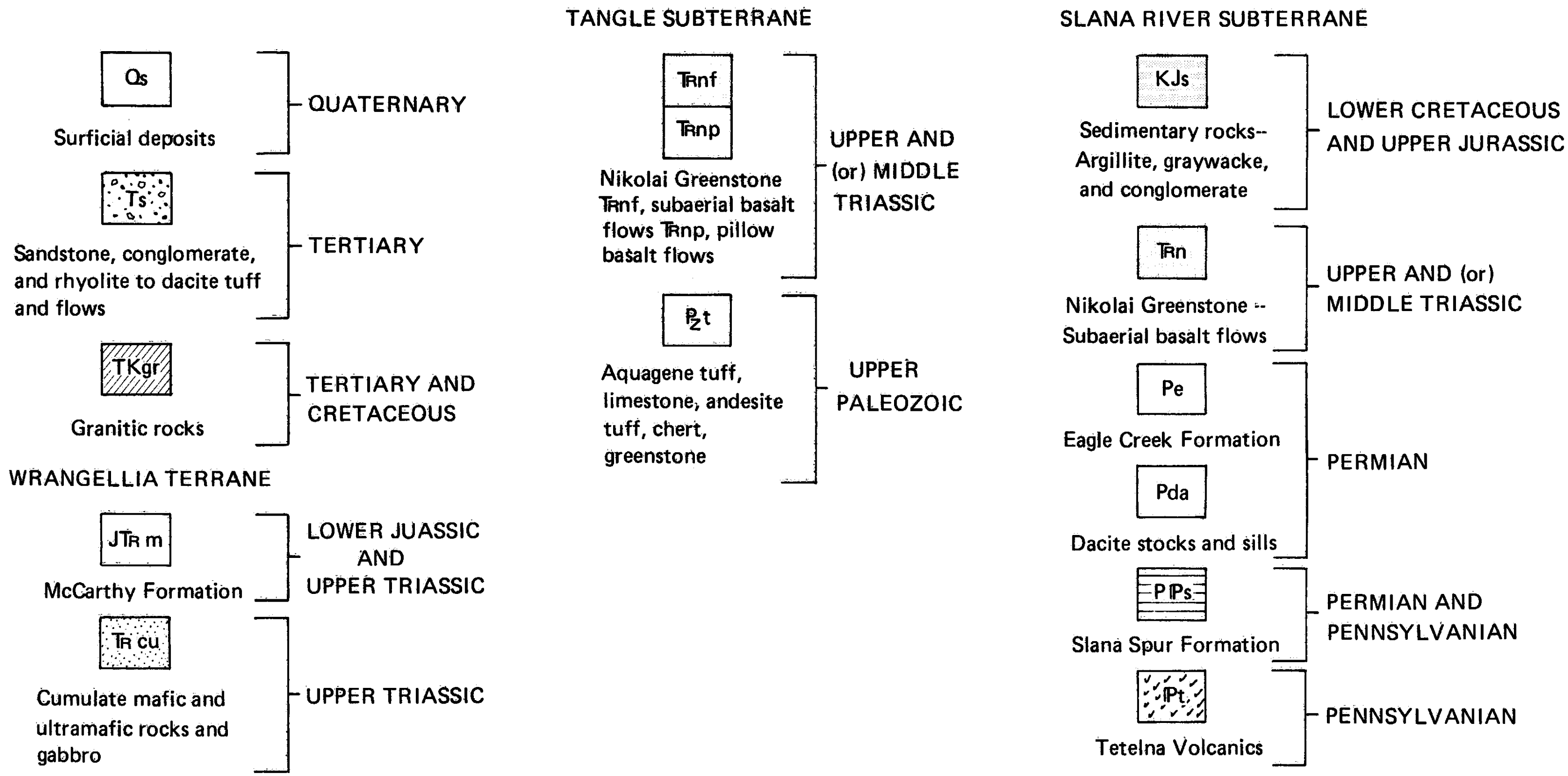


Figure 37.—Simplified geologic map of the southern Mount Hayes quadrangle, eastern Alaska Range, Alaska, showing location of major units of the Wrangellia terrane, and location of major types of mineral deposits and occurrences in Wrangellia. Adapted from Nokleberg and others (1982). Refer to table 12 for further explanation of mineral deposit and occurrence symbols.

Table 12.—Preliminary accretionary terrane model for metallogenesis of Wrangellia terrane, southern Mount Hayes quadrangle, eastern Alaska Range, Alaska

Tectonic event in model. Age of event	Geologic unit or process occurring during event	Mineral deposits or occurences forming during event. Symbol
Accretion of Wrangellia onto North America. Middle Cretaceous	Regional metamorphism and deformation; formation of late-stage quartz sweat veins and areas of epidote-chlorite alteration	Chalcopyrite, bornite, and pyrite in quartz veins and in altered areas in volcanic rocks *
Subduction and formation of island arc on leading edge of Wrangellia. Early Cretaceous and Late Jurassic	Deposition of volcanic rock and and flysch of Gravina-Nutzotin belt with intrusion of plutons	Chalcopyrite and pyrite in fractures and as disseminated grains in altered granodiorite and quartz diorite, and in skarn next to granitic rock +
Rifting of Wrangellia. Late and (or) Middle Triassic	Submarine and subaerial extrusion of basalts forming the Nikolai Greenstone; intrusion of sills of cumulate mafic and ultramafic rocks	Disseminated grains and lenses of chromite in cumulate mafic ultramafic rocks x
Subduction and late-stage formation of island arc. Permian	Intrusion of shallow andesite and dacite porphyries, dikes, and sills	Chalcopyrite and pyrite in massive sulfides in skarn and volcanic rock adjacent to porphyry chalcopyrite and pyrite as disseminated grains and in veins in porphyries o
Subduction and initial formation of island arc. Permian and Pennsylvanian	Submarine volcanism and sedimentation forming the Slana Spur Formation and Tetelna Volcanics	Chalcopyrite, bornite, sphalerite, and pyrite as disseminated grains in fractures and in hydrothermally altered andesite and dacite flows and volcaniclastic rocks =

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Reviewed by John P. Albers and Henry C. Berg.

FAULT ZONE GEOMETRY OF THE 1979 ST. ELIAS, ALASKA, EARTHQUAKE

By Robert A. Page, Michael H. Hassler, Christopher D. Stephens, and Edward E. Criley

On February 28, 1979, a large earthquake with a body-wave magnitude (m_b) of 6.4 and a surface-wave magnitude (M_S) of 7.1 occurred beneath the Chugach and St. Elias Mountains near Mount St. Elias. Focal mechanisms for the earthquake indicate reverse slip on a low-angle, north-dipping fault or fault system (Lahr and others, 1980; Perez and Jacob, 1980; Hasegawa and others, 1980). Rupture initiated beneath the Jefferies Glacier at a depth of 13 km (Stephens and others, 1980), with an uncertainty of about 5 to 10 km, and propagated south and east in an episodic fashion (Boatwright, 1980; Hasegawa and others, 1980). Aerial reconnaissance of the meizoseismal region revealed extensive snow avalanches in the central high core of the St. Elias Mountains, landslides in the foothills, and ground cracks in unconsolidated deposits along the margin of the Malaspina Glacier, but neither the aerial reconnaissance nor subsequent field investigations of mapped faults south of St. Elias massif disclosed evidence of fresh fault displacements (Lahr and others, 1979; Plafker, 1980).

Although many gross features of the earthquake source and rupture process could be satisfactorily resolved from seismic data recorded at regional and teleseismic distances, the configuration and depth of the fault system could not.

Buried fault surfaces that rupture in large earthquakes often can be resolved from the spatial distribution of accurately located aftershocks. The 1979 earthquake occurred near the periphery of the regional seismograph network operated by the U.S. Geological Survey in coastal southern Alaska. The broad distribution of aftershock epicenters (fig. 38, top) located with data from that network and nearby Canadian stations at distances of 100 to 250 km is consistent with the direction and extent of rupture inferred from teleseismic studies of the main shock (Boatwright, 1980; Hasegawa and others, 1980). Boatwright (1980) suggested that the dense concentration of aftershocks along the United States-Canadian border at latitude 60°15'N might be related to a splay or offset in the main fault surface that acted as a transient barrier to rupture during the earthquake. Vertical cross sections through the aftershock zone (fig. 38, bottom) show that the shocks were all less than 20 km deep; however, the precision of depths determined from the regional data is not sufficient to define the causative low-angle fault or fault system.

To resolve the geometry of the fault system, at least locally near the center of the aftershock zone, we deployed a temporary array of four 3-component seismographs near the dense cluster of epicenters (fig. 38, top). The stations were configured in a quadrilateral with a maximum diagonal of 8 km (fig. 39, top). Of the permanent stations in the regional seismograph network, the closest was about 25 km southwest of the center of the quadrilateral; the two next nearest stations were about 35 km east-southeast and west-southwest. The four-station array recorded hundreds of aftershocks, but only those that were recorded well and occurred within 25 km of the temporary stations were studied in detail. The P-wave velocity model and associated station traveltime corrections proposed for the region by Stephens and others (1980) were initially used to locate the events. The model features a laterally homogeneous crust with a linearly increasing velocity from 5.0 km/sec at the surface to 7.8 km/sec at 32 km depth; a half-space mantle with a constant velocity of 8.2 km/sec; and a P:S-

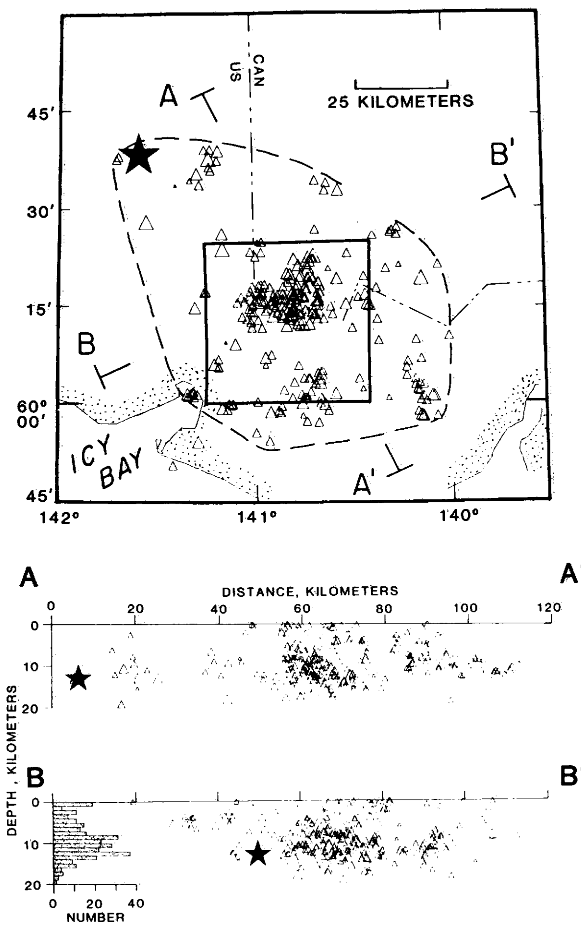

Figure 38.—Map and cross sections showing hypocenters of the 1979 St. Elias mainshock (star) and 287 aftershocks (triangles) that occurred between February 28 and March 31. Hypocenters determined from data recorded by regional seismograph stations (after Stephens and others, 1980). (Top) Map of epicenters showing location of cross sections and area included in figure 2. Symbol size is proportional to magnitude. Nearly all shocks are larger than magnitude 2.5. Dashed line defines boundary of aftershock zone. (Bottom) Cross sections of hypocenters shown in map and histogram of focal depths (left side of lower section).

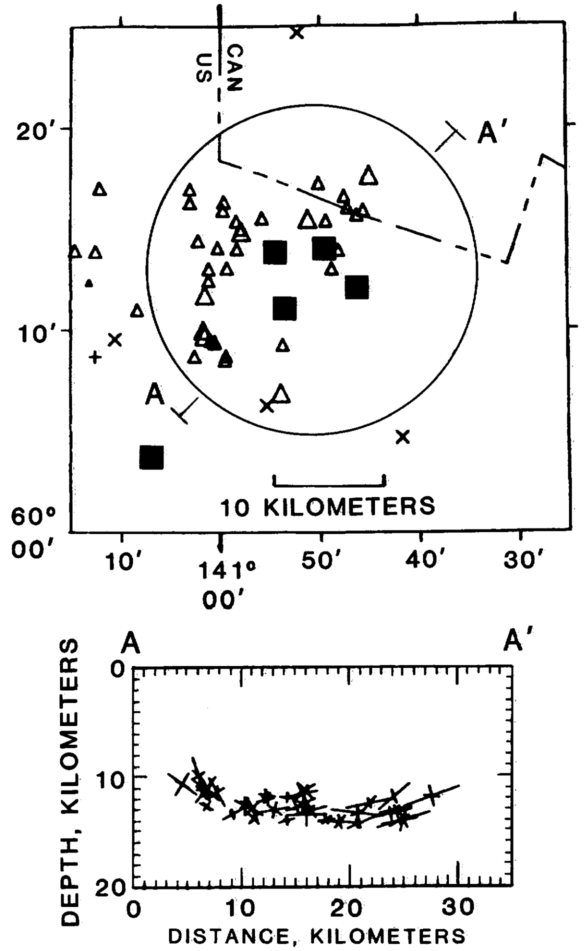

Figure 39. —Map and cross section showing hypocenters of 45 St. Elias aftershocks that occurred between July 26 and August 15. Hypocenters determined from data recorded by temporary four-station array (solid squares at center of map and by neighboring stations of the coastal southern Alaska seismograph network. (Top) Map of epicenters showing station locations (solid squares). Symbol type indicates depth range: +, 0-4.9 km; x, 5.0-9.9 km; triangle, 10.0-14.9 km. Symbol size is proportional to magnitude over range 0.0-2.9. (Bottom) Northeast-southwest cross section of hypocenters, showing projection of principal axes of one-standard-deviation confidence ellipsoids. View direction is northwest. Only those events lying within the 15-km radius circle shown in the map (top) are included.

wave velocity ratio of 1.78. Traveltime residuals for shocks located with this model were negative for P waves to distances of 200 km, but positive for S waves at distances less than about 35 km. Using 12 shocks that were especially well recorded at the four array stations and the two closest network stations, several alternative layered velocity models were explored to see which yielded the smallest traveltime residuals. A significant azimuthal variation in traveltimes at distances in the approximate range 100-135 km was observed that suggests the presence of a major discontinuity in crustal structure in the St. Elias-Icy Bay area, in which ray paths to stations to the southeast traverse a higher velocity structure than paths to the west and northwest. Of the several models considered, the preferred one has a constant-velocity layer of 5.0 km/sec over a 7.2 km/sec half-space, with the interface at 15 km to the southeast of the array and at 20 km to the west and northwest. The inferred discontinuity in structure lies near the Pamplona zone, a broad zone of late Cenozoic onshore and offshore folds and thrust faults that separate the Yakutat and Wrangell blocks (Lahr and Plafker, 1980).

Figure 39 (top) shows the epicenters of 45 shocks recorded between July 26 and August 15 that were located with the preferred velocity model. Only those events are shown for which the estimated (one standard deviation) uncertainties in the computed epicenters and focal depths are less than 3 km. The events occurring within the 15-km radius circle are more accurately located. All but one of the events inside the circle lie in the depth range 9-15 km. When projected onto a vertical plane, the hypocenters within the 15-km circle define a narrow tabular zone with little (less than 5°) or no dip (fig. 39, bottom). Within the uncertainty in the data, the events could originate on a single horizontal fault plane at a depth of 11 to 14 km. Alternatively, they could originate in a horizontal or subhorizontal zone of faulting, but the thickness of the zone could be no more than 2 or 3 km. Focal depths of less than 9 km were determined for two of the shocks located outside the 15-km circle; however, the depths have relatively large uncertainties. These shallow hypocenters are interpreted as scatter in the data rather than as evidence that faulting extends to shallow depths in the study area.

Thus, at the center of the rupture zone of the St. Elias earthquake, the distribution of accurately located hypocenters indicates a single fault surface or thin fault zone, approximately horizontal and planar, lying at a depth of about 11 to 14 km. Within the resolution of the data, there could be an offset or irregularity in the main fault surface as large as 1 or 2 km, which could have been a temporary barrier to rupture in the mainshock as postulated by Boatwright (1980). However, there is no evidence for an upward splay off the main fault trace or for an imbricate system of faults. Further analysis is planned to examine whether the defined fault surface is the boundary between the subducted Pacific plate and the overlying Wrangell block and to resolve more clearly the gross features of crustal velocity structure in this region.

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Reviewed by W. L. Ellsworth and W. H. Bakun.

STABLE ISOTOPE STUDY OF QUARTZ VEINS IN THE PORT VALDEZ GOLD DISTRICT

By W. J. Pickthorn

Gold mineralization in the Port Valdez district occurs principally in high-grade fissure veins within greenschist facies rocks of the Cretaceous Valdez Group. The Valdez Group, composed mainly of graywacke and argillite, is the major component of the Chugach terrane which was accreted to the Gulf of Alaska margin during Late Cretaceous and early Tertiary time (Plafker and others, 1977). During accretion, these rocks were metamorphosed and deformed into steeply dipping, east-west trending folds. Both barren and mineralized quartz veins are commonly found in the Port Valdez district. Gold-bearing quartz veins occur principally along fractures and shears that crosscut the local structure and fabric. The barren quartz veins normally parallel the local structure but in places strongly resemble the mineralized veins (Pickthorn and Silberman, 1984). Oxygen isotope analyses of these two types of quartz veins and hydrogen isotope analyses of their inclusion waters were performed in order to determine the origins of the fluids from which the veins formed. A quartz vein from a small mineralized granitic intrusion was also analyzed.

To determine the oxygen isotopic composition of the hydrothermal fluid the temperature of vein formation must be known or assumed. For the barren quartz veins, a formation temperature of 340°C was obtained from quartz-calcite fractionation (Clayton, O'Neil, and Mayeda, 1972; O'Neil, Clayton, and Mayeda, 1969). Petrographic and fluid inclusion examination of sample 79BS017 showed that the quartz and calcite are intergrown and to have the same freezing-point-depression and homogenization temperatures (Pickthorn, 1982). The temperature of formation of the mineralized veins was assumed to have been approximately 220°C based on fluid inclusion and petrographic evidence (Pickthorn and Silberman, 1984). Oxygen and hydrogen isotope compositions of the quartz veins and for calculated waters in equilibrium with the quartz at the assigned temperatures are given in table 13 and plotted in figure 40. $d^{18}0$ values for the waters were calculated using the quartz-water fractionation equation of Clayton, O'Neil, and Mayeda (1972).

The isotopic composition of waters calculated to be in equilibrium with the barren quartz veins plot within the area of overlap between the metamorphic and magmatic water regions (fig. 40). Structural and petrographic evidence indicate that these veins formed from a metamorphic fluid during and in response to regional metamorphism (Pickthorn, 1982; Robin, 1979). The calculated oxygen isotope formation temperature of 340°C is within the range expected for the low greenschist facies metamorphism of the surrounding country rock, which is in agreement with the proposed metamorphic origin. Data points for the mineralized quartz veins all plot outside the metamorphic water region (fig. 40), suggesting that meteoric water was a significant component of the hydrothermal fluid. Fluids comprised dominantly of meteoric water are unlikely for two reasons: (1) variations in dD for samples from the same vein (79BSO11Y, 79BSO11Z) suggest varying fluid composition (table 13), and (2) too large of an oxygen isotope shift is required of the meteoric water.

Table 13.—dD values for fluid inclusions, measured $d^{18}0$ values for quartz and calculated $d^{18}0$ values for H_20 in equilibrium with the quartz at the assigned formation temperatures, Port Valdez gold district, Alaska

$D^{18}O$ Measured	$d^{18}O$ H_2O calculated	dD Fluid inclusion
Metamorphic quartz vein samples. Temperature 340°C		
+15.1	+8.9	-49
+14.1	+7.9	-41
+14.6	+8.4	-40
+15.1	+8.9	-46
Mineralized quartz veins; metamorphic rock hosted. Temperature approximately 220°C.		
+13.0	+2.0	-63
+14.1	+3.1	-79
+12.9	+1.9	-71
+13.4	+2.4	-81
+13.6	+2.6	-86
Mineralized quartz vein from Rough Tough intrusion. Temperature approximately 450°C.		
+16.2	+12.7	-53

At the Rough Tough mine location, (fig. 41), mineralization occurs in quartz veins in a small granitic intrusion and the immediate surrounding hornfels. Similar plutons in the Valdez Group have been determined to be anatectic in origin and genetically related to higher grade metamorphism at depth (Hudson and others, 1979). Intrusive bodies of this type generally form at temperatures above 600°C (Winkler, 1976); however, mineralization occurred at a lower temperature, probably in the range of 450° to 500°C. Oxygen and hydrogen isotope analysis of a mineralized vein sample from the Rough Tough mine indicates that the mineralization fluid was dominantly, if not completely, metamorphic water.

The mineralized veins not associated with intrusive rocks are all post-metamorphism. Pickthorn (1982) and Mitchell and others (1981) have proposed models in which post-metamorphic uplift and dilation of the Valdez Group allowed meteoric water to circulate deeply along fractures and faults. Hydrothermal convection cells were probably formed, heated, and driven by the still hot high-grade metamorphic rocks at depth. The gold and other vein constituents may have been leached from the metamorphosed sedimentary rocks and deposited in fissure veins nearer the surface. Mineralization at the Rough Tough mine was probably related to a metamorphic fluid derived from the same source region as the Rough Tough pluton.

Of particular importance from an exploration standpoint is the difference in measured d values for the barren and mineralized quartz veins. The mineralized quartz veins have slightly lower $d^{18}0$ and significantly lower dD values compared to those of the barren veins.

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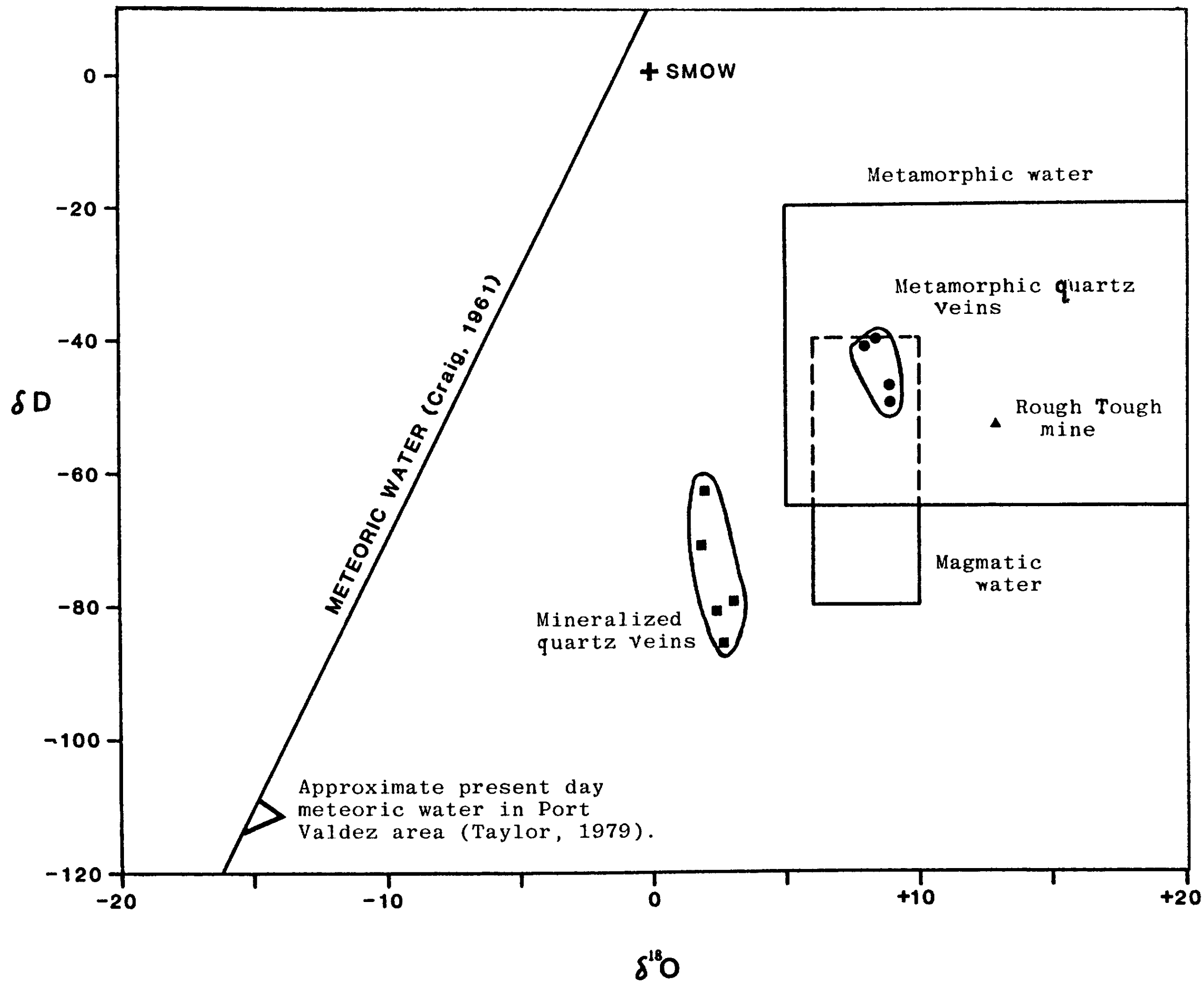

Figure 40.—dD/d^{18}0 diagram showing meteoric, metamorphic, and magmatic water regions, and calculated compositions of waters in equilibrium with barren and mineralized quartz vein samples. Diagram modified from Taylor, 1979.

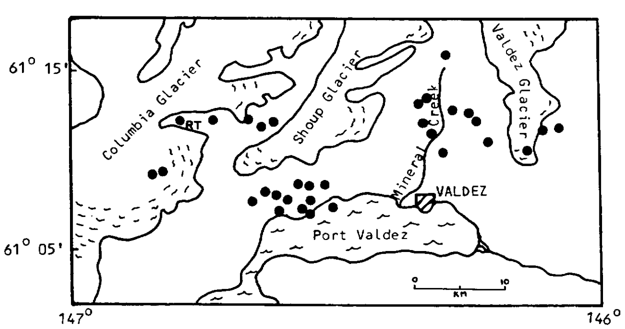

Figure 41.—Location and distribution of gold occurrences in the Port Valdez district, southern Alaska. RT, Rough Tough mine.

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PRELIMINARY RECONNAISSANCE SULFUR ISOTOPE GEOCHEMISTRY OF MASSIVE SULFIDE OCCURRENCES IN THE PRINCE WILLIAM SOUND DISRICT

By W. J. Pickthorn and Steven W. Nelson

Several small "Cyprus-type" and stratabound massive base-metal sulfide deposits are localized in the mafic volcanic complex of Knight Island and sedimentary rocks of Latouche Island. These deposits were formed in and on the ocean floor during a period of active volcanism in the early Tertiary (Tysdal, 1978; Tysdal and Case, 1979). A preliminary reconnaissance sulfur isotope study of several of these deposits and other sulfide occurrences in the Prince William Sound area was made in order to discern their origin and depositional environments. No detailed geochemical or isotopic work has been done on any of these occurrences.

In the submarine environment magmatic sulfur and reduced ocean water sulfate are the only reasonable sources of sulfur for the massive sulfide deposits. Sulfide minerals deposited from magmatic sulfur generally have d^{34}S values near or slightly above 0 (Ohmoto and Rye, 1979). d^{34}S values of sulfides formed from reduced ocean water sulfate are variable and reflect the depositional environment and method of sulfate reduction (Schwarcz and Burnie, 1973). The d^{34}S values and host bedrock types for the samples analyzed are given in table 14.

Table 14.—d^{34}S values for sulfide minerals from massive sulfide deposits and bedrock samples from the Prince William Sound area

Sample	Description	Mineral	d^{34}S
5A	Beatson mine; bedrock silicified sandstone and shale	pyrite	+7.8
11A	Duke mine; disseminated pyrite in slate and graywacke	pyrite	+11.0
22A	Quartz vein in sheared mafic mafic dike	pyrite chalcopyrite	+6.5 +5.8
45A	Quartz vein in pillow basalt	pyrite chalcopyrite	+5.3 +4.9
49B	Rua Cove mine; volcanic host host rocks	chalcopyrite	+4.2
137B	Sheeted dike complex	pyrite	+0.9
2A	Jeanie Point area	pyrite	+23.2
35D	Disseminated pyrite in fossiliferous and sandy limestone.	pyrite	+14.4

Four samples of sulfides in mafic volcanic host rocks from Glacier and Knight Islands were analyzed (fig. 42). Sample 49B is massive chalcopyrite ore from the Cooper Bullion mine, Rua Cove, eastern Knight Island (fig. 42). Bedrock at the Rua Cove mine consists of fine-grained greenstone, pillow basalt, and altered quartz diorite (Richter, 1965). Detailed examination of this deposit has shown it to be of the "Cyprus-type" and volcanogenic in origin (R. A. Koski, U.S. Geological Survey, written commun., 1982). The d^{34}S value of the Rua Cove ore is +4.2 which is within the range expected for magmatic sulfur (Ohmoto and Rye, 1979). Samples 22A and 45A are sulfides closely associated with quartz veining and hydrothermal alteration of pillow basalt in western Knight Island. Similar quartz veining at the Rua Cove mine is considered

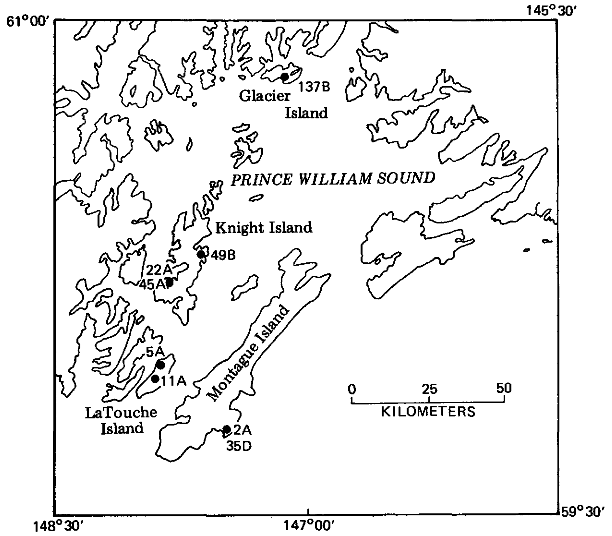

Figure 42.—Prince William Sound region showing localities of samples used in this report.

to represent the feeder system for the ore forming fluids (R. A. Koski, written commun., 1982). d^{34}S values for these samples are within the range of values normally found for magmatic sulfur but their high values may indicate a partial involvement of ocean water sulfate. Oxygen isotope analyses of quartz from these two rocks are approximately +11 indicating that the quartz may have been deposited from magmatic hydrothermal fluids.

Samples of sulfide deposits in the sedimentary rocks of Latouche Island have d^{34}S values that fall within the range of sulfides formed by either thermal or biogenic reduction of ocean water sulfate in a closed basin. The high d^{34}S values of sedimentary pyrite from the Jeanie Point area of Montague Island (fig. 42) and the presence of pyrrhotite in the ore deposits of Latouche Island suggest that the area was dominately a closed-system reducing environment. This does not, however, preclude the possibility that these deposits formed by thermal reduction of ocean water sulfate. Sulfide mineralization in the sedimentary rocks of Latouche Island occurs as stratiform massive and disseminated deposits. The massive deposits are usually accompanied by silicification, sericitization, and kaolinization of the slate and graywacke country rock (Stejer, 1956). These deposits may represent sites of submarine thermal springs that would not only provide a source of reduced sulfate but also of metallic ions leached from the ocean floor sediments and underlying rocks. Sample 5A is massive pyrite from the Beatson mine (fig. 42, table 14). Bedrock hosting the disseminated sulfides is for the most part unaltered, suggesting a source that is dominated by biologically reduced sulfate. Sample 11A is dissimated sulfide from near the Duke mine.

The Gulf of California is an area in which sulfides are presently being deposited by mechanisms similar to those proposed in this paper and may represent a modern analog to the Prince William Sound region.

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Reviewed by J. R. O'Neil and W. J. Nokleberg.

SHIELD VOLCANOES IN THE WRANGELL MOUNTAINS, ALASKA

By D. H. Richter, J. G. Smith, J. C. Ratté, and W. P. Leeman[1]

Tanada Peak and Capital Mountain along the northern fringe of the extensive Wrangell Mountains volcanic field in south-central Alaska (fig. 43) are remnants of two Pleistocene andesitic shield volcanoes. The volcanoes are characterized by apparently non-explosive summit calderas and may have been similar in form to active Mount Wrangell, the only heretofore recognized shield volcano in the volcanic field. Both volcanoes have had large parts of their edifices removed by erosion, leaving flat-lying caldera-fill lavas exposed on the higher elevations of the present structures.

Tanada volcano (lat. 62°17'N., long. 143°34'W.), which has a present summit elevation of 2,807 m, formed a large shield covering an area of probably more than 400 km^2 (fig. 44). Much of the original shield has been stripped away, and the volcano's

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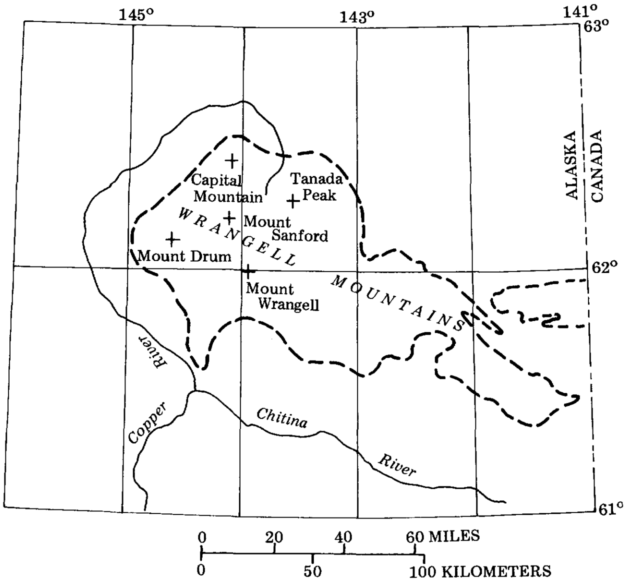

Figure 43.—Part of south-central Alaska showing extent of Wrangell Mountain volcanic field (dashed line) and the location of Tanada and Capital volcanoes.

southern flank is covered by flows and pyroclastic deposits from younger eruptive centers. The volcano was built on a thick sequence of older andesitic flows, flow breccias, and laharic deposits that generally dip gently to the southeast. K-Ar determinations on a pre-Tanada pyroxene andesite and a post-Tanada olivine basalt (table 15) suggest that the entire Tanada volcano was built and largely destroyed in the Pleistocene. Tanada shield lavas, consisting chiefly of thin (<10 m) andesite flows and of only minor intercalated lahars and pyroclastic deposits, dip 1° to 20° away from the summit area. The oval-shaped summit caldera, 8.1 km long by 5.6 km wide, is filled with more than 900 m of flat-lying massive andesitic flows and dacitic agglutinates. The agglutinate flows, as much as 50 m thick, occur in the upper part of the caldera-fill section and probably represent some of the late eruptions in the caldera. Most of the exposed caldera wall is composed of the older andesite sequence; only on the east side of the volcano and locally elsewhere, where erosion has been less extreme, are shield lavas still present in the wall. The caldera wall dips 20° to 50° inward and is mantled locally by thin pyroclastic beds and rubbly breccias. A number of andesitic dikes intrude the

Table 15.—Potassium-argon age determinations of pre-Tanada pyroxene andesite and post-Tanada olivine basalt, Wrangell Mountains, Alaska

Field Number	Material dated	Percent K_2O	Average K_2O	$^{40}Ar_{RAD}$ (moles/gm $\times 10^{-12}$)	$\frac{^{40}Ar_{RAD}}{^{40}Ar_{Total}}$	Calculated age $\pm$ analytical error (millions of years)
81ASJ77BN	Whole rock, crushed; treated with HNO_3	1.229 1.226	1.2275	2.883	15	1.63 $\pm$ 0.05
81ASJ77BF	Whole rock, crushed; treated with HNO_3 + HF	1.238 1.225	1.2315	3.531	33	1.99 $\pm$ 0.08
Sample description and location: fine-grained glassy olivine-free clinopyroxene andesite. Sample is from the center of an isolated pillow within a thick deposit of monolithologic hyaloclastite and interbedded broken pillow breccia layers and autointrusive flow tounges. Nabesna B-6 15-minute quadrangle, in bed of unnamed creek at 4,000-ft elevation, 62°20.27'N, 143°33.26'W						
81ASJ86N	Whole rock cruched; treated with HNO_3	1.259 1.241	1.250	1.749	15	0.972 $\pm$ 0.049
Sample desription and location: fine-grained vesicular olivine basalt flow. Most olivine phenocrysts are less than 1 mm in diameter. Sample is from a flow which is part of a post-Tanada cinder and lava cone. Nabesna B-6 15-minute quadrangle, at 6,400 ft elevation between peaks 6760 and 6715, 62°21.2'N, 143°33.1'W.						

