DALTON HIGHWAY, YUKON RIVER TO PRUDHOE BAY, ALASKA

Bedrock geology of the eastern Koyukuk basin, central Brooks Range, and eastcentral Arctic Slope

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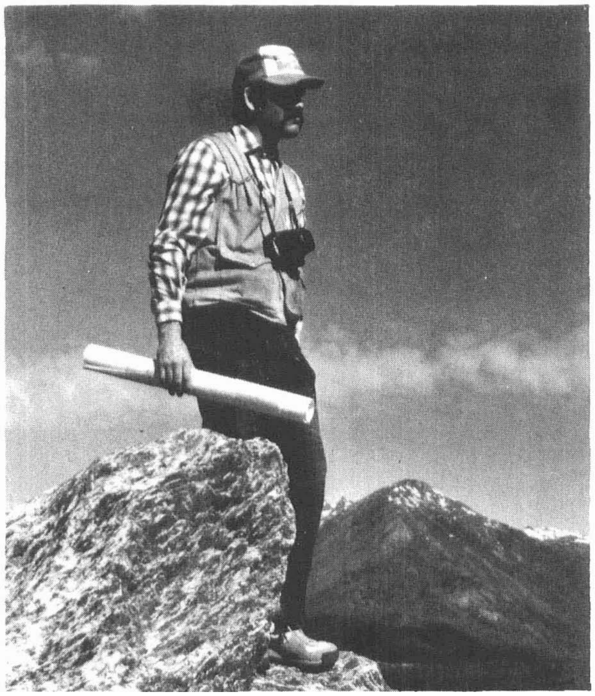

Dedication

Dr. John Thomas Dillon
(1947-1987)

John Dillon spent the **1987** field season in the mountains of the Arctic National Wildlife Refuge with his father, Steven Patrick Dillon, who served as John's field assistant for the summer. John finished his field work in late July, loaded his plane, and left the northeastern Brooks Range bound for Fairbanks. When his plane failed to reach Fairbanks on schedule, a massive search effort was conducted by members of the Civil Air Patrol, U.S. Air Force, Alaska Division of Geological and Geophysical Surveys, ARCO Alaska, Inc., U.S. Geological Survey, and the University of Alaska. John and his father were found by his colleagues seven days later at a crash site in the Brooks Range.

John came to Alaska in **1977** to join the geological staff of the Alaska Division of Geological and Geophysical Surveys and was appointed affiliate professor of geology at the University of Alaska. John soon fell in love with the Brooks Range and spent most of his Alaska career studying the geochronology and tectonic history of the mountain system. John tackled the complex 'schist belt' of the southern Brooks Range and rocks of the Angayucham and Ruby terranes to the south. His reports and maps summarize the ten years he explored the region, making significant contributions to our understanding of the evolution of the Brooks Range and its adjacent terranes.

John Dillon could be as inspiring and as intense as the mountains he wandered. He was unstoppable in the field. His 20-hour field days and late-night traverses became the subject of campfire stories told with dismay by his graduate students. When funds for helicopter support of his projects were low, John compensated by using imported South American llamas as pack animals. Among the constant dangers we all faced was the threat of being trapped by John's intense gaze--from which there was no escape--and being inflicted with hours of nonstop Angayucham tectonics. The trap was always tolerable. There was no doubt this man loved what he was doing and was having great fun. It was contagious.

John's colleagues have formally proposed to the U.S. Board on Geographic Names that a prominent peak along the Dalton Highway in the southern Brooks Range be named Dillon Mountain. The mountain is a fine monument to John's memory. It is a part of the land he loved, a remnant of an ancient seafloor thrust to prominence within a land of prominent mountains, close to the spirits of Doonerak and Angayucham. John would like that. To those of us who walked this country with John, he remains forever young, forever a part of the mountains, capable of touching and inspiring beyond his lifetime.

This volume contains the last of John's writings on the geologic evolution of the southern Brooks Range and adjacent regions. He contributed substantially to the production of this guidebook, and it was among his final projects. For his contributions to Alaska geology, his scientific insight and honesty, and for the ways he touched our lives, we dedicate this volume to the memory of John Thomas Dillon.

R. Keith Crowder

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CHAPTER 10.

STRUCTURE AND STRATIGRAPHY OF THE SOUTHERN BROOKS RANGE AND NORTHERN KOYUKUK BASIN NEAR THE DALTON HIGHWAY

By J.T. Dillon[1]

INTRODUCTION

The southern Brooks Range and northern Koyukuk basin are divided into four fault-bounded tectono-stratigraphic or lithotectonic terranes: the Ruby, Mosquito, Angayucham, and Arctic Alaska terranes (Jones and others, 1981; Silberling and Jones, 1984, Jones and others, 1987). The structurally highest lithotectonic terrane is the upper Paleozoic and lower Mesozoic oceanic Angayucham terrane, which was thrust onto the continental rocks of the Ruby, Mosquito, and Arctic Alaska terranes. These terranes are here subdivided into 13 east-west-trending, fault-bounded belts 10 to 30 km wide (fig. 139; sheet 2). The term 'fault panel' is adopted here for these fault-bounded regions or belts. Fault panels encompass areas of varying stratigraphy, structural style, and metamorphic grade and represent subdivisions of the lithotectonic terranes. Most fault panels are elongated east-west and are bounded by thrust faults, but the panels of the Angayucham terrane are bounded on the south by high-angle, strike-slip faults, Most of the fault panels were probably formed during Jurassic to Neocomian convergence and thrusting, but some probably developed during Late Cretaceous plate rotation and strike-slip faulting.

Sheet 2 (in pocket) is a gravity and aeromagnetically modeled geologic cross section that parallels the Dalton Highway in the Middle Fork Koyukuk and Dietrich River valleys in the Wiseman and Charidalar Quadrangles of the southern Brooks Range. The cross section illustrates the generalized geologic relationships along the Dalton Highway. It is based on recent DGGS and U.S. Geological Survey 1:63,360-scale mapping near the