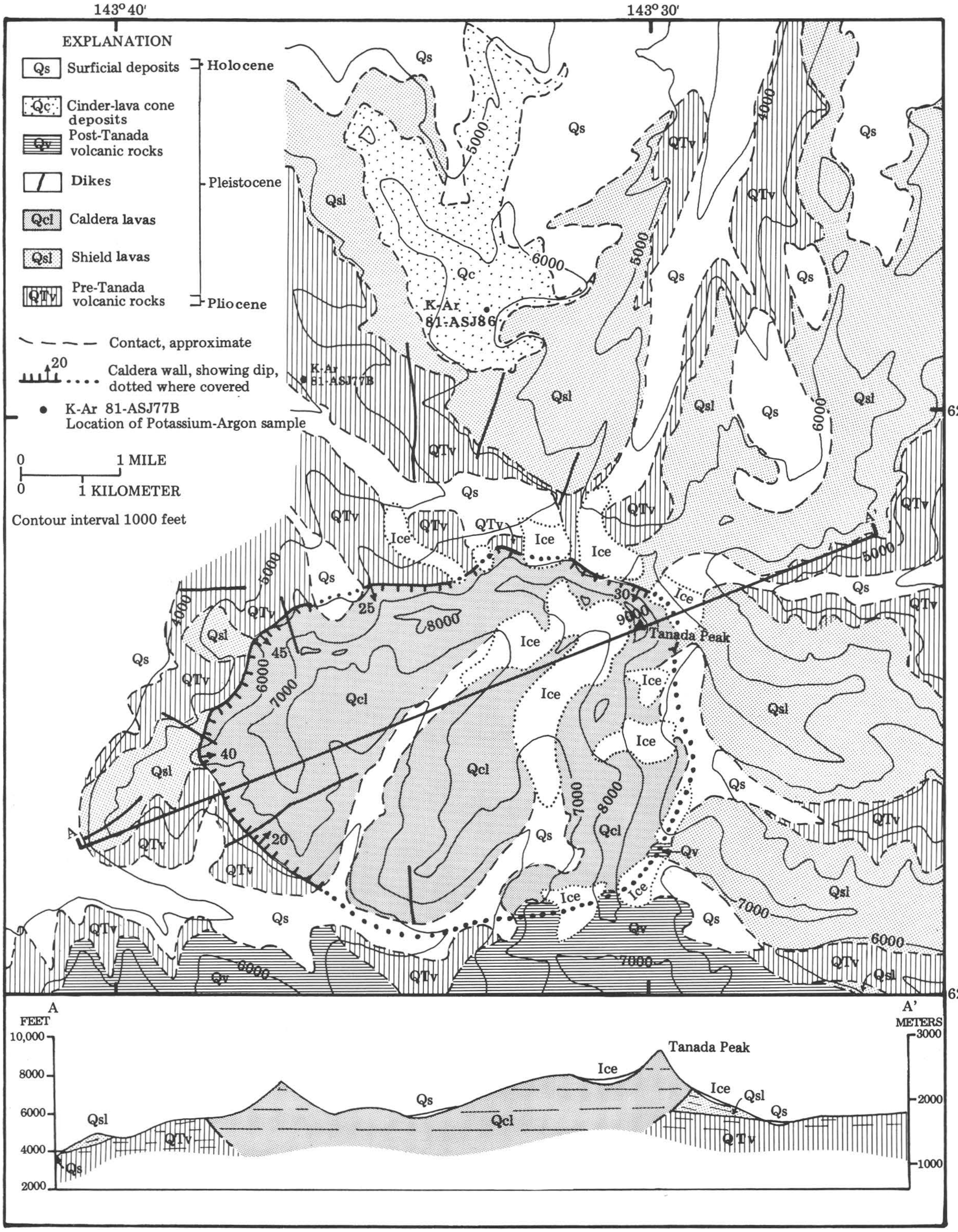

Figure 44.—Generalized geologic map and cross section of Tanada volcano. Topographic bases modified from Nabesna B-6 and B-5 quadrangles, Alaska (1964).

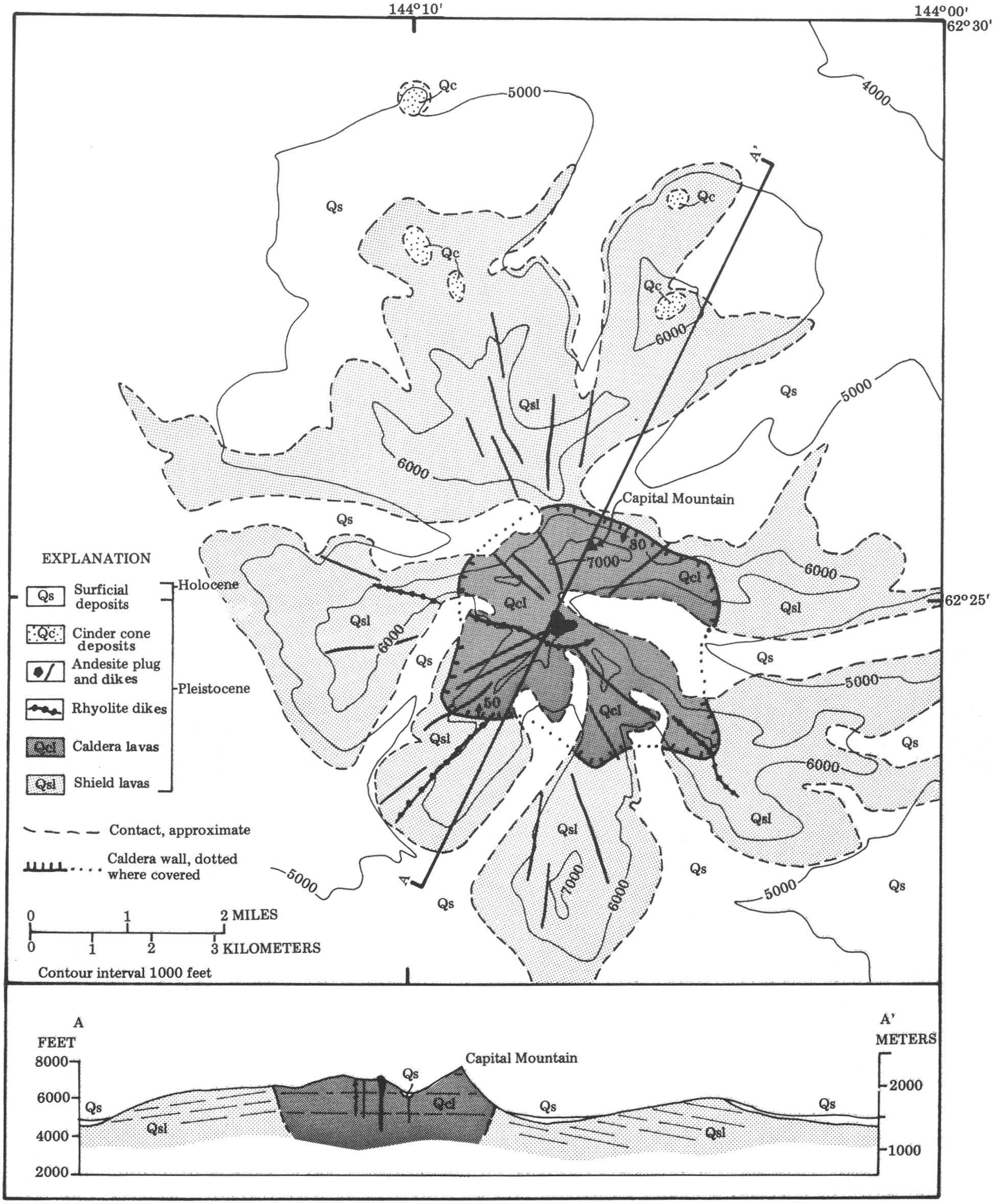

Figure 45.—Generalized geologic map and cross section of Capital volcano. Topographic bases modified from Gulkana B-1 quadrangle, Alaska (1964).

structure, and a rift-like chain of andesitic cinder-lava cones covers parts of the volcano's north flank.

Capital volcano (lat. 62°15' N., long. 144°08' W.), 32 km northwest of Tanada volcano and 2,320 m high, is a smaller shield that originally covered an area of about 200 km^2 (fig. 45). It has a roughly circular summit caldera 4 km in diameter. The shield consists chiefly of lava flows and of minor intercalated lahars and pyroclastic deposits that dip 3° to 25° away from the caldera. The caldera, where walls dip 50° to 80° inward, is filled with flat-lying massive flows, as much as 40 m thick. Locally, talus breccias, flow breccias, and irregular masses of pillow lavas occur as moat deposits between the caldera wall and caldera-fill flows. An andesitic plug, 300 m in diameter, intrudes the caldera-fill flows and forms a prominent spire rising 100 m above the present surface. A spectacular swarm of hundreds of nearly vertical andesitic dikes and a few rhyolitic dikes radiates from the general area of the plug. The dikes average less than 2 km long and 4 m thick. The base of the shield is not exposed. Flows from Mount Sanford and some of its flank eruptive centers locally overlie the southern edge of the shield, and a number of young andesitic cinder cones are scattered over the north flank of the shield. A K-Ar date on plagioclase from one of the rhyolitic dikes has yielded an age of 1.09 ± 0.17 m.y. (M. A. Lanphere and B. Dalrymple, U.S. Geological Survey, unpublished data).

The shield and caldera lavas of the two volcanoes and most of the associated dikes consist largely of a variety of aphanitic and porphyritic andesites. Porphyritic varities are chiefly olivine and olivine-hypersthene andesites; augite-hypersthene andesites are present but are apparently minor. The rhyolitic dikes of Capital and late caldera-fill dacitic agglutinates of Tanada suggest that more silicic magmas were beginning to evolve when eruptive activity ceased. Petrochemical studies of the rocks are in progress.

Reviewed by Fred Barker and B. L. Reed.

ESTIMATED PHYSIOLOGICAL TOLERANCE RANGES FOR SELECTED ELEMENTS IN SOILS, CAPPS COAL FIELD

By R. C. Severson and L. P. Gough

Data about the geochemistry and biogeochemistry of native (undisturbed) plants and soils, respectively, in areas to be stripmined, are essential for objective evaluation of the rehabilitation potential of an area. Such data can also be used after development to assess the changes in the chemical composition of plants and soils resulting from the rearrangement of rock strata and the disruption of soil development and natural plant communities.

Figure 46.—Location of the Capps coal field, Alaska.

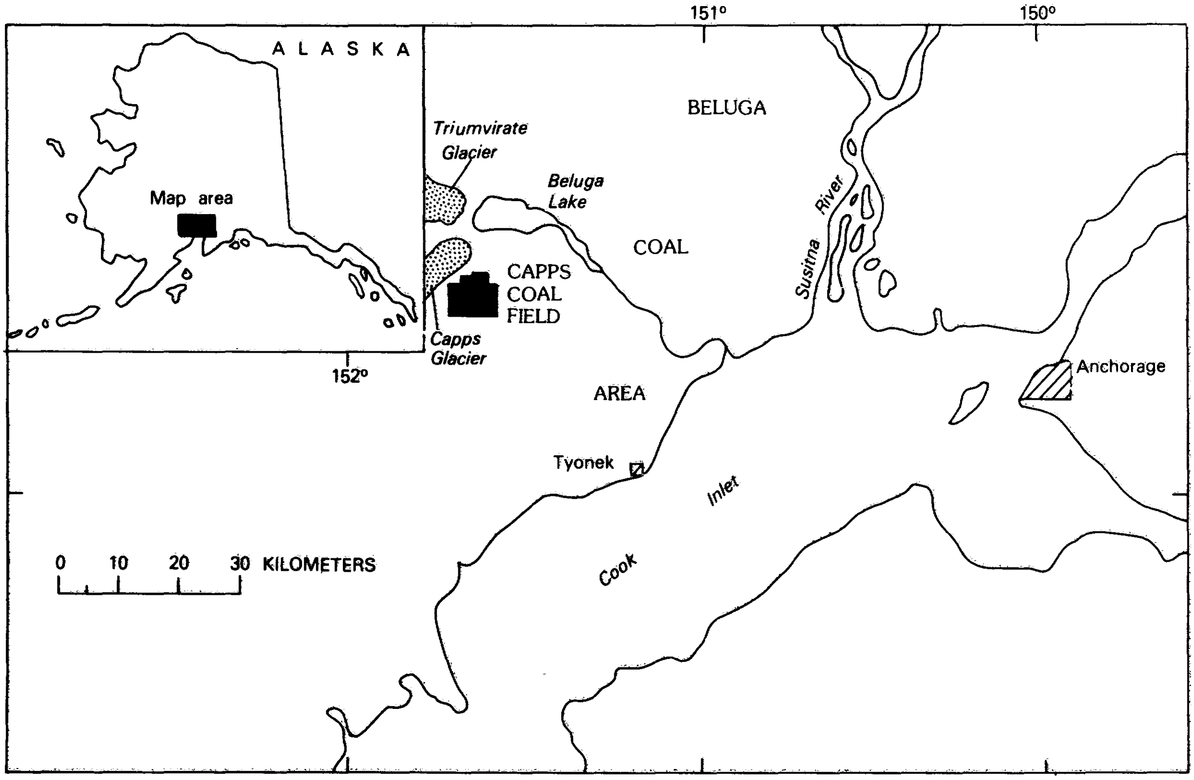

Table 16.—Parameters and methods of analysis for plant and soil materials from the Capps coal field

Parameter	Method
	Soil Materials
Al, Cd, Co, Cu, Fe, Mn, Ni, P, Pb, and Zn	ICP (induction coupled plasma) analysis of AB-DTPA (ammonium bicarbonate-diethylene-triaminepentaacetic acid) extract of <2 mm soil material
pH----------------------------	Selective ion electrode
B-----------------------------	Direct current plasma analysis of hot-water extract of <2 mm soil material
Ca, Cl, K, Mg, Na, SO_4, SC (specific conductance), and SI (saturation index)	Various analytical methods including ICP, selective ion electrode, and conductivity bridge for water-saturtion extract of <2 mm soil material
Al, As, Ca, Cd, Ce, Co, Cr, Cu, Dy, Fe, Ga, Gd, K, La, Li, Mg, Mn, Mo, Na, Nb, Nd, Ni, P, Pb, Sc, Sm, Sn, Sr, Ta, Ti, Th, U, V, Y, and Zn	ICP analysis of an acid digestion for total analysis of <100 mesh soil material
	Plant Materials
Cd, K, and Na------------------	AAS (atomic absorption spectrophotometry) on digested plant ash
Al, Au, B, Ba, Bi, Ca, Ce, Co, Cr, Cu, Dy, Er, Fe, Ga, Ge, La, Li, Mg, Mn, Mo, Nb, Nd, Ni, P, Pb, Pr, Sn, Sr, Th, Ti, V, W, Y, Yb, Zn, and Zr	ICP analysis on digested plant ash
As----------------------------	AAS on plant dry material
Se----------------------------	Fluorometric analysis on plant dry material
U-----------------------------	Fluorometric analysis on digested plant material
S-----------------------------	Turbidimetric analysis on plant dry material

The Beluga coal area is estimated to contain 2.1 billion tons of subbituminous coal (Barnes, 1966). In this area, the Capps coal field (fig. 46) was selected for study because the coal beds are near the surface, the geologic structure of the area is uncomplicated (Schmoll and others, 1981), and the probability of stripmining in the near future is great (Placer Amex, 1977).

During July and August of 1980 and 1981, ten 1-km-square cells were positioned to cover the Capps coal field and, at nine locations in each cell, plant and soil samples were collected. The sampling locations were randomly selected according to an analysis-of-variance sampling design to permit estimation of chemical variation among and within the cells. The data obtained can be used to: (a) indicate the appropriate scale for geochemical maps, (b) estimate the minimum number of samples necessary to prepare stable map patterns, (c) compute geochemical baselines, and (d) examine the

Table 17.— Observed range in concentration (in parts per million unless otherwise indicated) of six elements in samples of two soil zones and three plant species (dry weight basis) from the Capps coal field

Sample	Al	Cu	Fe	Mn	P	Zn
AB-DTPA[1] Soil Extract						
Upper zone (n=90)	19 - 240	1.9 - 9.3	160 - 560	3.8 - 64	6 - 60	0.5 - 8.2
Lower zone (n=90)	32 - 760	1.1 -15	140 - 750	0.7 - 21	7 - 50	0.5 - 6.9
Total Content in Soil						
Upper zone (n=90)	7.2 - 10%	22 - 47	3.8 - 5.9%	860 - 1,300	70 - 110	61 - 98
Lower zone (n=90)	4.0 - 9.5%	11 - 120	1.3 - 18%	220 - 2,800	30 - 1,500	30 - 110
Calamagrostis canadensis (n=26)	<16 - 36	5 - 15	<16 - 84	200 - 720	0.16 - 0.52%	29 - 72
Festuca altaica (n=64)	<16 - 400	3.1 - 13	<16 - 260	170 - 690	0.19 - 0.45%	4.7 - 80
Salix pulchra (n=90)	16 - 220	2.1 - 12	<16 - 130	52 - 340	0.072 - 0.33%	48 - 120

[1] Ammonium bicarbonate-diethylenetriaminepentaacetic acid.

relations between soil geochemistry and plant uptake. The soil and plant parameters measured are given in table 16.

At each sample location a soil-profile pit was hand excavated and examined, and soil and plant samples were collected.

Soils.—Two samples were collected in each pit. The first sample consisted of a gray medium-sand-sized volcanic ash layer at the surface (2 to 8 cm in thickness), and the second consisted of a composite of material from below the first sample to a depth of 40 cm. At nearly all sampling locations, this composite sample included three layers of volcanic ash; at some locations it included four layers. The uppermost ash layer was a 5- to 20-cm-thick zone of dark reddish-brown coarse sand texture. The lower two ash layers were light yellowish-brown material, the upper one being silt loam in texture and the lower one having a very fine sand texture. These soils are tentatively classified as cryandepts. Because of rodent burrowing, frost heaving, and hummock formation, the layers of volcanic ash were commonly intermixed with each other and(or) convoluted and were not horizontally zoned in a uniform manner. Each sample consisted of about 1 kg of material.

Vegetation.—A willow and a grass sample were collected near the soil pit at every site. The willow sample consisted of the terminal 10 cm of young branches of several Salix pulchra individuals. The taxonomy of this species is unsettled (Viereck and Little, 1972; Argus, 1973), and its morphology varies considerably among sites. We feel confident that the material we collected is S. pulchra, even though some of the diagnostic features used in the field varied considerably among locations. Festuca altaica was the grass most frequently collected. It is a common inhabitant of the crests and sides of frost-heave hummocks. Where Festuca was not found, a substitution was made with Calamagrostis canadensis (bluejoint), which inhabits areas that are protected by snow cover in the winter but is not characteristic of communities associated with deep, late-lying snow fields.

The stage of maturity of the plants varied among sites and probably influenced, to an unknown degree, the levels of some of the element concentrations measured. Individual willows possessed pre-flowering capsules, while some were releasing seeds. Most of the grasses were in the flowering stage, although some samples from lower elevations were composed of individuals with ripening seeds.

As an example of the data available, table 17 lists the observed range in the concentration of six biologically important elements as determined in a soil extract, in a total digestion of the soil, and in the dry material of the three plant species. It is assumed that the extractable and total concentrations measured in the soil represent physiologically acceptable ranges within which, in general, the normal growth of these three species and the herbaceous tundra vegetation (that is, characteristic of the Capps study area) occurs. These soil concentration ranges, therefore, may be used in judging the feasibility of revegetating areas disturbed by mining, and similar operations when native vegetation is to be used in the rehabilitation effort. The range of element concentrations in plant materials is also a measure of physiological tolerance. However, additional work is needed to determine whether or not the soil extractable levels are a good measure of element availability for the uptake and translocation of these elements in plants.

Preliminary results of the geochemical data, as it relates to the rehabilitation potential of soil, are given in Severson and Gough (1983). A complete tabular listing of analytical results for both the soil and the plant parameters (summarized in table 16) is given in Gough and Severson (1983).

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Reviewed by J. B. Cathrall and Dawn Madden.

SEISMICITY ALONG SOUTHERN COASTAL ALASKA: OCTOBER 1981-SEPTEMBER 1982

By Christopher D. Stephens, John C. Lahr, and Robert A. Page

Among the interesting earthquakes recorded between October 1, 1981, and September 30, 1982, are four shocks that triggered strong-motion instruments, a pronounced earthquake sequence near Icy Bay in the aftershock zone of the St. Elias earthquake, and two moderate-sized earthquakes that occurred along a section of the Fairweather fault zone near Yakutat Bay. In addition, an increased emphasis on locating small crustal earthquakes along the volcanic axis west of Cook Inlet is providing significant new data for the study of the local tectonic and volcanic regime.

The main features in the distribution of seismicity along southern coastal Alaska are shown in figure 47. For earthquakes larger than coda-duration magnitude 2 (fig. 47a), the pattern is dominated by Benioff zone activity beneath and north of Cook Inlet. Relatively little crustal activity in this magnitude range is observed west of about 148° W. longitude. The most prominent shallow (depth less than 30 km) seismicity occurred north of Icy Bay in the eastern half of the seismograph network and is due to continuing aftershock activity from the large (7.1 M_s) 1979 St. Elias earthquake (Stephens and others, 1980). Aside from this aftershock activity, only a few earthquakes of magnitude 2 and larger occurred in or near the Yakataga seismic gap, located north of the 1,000-fathom contour approximately between the longitudes of Kayak Island and Icy Bay (Lahr and others, 1980; McCann and others, 1980). Twenty of the shocks shown in figure x1 had body-wave magnitudes[1] of 4.5 and larger. About half these events occurred within the Cook Inlet Benioff zone at depths below 60 km. Of the four largest shocks, all with body-wave magnitudes of 5.0, two occurred in the Cook Inlet Benioff zone and two occurred in the aftershock zone of the St. Elias earthquake.

In contrast to the larger events, the distribution of earthquakes less than magnitude 2 (fig. 47b) is biased toward higher concentrations of events in areas where special emphasis is placed on locating small shallow earthquakes. Along the Kenai Peninsula and in northern Cook Inlet around Anchorage the crustal activity is confined to depths of less than about 20 km and is generally not well correlated with mapped fault traces (Lahr and Stephens, 1982; Stephens and others, 1982). Along the volcanic axis from Mt. Spurr to Mt. Iliamna, nearly all of the shocks are located at depths between 5 and 20 km with two notable exceptions. First, most of the earthquakes located within 5 to 10 km of the active volcanoes—Spurr, Redoubt, and Iliamna—have depths less than 5 km. Second, focal depths range from the surface to about 10 km for a pronounced cluster of shocks about 15 km south of Mt. Spurr between the Chakachamna and McArthur rivers.

East of Prince William Sound the seismicity for magnitude less than 2 (fig. 47b) is more extensive and diffuse than that for the larger events. The pattern is dominated by aftershock activity from the 1979 St. Elias earthquake. A tight cluster of epicenters is apparent near the Copper River Delta, and a more diffuse concentration is seen around Waxell Ridge between the Copper River Delta and Icy Bay. Offshore activity is scattered, and, apart from six events near 144° W. longitude, appears to be roughly bounded by the 1,000-fathom contour in the eastern Gulf of Alaska. Diffuse activity also was observed about 100 km north of the St. Elias aftershock zone scattered about the Duke River fault.

On May 2, 1982, a prominent earthquake sequence was initiated within the aftershock zone of the St. Elias earthquake by a magnitude 5.0 m_b (5.1 M_s) shock a few kilometers east of Icy Bay. A second magnitude 5.0 m_b (4.7 M_s) event occurred on May 3 about 3 km southeast of the first shock and

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[1] Surface-wave (M_s) and body-wave (m_b) magnitudes taken trom the U.S. Geological Survey Preliminary Determination of Epicenters (PDE).

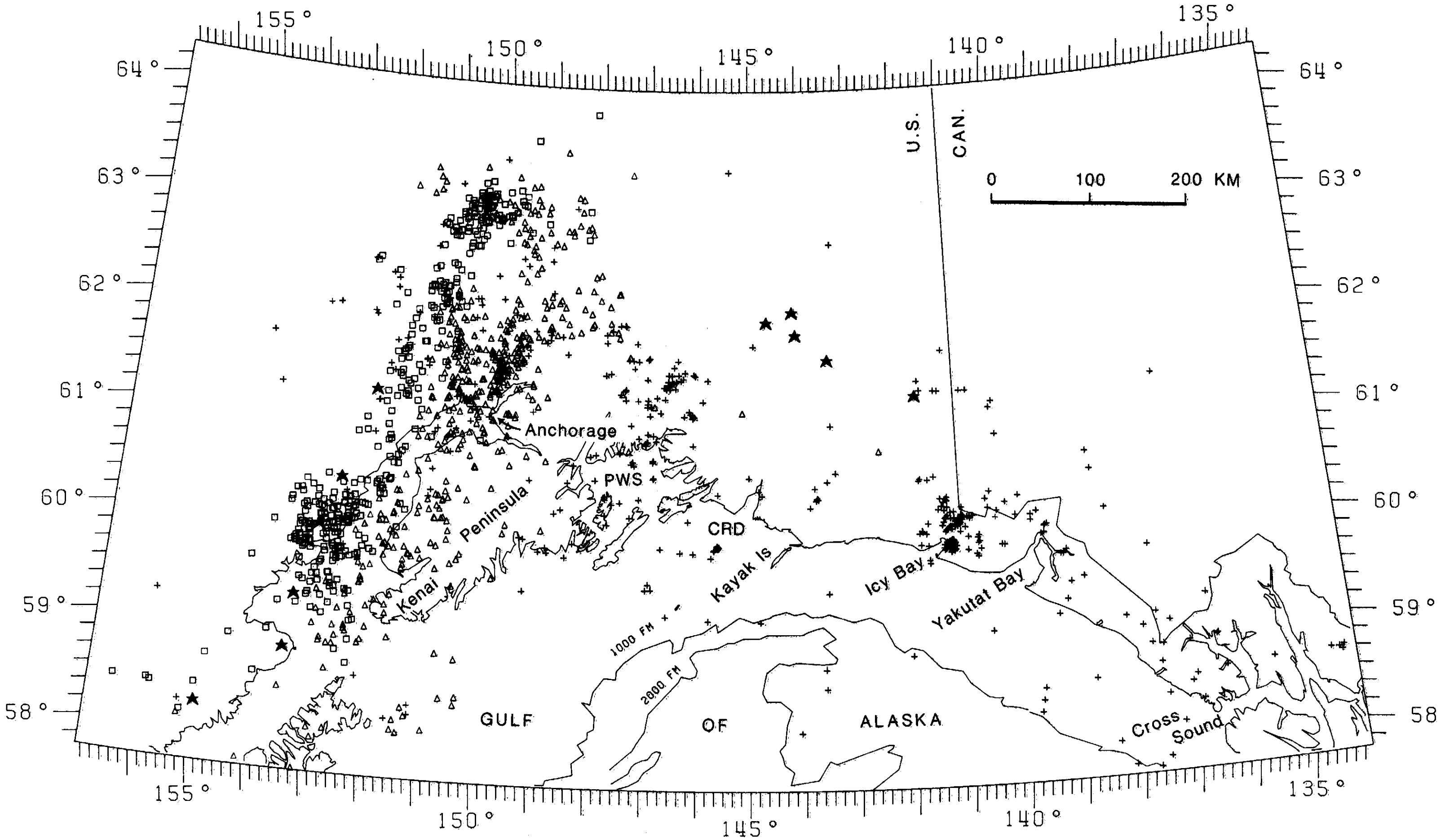


Figure 47a.—Epicenters for 1,302 earthquakes of coda-duration magnitude 2 and larger that occurred in southern Alaska between October 1, 1981, and September 30, 1982. Symbols indicate depth as follows: plus - less than 30 km; triangle - 30.0 to 69.9 km; square - 70 km or greater. Solid stars indicate Quaternary volcanoes. PWS - Prince William Sound, CRD - Copper River Delta.

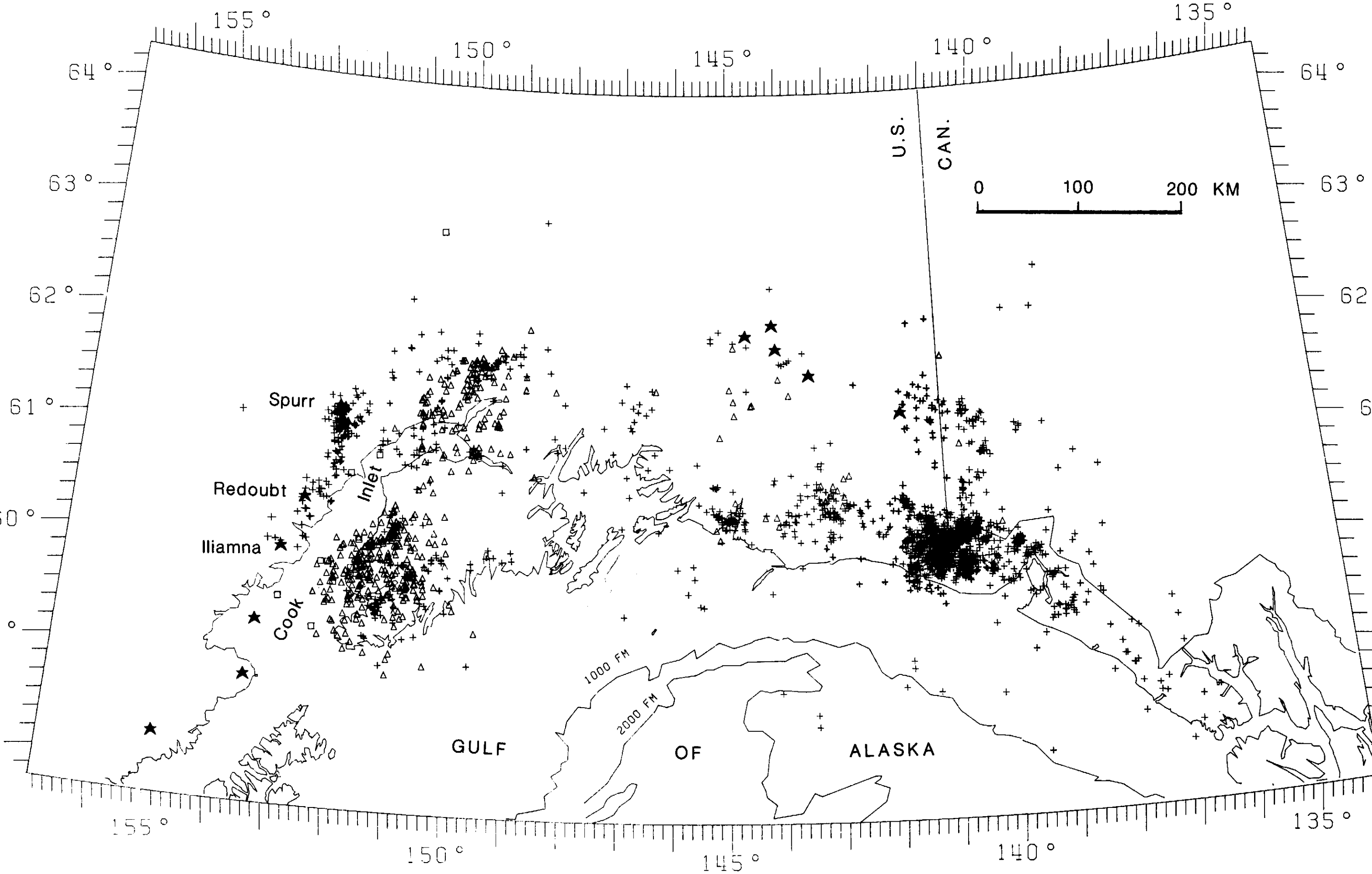
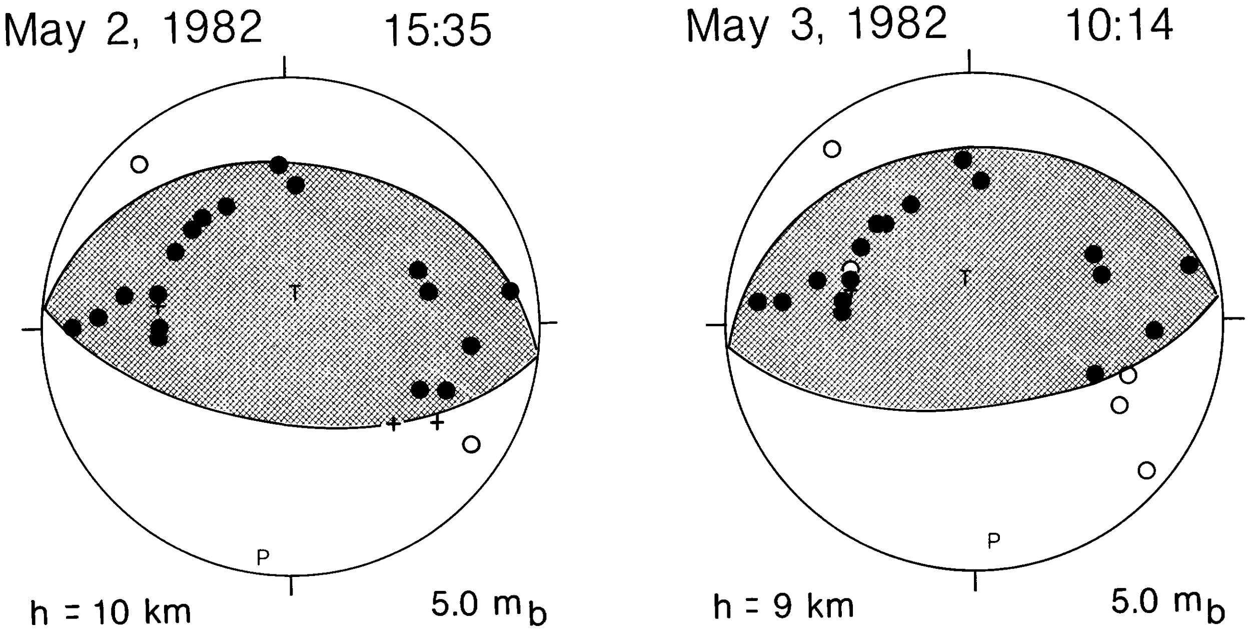


Figure 47b.—Epicenters for 3,853 earthquakes of coda-duration magnitude less than 2 that occurred in southern Alaska between October 1, 1981 and September 30, 1982. Symbols as in figure 47a.



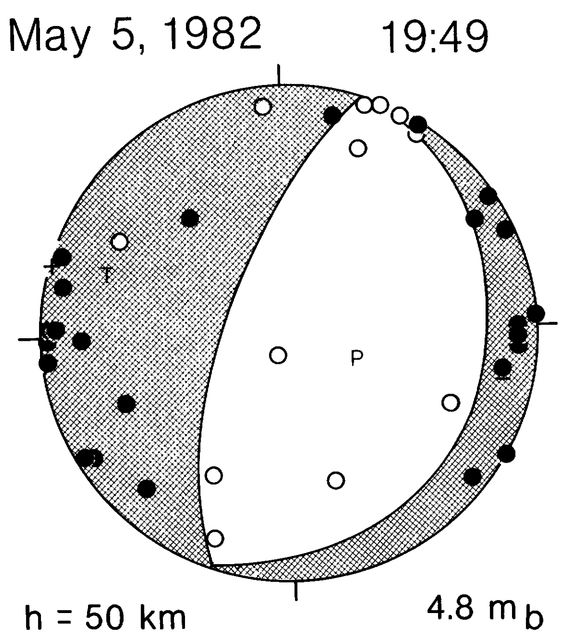


Figure 48.—Focal mechanisms determined from P-wave first motions of selected earthquakes. Compressions (solid circles, or + for less reliable readings) and dilatations (open circles, or -) are projected onto an equal area projection of the lower hemisphere. Compressional quadrants are shaded. Symbols 'P' and 'T' indicate axes of maximum and least compressive stress.

triggered a strong-motion instrument at a site about 26 km away. The maximum horizontal acceleration recorded was 0.02 g (1 g = 980 cm/sec^2). In the 3-day period following the first shock 324 aftershocks of coda-duration magnitude 1 and larger were located, and 124 others occurred by the end of May. The two largest shocks on May 2 and May 3 were located at depths of about 10 km and have focal mechanisms determined from P-wave first motions that are compatible with low-angle thrusting on northward-dipping planes (fig. 48). Considering the uncertainties in the velocity model used in this area, the focal mechanisms are remarkably similar to that determined for the St. Elias mainshock, which occurred about 100 km to the northwest (Lahr and others, 1979). Also, the depths determined for the two largest events are similar to those found by Page and others (this volume) for a selected group of aftershocks about 30 km to the northeast that have good depth control. These observations suggest that the May 1982 sequence may have occurred along the same buried fault system that ruptured during the 1979 St. Elias earthquake.

In contrast to the prominent sequence following the two shocks near Icy Bay, a magnitude 4.8 m_b earthquake that occurred on May 5, 1982, about 20 km east of Anchorage at a depth of 50 km was an isolated event. This shock was felt at Modified Mercalli Intensity V in Anchorage, Eagle River, and Palmer (PDE) and triggered 10 strong-motion instruments between Talkeetna and Whittier. The peak ground acceleration recorded was 0.08 g at the U.S.G.S. building in Anchorage (A. G. Brady, U.S. Geological Survey, oral commun., 1982). A focal mechanism determined from P-wave first motions (fig. 48) has a steeply dipping pressure axis and an ESE-WNW oriented tension axis aligned approximately downdip with respect to the Benioff zone. Earthquakes as small as magnitude 1.5 are commonly located in this area, but none was located with 10 km of the epicenter of the May 5 shock for periods of at least one month preceding and following the event. The lack of any detectable aftershock activity for the May 5 shock is consistent with the observations of Page (1968) that subcrustal earthquakes, in contrast to shallower events, usually do not have prominent aftershock sequences.

Two other earthquakes of note occurred in August, 1982, about 7 km northeast of Yakutat Bay near the mapped trace of the Fairweather fault. The earthquakes, with magnitudes of 4.6 m_b and 4.7 m_b, occurred 18 days apart but were located within a few kilometers of each other at depths of 8 and 9 km. Both events triggered a strong-motion instrument at Bancas Point, about 22 km away. A peak horizontal acceleration of 0.04 g was recorded for the first shock, while the second registered 0.10 g. These are the largest earthquakes known to have occurred along the section of the Fairweather fault zone northeast of Yakutat Bay since at least October 1974 when the regional seismic network was expanded to the current density of stations. Reliable focal mechanisms for the two earthquakes could not be determined from P-wave first motions, primarily because of uncertainties in the local seismic velocity structure. However, there is good evidence in the patterns of first motions that the two events had focal mechanisms significantly rotated with respect to each other, since distinct and opposite first motions were recorded at 10 out of 20 common stations. Considering the spatial proximity of the two events, this variation in the focal mechanisms suggests complexity in the fault system that generated the shocks. Similar evidence for fault complexity north of Yakutat Bay can also be found from pairs of earthquakes of about magnitude 4 m_b that occurred in April 1982 and in September 1981.

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Reviewed by John Kelleher and R. A. White.

SOUTHWESTERN ALASKA

(Figure 49 shows study areas discussed in this section.)

PRELIMINARY OBSERVATIONS ON FUMAROLE DISTRIBUTION AND ALTERATION, VALLEY OF 10,000 SMOKES, ALASKA

By Terry E. C. Keith

Fumaroles in the ash-flow deposit erupted at Novarupta on June 6-8, 1912, were responsible for the discovery and naming of the Valley of 10,000 Smokes (VTTS) in what is now Katmai National Park in southwestern Alaska (fig. 49). During a 60-hr period, about 15 km^3 of magma was erupted, producing approximately 20 km^3 of air-fall tephra and 11-15 km^3 of ash-flow tuff (Hildreth, 1983). The initial ejecta were 91-98 percent rhyolite (77 ± 0.6 percent SiO_2), but increasing amounts of dacite (66-64.5 percent SiO_2) and andesite (61.5-58.5 percent SiO_2) dominated the later half of the eruptive sequence (Hildreth, 1983). The exposed tuff in the lower part of VTTS, 10-20 km from the source, is a predominantly rhyolitic, nonwelded ash-flow deposit; in the upper VTTS, the exposed ash-flow deposit consists of a moderately welded, rhyolite-poor mixture of dacite and andesite. Air-fall units consisting of more than 98 percent dacite overlie the entire ash-flow deposit and become thicker and coarser up-valley toward the source. Fe-Ti-oxides analyzed by Hildreth (1983) yielded the following pre-eruptive magma temperatures: rhyolite (805°-850°C), dacite (855°-955°C), and andesite (955°-990°C). Temperatures decreased rapidly on mixing with air during eruption but remained hot enough for welding to occur in the upper VTTS.

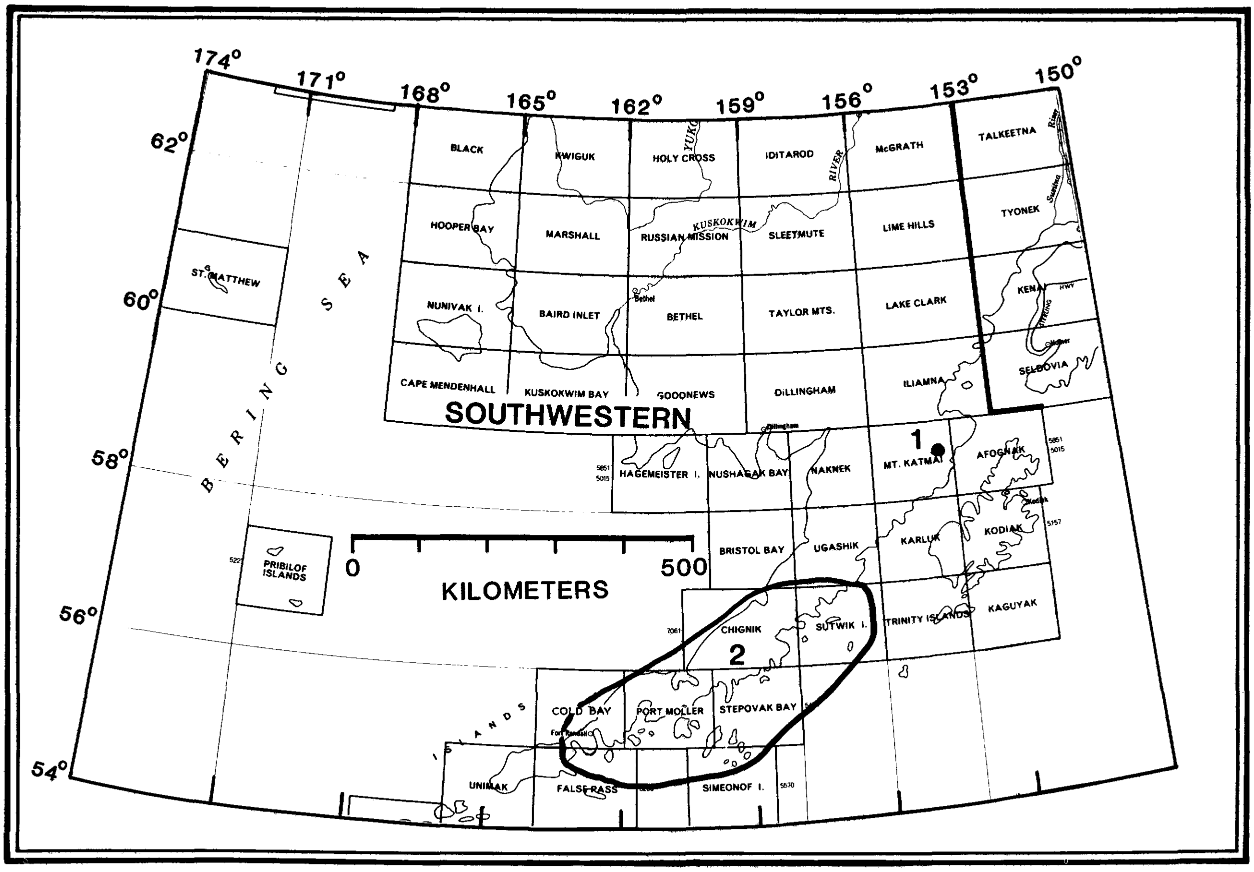

Figure 49.—Areas in southwestern Alaska discussed in this circular. Authors and inclusive pages of their articles are: (1) Keith, p. 82-85; (2) Nilsen, p. 85-88.

Fumaroles began to form immediately when the hot ash-flow tuff contacted water and ice of pre-existing rivers, glaciers, marshes and ponds that were sustained by continuing runoff from adjacent slopes onto the permeable ash-flow surface. Cooling of the ash-flow deposit took several decades. During the early expeditions (1916-1923), the highest fumarole temperature measured was 645°C in 1919 (Allen and Zies, 1923). Most of the fumaroles had died out by the 1930s, but a weakly steaming fumarole at the southwest end of Baked Mountain remained active in 1981, and several areas of warm, fumarolic seeps persisted around Novarupta in 1982.

Two principal types of fumaroles developed. Most were rooted in the ash-flow sheet itself and cooled as the ash-flow tuff cooled. Fumaroles near the Novarupta vent, however, are situated along subsidence fractures and remain warm because of residual heat from the Novarupta magma reservoir.

The configuration and distribution of fumaroles in the ash-flow sheet were related to the distribution of heat in the deposit, cooling fractures in welded parts of the tuff, water sources in contact with the hot deposit, and zones of permeability that controlled fluid flow within the tuff. Fumarolic deposits today are conspicuous above the buried pre-1912 river channels of Windy and Knife Creeks, River Lethe, and where major side streams drain onto the valley floor from bordering slopes. In the lower valley where the tuff is nonwelded and overlain by relatively thin air-fall layers (Hildreth, 1983), fumaroles developed by fluids that rose more or less vertically and formed single vents or circular groups of vents at the surface. Once a fumarole developed a channel to the surface, the fluid-ash reactions cemented the enclosing tuff, tending to reinforce the channel. In the upper valley, welded ash-flow tuff underlies a thin layer of nonwelded ash-flow and thick air-fall deposits. Cooling-contraction fractures in the welded tuff promoted lines of vertical fumaroles that locally connected with subhorizontal permeable zones in the superjacent deposits. Some fumaroles reached the surface only locally, but others formed long, linear fumarolic vents (Griggs, 1922).

Cementation by fumarolic deposits has helped to preserve the configuration and distribution of the fumaroles as usually colorful and long, resistant rubbly outcrops or isolated mounds throughout the valley. Detailed observations on the geochemistry of fumaroles active in 1919 are given in Allen and

Table 18.—Trace element chemistry and minerals identified in two zoned fumarolic deposits; samples A-E, A-H listed in order from vent outward; AFH, aluminum fluoride hydrate; —, not detected. Semiquantitative spectrographic analyses by C. Heropoulos; F, Cl, and total S by S. Neil, G. Mason, J. Graves, and E. Engleman

	79 KAT 15					79 KAT 34							
	A	B	C	D	E	A	B	C	D	E	F	G	H
	AFH Hematite	Opal Gypsum AFH	Opal AFH Hematite	Plagio-clase Magnetite	α-Cristo-balite	AFH Hema-tite	Fluorite Kaolinite Hematite	AFH Fluorite	Smectite Kaolinite AFH	Smec-tite AFH Kaolinite	α-Cristo-balite Hematite	Hematite Magnetite Smectite	Alunite Kaolinite Smectite Hematite
	parts per million												
As	1,000	3,000	2,000	30	100	200	100	300	700	70	50	500	150
B	700	1,500	700	15	7	150	5	150	7	50	30	20	70
Ba	150	70	100	500	10	50	150	100	300	200	500	300	300
Bi	5	3	10	3	---	1	5	5	20	10	20	10	---
Co	2	---	1	7	---	---	30	3	10	20	20	20	7
Cr	15	10	20	10	1.5	5	15	7	10	10	15	15	20
Cu	7	15	100	15	3	5	20	7	20	15	70	15	7
Mo	10	50	20	---	---	---	---	---	---	---	10	---	---
Pb	50	300	200	20	7	5	50	10	7	10	100	20	10
Sb	30	50	30	---	---	2	10	5	20	20	5	2	10
Sn	500	1,000	1,000	5	3	10	50	15	5	7	10	15	100
Sr	70	70	30	300	2	30	200	30	300	300	500	500	300
Zn	70	70	150	150	1.5	1	20	10	10	20	200	20	15
Ga	10	7	30	20	1	10	10	7	10	10	30	20	30
Tl	20	20	50	1	---	10	1.5	50	1	1	10	10	---
Hg	---	---	---	---	3	---	---	1	1	---	1	---	---
Se	---	---	---	---	---	---	---	20	50	---	---	---	---
	percent												
Cl	1.82	1.99	1.44	0.17	0.38	1.42	0.045	1.500	0.355	0.66	0.002	0.065	0.236
F	9.35	9.73	10.70	0.07	0.47	38.5	9.2	14.5	1.7	3.1	0.07	5.4	1.6
S_{total}	0.49	0.95	0.46	0.38	0.09	0.5	0.158	0.229	0.200	0.175	0.055	0.071	1.0

Zies (1923), but few observations were made as the fumaroles subsequently waned and died out. Studies of fumarolic activity in the near-surface vent region of active volcanoes in the Cascade Range of the northwestern United States show that many of the initial deposits are noncrystalline and unstable (Keith and others, 1981). Observations by Allen and Zies (1923) indicate that this was also the case at the VTTS. As the initial fumarolic deposits cooled, reactions took place between changing fluids and the enclosing ash deposits. Water-soluble and unstable components were removed by meteoric waters, leaving an assemblage of insoluble, resistant minerals.

Many elements and compounds that were concentrated and concentrically zoned around fumarolic orifices by early activity maintain their zonal distribution. Distinctive black or deep-red deposits in fumarolic vents are hematite or magnetite; brilliant brick-red to orange and yellow colors arranged radially outward from the orifices can be related to diminishing amounts of iron hydroxide staining the ash. High concentrations of As occur in yellow-orange outer zones of some fumaroles. Mn is usually abundant in very black earthy deposits. White deposits at the surface are commonly opal, α-cristobalite, aluminum fluoride hydroxy hydrate (AFH), and gypsum. AFH is a common material in the VTTS fumarolic deposits and is easily mistaken for silica. AFH appears to be a late deposit, probably formed by reaction of hydrofluoric acid with pumice during waning fumarolic activity. Fluorite is also common as a late-formed mineral.

Selected trace-element compositions and mineralogy for two zoned fumaroles are given in table 18. In each fumarole, sample A is from the vent wall and successive samples were progressively farther from the vent. The samples consist of fumarolic encrustations with a small amount of altered ash contamination. 79 KAT 34 is a fumarole cross-section exposed in the upper valley in a gully between Baked and Broken Mountains. Samples were from altered dacitic air-fall deposits about 3 m above the welded dacitic and andesitic ash-flow tuff; the samples spanned a distance of about 2 m, from the vent outward to apparently unaltered ash. 79 KAT 15 is from a fumarole near the south side of the lower valley that vented through rhyolitic ash-flow tuff and several centimeters of overlying dacitic air-fall deposits. Samples were collected near the surface in the resistant fumarolically altered and cemented air-fall deposits over a lateral distance of about 1 m. Concentrations of trace elements reflect the concentric zoning around fumarole orifices. Particularly noteworthy are concentrations of As, Bi, Sn, Tl, Sb, and Se (table 18). Trace amounts of Ag and Au were locally detected.

Along subsidence fractures in the thick dacitic tephra near the Novarupta lava dome are a few water-dominated fumaroles, which in 1982 had a maximum temperature of 82°C. The tephra has been altered to red, white, and gray mud consisting of smectite, kaolinite, alunite, and amorphous FeS_2. Hg contents of 15-20 ppm and Sb up to 15 ppm were locally detected in the warm, altered mud. Opal, native sulfur, goethite, and amorphous $Fe(OH)_3$ occur locally in cold outer zones around these active fumaroles.

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Reviewed by E. W. Hildreth and M. H. Beeson.

MIOCENE BACK-ARC TIDAL DEPOSITS OF THE BEAR LAKE FORMATION, ALASKA PENINSULA

By Tor H. Nilsen

The Miocene Bear Lake Formation, which crops out on the Alaskan Peninsula between Port Heiden and Pavlof Bay (fig. 50), is a thick sequence of dominantly tidal deposits on the inner margin of the Aleutian arc. Reconnaissance examination of four sections previously measured by Lyle and others (1979) permitted preliminary determination of its depositional and tectonic framework (fig. 50).

Burk (1965) mapped the Bear Lake Formation, measured a few sections, and designated the presumed lower part of it, which is best exposed in southern outcrops, the Unga Conglomerate Member. Wisehart (1971), from a study of the stratigraphy, sedimentology, and petrography of the Bear Lake Formation, concluded that it was deposited mainly on tidal flats and in tidal channels behind offshore bars; he inferred that some fluvial deposits were preserved to the south, chiefly on Unga Island. Galloway (1974) discussed diagenetic aspects, and Lyle and others (1979) petroleum-related aspects of the Bear Lake Formation.

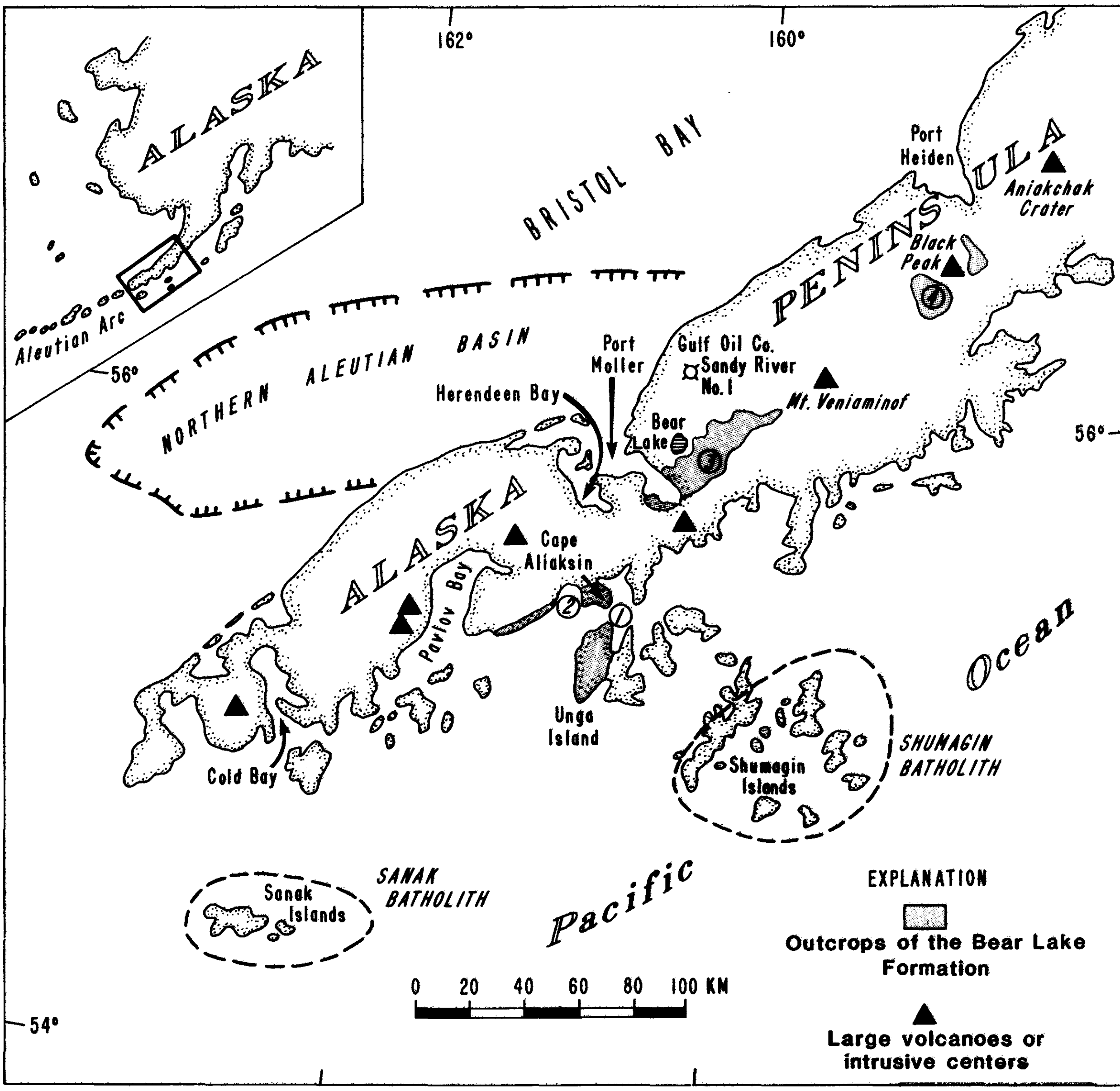

Figure 50.—Index map of southwestern Alaska Peninsula showing outcrops of the **Bear Lake Formation** and key geographic areas. Geology modified from Burk (1965) and Detterman and others (1981). Circled numbers refer to stratigraphic sections measured by Lyle and others (1979) of the Bear Lake Formation: 1, White Bluff, northern Unga Island; 2, northwest side of Cape Aliaksin; 3, type area south of Bear Lake; and 4, Cathedral Creek, south of Black Peak.

Detterman and others (1981) mapped the Bear Lake Formation in the Chignik and Sutwik Islands 1:250,000 quadrangles.

The Bear Lake Formation may be as thick as 1,600 m in outcrop (Burk, 1965; Detterman and others, 1981) and is 2,368 m thick in the Gulf Sandy River Federal No. 1 well (J. G. Bolm, Minerals Management Service, written commun., January 1983). The formation is probably present in subsurface to the northwest beneath Bristol Bay, where the North Aleutian Basin of Chase and others (1980) extends for almost 200 km north and west from Port Heiden. It is middle and late Miocene in age on the basis of megafossil assemblages (Louie Marincovich, Jr., U.S. Geological Survey, oral commun., January 1983). The Bear Lake Formation rests disconformably to the southwest on volcanogenic marine and nonmarine Oligocene mudstone and siltstone of the Stepovak Formation of Burk (1965) and to the northeast on

Eocene and Oligocene volcanogenic rocks of the Meshik Formation. It is overlain with angular unconformity by Pliocene volcanic rocks of the Milky River Formation (Detterman and others, 1981).

The Bear Lake Formation is well exposed in several places, particularly in the Black Peak and Bear Lake areas, where as much as 1,000 m of gently dipping sandstone, conglomerate, and mudstone crop out on very steep slopes. In contrast to most Tertiary units on the Alaska Peninsula, the Bear Lake Formation contains few, if any, interbedded volcanic rocks, and its sandstone contains more quartz, locally as much as 65 percent of the framework grains (Lyle and others, 1979). Conglomerate clasts are mostly volcanic rocks, but rounded clasts of granitic rocks as large as 25 cm in longest dimension were observed.

Coarse- and fine-grained tidal deposits dominate most of the stratigraphic sections of the Bear Lake Formation. Megafossils are locally abundant. Trace fossils are common in mudstone and siltstone and rare in sandstone. The largest conglomerate clasts, which are located to the southwest, southeast, and northeast, were probably deposited close to volcanic highlands.

Fluvial deposits were observed only at the base of the section on northern Unga Island. However, Burk (1965) and Wisehart (1971) concluded that deposits on central and southern Unga Island, which I did not see, were also of fluvial origin. They also considered the sequence on and adjacent to Unga Island to be the oldest part of the Bear Lake Formation, possibly as old as Oligocene.

The presence of debris-flow deposits of probable nonmarine origin in the Black Peak, Bear Lake, and Cape Aliaksin areas suggests that land areas with some relief were near, probably adjacent to areas of tidal sedimentation. The tops of the coarse-grained, reverse-graded, matrix-supported debris-flow units are locally reworked and contain fossil oysters and other mollusks.

The Bear Lake Formation generally consists of alternating 5- to 25-m-thick intervals dominated by sandstone and shale. The sandstone-dominated intervals are characterized by large-scale trough and tangential-tabular cross-strata, herringbone cross-strata, shale drapes on cross-strata, reactivation surfaces, abundant channeling, reversely oriented conglomerate-clast imbrication, reversely dipping cross-strata, superposition of small-scale cross-strata or current ripple markings on large-scale cross-strata with reversal of flow directions, scattered intact megafossils, local coquinas, and local burrowing that includes Ophiomorpha. The shale-dominated intervals are characterized by flaser bedding, current and oscillation ripple markings, starved ripple markings, abundant small-scale bioturbation, load casts, abundant mica and plant fragments, and synsedimentary slumps.

In most sections of the Bear Lake Formation, the sediments become finer grained upward. The uppermost parts of the sections appear to contain deposits that are more characteristic of open-marine shelf sedimentation. Fluvial deposits are present at the base of some sections, and land-derived debris-flow deposits are common in the lower half of the sections near Bear Lake and Black Peak. These relations suggest deposition in association with a marine transgression; however, Vail and others (1977) indicate worldwide eustatic lowering of sea level in the middle and late Miocene.

Paleocurrents from coarse-grained tidal channel deposits suggest sediment transport dominantly toward the southwest and north in the Black Peak area and northwest in the Bear Lake area. In the Cape Aliaksin area, medium and large-scale cross-strata in coarse conglomerate yield directions of sediment transport toward the south, and crude imbrication of clasts in an interbedded debris-flow deposit suggests sediment transport toward the north. Fluvial deposits on northern Unga Island were transported to the northwest.

The depositional setting for the Miocene Bear Lake Formation probably partly resembled the modern setting of the Herendeen Bay-Port Moller and the Port Heiden areas (fig. 50). Deposition in these bays is dominated by tidal currents, which move sands in and out of shallow, areally complex tidal channels and flats. The modern bays are bounded on the northwest by narrow inlets and prominent spits and bars that separate them from the strongly tide-influenced open-shelf sedimentation of Bristol Bay. The semi-enclosed southeastern margin of the Miocene Bear Lake basin was clearly much broader and larger than that of the modern bays. The Bear Lake basin extended offshore to the northwest, and tidal energy in it may have been stronger.

The Bear Lake basin was bordered by volcanic uplands of the Aleutian arc to the southeast and thus developed in a back-arc setting (fig. 51).

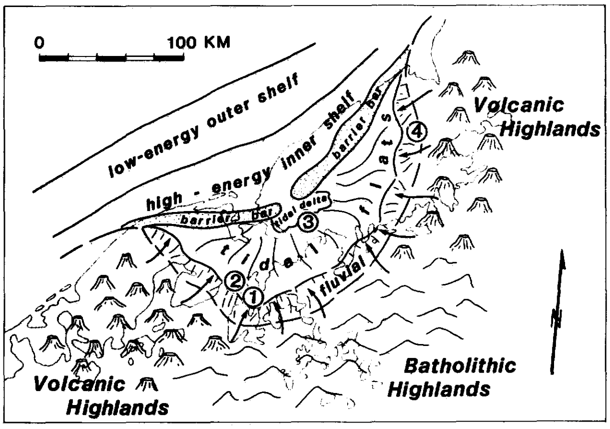

Figure 51.—Paleogeographic map for the Bear Lake Formation. Numbers show locations of measured sections of Lyle and others (1979).

Clasts probably derived from the Shumagin batholith and abundant fragments of Jurassic and Cretaceous sedimentary rocks in the Bear Lake Formation suggest that these units were exposed in the source areas. The relatively high quartz content of sandstone from the Bear Lake Formation probably results from a hiatus in volcanic activity during the Miocene, derivation of quartz from the upper Mesozoic quartz-rich sedimentary rocks and exposed batholiths, and high degree of physical abrasion occurring in the tidal environments, resulting in attrition of less-resistant grains. The fining-upward transgressive nature of the Bear Lake Formation sections may have resulted from subsidence of the Aleutian arc during an interval of relative volcanic quiesence.

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Reviewed by R. L. Detterman, H. E. Clifton, Louie Marincovich, Jr., and J. G. Bolm.

SOUTHEASTERN ALASKA

(Areas described in this section are shown in figure 52.)

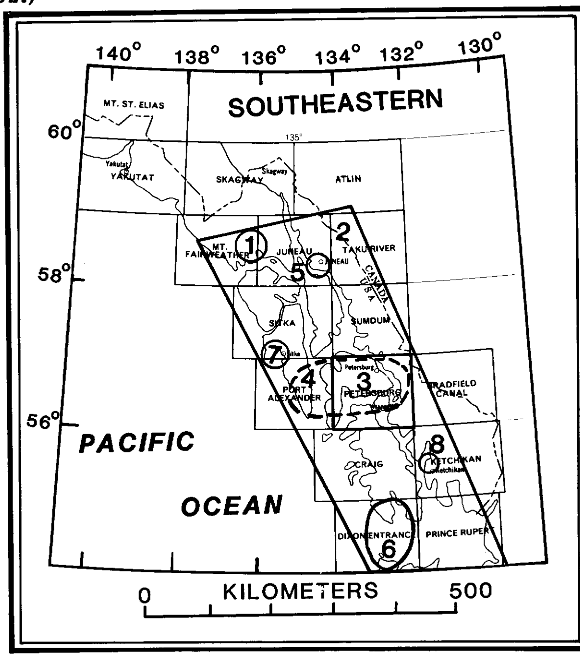

Figure 52.—Areas in southeastern Alaska discussed in this section. Authors and inclusive pages of their articles are: (1) Barnes, p. 88-90; (2) Brew and Ford, p. 90-93; Karl and Brew, p. 108-111; (3) Burrell, p. 93-96; Webster, p. 116-118; (4) Filipek, p. 96-100; (5) Gehrels, Brew, and Saleeby, p. 100-102; Himmelberg, Ford, and Brew, p. 105-108; (6) Gehrels, Saleeby, and Berg, p. 102-105; (7) Riehle and Brew, p. 111-116; and (8) Zen and Hammarstrom, p. 118-123.

NO MEASURABLE GRAVITY CHANGE AT GLACIER BAY REGIONAL UPLIFT AREA

By David F. Barnes

The regional uplift of southeastern Alaska near Glacier Bay is one of North America's better documented examples of aseismic vertical crustal movement. The magnitude of the movement (almost 4 cm/yr in some places) should be large enough to cause a detectable gravity change during the 10- to 15-yr record of repeated gravity measurements. However, such changes are still too small to measure. Most gravity data were collected during regional reconnaissance surveys, which lack the precision desired for studies of temporal gravity changes. Therefore, a longer measurement interval and/or improved measurement procedures and transportation may be needed to verify any preliminary conclusion. This lack of a detectable gravity change could suggest that tectonic processes (Jachens, 1978; Barnes and Strange, 1979) may better explain the uplift than isostatic rebound, the process favored by Hicks and Shöfnos (1965) and Hudson and others (1980). The nature and limitations of the available data are summarized below.

The uplift was first defined by Hicks and Shöfnos (1965) after study of records from permanent and temporary tide gauges during a 50-yr period ending in about 1960. These measurements showed an uplift zone about 500 km long and 150 km wide centered at Glacier Bay and uplift rates as high as 4 cm/yr at Bartlett Cove. In 1979 and 1980, Hudson and others (1980) revisited many of these tide gauge stations. By releveling between the tidal bench marks and high-tide level, they showed that the uplift was continuing and that the location of most rapid emergence had shifted 60 km northward to the upper part of Glacier Bay, where the rate between 1960 and 1980 exceeded 3 cm/yr. The approximate coincidence of the zones of most rapid uplift and most rapid deglaciation suggests isostatic rebound as a cause of the uplift. Hudson and others (1980), however, argued that the extent of the uplift over previously unglaciated areas suggests, at least in part, a tectonic origin that probably involves movement on the Queen Charlotte-Fairweather fault system.

The earliest repeatable gravity measurements in southeastern Alaska were made by Thiel and others (1958) in 1956, and significant amounts of additional data were obtained by the Coast and Geodetic Survey (Rice, 1969) and Geological Survey during the 1960s. Thus, 10-25 years of useful data could be available for measurement of a gravity change. However, the older measurements may involve larger meter drifts than those obtained in 1969 and later. An uplift of 4 cm/yr should cause a Bouguer gravity change of nearly 0.08 mGal in a 10-yr period, and the free-air change would be even larger. The problem of detecting a gravity change associated with the uplift can be best considered in two parts. The first is the gravity change at Juneau, which is the principal transportation and operation center for gravity surveys in northern southeastern Alaska. The second part is the possible gravity change between Juneau and Glacier Bay or other places where tide data suggest the uplift is most rapid.

More than 45 years of tide gauge records from Juneau indicate a fairly continuous uplift there of 1.5 cm/yr, but the accompanying Bouguer change of 0.07 mGal has not been detected in the 25 years of gravity data, even though many gravimeter readings have been made at almost the same spot at Juneau airport since 1956. However, all these readings are relative, and the ties to stations distant from the uplift area are few and questionable. The primary commercial air routes serving Juneau are from Sitka, Yakutat, Whitehorse, Ketchikan, Seattle, Cordova, Anchorage, and Fairbanks. The first two are in the uplift area, and most of the others are in areas known to be affected by tectonic activity. The most stable site is probably Ketchikan, where a new airport has replaced the older field at Annette Island. Five ties from Juneau to Annette Island in the 1960s range from 240.35 to 240.49 mGal, which does not permit measuring a change smaller than 0.10 mGal. Furthermore, the new Ketchikan airport has not yet been adequately tied to Juneau or the old field at Annette Island.

The gravity change accompanying the uplift of Glacier Bay relative to Juneau is easier to study than the Juneau uplift. This change averages 2.0-2.5 cm/yr, or a relative Bouguer change of nearly 0.05 mGal in only a 10-yr period. In 1956 Thiel made one measurement at Gustavus (the Glacier Bay airport), but his measurement was reported to only a tenth of a milligal. His description of the station location may not permit precise reoccupation. In 1964, Rice's (1969) observer tried to reoccupy the station, as did I in 1972, but these measurements differ by more than 2 mGal from the original observation. However, the agreement between the two measurements is 0.02 mGal, which suggests that we occupied the same spot and that there is very little change in the 8-yr interval.

Rice also reported a 1964 tie using a float-plane flight between Juneau and Bartlett Cove on which the drift, and thus the possible uncertainty, was 0.08 mGal. This station was reoccupied several times between 1972 and 1976. These measurements have a standard deviation of 0.02 mGal and an observed gravity that differs from Rice's drift-corrected measurement by +0.01 mGal.

Geological Survey studies at Glacier Bay began in 1969 when a station at Gustavus airport was tied to Juneau and Haines with two meters and a discrepancy of 0.06 mGal. The Juneau-Gustavus difference was remeasured six times in 1978 and three times in 1980, with an uncertainty (standard deviation divided by square root of number of measurements) of 0.02 mGal. The means of the 1969 and 1978-80 measurements differ by less than 0.01 mGal and again suggest very little change at Gustavus. This station and its reference mark have also been tied to stations at Bartlett Cove with estimated uncertainties of about 0.02 mGal. New base stations in the National Monument were also established with multiple ties (usually 3 to 6) in 1972 and 1976, but small boat transportation contributed to estimated uncertainties of 0.03-0.06 mGal. Most of these stations could be reoccupied using float-planes to improve their precision and/or to measure possible changes. This network provides fair areal coverage, but the stations are still too new for measurement of possible gravity change.

Although the estimated uncertainty of the early measurements is largerthan 0.05 mGal, the available data suggest that the gravity in this uplifted region is changing very slowly and has probably been no more than about 0.00 ± 0.02 mGal in a 10-to 15-yr period. Most isostatic mechanisms involve an elastic flexure of the crust with concurrent movement of subcrustal material that has a probable density of about 3.35 g/cm^3. Uplift involving movements of such subcrustal materials would be accompanied by calculated changes of 0.2 mGal/m, or a gravity decrease of about 0.04 to 0.05 mGal for a 20-25 cm uplift in a 10-yr period. The measured changes are close to zero and could even suggest a gravity increase of about 0.01 mGal. The measured changes are thus about 0.05-0.06 mGal smaller than the theoretical changes related to isostatic uplift. This discrepancy is only slightly larger than the 0.02 0.03 mGal estimated uncertainty of 0.02-0.03 mGal in the data, but it does suggest that the isostatic model is not the best explanation for the uplift and that the gravity change should be carefully monitored for at least another 10 years.

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Reviewed by R. C. Jachens and D. A. Brew.

TECTONOSTRATIGRAPHIC TERRANES IN THE COAST PLUTONIC-METAMORPHIC COMPLEX, SOUTHEASTERN ALASKA

By David A. Brew and Arthur B. Ford

The Coast plutonic-metamorphic complex is a major geologic feature that extends the length of

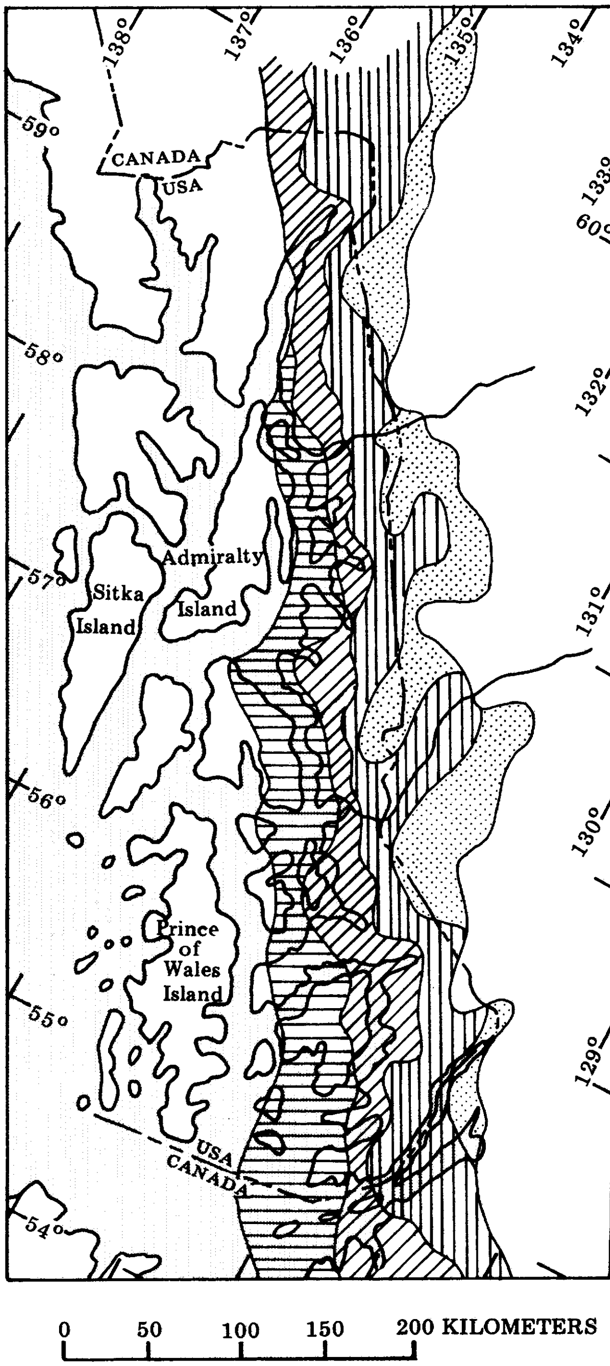

Figure 53.—Northern Coast plutonic-metamorphic complex, southeastern Alaska and northwestern British Columbia, showing zones as follows: horizontal lines - western metamorphic zone (WMZ), diagonal lines - central metamorphic zone (CMZ), vertical lines - central granitic zone (CGZ), stippling - eastern metamorphic zone (EMZ). After Brew and Ford (1984).

the Coast Mountains in southeastern Alaska (figs. 52 and 53) and British Columbia; it consists of a central belt of granitic and gneissic rocks and the adjoining related metamorphic rocks. It is subdivided (fig. 53) into the western metamorphic (WMZ), central metamorphic (CMZ), central granitic (CGZ), and eastern metamorphic (EMZ) zones (Brew and Ford, 1984). The deformation, metamorphism, and intrusion that formed the complex took place in middle Cretaceous through Eocene time (Gehrels and others, 1983; Brew, 1983).

In the first tectonostratigraphic terrane analysis of any part of Alaska, Berg and others (1972) included the westernmost part of the complex in their Gravina-Nutzotin belt and the adjacent Taku-Skolai terrane. This was later revised (Berg and others, 1978) so that the complex was covered by part of the Gravina-Nutzotin belt and by the Taku, Tracy Arm, and part of the Stikine terranes (fig. 54). They stated (on sheet 1) that the terranes were

> "...bounded by known and inferred faults. Each terrane is characterized by distinctive stratigraphic sequences that differ substantially from those of adjoining terranes. The depositional and structural histories recorded in each terrane are so different that large-scale tectonic juxtaposition seems to be demanded."

These same terranes were among those later characterized as "suspect" by Coney and others (1980, p. 329) "...because we cannot be certain of their paleogeographic setting with respect to North America through much of Phanerozoic time." Jones and others (1981, p. 1) provided a succinct definition of a tectonostratigraphic terrane

> "...a fault-bounded geologic entity, usually of regional extent, that is characterized by a distinctive stratigraphic sequence or rock assemblage that differs markedly from those of nearby, partly or entirely coeval neighbors."

Our recent studies in the complex have required that we evaluate relations of its granitic and metamorphic rock units against the definitions and maps presented in the above-mentioned reports.

The distribution of the zones of the Coast plutonic-metamorphic complex is shown in figure 53. The features of the zones, including dominant rock types, inferred nature and ages of non-intrusive protoliths, metamorphic facies series, nature and ages of intrusive components, and contacts between the zones are described by Brew and Ford (1984). The distribution of the tectonostratigraphic terranes of Berg and others (1978) in the complex is shown in figure 54. The original descriptions of the terranes and of their contacts are given by Berg and others (1978).

Comparison of figures 53 and 54 shows that the western contact of the WMZ roughly coincides with

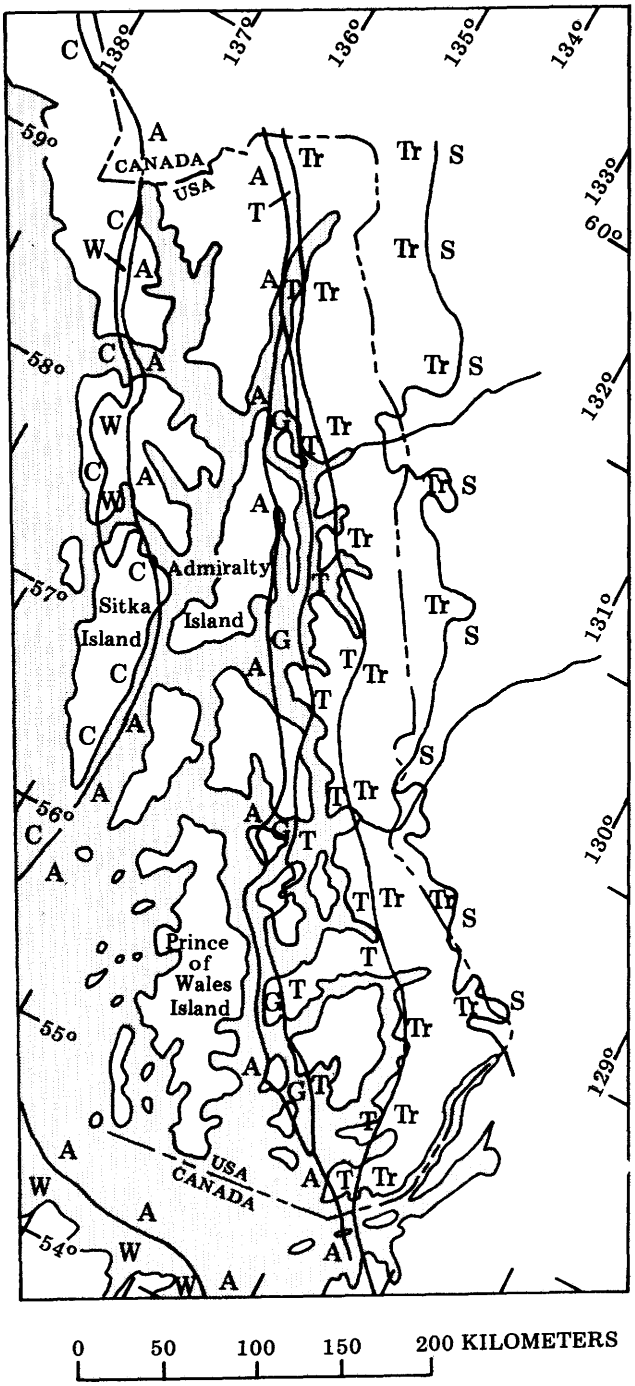

Figure 54.—Tectonostratigraphic terranes of southeastern Alaska. After Berg and others (1978) and Jones and others (1981) showing the terranes as follows: C - Chugach, W - Wrangellia, A - Alexander, G - Gravina-Nutzotin belt, T - Taku, Tr - Tracy Arm, and S - Stikine.

the western limit of the Gravina-Nutzotin belt (Berg and others, 1978). The Gravina-Nutzotin and Taku terranes together correspond almost exactly to the WMZ. The Tracy Arm terrane coincides closely (except for the uncertain eastern boundary) with the CMZ, CGZ, and EMZ. The close correspondence between the zones defined on the basis of metamorphic and intrusive features and the terranes suggests that the latter were also defined primarily on those features. The relations between the ages of metamorphism and intrusion and the ages of the protoliths are therefore critical

Brew and Ford (1984) and Gehrels and others (1983) established the ages of the metamorphic and intrusive features of the complex as ranging from middle Cretaceous into Miocene time, with the evidence for the older events preserved only in the WMZ. One current plate-tectonic model for the northeastern Pacific region (Monger and others, 1982) includes generation of the complex as a tectonic and metamorphic "welt" formed during accretion of the amalgamated Wrangellia and Alexander terranes against the earlier emplaced Stikine and other terranes in latest Cretaceous time, shortly before accretion of the Chugach terrane on the oceanward side of the Wrangellia and Alexander terranes. Other tectonic scenarios either avoid details concerning time of accretion or suggest either a pre-Late Jurassic age or a Late Cretaceous age (both in Berg and others, 1972). Thus, all the metamorphic and intrusive events that define both the zones of the complex and the Taku and Tracy Arm terranes of Berg and others (1978) are later than or concurrent with the proposed times of accretion and are not evidence of contrasting pre-accretionary stratigraphic, metamorphic, or structural histories. Given the available lithologic and age information, it appears that these terranes were defined mainly on relatively young metamorphic and intrusive features and not on the basis of reconstructed contrasting coeval lithostratigraphy of the protoliths.

Stratigraphic evidence from the premetamorphic and preintrusive rocks of the complex was examined to see if it supports differentiation of terranes (Brew and Ford, 1983). Ten recently mapped areas in southeastern Alaska, covering parts of the Alexander, Gravina-Nutzotin, Taku, Tracy Arm, and Stikine terranes, were studied. The results show that the Tracy Arm terrane could be produced by metamorphosing the lower part of the Alexander terrane and the Taku by metamorphosing the upper part of the Alexander and(or) Stikine terranes. The available fossil, structural, and paleomagnetic evidence provides no convincing arguments (Brew and Ford, 1983) to support differentiation of the terranes proposed by Berg and others (1978), with the exception of the Gravina-Nutzotin belt, which is clearly different. The Taku and part of the Stikine terranes appear to be the metamorphic equivalents of the late Paleozoic and younger parts of the Alexander terrane, the Tracy Arm terrane is the equivalent of the lower and middle Paleozoic part of the Alexander terrane, and the Alexander and Stikine terranes are one and the same.

The detailed and reconnaissance geologic mapping now available (Brew and Grybeck, 1984; Ford and Brew, 1973, 1977; Brew and Ford, 1977; Brew, Ovenshine, Karl, and Hunt, 1984; Berg and others, 1978; Berg and others, 1976; Berg and others, 1977; Elliott and Koch, unpublished map of the Bradfield Canal quadrangle; Smith, 1977) shows that only a few of the contacts of the Taku and Tracy Arm terranes proposed by Berg and others (1978) are recognizable faults and that none of those are inferred to have large displacements.

In sum: (1) the Taku and Tracy Arm suspect terranes differ mainly in features formed in post-accretionary events but probably had different protoliths; (2) those protoliths are recognizable in the nearby Alexander and Stikine terranes; (3) the Taku and Tracy Arm terranes do not meet the terrane criteria and thus neither exist nor have the tectonic significance ascribed to them; and (4) the complex is a linear deformational, metamorphic, and intrusive feature that developed in a single large terrane that consists of what has been called the Alexander and Stikine terranes.

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Reviewed by W. J. Nokleberg, Béla Csejtey, Jr., and W. H. Nelson.

LATE CRETACEOUS PLUTONIC ROCKS, PETERSBURG QUADRANGLE, SOUTHEAST ALASKA

By Peter D. Burrell

Continuing field and petrologic studies have led to the identification of a mineralogically distinct suite of plutonic rocks in the Petersburg 1:250,000 quadrangle which intruded pelitic schists, semischists, and phyllites of the Gravina-Nutzotin Belt (Berg and others, 1972) in early Late Cretaceous time. These rocks are exposed on small islands near the mouth of the Stikine River and on Zarembo, Woronkofski, Etolin and Wrangell Islands (figs. 52, 55). They are correlated with previously studied plutons on Mitkof Island and Lindenberg Peninsula (Burrell 1984). All are part of the Admiralty-Revillagigedo plutnic belt of Brew and Morrell, 1980.

Calculated modal compositions fall mainly in the granodiorite and tonalite range (fig. 56). Most granodiorites crop out on Woronkofski, Etolin, and part of central Wrangell Island to the west of the tonalites, which make up the Stikine River mouth islands and most of Wrangell Island. Modes of the Zarembo Island samples range from granodiorite to quartz diorite.

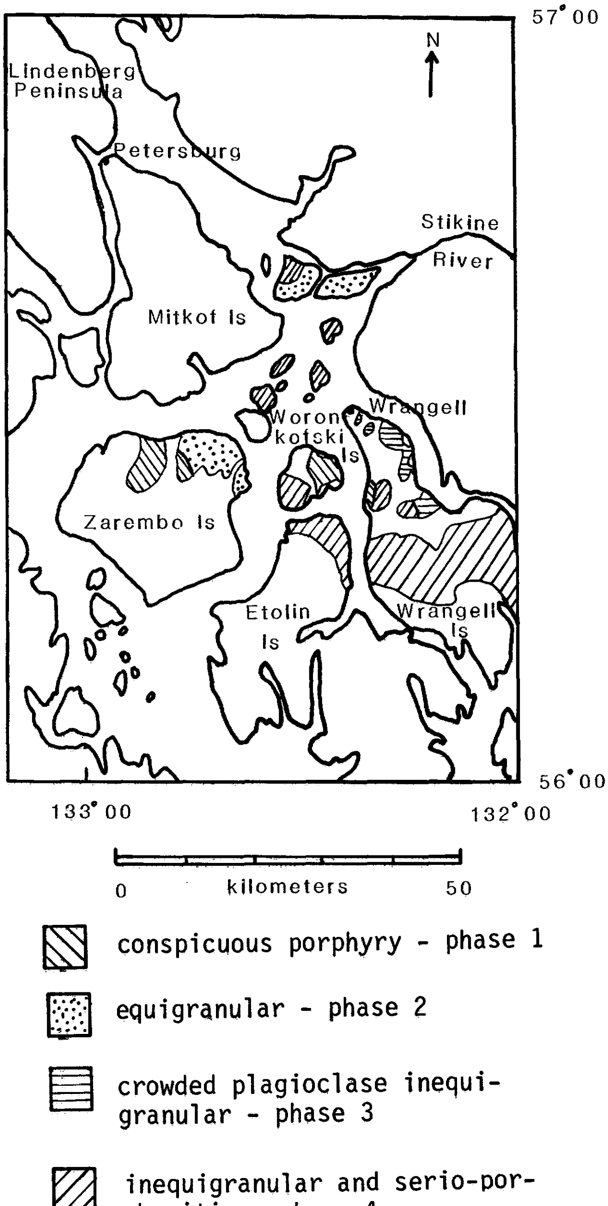

Figure 55.—Map of the eastern Petersburg quadrangle showing textural distribution of Late Cretaceous plutonic rocks discussed in this article.

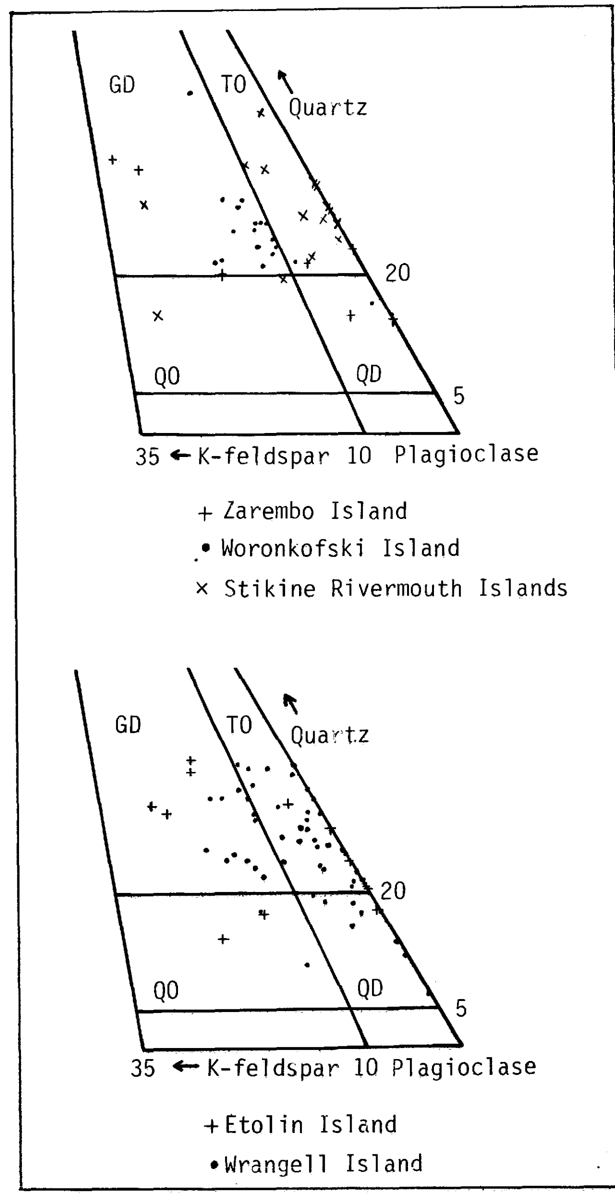

Figure 56.—Composition diagrams showing IUGS modes (Streckeisen, 1973) of Late Cretaceous plutonic rocks in the Petersburg quadrangle. QD, quartz diorite; QO, quartz monzonite; TO, tonalite; GD, granodiorite.

The textural classification is based on the nature and habit of plagioclase and mafic minerals Four textural phases exist, as follows (see also fig. 55):

(1) A conspicuous porphyry with subhedral to euhedral plagioclase phenocrysts (up to 12 mm) and with interstitial fine- to medium-grained mafic minerals (up to 2 mm). The mafic minerals are mostly discrete hornblende crystals, with minor primary epidote and biotite. Secondary epidote is common. The color index has two distinct ranges of 17-21 and 35-36.

(2) A fine- to medium-grained equigranular phase with nondescript subhedral plagioclase and mostly discrete mafic minerals. In some localities a seriate variation in plagioclase produces a sparsely porphyritic texture with phenocrysts up to 5 mm. Hornblende is the dominant mafic mineral. Biotite and primary epidote are also present. The color index is 21-26.

(3) A very fine- to medium-grained crowded plagioclase inequigranular phase having closely spaced euhedral plagioclase laths. Biotite,

clinozoisite, garnet, muscovite, and quartz are interstial to plagioclase and occur in small clots. The color index is 20-29. This phase occurs only in the northern Wrangell Island bodies.

4) A foliated seriate plagioclase inequigranular phase with an inhomogeneous textural aspect due to variable grain size and mafic mineral content. Subhedral and euhedral plagioclase (up to 5 mm) occurs with quartz, mafic clots, and sparse potassium feldspar. Other minerals include biotite, hornblende, clinozoisite, local pyroxene, and local garnet. The grain sizes are fine grained (less than 2 mm) and fine to medium grained (1-5 mm). The color index ranges from 26-41. This inequigranular phase grades to a serio-porphyritic phase with the addition of 10-40 percent plagioclase phenocrysts (up to 9 mm).

Mineralogically, this suite of plutons is characterized by oscillatory zoned plagioclase with inclusions of epidote or clinozoisite and with minor sericite alteration. Plagioclase less commonly has inclusions of garnet. Quartz grains are interstitial. Potassium feldspar, where present, is interstitial to all other minerals and occurs in either an even distribution or as sporadic clots. In several thin sections muscovite occurs as large flakes intergrown with biotite, suggesting it may be a magmatic mineral. Sphene, apatite, and allanite are accessory minerals. Mafic minerals include hornblende (some poikilitic), biotite, and local pyroxene.

Epidote, clinozoisite, and garnet are present as primary igneous minerals. Epidote and clinozoisite are mutually exclusive. One or the other is found as euhedral to subhedral crystals, commonly as inclusions in plagioclase and biotite. These epidote group minerals are rarely twinned and zoned. Epidote also occurs as a secondary mineral. Garnet ranges from euhedral to anhedral; some have inclusions of mafic minerals

Recent potassium-argon age determinations from a tonalite (textural phase 2) on the east coast of Zarembo Island gives biotite and hornblende ages of 90.4 m.y. and 93.0 m.y., respectively. A foliated inequigranular tonalite (textural phase 4) from the west coast of central Wrangell Island gives biotite and hornblende ages of 83.8 m.y. and 91.6 m.y. (Marvin Lanphere, U.S. Geological Survey, written commun., 1982). The hornblende ages are preferred. These results give the same general age as reported earlier for similar rocks on Mitkof Island (Burrell, 1983).

The Admiralty-Revillagigedo belt is a northwest trending belt of mineralogically unique Cretaceous plutonic rocks cropping out in the Petersburg quadrangle and the neighboring Sumdum (Brew and Grybeck, 1984), Bradfield Canal (R. D. Koch, U.S. Geological Survey, oral commun., 1982; Smith and others, 1977), Craig (Berg and others, 1976), and Ketchikan quadrangles (Berg and others, 1978; Smith and others, 1977; Smith and Diggles, 1981). The presence of epidote and garnet is unusual and may be critical to our interpretation of the emplacement history of these plutons. Zen and Hammarstrom (1983) suggest that garnet inclusions in plagioclase from a tonalite at Bushy Point, Ketchikan quadrangle, formed in the lower crust. They explain this occurrence by rapid ascent of a lower crustal rock, followed by decompression anatexis and crystallization of plagioclase around the garnet. Crystallization of primary epidote took place during subsequent magma evolution.

Plutons on Mitkof Island and Lindenberg Peninsula have compositions, textures, and mineralogies (Burrell, 1984) similar to those described in this report. The similar textural phases are the seriate plagioclase inequigranular phase (textural phase 4) and the conspicuous porphyry (textural phase 1). The characteristic epidote group minerals are present and show the same relations to plagioclase and the mafic minerals. This evidence, together with the field and available age data, correlates the Mitkof Island and Lindenberg Peninsula bodies with this suite of plutons, thereby defining the belt of Late Cretaceous plutonic rocks in the Petersburg quadrangle.

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Reviewed by R. D. Koch and E-an Zen.

THE POSSIBLE SIGNIFICANCE OF DIAGENETIC REMOBILIZATION OF ORE ELEMENTS IN THE MUD-FLAT ESTUARINE ENVIRONMENT

By Lorraine H. Filipek

A study was begun in the Petersburg quadrangle in September 1981 to determine what physical and chemical changes in ore element species occur in the transition from a riverine (fresh-water) environment to an estuarine environment. The purposes of this study are: (1) to determine mechanisms that might be responsible for concentrating ore elements in estuarine environments, and (2) to determine whether analytical data from samples taken at the freshwater-seawater interface will identify anomalously mineralized areas.

Sediments and water samples were collected from streams and mud flats on Zarembo, Kupreanof, and Kuiu Islands (fig. 57) in a volcanogenic massive sulfide belt (Berg, 1980). On the mud flats, only a thin (about 1 mm) veneer of brown oxidized mud generally occurs over black reduced H_2S-rich mud. Separate samples were collected of the oxidized and reduced muds. Water samples were analyzed for conductivity, pH, and selected dissolved major and trace elements. (See tables 19 and 20.) Sediment samples were subjected to a six-step extraction procedure designed to isolate the following (operationally defined) geochemical phases: (1) humic-metal complexes; (2) exchangeable, carbonate, and acid-soluble sulfide phases; (3) hydrous Mn oxides; (4) resistant organic compounds, pyrite, and other acid-insoluble sulfides; (5) hydrous Fe oxides; and (6) "lithogenous" (residual crystalline) phases.

Preliminary results on both water and sediment samples are noteworthy. Dissolved ([0.45) riverine Fe (table 19) is at least an order of magnitude higher than would be expected for Fe^{3+} in equilibrium with atmospheric O_2, suggesting that most of the "dissolved" Fe is actually colloidal hydrous Fe oxides or Fe-humic complexes. Decreased "dissolved" Fe concentrations in estuarine samples give further evidence of the originally colloidal nature of the Fe because colloidal Fe oxide and humic complexes are known to flocculate upon estuarine mixing (Sholkovitz and others, 1978). Unlike Fe, the concentrations of the ore elements Cu, Zn, and As in the riverine water samples from the mineralized drainages are generally less than or equal to average global river concentrations of these elements and give little indication of mineralization. However, the concentrations of all three ore elements increase significantly in the estuarine stretches of the rivers (table 19). This distribution of concentrations contrasts with previously published results of metal behavior during estuarine mixing, which show that dissolved trace metals tend to decrease due to estuarine flocculation of the metals that exist in river water, in part as colloids in association with colloidal humic acids and hydrous Fe oxides (Sholkovitz, 1978). The increase in concentration suggests that an additional process is active for ore elements in the mud flats. Finally, dissolved Mn concentrations show no pattern.

Total concentrations of ore elements in the riverine and mud-flat sediments (table 20) do not differ appreciably from average global concentrations in sediments; therefore, total concentration values are not effective indicators of mineralization. Relative partitioning within the nonlithogenous phases (extractions 1-5) of Cu and Zn does not seem to vary much with environment. However, concentrations of all these phases tend to decrease substantially for Cu (and to a lesser extent for Zn) in the samples of subsurface reduced mud-flat sediments compared with the samples of riverine and surficial oxidized mud-flat sediments. These results suggest that diagenetic reactions in the reduced sediments release ore elements into the overlying oxidized sediments and water column. In addition, humic-Fe complexes and hydrous Fe oxides decrease in magnitude in the subsurface reduced sediments. In contrast, high concentrations of Mn oxides and other nonlithogenous forms of Mn are found only in the riverine samples, suggesting that the environments of even the surficial mud-flat sediments are too reducing for production of Mn oxides.

Previous geochemical studies in this area by John Cathrall (U.S Geological Survey, unpub. data) showed similar low concentrations of Cu and Zn in the stream sediments and indicated that analysis of heavy-mineral concentrates was necessary to delineate mineralization. Both Cathrall's work and the present results suggest that Cu and Zn undergo little chemical weathering before entering the estuarine environment and (or) that the abundant

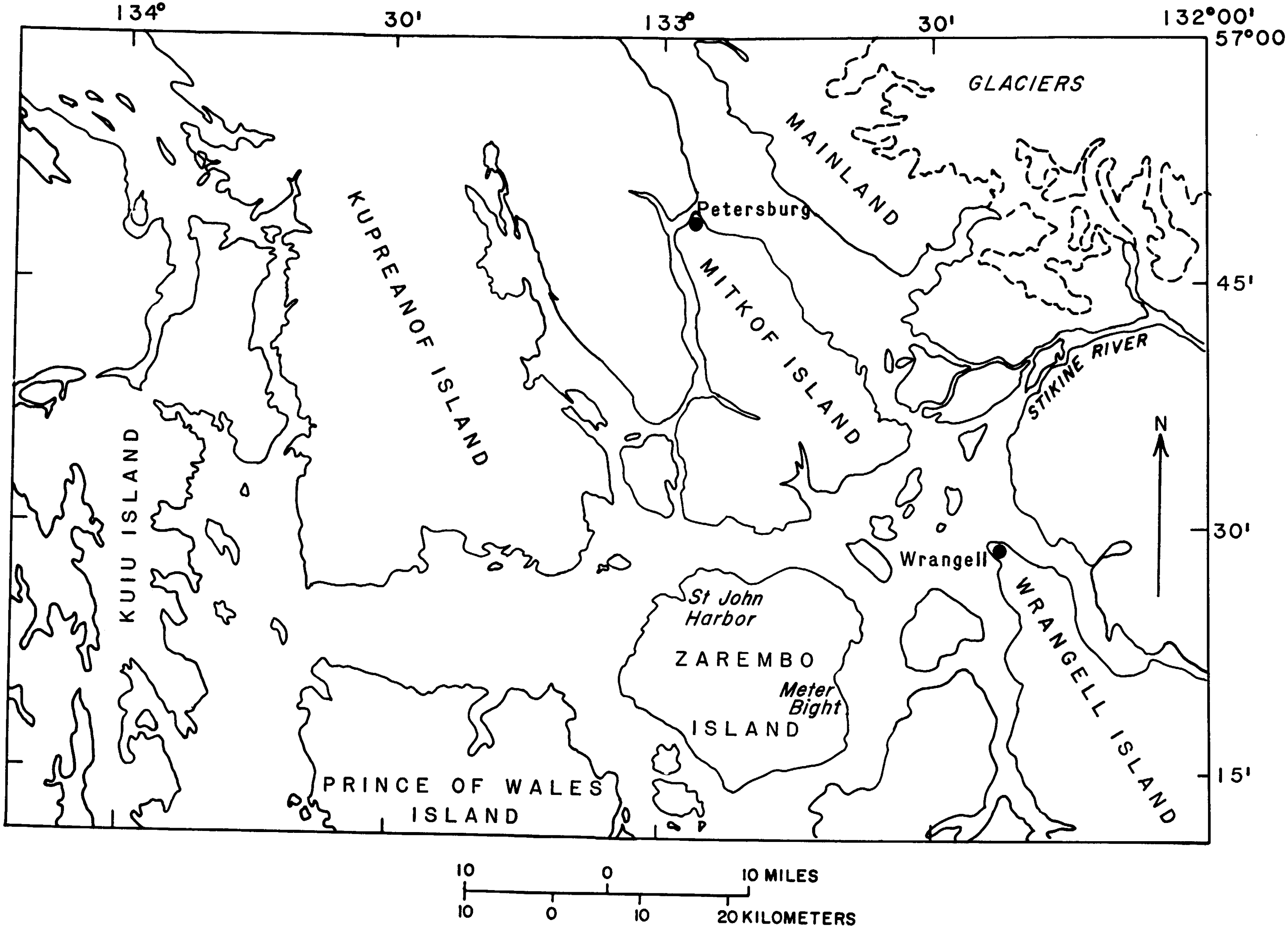

Figure 57.--Map of study area, including most of the Petersburg quadrangle and the easternmost part of the Port Alexander quadrangle.

Table 19.—Concentrations (in micrograms per liter) of trace elements in water samples from the Petersburg quadrangle

[All results are for waters filtered at 0.45 , except where noted for Cu concentrations. Kp, Kupreanof Island; Z, Zarembo Island; and K, Kuiu Island. Analyses performed by D. Preston]

	Conductance (μmho)	Fe	Mn	Cu 0.45	Cu 0.22	Zn	As
			Estuarine water				
Kp 5	11,000	100	40	15.6	--	23.2	12.0
Z 3	8,000	100	30	10.1	--	18.0	13.8
Kp 25	2,500	100	20	3.1	1.5	9.3	1.5
Z 50	2,300	200	25	2.6	--	15.2	<1
K 31	1,100	300	25	1.6	--	10.2	1.1
Average marine[1]		3	2	3		5	2.3
			Riverine water				
Z: 10	40	200	25	1.3	0.9	3.3	<1
39[2]	200	200	20	3.8	--	15.0	<1
41	15	200	15	3.2	--	19.0	<1
42	100	200	20	1.8	--	7.3	<1
43	70	300	30	6.2	--	36.0	<1
48	80	100	25	2.0	--	8.5	<1
K: 27	100	100	5	1.4	--	4.2	<1
28	260	300	25	1.2	--	3.0	<1
29	90	400	35	1.6	--	3.7	<1
30	175	500	35	1.7	--	4.0	<1
Kp: 6	20	300	5	1.5	--	9.3	1.2
7	17	200	20	1.4	0.8	5.1	1.1
12	90	50	20	1.2	--	7.3	<1
14	60	200	30	1.2	--	2.8	<1
26	65	200	25	1.8	--	3.2	<1
13[3]	12	600	30	1.7	0.8	12.9	<1

[1]Riley and Chester (1971)

[2]Immediately downstream from Cu-Zn prospect

[3]Swamp

rains of the area dilute and remove weathering products from the riverine environment. Once the stream sediments reach the estuary, however, the ore elements appear to undergo significant diagenetic remobilization, a process that releases dissolved Cu, Zn (and As?) into the overlying water column. If it can be shown that (1) mud-flat sediments are generally locally derived and (2) there are few other complicating processes, then sampling of water in estuarine stretches of drainages could offer a simple and effective means of enhancing anomaly contrasts in reconnaissance surveys in this type of environment. Studies to resolve these two issues are presently under way.

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Table 20.—Selected extraction results (in parts per million) for sediments from Zarembo Island

[For the mud-flat sediments: A, surficial oxidized; B, subsurface reduced. Σn-1, sum of extractions 1-5 (nonlithogenous). Total, sum of all six extractions. Analyses performed by L. Filipek and D. Ditmore]

	Fe				Mn		Cu		Zn	
	Total	Σn-1	Humic	Oxide	Total	Σn-1	Total	Σn-1	Total	Σn-1
Meterbight										
River:										
41	18,500	9,080	1,930	5,190	1,370	626	30.3	15.8	73	22.0
43	19,600	9,410	1,740	5,240	1,660	538	18.2	16.2	120	65.0
Mud flat:										
52A	17,000	8,480	1,270	4,840	763	128	13.3	12.8	68.2	25.2
52B	20,400	5,760	784	2,890	574	179	16.3	6.8	58.4	15.0
53A	18,600	7,660	634	4,600	1,620	121	36.7	19.2	80.0	21.0
51B	22,700	6,310	552	3,320	705	325	20.0	6.9	70.9	24.0
44A	17,000	7,280	456	4,760	1,290	88	23.4	14.4	66.0	21.0
44B	20,800	5,150	325	2,780	458	122	16.7	9.8	57.4	19.3
45A	12,900	8,210	1,080	5,130	989	514	30.0	25.0	99.7	46.2
45B	25,400	6,760	694	3,590	762	308	23.5	9.2	75.9	24.5
54A	17,700	7,990	596	4,950	1,160	228	33.5	28.0	88.1	33.6
54B	19,900	4,160	363	1,150	267	109	15.0	7.4	56.1	13.8
St. John Harbor										
River:										
9	19,200	9,320	1,630	5,130	1,610	899	16.1	15.6	145.1	72.6
11	22,000	13,100	2,510	7,200	2,100	1,430	30.5	27.0	119.1	67.6
10	18,800	9,480	1,880	5,040	1,050	292	21.9	16.4	87.2	35.2
Mud flat:										
2A	20,100	10,700	1,450	4,980	836	226	34.1	23.6	85.2	39.2
2B	36,600	11,800	658	5,380	960	316	25.8	17.8	97.6	45.6
1A	19,700	9,540	1,040	4,830	1,160	147	22.8	20.8	80.5	31.0
1B	28,000	8,950	237	3,990	757	206	19.6	13.4	75.2	39.1
3A	19,500	10,100	1,330	5,040	841	121	26.1	21.6	93.4	44.4

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PROGRESS REPORT ON U/PB (ZIRCON) GEOCHRONOLOGIC STUDIES IN THE COAST PLUTONIC-METAMORPHIC COMPLEX EAST OF JUNEAU, SOUTHEASTERN ALASKA

By George E. Gehrels, David A. Brew, and Jason B. Saleeby[1]

Geologic mapping (Ford and Brew, 1973, 1977; Brew and Ford, 1977) and U/Pb (zircon) geochronologic studies have delineated two major plutonic suites in the Coast plutonic-metamorphic complex near Juneau (fig. 52). This report describes the age relations and structural characteristics of the plutons that comprise the suites and discusses the structural evolution of the Coast plutonic-metamorphic complex during their emplacement.

The central part of the Coast plutonic-metamorphic complex east of Juneau consists of a western belt of elongate and highly foliated tonalitic Plutons, an eastern belt of large, non-deformed, granodiorite plutons, and an intervening belt of schist, gneiss, marble, plutons, and migmatite (fig. 58). Plutons in the western belt include the Mount Juneau, Carlson Creek, Lemon Creek Glacier, and Mendenhall bodies, which are bounded and generally separated by amphibolite-facies metamorphic rocks (Brew and Ford, 1981). These plutons are referred to as sills because they form thin but laterally continuous sheets that are generally concordant with the steeply dipping foliation in adjacent metamorphic rocks. Toward the northwest, the Mount Juneau pluton swings to the west across foliation and terminates in metamorphic rocks (Ford and Brew, 1981), and the Carlson Creek pluton interfingers with the Lemon Creek Glacier pluton which probably pinches out before reaching the southern end of the Mendenhall pluton (Ford and Brew, 1973). The Mendenhall pluton is a large sill-like body that continues more than 80 km to the northwest (Brew and Morrell, 1980). South of Taku Inlet, the Mount Juneau and Carlson Creek plutons are inferred to merge and form a single sill that extends over 120 km to the southeast (Brew and Ford, 1981).

The sill-like plutons of the western belt generally consist of medium- to coarse-grained biotite-hornblende tonalite and quartz diorite, with minor sphene and garnet. Alignment of the mafic minerals and of widespread mafic inclusions in these rocks defines a penetrative foliation and, locally, a steeply plunging lineation. These fabrics and also a gneissic layering are generally parallel to the margins of the bodies and to the foliation in the adjacent metamorphic rocks. An important exception to this parallelism occurs on the north side of the Mount Juneau pluton (fig. 58), where an internal gneissic layering parallels the foliation in the country rock, but the contact of the pluton cuts across regional metamorphic isograds and local foliation surfaces. These structural relations and the recognition of metamorphic mineral growths in the plutonic rocks suggest that the sills were emplaced during the latter phase of a regional metamorphic and deformational event in the Coast plutonic-metamorphic complex (Ford and Brew, 1981).

Two zircon size-fractions of a foliated but homogeneous phase of the Carlson Creek pluton have yielded a concordant age of 67 ± 2 m.y. A similar tonalite sill along the western side of the Coast plutonic-metamorphic complex in Thomas Bay, 160 km to the southeast, yielded a concordant age of 64 ± 2 m.y. based on three size-fractions. Thus, the ages of the tonalitic sills along the west side of the Coast plutonic-metamorphic complex near Juneau are probably in the 62-69 m.y. range.

Plutons along the eastern side of the Coast plutonic-metamorphic complex consist of large bodies of granodiorite that have not been significantly metamorphosed or deformed. In the Juneau area, the main pluton is the Turner Lake body, which is a medium- to coarse-grained sphene-hornblende-biotite granodiorite that crops out over an area of at least 2,600 km^2. South and north of the area shown in figure x2, the Turner Lake pluton cuts across amphibolite-facies meta- morphic rocks of the central belt, and apparently intrudes the tonalitic sills of the western belt. U/Pb analyses of three zircon size-fractions from the Turner Lake body have yielded data that give a lower-intercept age of 50 ± 2 m.y. Moderate discordance in the U/Pb systematics is presumably due to inheritance of older radiogenic Pb.

Between the two main plutonic suites is a complex belt of locally gneissic plutons, migmatite, and amphibolite-facies schist, paragneiss, and marble. The main pluton in this belt is the composite Annex Lakes-Flat Point body, which has characteristics intermediate between the sill-like Mount Juneau and Carlson Creek tonalite bodies and the non-deformed Turner Lake-type granodioritic plutons. The large outer zone of this body (the Annex Lakes pluton) has a gneissic foliation and a slightly elongate map pattern like that of the suite of tonalite sills. In this pluton, however, the foliation is thought to be a product of syn-emplacement processes rather than a superimposed deformational event. The core of the body (the Flat Point pluton) consists of homogeneous fine- to

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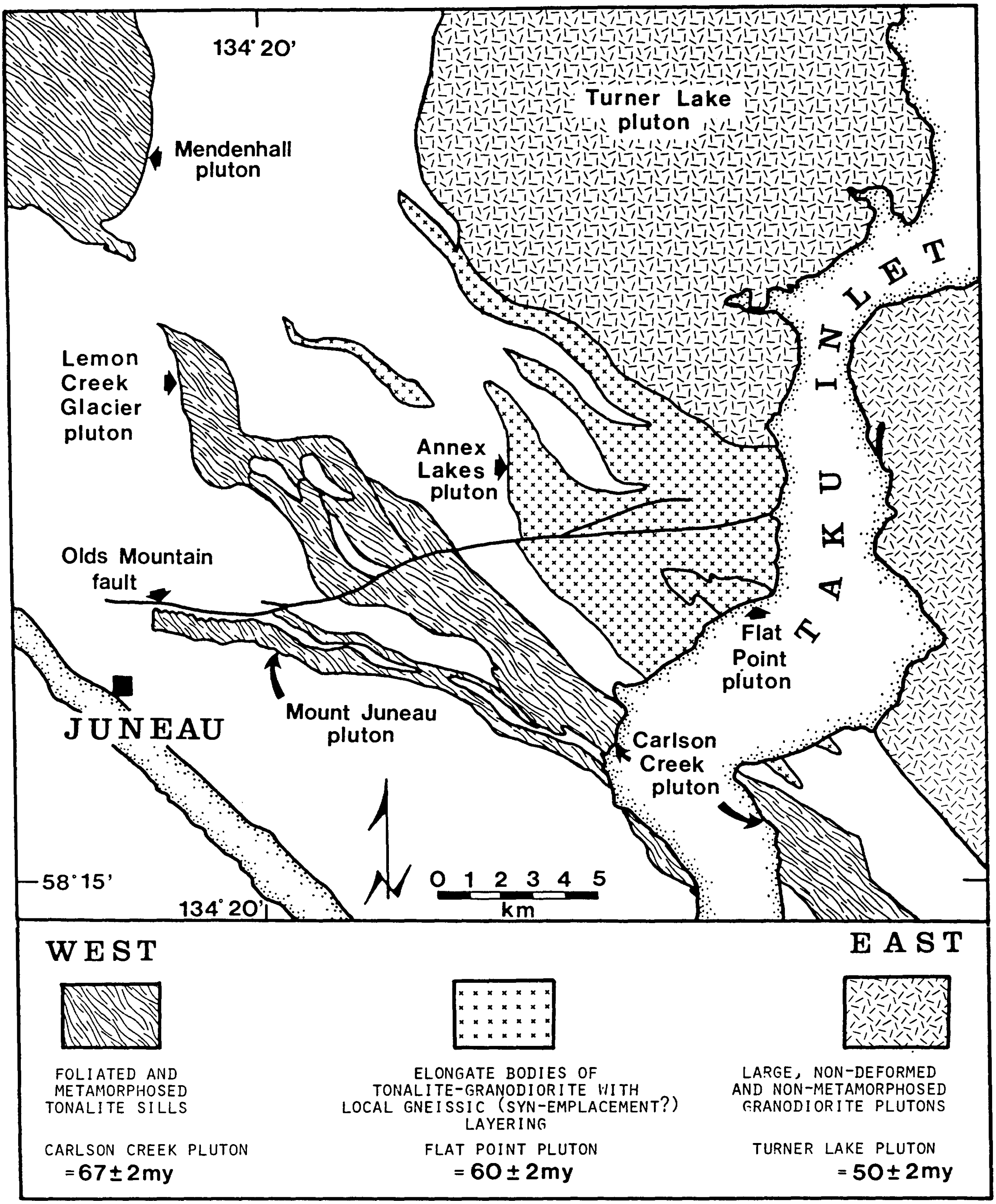

Figure 58.—Sketch map of the main plutonic suites in the Coast plutonic-metamorphic complex near Juneau, Alaska (from Ford and Brew, 1977).

medium-grained biotite-hornblende tonalite and granodiorite. The mineralogy and fabric in this core zone are generally similar to those in the Turner Lake body. Two zircon size-fractions from this core zone yielded a lower-intercept age of 60 ± 2 m.y. Discordance in this sample is also presumed to be due to inheritance of older radiogenic Pb.

Variations in age and emplacement style of the plutons east of Juneau (fig. 58) are consistent with the following interpretations: (1) the 62- to 69-m.y.-old tonalite bodies (Mount Juneau and Carlson Creek) along the western side of the Coast plutonic-metamorphic complex were emplaced during the latter phases of a major deformational and metamorphic event, (2) the intensity of this deformation and metamorphism decreased significantly prior to the emplacement of the 60 ± 2-m.y.-old pluton (Annex Lakes-Flat Point), (3) 50 ± 2 -m.y.-old granodiorite pluton(s) (Turner Lake) were emplaced after the deformational and metamorphic event had ceased, (4) the deformed and metamorphosed tonalite sills and their amphibolite-facies country rocks were probably uplifted prior to the emplacement of the undeformed and unmetamorphosed Turner Lake-type granodiorites, and (5) the thermal effects of the 50 ± 2-m.y.-old Turner Lake-type plutons did not produce a noticeable discordance in the U/Pb systematics of the older plutonic rocks.

Significant events in the Late Cretaceous and early Tertiary evolution of the Coast plutonic-metamorphic complex may have included: emplacement of the tonalite sills into a regime of active deformation and metamorphism during latest Cretaceous and early Paleocene time; uplift of at least the western part of the Coast plutonic-metamorphic complex and a decrease in the intensity of deformation and metamorphism during Paleocene time; and post-tectonic emplacement of the early and middle Eocene granodiorite plutons into a regime that was not undergoing regional metamorphism.

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Reviewed by M. A. Lanphere and R. A. Loney.

PROGRESS IN GEOLOGIC STUDIES OF SOUTHERN PRINCE OF WALES ISLAND, SOUTHERN ALEXANDER TERRANE, SOUTHEAST ALASKA

By George E. Gehrels, Jason B. Saleeby, and Henry C. Berg[1]

Geologic mapping during the summer of 1982 has shown that southern Prince of Wales Island (fig. 52) is underlain by two distinct lithologic packages which are separated by a major low-angle fault (fig. 59). Beneath and west of the fault are pre-Middle Ordovician metasedimentary and metavolcanic rocks of the Wales Group. Above and east of the fault is a complex of Ordovician and Silurian plutonic, volcanic, and marine sedimentary rocks, and a superjacent sequence of Devonian strata. Middle Cretaceous granitic plutons occur in both of these packages and locally cross-cut the fault that separates them (Turner and others, 1977; Redman, 1981). Northwest- to north-striking high-angle faults in the area have up to several kilometers of primarily left-lateral displacement.

Several structural and depositonal relationships within and between the units on southern Prince of Wales Island have significant implications for the geologic history and mineral resource potential of southern southeast Alaska:

(1) Several massive sulfide deposits have been recognized in Ordovician and Silurian volcanic and sedimentary rocks in the Nichols Bay region of southern Prince of Wales Island and in the Barrier Islands (fig. 59; Gehrels, Berg and Saleeby, 1983). The predominant sulfide mineral is pyrite, with minor occurrences of sphalerite and arsenopyrite. The occurrence of these sulfides as rinds around

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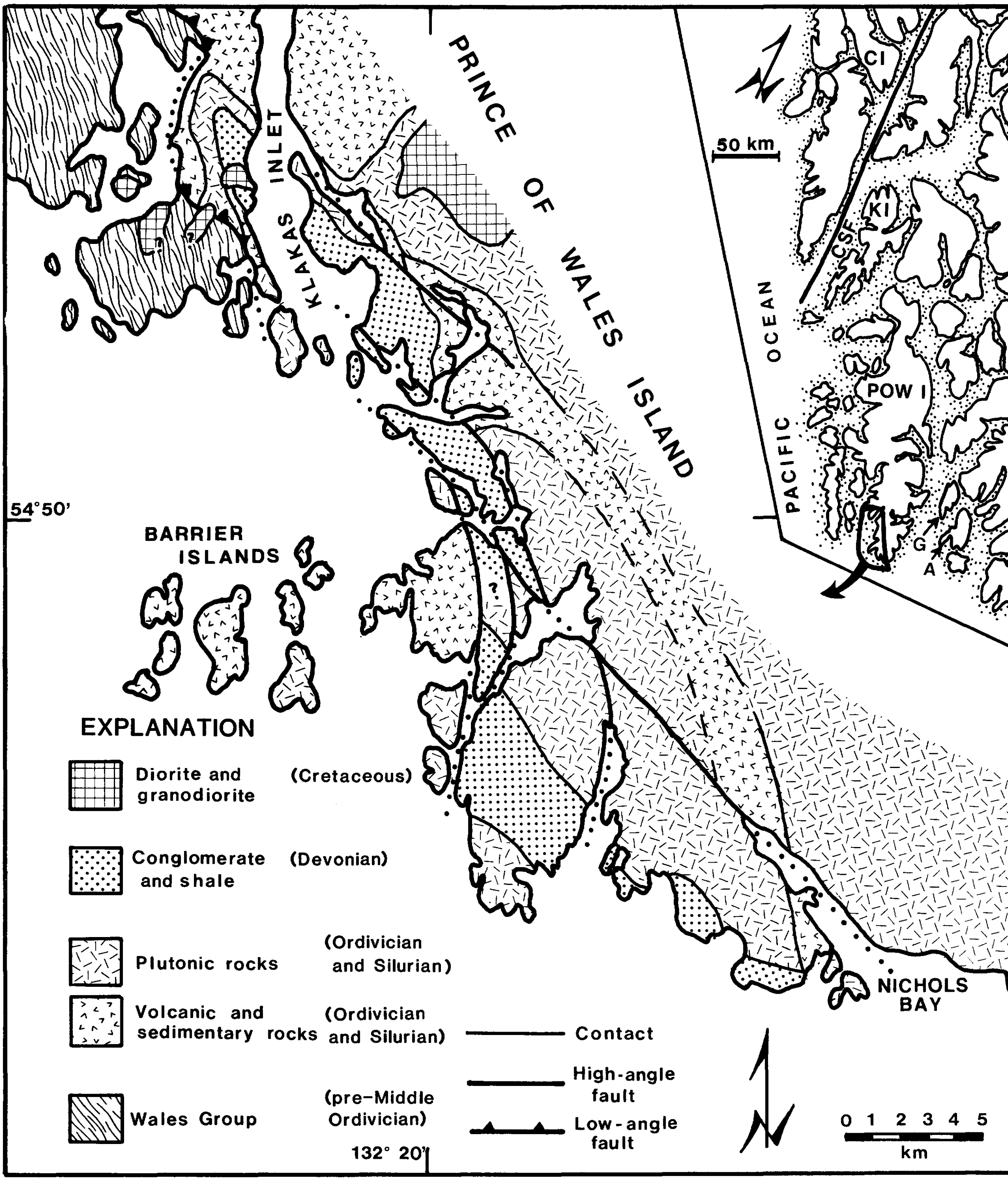

Figure 59.—Geologic sketch-map of southern Prince of Wales Island. On inset map: CI, Chichagof Island; KI, Kuiu Island; CSF, Chatham Strait fault; POWI, Prince of Wales Island; G, Gravina Island; A, Annette Island.

pillow-like forms in silicic-intermediate volcanic rocks and as stratiform layers in siliceous shale and graywacke interbedded with the volcanic rocks suggests that the mineral deposits are a product of volcanogenic exhalative processes. This type of deposit has not previously been reported in Ordovician and Silurian rocks on southern Prince of Wales Island.

(2) The relation between the pre-Middle Ordovician Wales Group and the Ordovician through Devonian rocks on central and southern Prince of Wales Island has been a controversial subject. North of the study area the boundary between the two packages is reported to be a low-angle fault that cuts Devonian strata and is cut by middle Cretaceous plutons (Turner and others, 1977; Herreid and others, 1978; Redman, 1981; and Eberlein and others, 1983). On the other hand, Eberlein and others (1983) report that, in the Klakas Inlet area (fig. 59), the Wales Group is depositionally overlain by Devonian strata and is intruded by Ordovician and Silurian plutons. Our mapping in this area has shown, however, that the Devonian strata are not in contact with the Wales Group and that the Wales Group and the Ordovician and Silurian plutonic rocks are separated by a low-angle fault. The oldest demonstrated relation between the two packages in this area is therefore the low-angle fault that separates them, which moved between Middle Devonian and middle Cretaceous time. Earlier movement could have occurred on this fault, but evidence for such movement has not been recognized.

(3) The Ordovician to Devonian rocks record a Late Silurian and Early Devonian deformational, uplift, and mountain-building event in the southernmost Alexander terrane of southeastern Alaska. We refer to this event as the Klakas orogeny because the various manifestations are best developed in the Klakas Inlet area (fig. 59)(Gehrels, Saleeby, and Berg, 1983). Deformational fabrics produced by this event include kilometer-size domains of brecciation and shearing in rocks of Ordovician through Middle and perhaps Late Silurian age. These deformed rocks are overlain by Early and early Middle Devonian conglomerates that are not appreciably deformed. The Ordovician and Silurian rocks must therefore have been deformed, uplifted, and exposed at the surface during Late Silurian or Early Devonian time. The existence of kilometer-scale topographic relief in the area is recorded in a Devonian basal conglomerate which is locally more than a kilometer thick and consists of talus breccia, conglomerate with rounded boulders up to a meter in diameter, and cross-bedded and locally channelled sandstone. The overlying Lower and lower Middle Devonian strata record the cessation of this event by a transition from high-energy fluvial to quiet-water marine environments of deposition.

The Klakas orogeny marks a major event in the geologic history of other parts of the Alexander terrane as well (Gehrels, Saleeby, and Berg, 1983). On Annette and Gravina Islands (A and G, fig. 59), the Ordovician and Silurian complex was metamorphosed to greenschist and perhaps amphibolite facies, uplifted, and brought to the surface prior to the deposition of Middle and perhaps Lower Devonian strata. On central and northern Prince of Wales Island, the Ordovician and Silurian rocks were apparently not deformed during this event, but there is evidence of southward-increasing uplift during Late Silurian or Early Devonian time. Ovenshine and Churkin (1969) report that Lower Devonian strata on central and northern Prince of Wales Island were deposited as a clastic wedge that became thinner, finer grained, and more stratigraphically conformable to the north. These workers attributed this clastic wedge to "***Late Silurian to pre-Middle Devonian diastrophism in southern southeastern Alaska***" (p. 50).

There is little evidence of the Klakas orogeny on northern Prince of Wales Island and on Kuiu Island (Muffler, 1967), but across the Chatham Strait fault, the Silurian and Devonian rocks record a geologic history similar to that on southern Prince of Wales Island (fig. 59). Loney and others (1975) report that on northeastern Chichagof Island (fig. 59), Silurian syenites and trondhjemites were uplifted and exposed prior to the deposition of a thick section of Middle Devonian conglomerate and clastic strata. Correlation of various features across the Chatham Strait fault suggests that these rocks on northeastern Chichagof Island have been offset 150 km (Hudson and others, 1981) to 205 km (Ovenshine and Brew, 1972) from an original position west of Kuiu Island. However, similarities in the Silurian plutonic rocks (Brew and Morrell, 1980) and the nature and age of uplift and erosion between northeastern Chichagof Island and southern Prince of Wales Island suggest that these two areas may have been adjacent to one another during Silurian and Devonian time. The Chatham Strait fault and its northern and southern extensions may, therefore, have approximately 350 km of post-Devonian right-lateral displacement.

Acknowledgements

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Reviewed by David A. Brew and Martha L. Miller.

REACTION ISOGRADS IN PELITIC ROCKS OF THE COAST PLUTONIC-METAMORPHIC COMPLEX NEAR JUNEAU, ALASKA

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

A regional metamorphic terrane containing mineral assemblages that range from the prehnite-pumpellyite metagraywacke facies to the upper amphibolite facies is exposed along the western margin of the Coast plutonic-metamorphic complex in southeastern Alaska. Forbes (1959) documented the first appearance of the Barrovian index minerals biotite, garnet, staurolite, kyanite, and sillimanite in a transect of the metamorphic belt along Blackerby Ridge (fig. 60) More recently, index mineral isograds have been mapped over a broad area from Taku Inlet to Berners Bay and Lynn Canal (fig. 52) (Ford and Brew, 1973, 1977; Brew and Ford, 1977).

The isograds are based on the first occurrence of the particular index mineral, which can be highly dependent on bulk composition effects and thus do not necessarily indicate the same metamorphic conditions. Thompson (1957) introduced the concept of defining isograds on the basis of discontinuous reactions in order to minimize effects of bulk composition. Isograds so determined were defined as reaction isograds by Winkler (1979, p. 66). This report describes results of continued detailed field, petrographic, and mineral chemistry studies of the progressive metamorphism in the area of Juneau (Himmelberg and others, 1984), and specifically demonstrates that the previously determined index-mineral isograds along Heintzleman and Blackerby Ridges (fig. 60) in fact mark the occurrence of particular reactions and thus are true reaction isograds.

Following Thompson (1957), it has become customary to depict pelitic mineral assemblages on the AFM projection of the "ideal" pelitic system SiO_2-Al_2O_3-K_2O-FeO-MgO-H_2O. On the AFM projection, discontinuous reactions are indicated by distinct changes in the tie-line configuration of the diagram. Although the change in topology can be modeled by a chemical reaction, it does not necessarily imply that the reaction actually took place in the rocks being considered.

Listed in table 21 are mineral assemblages on Heintzleman and Blackerby Ridges that are appropriate to the AFM projection. In addition to the minerals listed, all the assemblages contain quartz and muscovite and may contain plagioclase and the accessory minerals magnetite, ilmenite, sphene, graphite, apatite, zircon, tourmaline, beryl, and a sulfide mineral. On the basis of textural

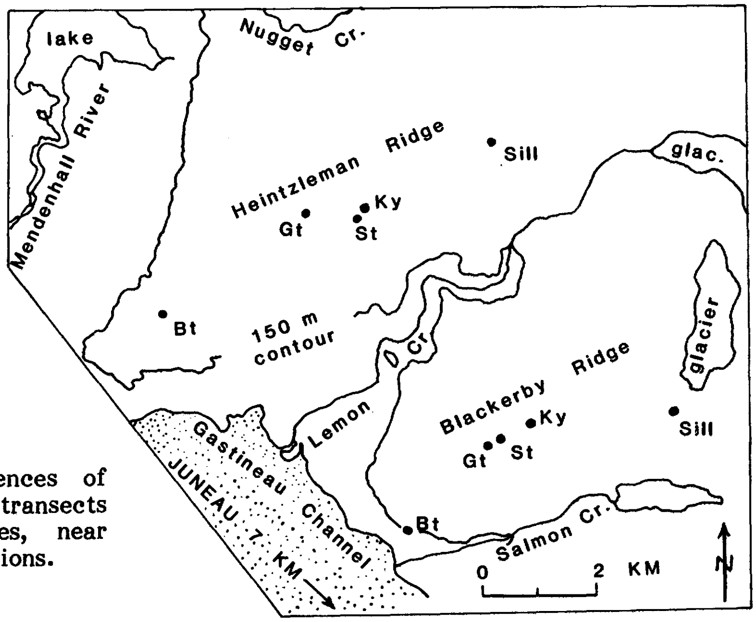

Figure 60.—Locations of first occurrences of metamorphic minerals in northeastward transects on Heintzleman and Blackerby Ridges, near Juneau. See table 21 for mineral abbreviations.

criteria, all phases in an assemblage are inferred to be in equilibrium. The order in which minerals are listed carries no implications.

Schematic AFM diagrams (projected from quartz and muscovite) constructed from the observed mineral assemblages listed in table 21 are illustrated in figure 61. The sequence of reactions and equilibria in the pelitic schists can be deduced from the AFM topologies of each zone. In the reactions that follow we have ignored the reaction coefficients. (See table 21 for abbreviations.)

The formation of staurolite, and thus the staurolite isograd which is the boundary between the garnet zone and the staurolite zone (fig. 61A, B), can be modeled by the discontinuous reaction

$$\mathrm{Chl + Mu + Gt = St + Bt + Qtz + W} \qquad (1)$$

which is reflected by breaking the chlorite-garnet tie line and forming the staurolite-biotite tie line. As schematically illustrated in figure 61B by the dashed chlorite-garnet tie line, once staurolite forms, the complete reaction assemblage, Chl-Gt-Bt-St, persists into the kyanite zone. This increased variance in nature has been reported in other areas (Carmichael, 1970; Guidotti, 1974; Novak and Holdaway, 1981) and consideration of garnet analyses (G. R. Himmelberg, unpublished data) suggests it is due to variable Mn in the garnet.

In other areas of progressive metamorphism (Carmichael, 1970; Guidotti, 1974; Novak and Holdaway, 1981), the formation of kyanite (or sillimanite) results from the model discontinuous reaction

$$\mathrm{Chl + St + Mu + Qtz = Ky + Bt + W} \qquad (2)$$

The change in the AFM diagram topologies on either side of the kyanite isograd in the Juneau area (fig. 61B, C) suggests that reaction 2 also can account for the kyanite isograd in these rocks. Note, however, that the phase assemblages in the natural system are more complex than those expected in the "ideal" system. Not only do the "four-phase" assemblages, Chl-Bt-Gt-St and Bt-Gt-St-Ky, occur over a significant field interval, but so does the "five-phase" assemblage Chl-Bt-Gt-St-Ky (sche-

Table 21.—Observed mineral assemblages[1] on Heintzleman and Blackerby Ridges appropriate to the AFM projection

[Chl, chlorite; Bt, biotite; Gt, garnet; St, staurolite; Ky, kyanite; Sill, sillimanite; Qtz, quartz; Mu, muscovite; W, water]

Garnet Zone	Middle Kyanite Zone
Chl - Bt	Bt - Gt
Bt - Gt	Bt - Gt - St - Ky
Chl - Bt - Gt	
Staurolite Zone	Upper Kyanite Zone
Bt - Gt	Bt - Gt
Bt - Gt - St	Bt - Ky
Chl - Bt - Gt - St	
Lower Kyanite Zone	Sillimanite Zone
Chl - Bt - Gt - St	Bt - Gt
Bt - Gt - St - Ky	Bt - Sill
Chl - Bt - Gt - St - Ky	Bt - Gt - Sill

[1]All assemblages also contain quartz and muscovite.

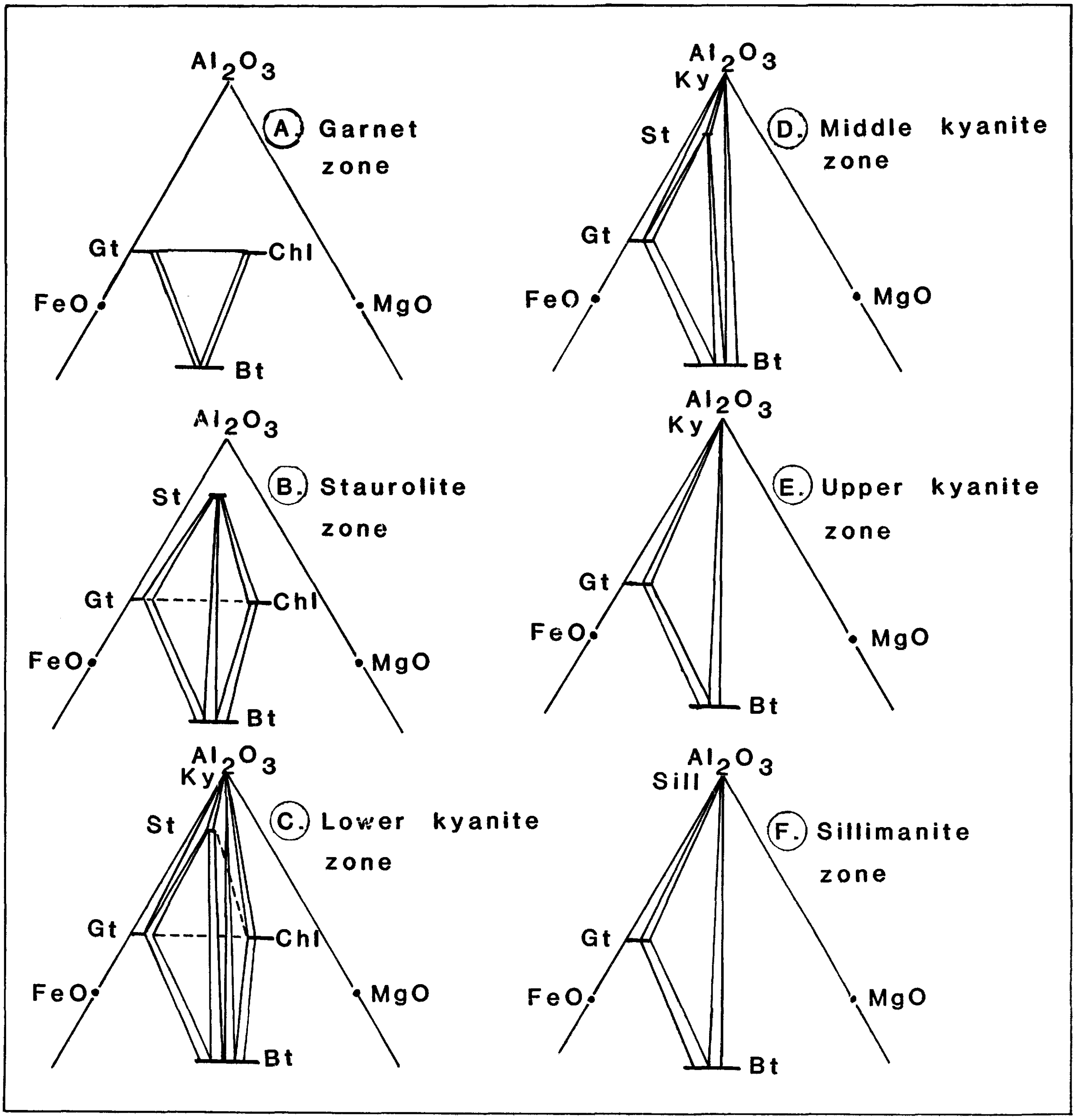

Figure 61.—Schematic AFM diagrams based upon observed mineral assemblages on Heintzleman and Blackerby Ridges. Dashed lines indicate existence of four- and five-phase assemblages. All assemblages include quartz and muscovite.

matically illustrated by dashed lines in figure 61C). Although the "four-phase" assemblages can be explained by Mn in garnet, other factors such as variability of Zn in staurolite, Na in muscovite, or X_{H_2O} are required to explain the increased variance of the "five-phase" assemblage. The "five-phase" assemblage also has been mapped by Guidotti (1974) in Maine, where he attributed the zone to increasing X_{H_2O} caused by reaction 2.

Based upon mineral assemblages, the kyanite zone is divided into lower, middle, and upper (fig. 61C, D, E). The boundary between the lower and middle kyanite zones (fig. 61C, D) essentially represents reaction 2 going to completion by the consumption of chlorite. Assemblages of Ky-Bt-Chl without staurolite have not been observed in the Heintzleman and Blackerby Ridge areas, reflecting bulk compositions left of the kyanite-biotite join and allowing the boundary between the lower and middle kyanite zones to be mapped as the chlorite-out isograd (Ford and Brew, 1973, 1977; Brew and Ford, 1977).

The middle kyanite zone is characterized by the "four-phase" assemblage Bt-Gt-St-Ky. The persistence of this assemblage can be modeled by the staurolite-consuming reaction

$$St + Mu + Qtz = Gt + Ky + Bt = W \quad (3)$$

which is a discontinuous reaction in the "ideal" system but continuous in nature because of factors previously mentioned.

The upper kyanite zone as defined by the topology of figure 61E is tentative. The lower boundary of the upper kyanite zone would be marked by reaction 3 going to completion. To date, however, the limiting assemblage Bt-Gt-Ky has not been observed in muscovite-bearing rocks. The limiting assemblage is common, however, in muscovite-absent rocks, some with a trace of staurolite, which might represent reaction 3 going to completion in K-deficient rocks. The absence of staurolite in sillimanite-bearing samples indicates that reaction 3 goes to completion before kyanite reacts to sillimanite and supports interpretation of an upper kyanite zone as defined in figure 61E. Until better defined, the boundary between the middle and upper kyanite zone can be approximated by the staurolite-out isograd.

The sillimanite isograd is based upon the reaction

$$Ky = Sill \quad (4)$$

Kyanite and sillimanite occur together in many specimens, but there is usually evidence of reaction. In one specimen where kyanite and sillimanite occur together without reaction textures, they do so in alternate bands on the scale of millimeters.

In summary, in the Heintzleman and Blackerby Ridge areas, the staurolite, kyanite, and sillimanite isograds as previously mapped can be modeled by reactions 1, 2, and 4, respectively. In addition, a new isograd modeled by reaction 3 is recognized in the kyanite zone.

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Reviewed by Cynthia Dusel-Bacon and R. A. Loney.

MIGMATITES OF THE COAST PLUTONIC-METAMORPHIC COMPLEX, SOUTHEASTERN ALASKA

By Susan M. Karl and David A. Brew

Migmatites in the central metamorphic belt of the Coast plutonic-metamorphic complex (Brew and Ford, 1984) in southeastern Alaska (figs. 52, 62) have generally consistent internal compositional and structural relations throughout the belt. In this study, one pre-intrusive and two minor and two major intrusive components are shown to combine to form three, and locally four, migmatite units; none of the intrusive and migmatite units is present everywhere in the Coast plutonic-metamorphic complex. The migmatite units are well exposed east of Chief Shakes Lake in the Petersburg quadrangle and south of Crater Lake in the Taku River quadrangle (fig. 62), and they are defined by the first appearance of successive neosomes from west to east. A detailed diagrammatic cross-section of a Crater Lake traverse is shown in figure 63; the relations are similar near Chief Shakes Lake.

The pre-intrusive component of the suite of migmatites consists of an amphibolite facies (hornblende-) biotite-quartz-feldspar schist and gneiss unit that occurs both in a continuous narrow belt in the western part of the Coast plutonic-metamorphic complex and as a paleosome in some migmatite units. The quartzo-feldspathic com-

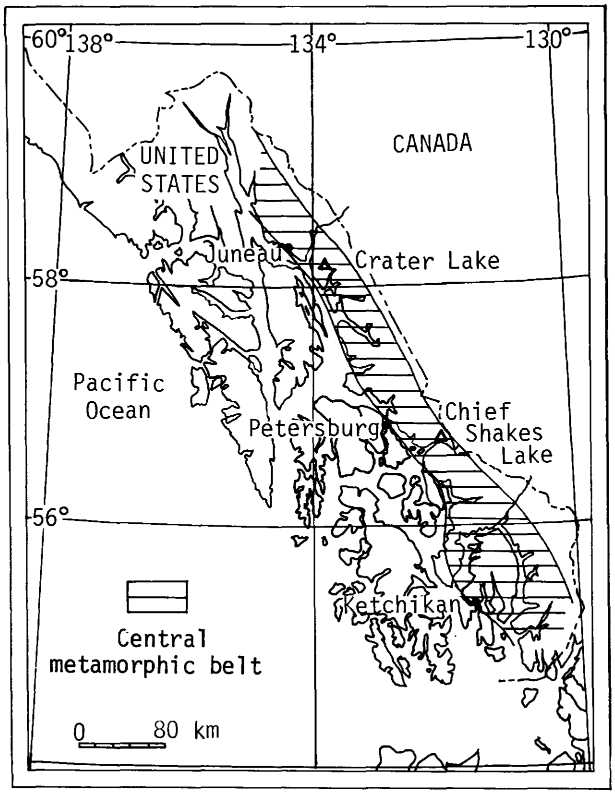

Figure 62.—Location of Crater Lake and Chief Shakes Lake in the central metamorphic belt of the Coast plutonic-metamorphic complex, southeastern Alaska.

position of this unit and the presence of intercalated carbonate layers suggest sedimentary protoliths for most of the rocks. The age of the protolith is not known, but it is inferred to be mostly Paleozoic (Brew and others, 1977; Brew and Ford, 1983b).

The narrow belt of schist and gneiss is intruded on its eastern edge by the first major intrusive component: the homogeneous foliated sphene-biotite-hornblende granodiorite, tonalite, and quartz diorite of the Coast plutonic-metamorphic complex sill belt (Brew and Morrell, 1980; Brew and Ford, 1981). This locally produces the oldest and westernmost of the four migmatite units (migmatite unit 1, fig. 63). The margins of the tonalite and granodiorite sills are foliated to gneissic and parallel the fabric in the schist and gneiss. Near the margins of the sills, seriate hornblende phenocrysts are aligned with the regional fabric, but in some sills they have a swirled and convoluted igneous fabric.

Local zones of intense deformation occur both at the margins and in the sills. In these zones, fine-grained mafic inclusions are streaked and schleiric, and in many places the inclusions have indistinct borders and contain feldspar porphyroblasts in various stages of development. In many of the inclusions, the feldspars straddle inclusion boundaries. Diked and deformed xenoliths of the quartzo-feldspathic schist and gneiss in the margins of the tonalitic sills constitute the first of the four migmatite units. Zircons from three tonalite localities in the sill belt have yielded ages near the Cretaceous-Paleocene boundary (Gehrels and others, 1984).

The second migmatite unit was produced by the invasion of the eastern part of the tonalite sill belt and associated migmatite, the first minor intrusive component, by a texturally heterogeous K-feldspar megacrystic hornblende-biotite grandiorite. This second generation migmatite is a compositionally and texturally complicated unit, typically consisting of approximately 2 percent calc-silicate inclusions, 8 percent quartzo-feldspathic inclusions, 15 percent seriate hornblende tonalite inclusions, 50 percent intermediate tonalitic gneiss and migmatite inclusions, and 25 percent K-feldspar megacrystic granodiorite neosome. On a small scale (centimeter to meter) migmatite unit 2 inclusions are locally bent, folded, and contorted, but at outcrop and larger scale, this migmatite is foliated and compositionally banded parallel to the trend of regional foliation.

The eastern side of the K-feldspar-megacrystic-granodiorite-neosome migmatite (unit 2) is locally invaded by the second minor intrusive component: leucocratic biotite granodiorite and granite that forms irregular apophyses, as well as intrusive masses up to tens of meters in dimension. These constitute the third migmatite unit. Inclusions of the older units described above are present in approximately the same proportions as in the second migmatite unit, but they are reduced in overall abundance relative to 40 percent leucocratic grandiorite neosome. The rocks are intensely deformed, and almost all vestiges of the regional structure are lost in this third migmatite unit. To the east, the abundance of inclusions in the leucocratic granodiorite diminishes, the boundaries of inclusions become less distinct, and the leucocratic granodiorite gradationally assumes a gneissic and K-feldspar porphyritic or porphyroblastic fabric.

This gneissic leucocratic granodiorite is the neosome of the fourth migmatite unit. Paleosomes include zones of agmatite composed of coarse-grained amphibolite or pyroxenite with amphibolitic reaction rims, zones of texturally varied gneisses of intermediate composition, and zones of quartzo-feldspathic schist and gneiss like that described earlier. The foliation and compositional banding in the leucocratic granodiorite are broadly warped and do not form consistent regional trends.

This gneissic and leucocratic biotite granodiorite neosome of the fourth migmatite unit grades eastward into the main phase of the homogenous sphene-hornlende-biotite granodiorite, east of the diagrammatic cross section shown in figure 63, which is the second major intrusive component. It constitutes the main phase of the central granitic

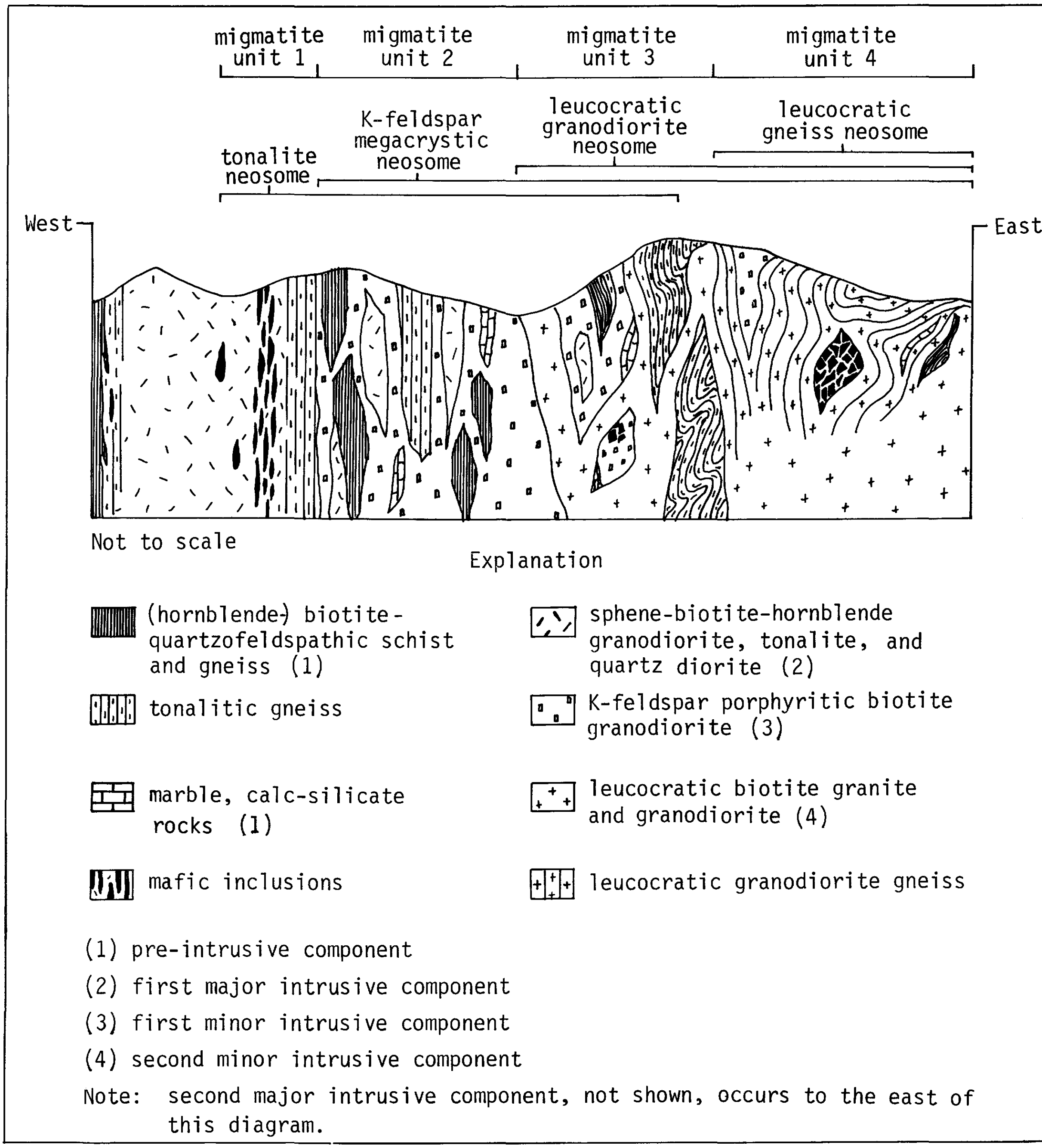

Figure 63. Diagrammatic cross-section of four migmatite units occurring within the central metamorphic belt of the Coast plutonic-metamorphic belt near Crater Lake in the Taku River quadrangle, southeastern Alaska. The units are defined by the first appearance of each successive neosome from west to east.

belt of the Coast plutonic-metamorphic complex. Biotite and hornblende from these granodiorite plutons have consistently yielded Eocene ages by K-Ar methods (Forbes and Engels, 1970; Brew and others, 1977; Smith and Diggles, 1981). Zircons have also given Eocene ages (Gehrels and others, 1984).

In summary, the migmatites of the Coast plutonic-metamorphic complex central metamorphic belt in southeastern Alaska contain evidence for at least four, and possibly five, discrete metamorphic and (or) igneous events during Paleocene and Eocene time. In probable order of their occurrences, these are: (1) the development of the fabric in the quartzo-feldspathic schist and gneiss unit; (2) the syn- or late-tectonic intrusion of the biotite-hornblende granodiorite, tonalite, and quartz-diorite sills, which parallel the fabric of the schist and gneiss; (3) the invasion by the K-feldspar megacrystic biotite granodiorite; (4) the subsequent intrusion of the leucocratic biotite granite and granodiorite, which may be a preliminary phase, based on gradational contacts and similar composition, and (5) the biotite granodiorites emplaced as the main phase of the central granitic belt of the Coast plutonic-metamorphic complex.

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Reviewed by R. A. Loney and Cynthia Dusel-Bacon.

EXPLOSIVE LATEST PLEISTOCENE(?) AND HOLOCENE ACTIVITY OF THE MOUNT EDGECUMBE VOLCANIC FIELD, ALASKA

By James R. Riehle and David A. Brew

The Mount Edgecumbe volcanic field on Kruzof Island, west of Sitka (fig. 64), consists of lava flows and domes ranging from 47 percent to 70 percent SiO_2 (Brew and others, 1969; Myers and Marsh, 1981; Kosko, 1981) that are mantled by unconsolidated ash and lapilli. Part of the composite cone of Mount Edgecumbe, as well as the caldera of Crater Ridge and several smaller cones (fig. 64), are interpreted to have formed during eruptions that produced the unconsolidated mantle (Brew and others, 1969). An early Holocene age of the culminating eruptions is implied by radiocarbon ages of 10,300 ± 300 yr for peat beneath an ash layer at Juneau (Heusser, 1960) and 8,750 ± 300 yr for wood atop ash at Sitka (Yehle, 1974). Because of the unmodified morphology of Mount Edgecumbe, Myers (1979) suggested that some activity may have occurred in the past thousand years; supporting evidence exists in an undated, thin ash layer near Sitka (Yehle, 1974).

This study focuses on the culminating explosive activity of the field. Previous work on the unconsolidated deposits is restricted to descriptions of two samples by Brew and others (1969). In this preliminary report, we describe the remarkably wide range of compositions of the airfall deposits and present several new radiocarbon dates. Field stations are shown in figure 64. About 200 samples have been sieved, the size frequencies determined, and clasts examined under binocular microscope. Thin sections, together with 17 whole-rock analyses and 24 microprobe analyses of glass separates (fig. 65), comprise the data base; all chemical analyses are on hand-picked lapilli judged to be juvenile. The petrographic classification used herein (mafic, andesitic, dacitic, rhyolitic) is based solely on whole-rock SiO_2 content and is adopted for brevity.

Mafic tephra (<56 percent SiO_2, whole-rock).—The oldest preserved tephra layers consist of scoriaceous mafic clasts that are dark brown or gray to reddish gray. Phenocrysts total 25 to 40 percent; plagioclase is most abundant, followed by subequal clinopyroxene and olivine. The mafic tephra is entirely of airfall origin and is thickest (several tens of meters) adjacent to scoria cones northeast of Mount Edgecumbe. Granulometry suggests that some mafic tephra may also have originated from vents southwest of Mount Edgecumbe. The thicknesses of mafic-tephra layers are 30-40 cm at Sitka, about 15 cm on Partofshikof Island (inset, fig. 64), and 80-100 cm near Shelikof Bay (81-92, fig. 64); mafic tephra does not occur on Gornoi Island (inset, fig. 64). Mafic tephra comprises up to several layers, implying multiple eruptions.

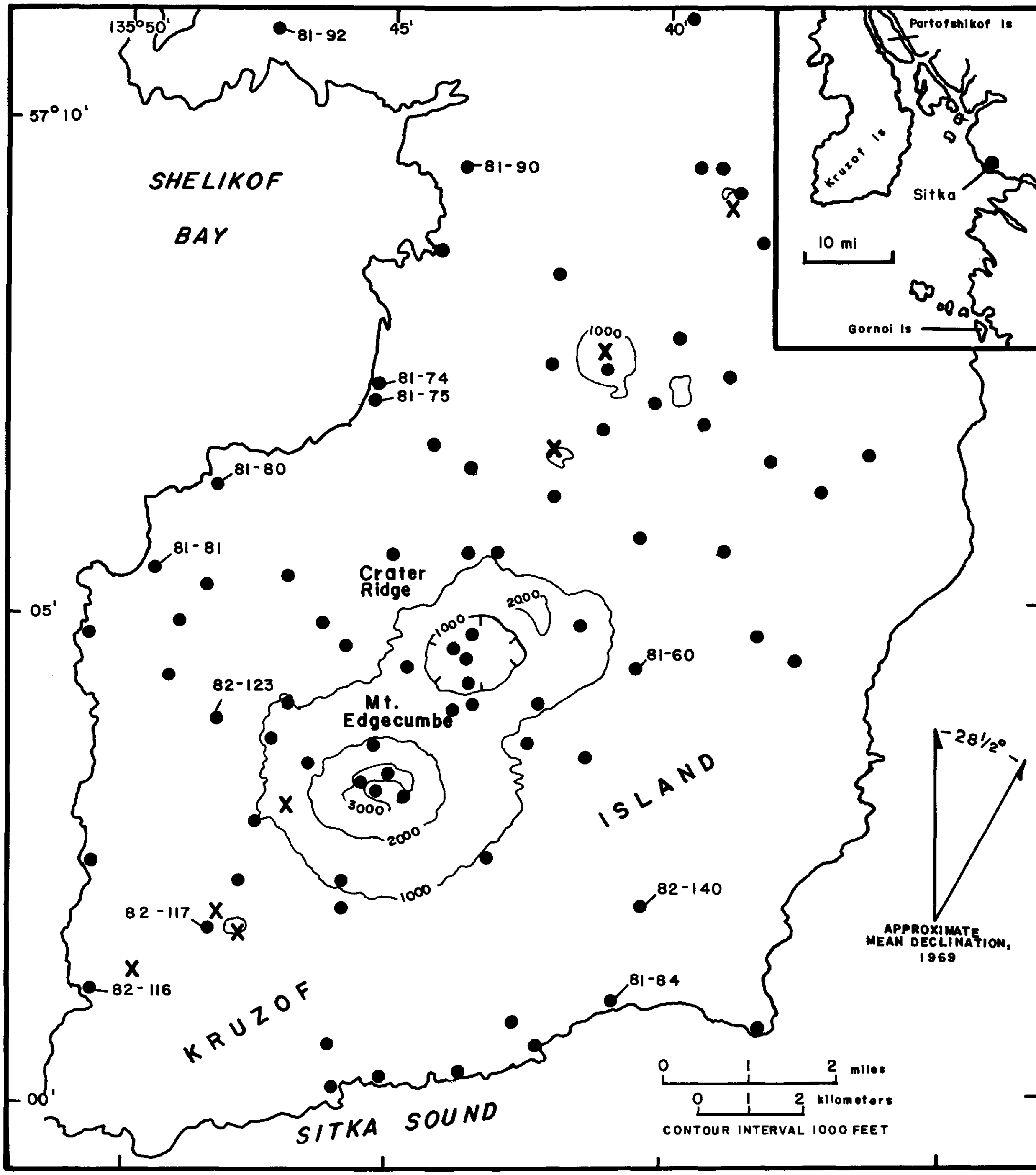

Figure 64.—Locations of stations occupied on Kruzof Island during 1981 and 1982. Smaller vents, primarily cinder cones, are shown by the symbol "x" (modified from Brew and others, 1969); the composite cone of Mount Edgecumbe and the caldera of Crater Ridge are outlined by the 1000-, 2000-, and 3000-ft contours. Labelled stations are those to which reference is made in the text. Base is from U.S. Geological Survey maps Sitka A-5 and A-6, 1951.

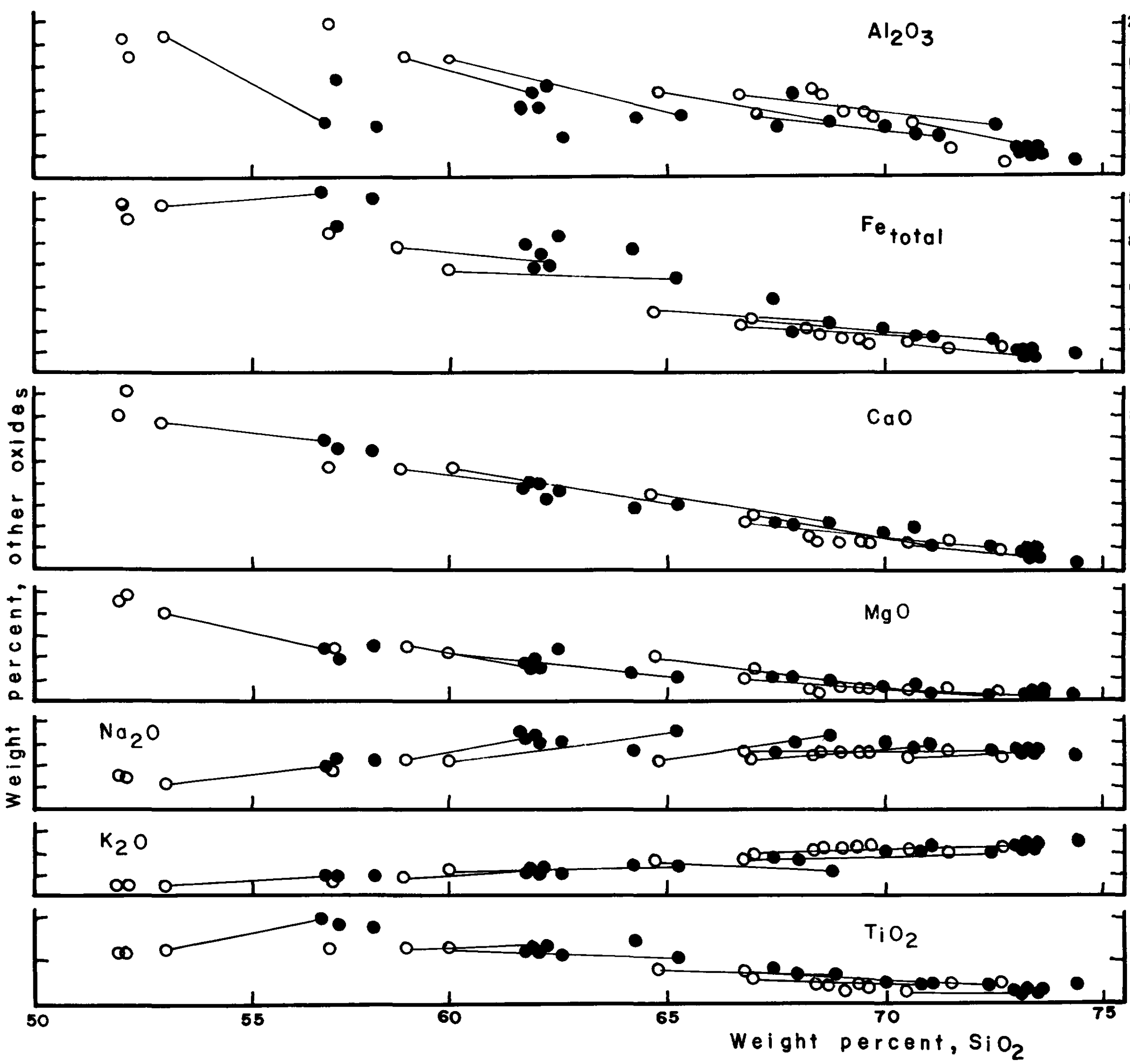

Figure 65.—SiO_2-variation diagrams showing compositions of 17 whole-rock samples (open circles) and 24 glass separates (solid circles), Mount Edgecumbe volcanic field. Most of the glass separates, analyzed by microprobe, probably include some microlites. Tie-lines are between whole-rock and glass analyses of the same sample. All analyses are of hand-picked pumice or scoria judged to be juvenile in origin. (Whole-rock analyses by Branch of Analytical Laboratories, USGS; analysts: A. Bartel, J. S. Wahlberg, J. Taggart, J. Baker. Microprobe analyses carried out on A. R. L. EMX-SM, USGS, Menlo Park; J. Riehle, analyst. All whole-rock analyses initially totalled between 99 and 101 percent, including P_2O_5, MnO, and loss on ignition; microprobe analyses initially totalled between 93 and 101 percent, including MnO. Values plotted here are recalculated to 100 percent on volatile-free basis).

Andesitic tephra (56-60 percent SiO_2, whole-rock).—Andesitic tephra comprises up to six layers of brown-gray to yellow-brown, vesicular airfall ash and lapilli which directly overlie mafic tephra. Phenocrysts total 20 percent; plagioclase exceeds subequal amounts of clinopyroxene and orthopyroxene. Andesitic tephra is locally interbedded with dacitic tephra (81-76, 81-90; table 22), implying either zoned eruptions from a single vent or separate vents. Maximum thicknesses (several tens of meters) occur southeast of Mount Edgecumbe; no significant sources of andesitic tephra appear north of Crater Ridge. The thicknesses of andesitic tephra are 20-30 cm at station 81-92, 10-15 cm on Partofshikof Island, and 35-50 cm at Sitka. A 2- to 4-cm-thick bed of fine ash on Gornoi

Table 22.—Generalized geologic columns selected to illustrate the variations in silica content and lithologic types among unconsolidated deposits of the Mount Edgecumbe volcanic field. Twenty-three of the 41 samples plotted on figure 65 are included. Locations of stations are shown on figure 64. SiO_2 contents of analyzed samples are shown at the right margin of each column; whole-rock analyses are underlined and glass separates are in parentheses. Total thickness of each station is given in parentheses.

81-60 (10 m)	81-81 (20 m)	81-75 (7.5 m)	81-90 (3 m)	81-76 (1.8 m) (Sitka)
multiple layers of airfall ash, including dense volcanic lithics and dacitic pumice	ash-flow deposit, dacitic pumice (73%)	lahar deposit	multiple layers of dacitic airfall pumice	black silt containing 1 or possibly 2 layers of very fine ash (73%)
lahar deposit	multiple layers of dacitic airfall pumice	multiple layers of dacitic airfall pumice	andesitic airfall scoria (61%)	pale grey pumice, probably colluvial (61%)
lahar deposit	lahar deposit	dark, fine air-fall ash	andesitic airfall scoria 57%	single graded airfall bed of dacitic ash and fine lapilli (73%)
airfall or ash-flow deposit; dacitic pumice 69%	fine, dark air-fall ash	vaguely stratified dacitic pumice, as airfall blocks to 25 cm diameter 66% (70%) (72%)	dacitic airfall pumice 60% (65%)	3 layers of dacitic air-fall ash (73%) (73%) (73%)
multiple layers of dacitic airfall pumice 68%	vaguely stratified dacitic pumice, as airfall blocks to 25 cm diameter	not exposed	multiple layers of mafic airfall scoria (57%)	several interbedded layers of dacitic pumice and andesitic scoria, airfall (67%) (62%) (64%)
ash-flow or lahar deposit; dacitic pumice 69%	not exposed		bedrock	several layers of mafic scoria, airfall (58%)
airfall or lahar deposit; dacitic pumice and dense volcanic lithics 69%				bedrock
not exposed				

Island is tentatively classified as andesitic tephra.

Dacitic tephra (60-72 percent SiO_2, whole-rock).—Dacitic tephra is predominant pumice with spherical to slightly elongate vesicles, ranging from yellow to pale brown, pink, or gray. Phenocrysts are less than 20 percent and consist, in decreasing order of abundance, of plagioclase, orthopyroxene, and clinopyroxene. Most dacitic tephra overlies the andesitic tephra, although some layers are interbedded (81-76, 81-90; table 22). A particularly thick (10 m) and coarse (up to 25 cm) unit of tephra crops out at Shelikof Bay (81-80, 81-81, fig. 64); welded tephra on the southwest flank of Mount Edgecumbe has the same composition (66 percent SiO_2, whole-rock), and both are probably airfall formed in a Plinian-type eruption of Mount Edgecumbe.

Dacitic pumice occurs in pyroclastic-flow deposits and with clasts of older volcanic rocks in lahar deposits on Kruzof Island. The thickest lahar deposits (several tens of meters) occur northwest and southeast of the ridge between Mount Edgecumbe and Crater Ridge caldera. Thin ash-flow deposits (1 to 3 m) crop out along the south shores of Kruzof Island and Shelikof Bay (81-84, 82-140, 81-81, 81-80; fig. 64). The sources of the ash flows and lahars are uncertain; clasts in the lahars are dominantly nonvesicular volcanic rocks which resemble older lavas truncated by Crater Ridge caldera. The simplest hypothesis is that the lahars originated in water-rich eruptions from Crater Ridge caldera (caldera lake?); if so, they may document its formation. Some tephra is compositionally similar to the ash-flow deposits, suggesting that the ash flows originated from an eruption column that vented from either Mount Edgecumbe or Crater Ridge caldera.

Dacitic tephra is 220 cm thick at station 81-92, 15-20 cm on Partofshikof Island, 1 m at Sitka, and 6-8 cm on Gornoi Island.

Rhyolitic tephra (>72 percent SiO_2, whole-rock).—A thin layer of tephra is separated by silt and peat from underlying tephra layers at several stations on Kruzof Island. The layer is a heterogeneous assemblage of dense volcanic rocks and pumice. The pumice contains 72 percent SiO_2 in whole-rock and 74 percent SiO_2 in the glass and is slightly more siliceous than samples classified herein as dacite (fig. 65). Maximum thickness (5 cm) and lapilli size (2 cm) occur southwest of Mount Edgecumbe (82-116, 82-117; fig. 64). The deposit is probably the result of a minor eruption of Mount Edgecumbe or Crater Ridge caldera; the pumice could be acces- sory material from an earlier unknown deposit incorporated in a largely phreatic eruption.

We found one tree trunk in a dacitic-pumice-bearing lahar deposit (81-74; fig. 64). The piece was vertical and had a frayed top and a flared base resembling the remains of a root system. We think that the tree was killed and transported by a lahar. A split sample yielded 9,180 ± 150 and 9,150 ± 150 yr B.P. (I-12,218; I-12,219). Horizontal wood fragments at the base of peat atop the tephras elsewhere on Kruzof Island yielded 5,690 ± 100 and 5,520 ± 100 yr B.P. (I-12,234; I-12,235); the rhyolite tephra was not identified at this locality. We interpret these results to mean that part of the dacitic tephra is greater than about 5,600 yr old.

Samples of peat from above and below the rhyolitic tephra (82-123; fig. 64) yielded 4,030 ± 90 and 4,310 ± 140 yr B.P., respectively (Beta-6004; Beta-6005); a single peat sample enclosing the rhyolitic tephra (82-116; fig. 64) yielded 5,760 ± 70 yr B.P. (Beta-6003). The discrepancy in ages suggests that there may have been two eruptions at about 4,200 and 5,700 yr B.P., which followed the last dacitic eruption after a hiatus. Supporting evidence for two eruptions of rhyolitic tephra is meager (for example, 81-76; table 22).

We have found no organic material in or beneath andesitic and mafic tephra. However, nowhere in 100 stations have we observed any evidence (erosional channels, soil profiles, peat) for a significant hiatus within or between layers of mafic, andesitic, and dacitic tephra. We believe that the entire tephra sequence, from mafic through dacitic, was erupted within a few hundred to a maximum of perhaps 2,000 years prior to 9,000 yr B.P. and that the rhyolitic tephra is the product of one or possibly two minor eruptions at about 5,000 yr B.P.

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Reviewed by L. J. P. Muffler and E. W. Hildreth.

PRELIMINARY REPORT ON A LARGE GRANITIC BODY IN THE COAST MOUNTAINS, NORTHEAST PETERSBURG QUADRANGLE, SOUTHEASTERN ALASKA

By John H. Webster

The lengthy geological feature known as the Coast plutonic-metamorphic complex extends from the British Columbia-Washington border to the Yukon Territory-Alaska boundary (Brew and Ford, 1984). This heterogeneous complex includes metamorphic rocks and nonfoliated equigranular granitic rocks to the northwest and schist, gneiss, migmatites, and elongated tonalite and quartz diorite plutons to the southwest. Preliminary petrographic studies of granitic rocks from a large body in the Petersburg D-1 quadrangle (fig. 52) show that they are lithologically and texturally similar to radiometrically dated granitoids along the International Boundary in the nearby Sumdum, Taku River, Bradfield Canal and Ketchikan quadrangles. The plutons are notable for their equant dimensions and, except for some weak primary foliations related to emplacement, outcrops generally lack fabric.

Much of the Petersburg D-1 quadrangle adjacent to the International Boundary is underlain by granitic rocks belonging to a single pluton (fig. 66). Thin-section and stained-slab studies show that most of the rocks are sphene-bearing biotite-hornblende granodiorite (fig. 67A), with some variation into the granite and tonalite fields. Seven of the 35 samples discussed herein are granitoid leucosomes from migmatites adjacent to the main body, which are presently interpreted to be related to the main body. Most of the rocks studied contain 5-20 percent potassium-feldspar, and plagioclase content ranges from 32 to 57 percent. Color indices range from 2 to 25, with a mean of about 12, and generally increase southwestward toward a narrow belt of plutons known as the Coast plutonic complex sill (Brew and Ford, 1981).

The granodiorites are medium to coarse grained, hypidiomorphic granular, and locally porphyritic. Biotite and hornblende are the dominant mafic minerals; biotite is generally more abundant than hornblende and is chloritized to some extent. Plagioclase is dominantly subhedral, but distinct euhedral grains are not uncommon. Albite is the most prevalent twin type, but Carlsbad and crosshatch twins are also present. Plagiocalse grains are normally zoned, each sample having a few oscillatory zoned grains. Based on flat stage optical determination, plagioclase composition ranges from An_{25} to An_{37}. Metamorphic inclusions are rarely seen on a microscopic scale. Sphene and apatite are the most abundant accessory minerals, and large euhedral sphene crystals are common. Myrmekite is ubiquitous.

The thin-section and slab studies suggest that the pluton in the Petersburg D-1 quadrangle is

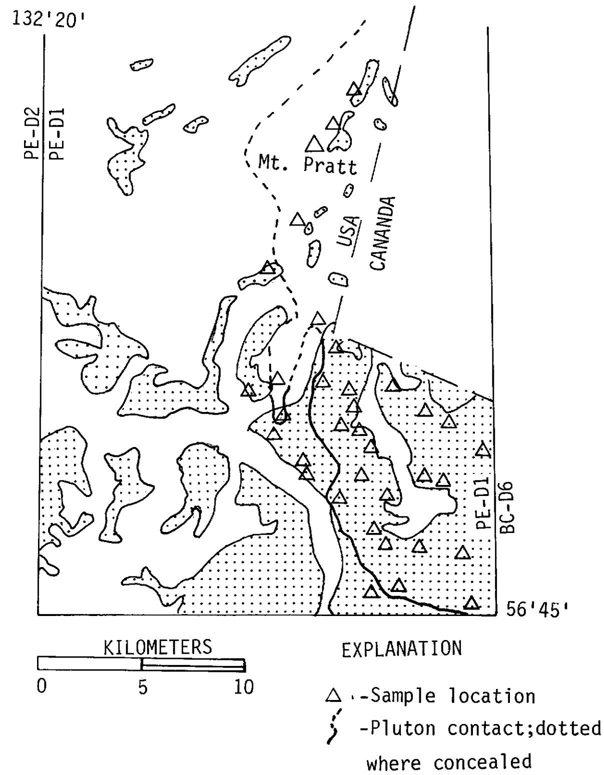

Figure 66.—Locations of samples studied from a large pluton near the International Boundary, Petersburg D-1 quadrangle, southeastern Alaska.

lithologically similar to granitic rocks along the International Boundary in the Taku River and Sumdum quadrangles to the northwest. Brew and others (1977) have described these granitic rocks as locally porphyritic, medium- to coarse-grained, nonfoliated, sphene-biotite-hornblende granodiorite, sphene-hornblende-biotite granodiorite, hornblende granodiorite, and sphene-biotite-hornblende granite. Color indices range from 10 to 20, and there is some modal variation to granite and tonalite (fig. 67B).

R. L. Elliott and R. D. Koch (U.S. Geological Survey, oral commun., 1983) have studied the eastward continuation of the pluton in the Petersburg D-1 quadrangle into the Bradfield Canal (D-6) quadrangle and have determined it to be biotite-hornblende granodiorite and granite. The eastern pluton has a distinct oval form and lacks any distinct internal fabric. Myrmekitic intergrowths are abundant, biotite and hornblende show partial alteration to chlorite, and color indices range from 4 to 12. The most common accessory minerals are euhedral sphene and opaque minerals. Modal analyses are spread equally in the granite and granodiorite fields (fig. 67C).

Smith (1977) has studied several granitic bodies to the southeast near Hyder in the Ketchikan

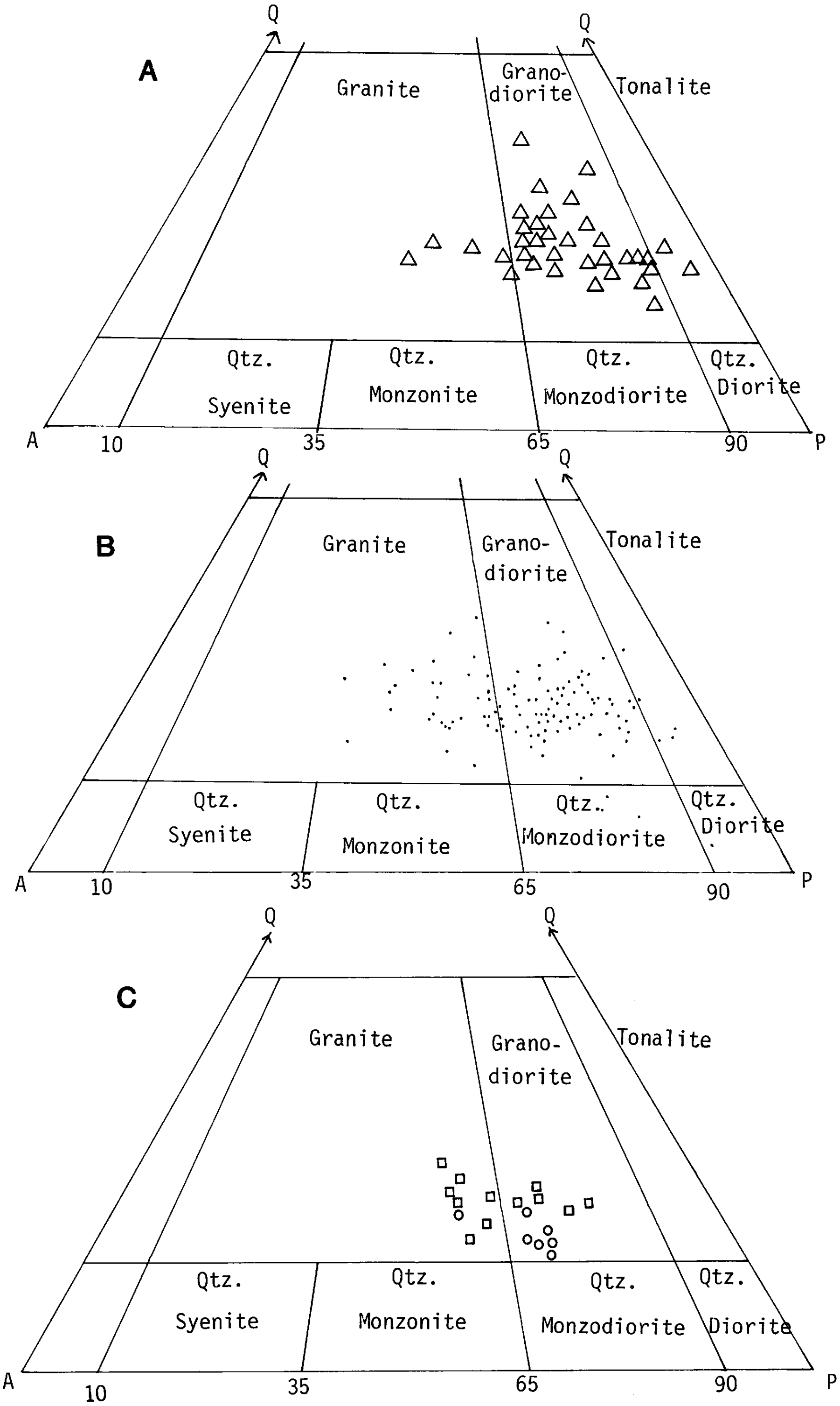

Figure 67.—Modal compositions of granitoids based on plutonic rock classification of Streckeisen (1973). (A) Modal compositions of granitoids from the D-1 quadrangle, Petersburg, Alaska; (B) Modal compositions of granitoids from the Taku River and Sumdum quadrangles, Alaska (from Brew and others, 1977); (C) Modal compositions of granitoids from Bradfield Canal quadrangle, Alaska. o - Boundary Granodiorite (Smith, 1977); □-D-6 granodiorite bodies (R. L. Elliott and R. D. Koch, U.S Geological Survey, unpub. data).

quadrangle. One body, the Boundary Granodiorite, is described as slightly inequigranular, hypidiomorphic, medium-grained, biotite-hornblende granodiorite (fig. 67C). Potassium feldspar is anhedral and interstitial and ranges in abundance from about 16 to 27 percent. Plagioclase content averages about 48 percent; the grains are euhedral to subhedral and commonly twinned, with thin twin lamellae and oscillatory zoning. Compositions in zoned grains range from An_{15} to An_{55}. Biotite occurs in excess of hornblende, and the biotite is partly altered to chlorite. The color index ranges from about 10 to 16. Sphene, opaque minerals, and apatite are accessory minerals. Internal fabrics are lacking.

An Eocene age is inferred for the as-yet undated, large granitic body in the Petersburg D-1 quadrangle. This is based on the above comparisons of lithology, field characteristics, and position in the Coast plutonic-metamorphic complex, all of which suggest that the body belongs to the same suite as K-Ar-dated granitic rocks in adjacent quadrangles. Smith and Diggles (1981) report 18 mineral-pair ages from various granodiorite bodies in the Ketchikan quadrangle, the majority of which are slightly to strongly discordant. Ages of the hornblende range between 55 and 50 m.y, whereas K-Ar dating on hornblende and biotite from the Boundary Granodiorite yielded ages of 50 m.y. (Smith, 1977). K-Ar age determinations from the large plutons near the International Boundary in the Sumdum and Taku River quadrangles have yielded concordant ages ranging from 49-54 m.y. (Brew and others, 1977). Gehrels and others (1983) report that a lithologically similar sphene-bearing, biotite-hornblende granodiorite pluton near Juneau gives a slightly discordant zircon date of 50 ± 2 m.y. Other work on the same granodiorite pluton (J. G. Smith, U.S. Geological Survey, unpub. data; Forbes and Engels, 1970) yielded K-Ar mineral pair ages of approximately 50 m.y. This suggests that K-Ar dating currently being done on the large granitic body in the Petersburg D-1 quadrangle will also yield an age of approximately 50 m.y.

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Reviewed by R. L. Elliott and A. B. Ford.

MINERALOGY AND A PETROGENETIC MODEL FOR THE TONALITE PLUTON AT BUSHY POINT, REVILLAGIGEDO ISLAND, KETCHIKAN 1° x 2° QUADRANGLE, SOUTHEASTERN ALASKA

By E-an Zen and J. M. Hammarstrom

The tonalite pluton at Bushy Point on Revillagigedo Island (fig. 68) was originally mapped (Berg and others, 1978) as Cretaceous porphyritic biotite granodiorite. Samples from an outcrop near Bushy Point were dated at 89.6 ± 2.7 m.y. (hornblende) and 82.3 ± 2.5 m.y. (biotite) by Smith and Diggles (1981, sample 70SJ379) using the K-Ar method. Petrographic and microprobe studies of the minerals in this tonalite reveal an interesting and complex story of magmatic evolution.

Samples of the garnet-bearing tonalitic pluton were collected from the dated outcrop and studied. Garnet occurs as orange-red, subhedral to euhedral grains and clusters a fraction of a millimeter in diameter. The grains are invariably sheathed in white plagioclase (fig. 69). Plagioclase is twinned, zoned (core about An_{40}, rim about An_{20}), and most grains contain some large intergrowths of skeletal zoisite. The rest of the rock includes a highly aluminous, green hornblende,

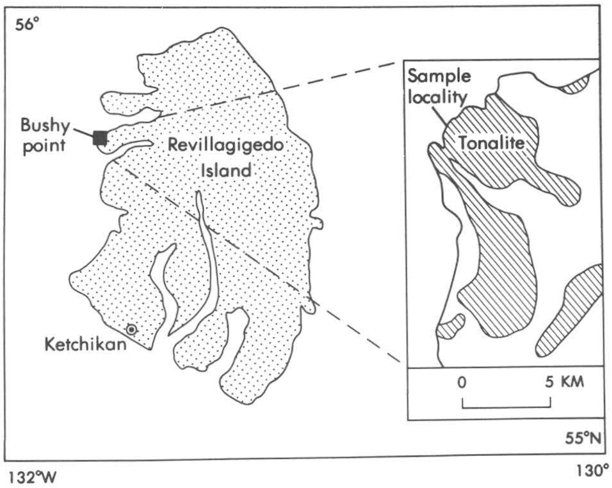

Figure 68.—Index map showing Revillagigedo Island and the sample locality of the tonalite at Bushy Point. Details shown in inset are from Berg and others (1978).

biotite, sphene, rare K-feldspar, quartz, and epidote. Textural relations of epidote to plagioclase and to biotite are typical of epidote of a magmatic origin (Zen and Hammarstrom, 1982). Representative microprobe results of mineral compositions analysed by Hammarstrom are shown in table 23.

The unusual composition of the garnet is noteworthy. Except for minor contents of spessartitic (spe) and pyropic (pyr) components, the bulk composition corresponds to nearly equal molar proportions of the grossular (gros) and almandine (alm) components. Such a garnet composition, if stable, would preclude the coexistence of anorthitic plagioclase and pyroxene (iron-rich clinopyroxene or "ferrosilite" or both). Bushy Point garnet compositions are plotted in terms of alm, gros, and pyr contents in figure 70, along with (a) analyses of garnet synthesized in laboratory experimental studies (open symbols and fields) on andesitic and tonalitic compositions at pressures ranging from 13 to 30 kbar and (b) naturally occurring alm-gros garnets (closed symbols). Compared to the Bushy Point garnets, most of the synthetic garnet compositions are richer in pyr (Allen and Boettcher, 1978; Allen and others, 1975). The Bushy Point garnets plot at the low temperature end of Stern and Wyllie's (1973) calculated garnet trend for gabbroic and tonalitic bulk compositions at 30 kbar and near the low temperature end of the field of compositions found by Green (1972) in experiments on hydrous andesite at pressures ranging from 13 to 36 kbar. This suggests that Bushy Point garnet cores must have formed at pressures not less than 13-15 kbar (minimum pressure for low temperature garnet stability in Green's experiments), or depths of about 40-50 km.

Additional support for a high-pressure origin for Bushy Point garnets comes from two well-studied

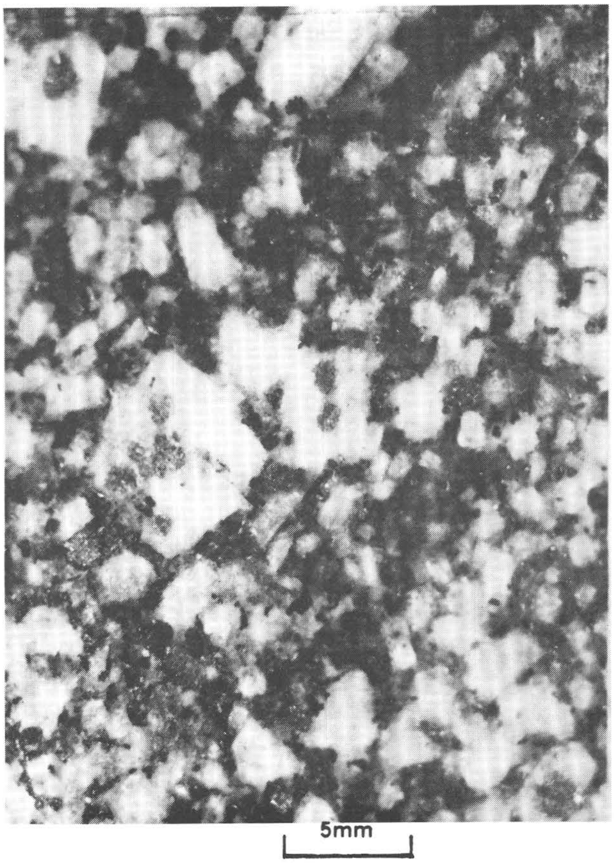

Figure 69.—Tonalite of Bushy Point; note the clusters of garnet set in fresh plagioclase. Scale divisions are in mm.

crustal xenoliths. Both of these xenoliths carry garnets; their compositions plot near the Bushy Point garnets (fig. 70). One xenolith is an amphibolite of presumed lower crustal origin from the Williams-4 kimberlite diatreme field in north-central Montana (Hearn and McGee, 1983). Another xenolith is an amphibolite included in a latite flow in Chino Valley, central Arizona (Arculus and Smith, 1979). Garnets from eclogitic glaucophane schists from Ecuador (Feininger, 1980), Oregon (Ghent and Coleman, 1973), and the western Alps (Ernst and Dal Piaz, 1978), all from high-pressure environments, have compositions similar to the Bushy Point garnet.

We interpret the garnet from Bushy Point as having formed initially at depths of at least 40-50 km, possibly as a very high-pressure magmatic rock. A possible paragenetic sequence for the Bushy Point rocks is illustrated in figure 71. We suggest that a depressed crustal block carrying garnet-bearing rock was rapidly uplifted. Melting resulted from the lowered pressure at the prevailing temperature and H_2O fugacity. At pressures less than about 12-13 kbar, garnet became unstable and began to react with the melt to form plagioclase and a more calcic garnet rim. Garnet is preserved

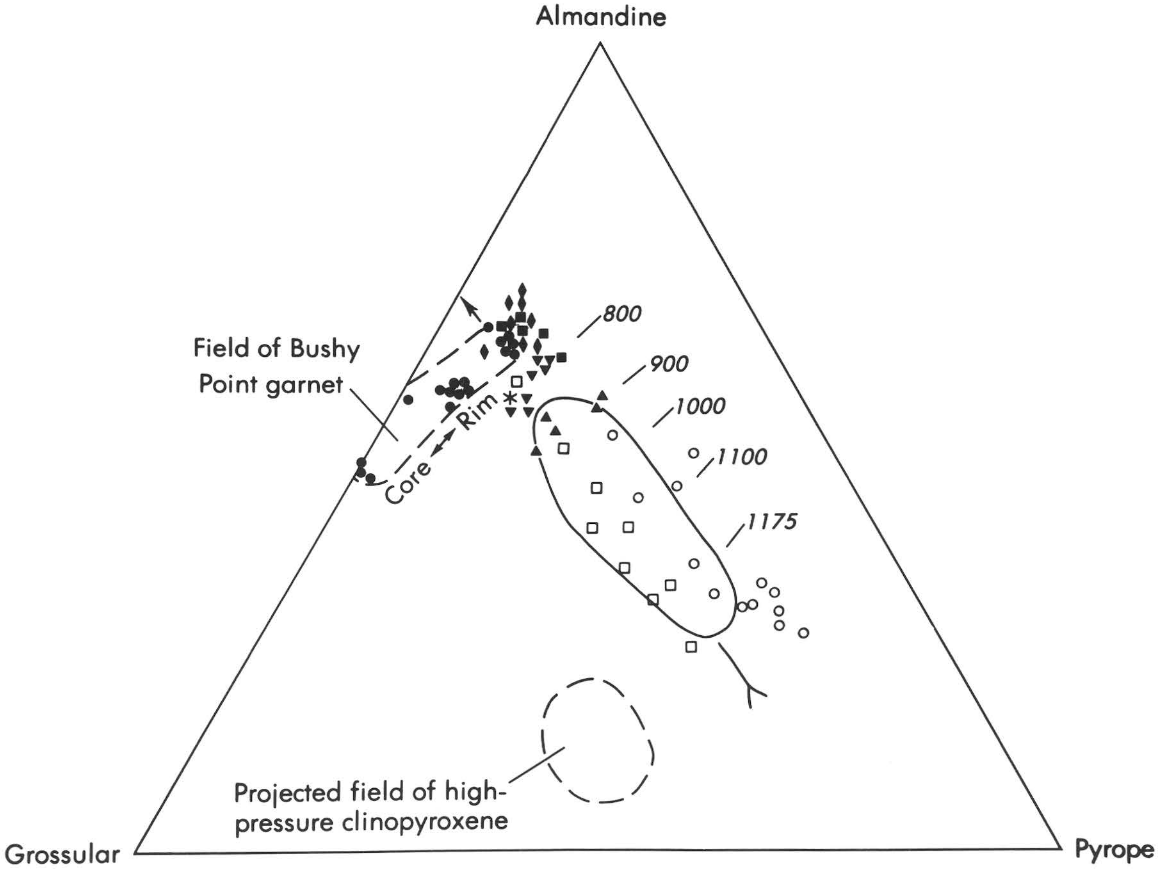

Figure 70.—Almandine-grossular-pyrope plot of garnets from the Bushy Point pluton, from eclogites and crustal xenoliths, and from experimental studies. Solid symbols represent natural garnet compositions; circles, Bushy Point garnets, this study; inverted triangles, low Mg-garnet from amphibolite inclusion, kimberlite diatreme, Montana (Hearn and McGee, 1983); asterisk, amphibolite inclusion from latite flows in Chino Valley, Ariz. (Arculus and Smith, 1979); upright triangles, Ecuador eclogite (Feininger, 1980); squares, Oregon eclogite (Ghent and Coleman, 1973); diamonds, Western Alps eclogite (Ernst and Dal Piaz, 1978). Open symbols represent garnet compositions produced in experimental studies; circles, 18-26 kbar data on andesitic and basaltic compositions (Allen and Boettcher, 1978; Allen and others, 1975); squares, 30 kbar data for tonalitic compositions (Stern and Wyllie, 1973); the arrow with temperature ticks represents Stern and Wyllie's calculated trend of garnet compositions at 30 kbar. The area defined by the solid line encloses the field of garnet (alm = almandine + spessartine) compositions produced in experiments on andesitic compositions by Green (1972). The dashed line encloses the field of clinopyroxene compositions from Allen and others (1975).

at lower pressures because of the sheathing by the plagioclase that resulted from anatectic melting. Subsequently, hornblende started to precipitate out of the magma. A possible schematic reaction for this paragenesis can be given:

$$\underset{\text{(core)}}{\text{garnet}} + \text{clinopyroxene} + \text{jadeite} + H_2O =$$
$$\text{plagioclase} + \text{hornblende} + \underset{\text{(rim)}}{\text{garnet}} + \text{quartz}$$

In balancing the equation, analysed Bushy Point mineral compositions are used for garnet and for hornblende (core of a slightly zoned, high-Al hornblende; table 23). The clinopyroxene is hypothetical and is assumed to have reacted out, its composition is taken from the experimental studies of Allen and others, (1975) (fig. 70). A jadeite component is included in the pyroxene formula; it is needed to form hornblende, and a clinopyroxene with about 20 percent jadeite component at the postulated depth is reasonable. At about 800°C and about 6-10 kbar (20-30 km) (Naney, 1983), epidote begins to crystallize from the silicate melt, together with hornblende of increasing Al content (about 2.2 Al per 23 oxygens). Then hornblende is resorbed, becoming metastable relative to epidote and biotite.

A peculiar feature of the Bushy Point rocks is the occurrence of an intragranular metamorphic

Table 23.—Microprobe data on minerals from the tonalite of Bushy Point

	Hornblende Core	Hornblende Rim	Biotite	Epidote	Garnet Core	Garnet Rim	Plagioclase Core	Plagioclase Rim
SiO_2	40.90	40.04	35.17	39.05	39.85	38.08	58.17	62.97
Al_2O_3	12.99	14.17	16.26	27.66	19.86	21.15	26.87	23.26
*FeO	19.96	21.40	23.81	6.70	22.90	18.83	0	0
MgO	7.21	6.06	8.12	0	1.43	0.53	0	0
CaO	11.32	11.22	0.02	24.20	12.49	16.82	8.43	4.37
Na_2O	1.55	1.52	0.21	0.01	0.02	0.01	6.22	7.52
K_2O	1.49	1.82	8.98	0.09	0	0	0.19	0.13
TiO_2	1.28	1.17	1.92	0.26	0.31	0.15	N.A.	N.A.
MnO	0.36	0.37	0.30	0.07	4.20	4.75	N.A.	N.A.
BaO	0	0	0.68	N.A.	N.A.	N.A.	0.08	0
F	0.10	0	0.22	N.A.	N.A.	N.A	N.A	N.A.
Cl	0.08	0.09	0.11	N.A.	N.A.	N.A.	N.A.	N.A.
Σ	97.25	97.86	95.80	**98.04	101.07	100.32	99.96	98.25
No. of oxygens for cation calculation	23		11	13	12		8	
Si	6.31	2.19	2.75	3.14	3.11	3.00	2.60	2.81
Al	2.37	2.58	1.51	2.62	1.83	1.96	1.41	1.22
Fe^{+2}	2.52	2.70	1.56	***0.43	1.50	1.24	--	--
Mg	1.66	1.40	0.95	0	0.17	0.06	--	--
Ca	1.87	1.86	0	2.08	1.05	1.42	0.40	0.21
Na	0.46	0.45	0.03	0	0	0	0.54	0.65
K	0.29	0.36	0.90	0.01	0	0	0.01	0.01
Ti	0.15	0.14	0.11	0.02	0.02	0.01		
Mn	0.05	0.05	0.02	0	0.28	0.32		
Ba	--	--	0.02					
F	0.05	0	0.05					
Cl	0.02	0.02	0.01					
				Alm	52	41	An 42	24
				Gross	33	46	Ab 57	75
				Py	6	2	Or 1	1
				Spess	10	11		

*Total iron as FeO   **Fe_2O_3 7.44; Σ 98.78   ***Fe^{+3}

texture in the plagioclase. The plagioclase has skeletal zoisite inclusions suggesting metamorphic recrystallization, yet field relations show that the pluton is post-metamorphic and intrudes the surrounding schist with a sharp cross-cutting contact. We interpret the zoisite to be the product of solid-state alteration as the plagioclase-bearing magma ascended through the crust to final crystallization at about 20-30 km depth.

If the model is correct, then the minerals in the pluton at Bushy Point give an indication of the minimum thickness of the crust about 90 m.y. ago, shortly after accretion of the Taku terrane (Jones and others, 1981). This crust must have been between 50 km and 80 km thick. The present exposure of the pluton indicates the amount of erosion and/or tectonic denudation that has occurred since then. This amount, estimated at 20-30 km on the basis of

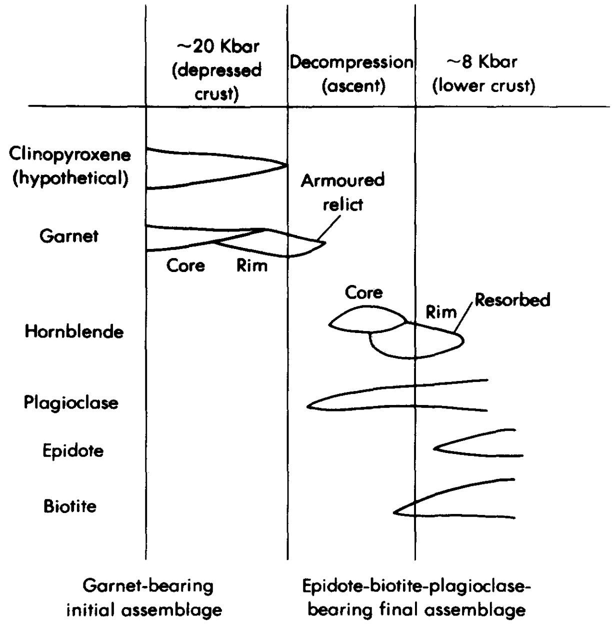

Figure 71.—Suggested paragenesis of minerals in the pluton at Bushy Point.

the magmatic epidote, is comparable to the amount of uplift and erosion estimated for the Coast Plutonic Complex in the area south of Portland Canal in British Columbia (Hollister, 1982). Our model also suggests an average paleogeothermal gradient of 10° -20°C/km, depending on whether a pressure of about 15 kbar or about 30 kbar is used for the stability of the garnet. The lower limit of 10°C/km is close to the value of the final gradient calculated by Hollister. The higher value for the gradient would be too high for the area studied by Hollister, as melting would occur at much shallower depths than the approximately 50 km postulated; however, the location of the pluton at Bushy Point, west of the coastal megalineament (Brew and Ford, 1978), might have had a different crustal structure and thermal profile. Finally, if the petrogenetic hypothesis is approximately correct, then a very large fraction of the lithospheric thickness was involved in the plutonism. If plutonism was related to subduction and possibly microplate accretion, that process is likely to have involved the entire lithosphere, and any cratonal accretion and continental growth did not result merely from the piling up of crustal thrust sheets according to the model proposed by Cook and others (1979) for the southern Appalachian piedmont.

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Reviewed by Joe Arth, Pete Toulmin, and Fred Barker.

OFFSHORE ALASKA

(Figure 72 shows areas discussed.)

GEOPHYSICAL AND GEOLOGICAL STUDIES OF THE BERING SEA SHELF

By A. K. Cooper and M. S. Marlow

During July 1982, approximately 2,900 nautical miles (nm) of geophysical data, as well as rocks from 14 dredge stations, were collected aboard the R/V S.P. LEE from the Bering Sea shelf (fig. 73). The first phase of the cruise (L9-82-BS) was a government-industry cooperative project (GICORP) with the Center for Marine Crustal Studies of the Gulf Oil Corporation. Its objectives were to collect data on deep crustal structure from seismic refraction profiles and to use these data for designing a future two-ship, multifold seismic-reflection/-refraction survey of the shelf. During this phase, 52 sonobuoys and approximately 1,500 nm of multichannel seismic-reflection, gravity, magnetic, and bathymetric data were collected. The second phase included dredging the outer edge of the continental shelf and collecting geophysical data between dredge stations.

Phase 1—Deep crustal refraction studies.—Unreversed seismic refraction profiles were recorded to offsets of 35-45 km using U.S. Navy and commercial sonobuoys. A large-volume (1,300-2,250 in^3) airgun array, fired at 50-m intervals, provided the seismic source. Preliminary analysis of the sonobuoy data has been done by the slope-intercept technique has provided crustal velocity-versus-depth information to depths of 12-14 km (fig. 74). The technique assumes that discrete layers of uniform velocity are present, an assumption that is not valid over long distances in areas of complex geology. Because only straight segments of the refraction arrivals have been used, the velocities and thicknesses of crustal layers close to the sonobuoy location are probably reliable. Future processing of the refraction data with intercept-ray parameter and ray-tracing methods will give better definition of the variations of velocity with depth and with horizontal distance along the transect. Several preliminary observations and conclusions can be made from two transects shown in figure 74:

(1) Crustal layers with high velocities (6.7-7.6 km/sec) are found at shallow depths (5-9 km) beneath large areas of the inner Bering shelf. High-velocity igneous and metamorphic rocks, such as layered gabbros, ultramafic rocks, schists, and amphibolites, occur near Hagemeister Island (transect 4-5; Hoare and Coonrad,1977). These or similar rocks may underlie the inner shelf near St. Matthew Island (transect 6-9) and be part of the igneous-metamorphic belt that is believed to extend across the shelf from Alaska to Siberia (Marlow and others, 1976).

(2) Large crustal depressions are associated wth Bristol and Navarin basins, and these depressions may be filled with as much as 6 km and 13 km, respectively, of Cenozoic and Mesozoic strata.

(3) The northern part of Bristol Basin is underlain by rocks with velocities (6.0-6.2 km/sec) similar to those for granitic rocks; these rocks may be a continuation of the granitic-plutonic belt in the Alaska Peninsula.

(4) An abrupt change in magnetic and gravity anomalies occurs along transect 6-9 on the eastern side of Navarin Basin; the change is seen in other profiles across the region and may mark the fault-controlled juxtaposition of two distinct terranes: a predominantly igneous-metamorphic terrane beneath the inner shelf (Nunivak arch) and a predominantly sedimentary terrane beneath Navarin Basin.

(5) Kuskokwim Bay is underlain by a crustal depression that is filled with at least 5 km of rocks that have velocities (4.1-5.5 km/sec) similar to those for Mesozoic volcanic and sedimentary rocks sampled in the onshore Bethel well, 150 km to the northeast. The offshore section may be a continuation of the onshore Kuskokwim Basin. Other small depressions, between Nunivak and St. Matthew Island, are suggested by undulations in the gravity data and by shallow high-velocity refractors; these depression may also contain Mesozoic sedimentary and volcanic rocks.

(6) Large magnetic anomalies of the inner shelf fall into two general categories: those of very high amplitude and high frequency (St. Matthew Island), and those of lesser amplitude and broader wavelength (Kuskokwim Bay). The first category may be caused by near-sea-floor intrusive and volcanic rocks (Hagemeister Island area), as well as volcanic flows (St. Matthew Island); the second category probably results from more deeply buried volcanic piles (Kuskokwim Bay) and intra-basement structures.

(7) The extensive volcanic flows that make up Nunivak Island appear to be confined to the island because large magnetic anomalies are not present

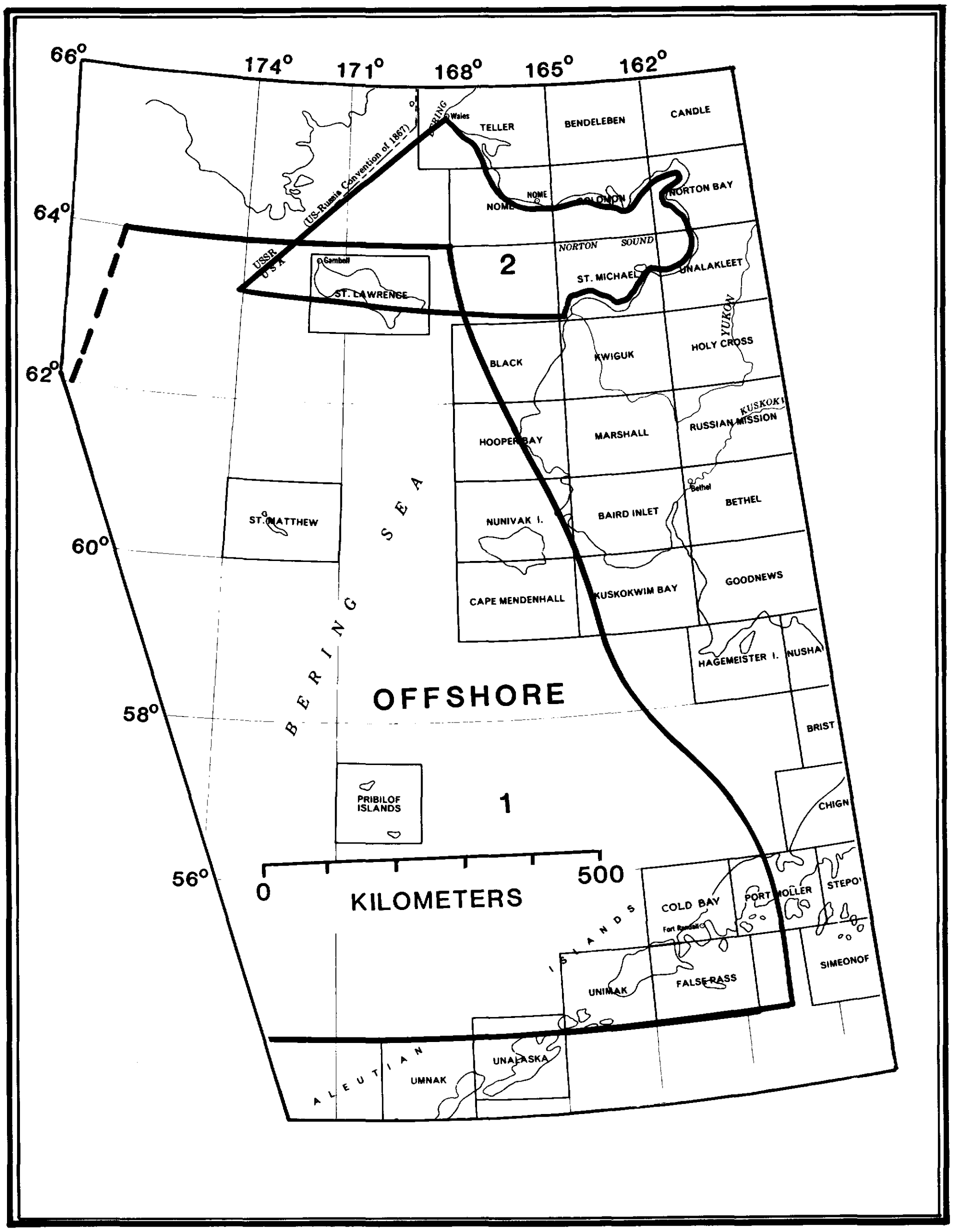

Figure 72.—Areas of offshore Alaska discussed in this section. Authors and inclusive pages of their articles are: (1) Cooper and Marlow, p. 123-127, and (2) Johnson, Mitchell, and Nelson, p. 127-133.

in the adjacent offshore areas.

Phase 2—Geological studies.—Prior to this study, dredge sampling along the northern Bering Sea shelf edge (Marlow and others, 1979) recovered Upper Jurassic and Upper Cretaceous limestone and sandstone, as well as lower Tertiary basalt, limestone, and mudstone from the acoustic basement; these rocks ae overlain by Cenozoic mudstone, sandstone, limestone, and tuff. During the 1982 cruise, dredge samples were collected at 14 nearby localities along the outer shelf edge (fig. 73). Rocks from beneath and above the prominent unconformity at the top of the acoustic basement surface were collected and analyzed.

Preliminary examination of the rock samples obtained during the 1982 cruise shows that:

(1) Limestone recovered at most dredge stations and throughout the sedimentary section is of secondary origin and was not deposited originally as carbonate sediment.
(2) A pinnacle of fresh basaltic rocks, discovered on the shelf in shallow water, cross-cuts and overlies the flat-lying sedimentary section at the shelf edge. These rocks may be syngenetic with Quaternary volcanic flows on Nunivak and the Pribilof Islands.
(3) Rocks with opaline silica (opal-CT) and quartz were recovered from the lower parts of the sedi-

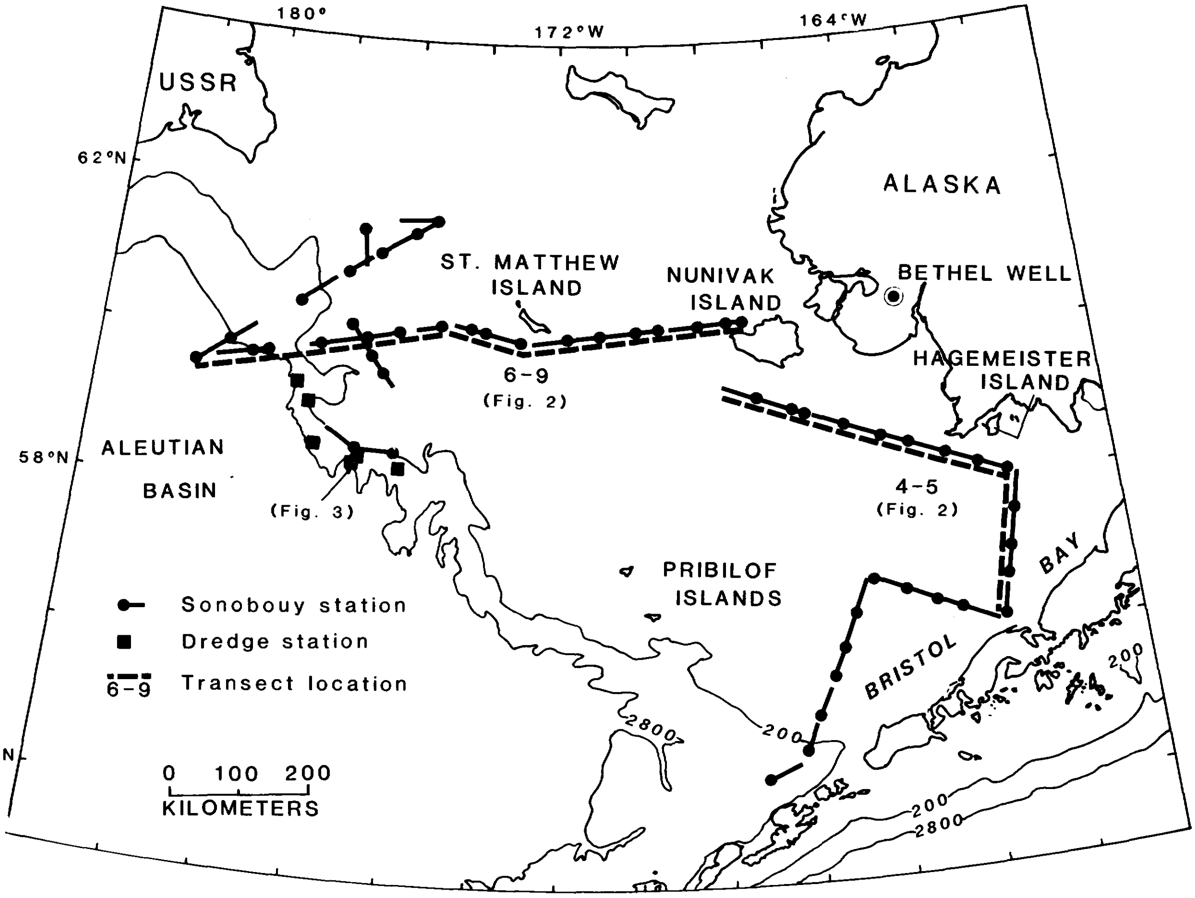

Figure 73.—Index map showing location of geophysical tracklines and geological sampling sites for cruise L9-82-BS. Water depth in meters.

mentary section and may be the result of diagenesis of the overlying diatomaceous (opal-A) mudstone and siltstone.

(4) Conglomerate, found at several sites and presumed to be from the unconformity directly above basement, may be of Oligocene age, based on its similarity to a sample recovered nearby in 1970 (Marlow and others, 1976).

(5) Volcanic and meta-igneous rocks, recovered from the basement in an area where a sample of Eocene (minimum age) basaltic andesite was obtained in 1978, do not appear to be widespread beneath the slope, based on magnetic and sample data; however, they may be more common beneath the shelf where refraction data indicate rocks having similar velocities (5.5-6.0 km/sec) are present.

(6) Correlation of refraction and geologic data at the shelf edge (fig. 75) indicates a sequence of presumably Neogene diatomaceous mudstone, sandstone, and limestone (1.6-4.5 km/sec) overlying an unconformity marked by conglomerate (4.9 km/sec) and underlain by other more highly indurated sedimentary and volcanic basement rocks (4.9-5.8 km/sec).

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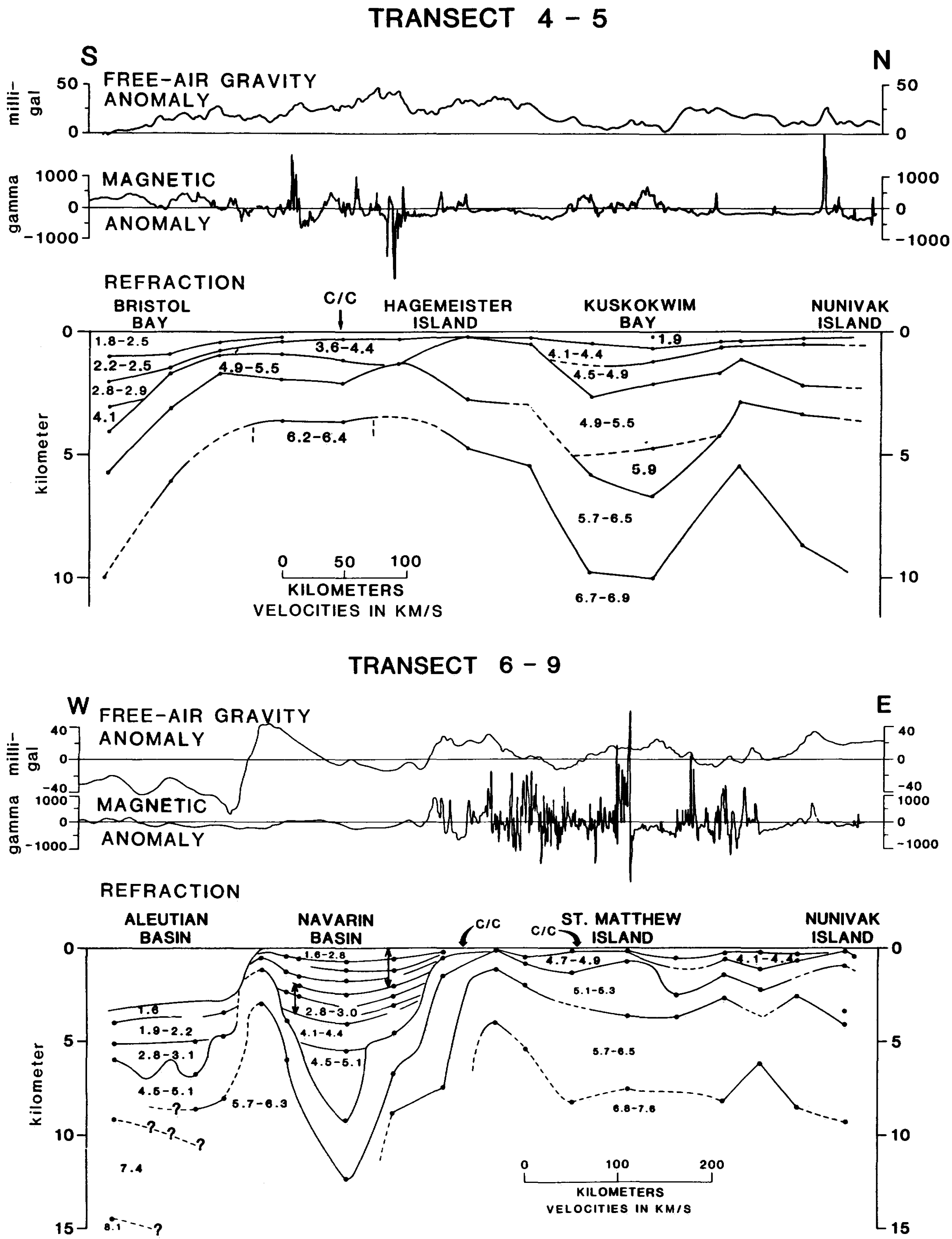

Figure 74.—Interpretation of two geophysical transects across the Bering Sea shelf that show preliminary velocities (km/sec) and thicknesses for crustal layers, based on sonobuoy refraction profiles located on figure 73.

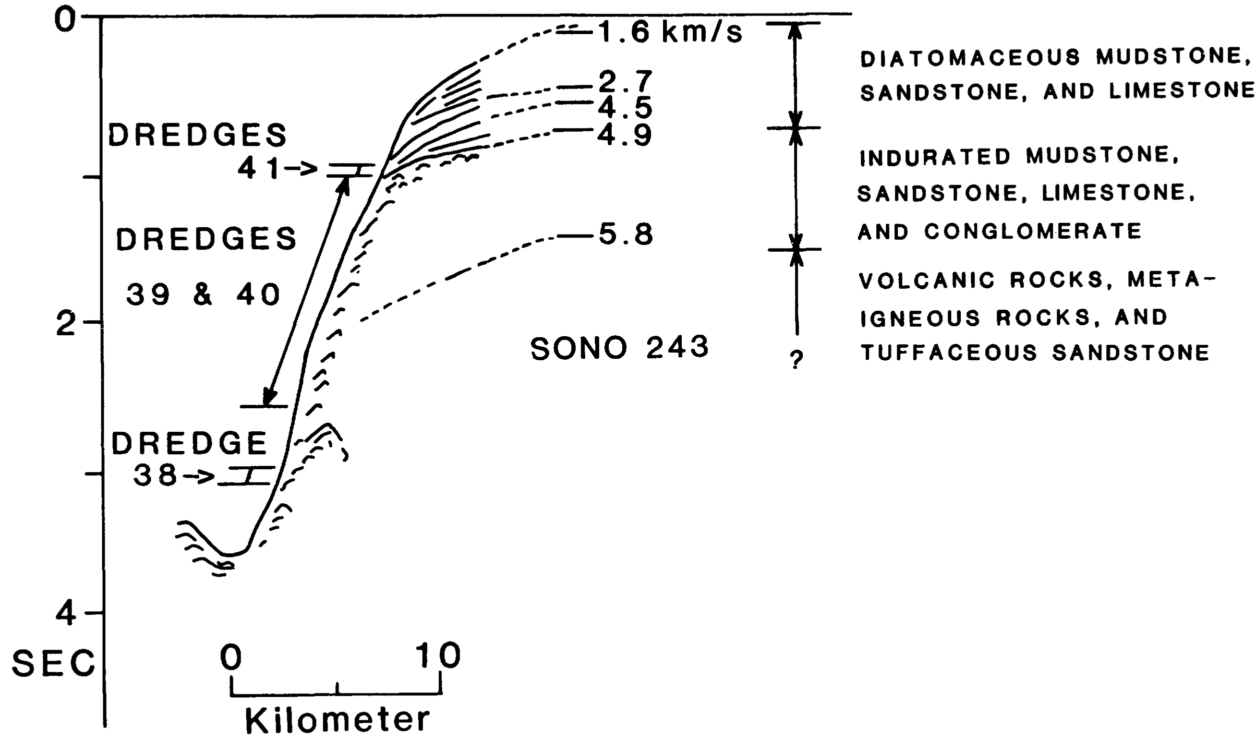

Figure 75.—Seismic-reflection profile showing rock types dredged along the profile and refraction velocities (km/sec) measured at a sonobuoy station 5 km from the end of the profile. See figure 73 for location.

GRAY WHALE FEEDING ECOLOGY AND SEA-FLOOR INTERACTION IN THE NORTHERN BERING SEA: PRELIMINARY RESULTS

By Kirk Johnson[1], Heidi-Lynn Mitchell, and C. Hans Nelson

The principal feeding grounds of the California gray whale (Eschrichtius robustus) are the Bering and Chukchi Seas. The gray whales migrate from winter breeding and calving areas in the lagoons of Baja California to this summer feeding region each year (Rice and Wolman, 1971). Here, they feed by sieving the sea-floor sediment, thus creating pits in the bottom topography. By assessing the nature and distribution of gray whale feeding pits on the sea floor and correlating them with whale sightings from aerial surveys together with the distribution and substrate affinities of the main prey species, this study will delineate the grey whale feeding grounds and will establish how the whales interact with the sea floor in the Chirikov Basin of the northern Bering Sea. Such knowledge will aid in making sound ecological decisions concerning resource exploitation on the Alaskan continental shelf.

Stomach content analyses of gray whales taken in these waters show that they feed primarily on gammaridean amphipods, mainly the tube-building ampeliscid amphipod Ampelisca macrocephala (Rice and Wolman, 1971). Although the precise mechanism of gray whale feeding is not well understood, they apparently roll on one side, mouth parallel to the sea floor, and use suction produced by the retraction of the large, muscular tongue in the mouth cavity to excavate patches of amphipod-rich sediment. The sediment is then expelled through the baleen on the opposite side of the head and the amphipods are retained (Nerini, 1984; Ray and Schevill, 1974). This feeding mechanism results in the formation of shallow oval pits (<40 cm deep) on the sea floor of the Bering Sea (Nerini and others, 1980). Similar depressions have recently been reported by S. Swartz (Univ. of California, Santa Cruz, oral commun., 1982) in the lagoons of Baja California and by J. S. Oliver (Moss Landing Marine Lab., written commun., 1982) off the coast of British Columbia.

Two types of data were used. The first consists of direct samples or bottom observations (box cores, grab samples, vibracores, underwater still photo-

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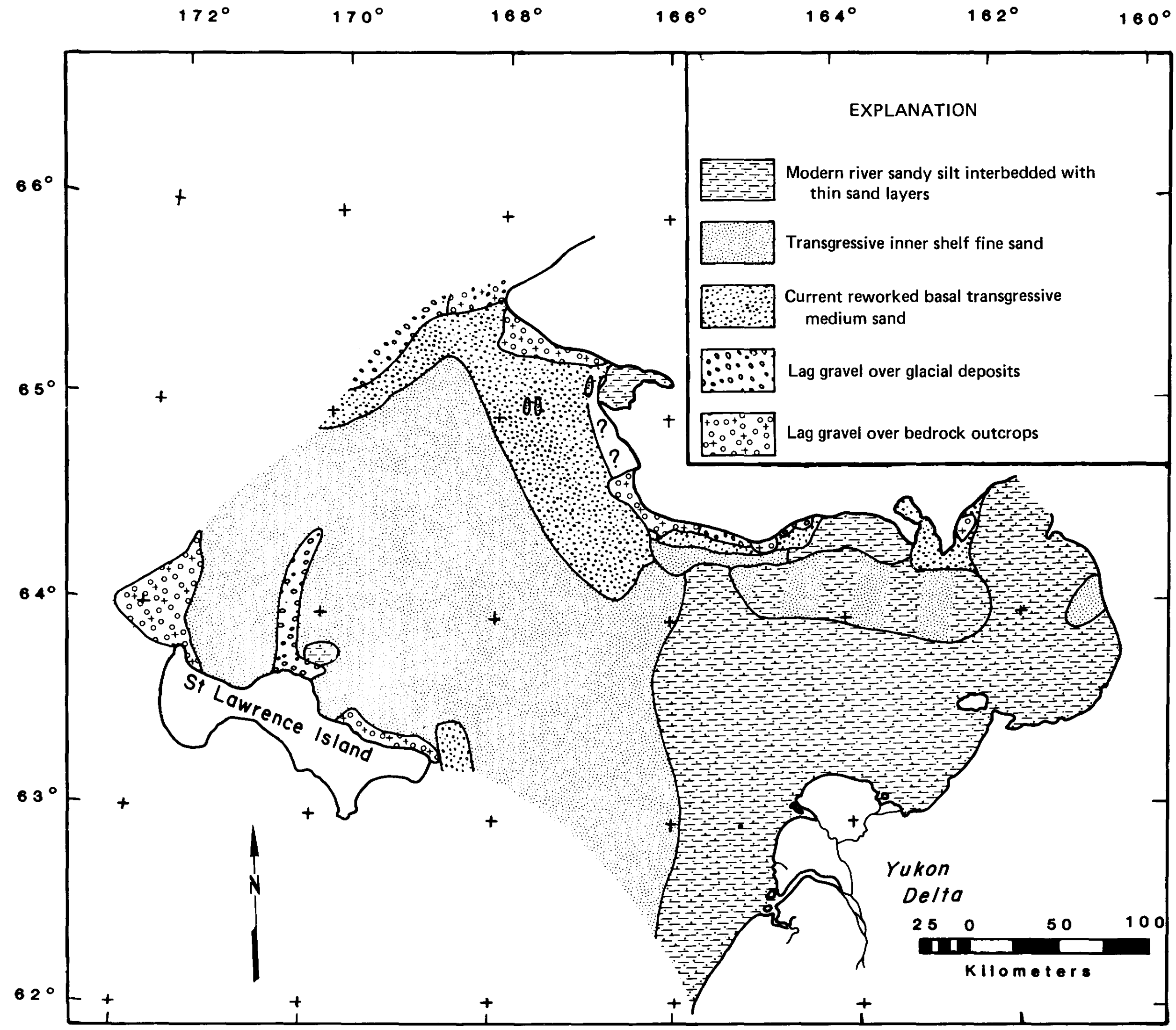

Figure 76.—Surficial geology of the northeastern Bering Shelf. In the central Chirikov Basin, transgressive inner-shelf fine sand predominates; coarser material occurs at the edges of the basin. After Nelson (1982).

graphs, underwater television, and SCUBA diver observations). Amphipod distribution data were gathered at 221 stations in Chirikov Basin by U.S. Geological Survey cruises from 1960 to 1980 and by Stoker (1978), Nerini (1980), Feder and others (1981), and Thomson (L. G. L., Ltd. [Toronto], written commun., 1983). Substrate and grain-size data were gathered from 683 stations in Chirikov Basin and Norton Sound by the Geological Survey. The second type of data is side-scan sonographs that were collected by side-scan sonar, a planographic mapping device that detects small-scale topographic relief on the sea floor. More than 4,500 line-km of side-scan data were collected by low-resolution (105 kHz) and high-resolution (500 kHz) systems. The two systems were then calibrated with each other and with measurements made by SCUBA divers. The relatively small size of the whale feeding traces places them at the lower margin of resolution of the 105-kHz system and therefore produces some margin of error. Nevertheless, the relative size and density of the different features can be compared.

Chirikov Basin was above sea level during the late Quaternary and early Holocene(?) until about 8,000-10,000 yr B.P. when the rising sea began to trangress over the limnic peats and silts of the emergent plain. Two sedimentary units were deposited: a thin (<10 cm) basal transgressive coarse sand and gravel unit, and an overlying 50-200 cm fine-grained inner shelf transgressive sand (Nelson, 1982). In Chirikov Basin, modern current patterns preclude deposition or erosion; thus, this homogeneous relict sand body has remained essen-

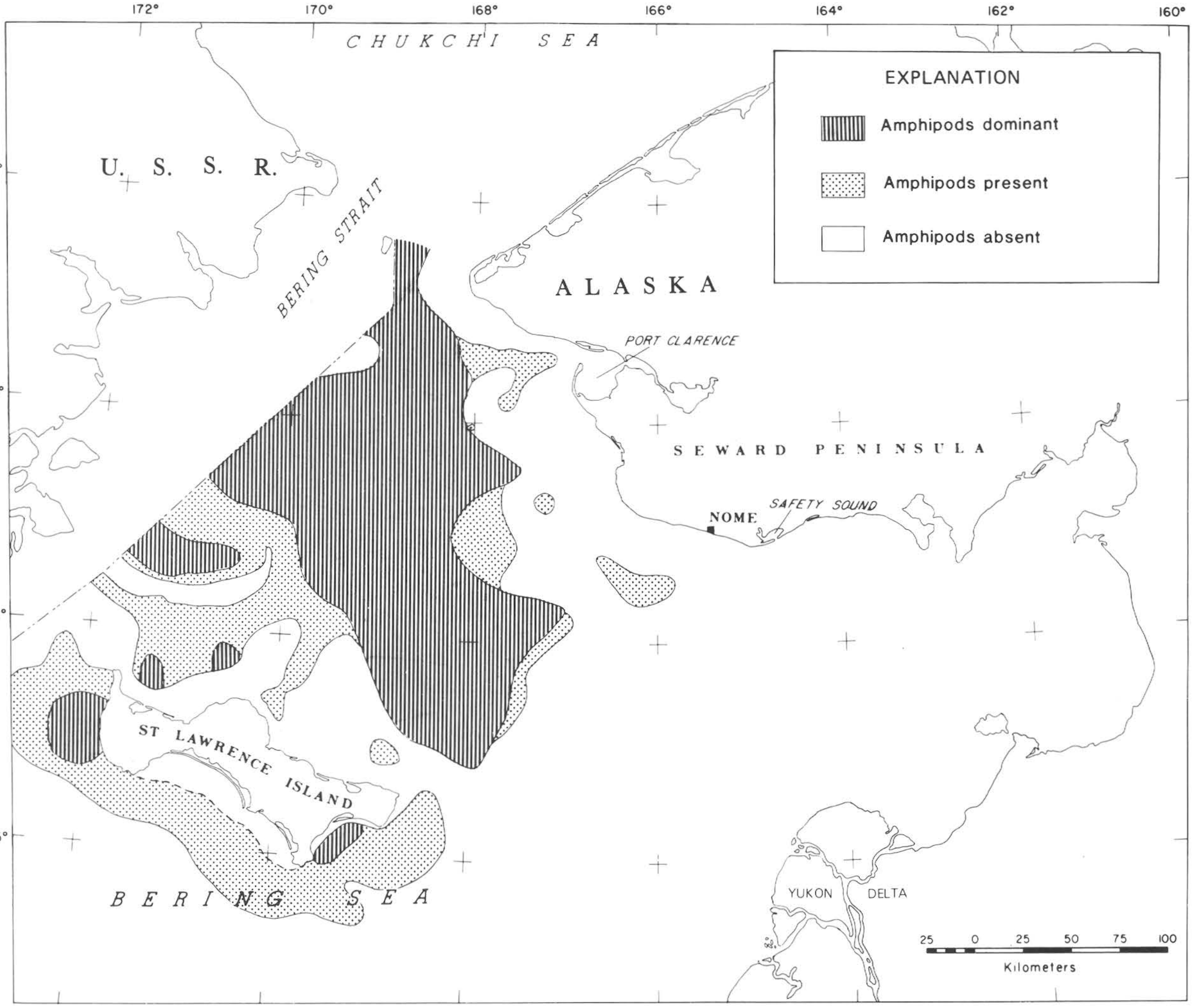

Figure 77.—Ampeliscid amphipod distribution in the Chirikov Basin. The regions of high density are in the center of the basin. Modified from Nelson and others (1981) with additional data from Stoker (1978), Nerini and others (1980), Feder and others (1981), and Denis Thomson (L. G. L., Ltd. [Toronto], written commun., 1983).

tially intact and unvaried since its deposition in the late Quaternary (fig. 76).

Sediment sample analyses, bioturbation studies, and mapping of amphipod distribution show that the area of highest amphipod concentration is closely associated with the Chirikov fine-sand body (Nelson and others, 1981; this study). As the sand body becomes coarser and more heterogeneous toward the margins of the basin, the amphipod density decreases drastically (figs. 76, 77). The fine sand [mean grain size of 3 phi (McManus and others, 1977)] provides the optimum setting for dense populations of the gray whale prey (Stoker, 1978). The dredging of the Chirikov sand body for construction of artificial drilling islands (Hess and Nelson, 1982) and the low tolerance of Ampelisca macrocephala to oil spills (Sanders, 1977) present a potential hazard for the gray whale food resource in Chirikov Basin.

During summer feeding periods, aerial surveys show the highest numbers of gray whales in the center of Chirikov Basin (Consiglieri and others, 1980; Moore and Ljungblad, 1984). This distribution closely matches that of the Chirikov sand body with its resident amphipod population. The benthic-feeding gray whales are frequently associated with sediment plumes in the water column resulting from the expulsion of sediment. Of 299 gray whales observed in the center of Chirikov Basin during early summer of 1981, 85 percent were

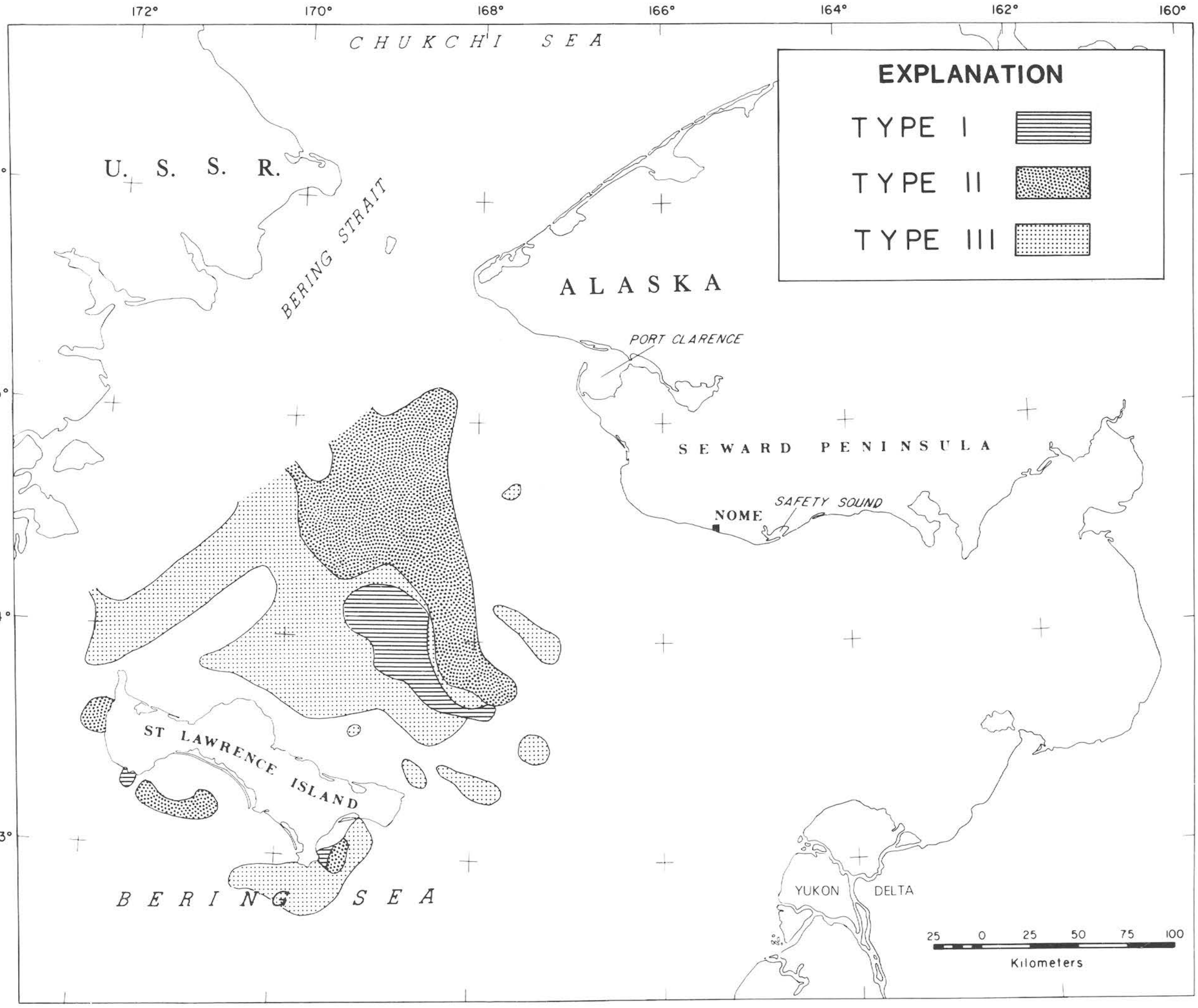

Figure 78.—Whale feeding depression distribution in the Chirikov Basin based on 4,500 line-km of sonograph records (Hess and Nelson, 1982; Larsen and others, 1980; Nerini, 1981; Denis Thomson, (L. G. L. Ltd. [Toronto], unpub. data).

associated with sediment plumes, implying extensive feeding in this area (Moore and Ljungblad, 1984).

Three types of feeding pits were described from side-scan records (table 24 and figs. 78, 79). They are interpreted as gray whale feeding traces because of their size, arrangement in ordered groups, and locations in areas of heavy whale feeding pressure and high prey concentrations. Type I features occur in groups, representing multiple suction feeding events, and are similar to fresh features described by Nerini (1984); they are associated with current-scour enlarged pits. Type II features are elongate pits that represent a combination of fresh feeding and current modification. Type III features are oval pits. Type I and II features are located in the center of the basin and correlate closely with areas of amphipod dominance (figs. 77, 78). Type III features occur along the margins of the basin where amphipods are present but less common.

Type I and Type II features are clearly produced by intensive whale feeding, whereas Type III features are of more questionable origin. The increasing coarseness of sediment along the basin margin may physically control pit morphology, as well as providing only a marginal amphipod habitat; this in turn may cause sporadic whale feeding behavior. However, non-whale origins of Type III features, such as gas expulsion craters (Nelson and others 1980), are unlikely in this area due to the absence of the surficial mud necessary to entrap biogenic gas (fig. 76). The Pacific walrus (Odobenus rosmarus) also feeds on benthic infauna, but walrus exploit different prey and their feeding pits are

Table 24.—Definition of the three types of whale feeding pits observed from side-scan sonographs; n = 1,027

Type	Description and origin	Average length (meters)	Average width (meters)	Average length/ width ratio	Shape class	Average percentage of sea-floor disturbed: Fresh and modified features	Average percentage of sea-floor disturbed: Fresh features
I	multiple suction feeding pits with current-enlarged pits	4.6	1.9	2.5	elliptic	12	2.5
II	fresh and partially tially modified fied elongate pits	4.1	1.3	3.2	narrow elliptic	10	4.1
III	oval pits	2.4	1.3	1.9	wide elliptic	5.7	2.1

much smaller than those of the gray whale (J. S. Oliver, Moss Landing Marine Lab., written commun., 1982).

The percentage of sea floor disturbed by feeding pits and by current-scour enlarged pits averaged 12 percent in Type I areas and 10 percent in Type II areas. The precentage of sea-floor disturbance in Type III areas averages 5 percent. A more accurate estimate of whale feeding pressure can be obtained by separating the pits into four size classes by area. Thus, fresh whale feeding pits can be isolated from the current-scour enlarged pits. Using 5.3 m^2 as an upper limit for the area of a fresh pit, the average percentage of sea-floor disturbance by recent whale activity can be determined for each of these areas. For Type I features, the average is 2.5 percent, for Type II features 4.1 percent, and for Type III features, 2.1 percent (table 24). Even though the rates of feature modification and amphipod generation are unknown, these numbers provide a first approximation of the degree to which the whales are using their food resource in Chirikov Basin.

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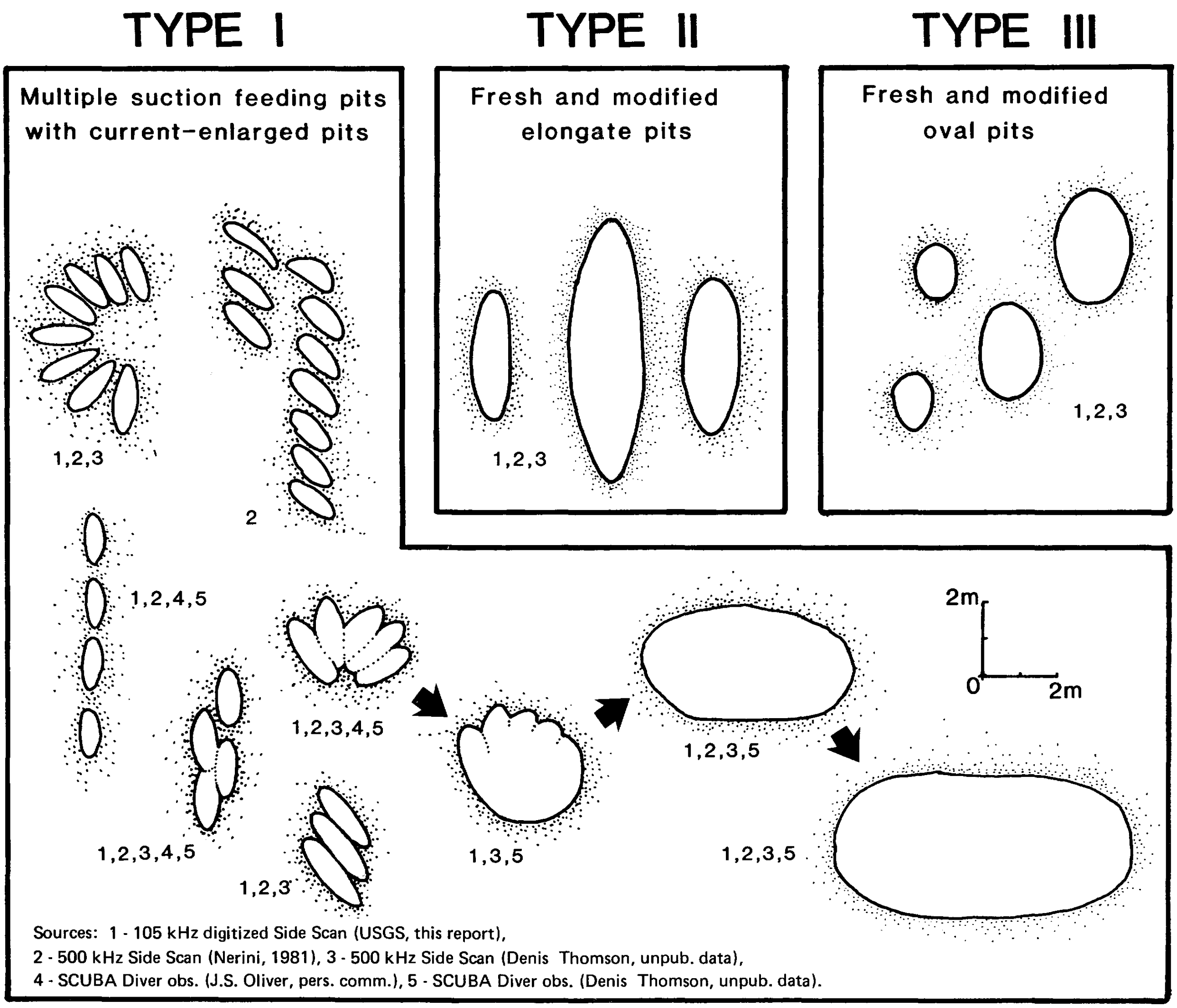

Figure 79.—Three types of whale feeding pits indentified through side scan sonar in Chirikov Basin.

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LISTING OF CROSS REFERENCES BY SUBJECT MATTER

Geochemistry

Geochronology

Geophysics

Quaternary geology

Rock alteration; diagenesis

Structural geology; stratigraphy

Structural geology; stratigraphy--Continued

Tectonics

Tectonics--Continued

Volcanology

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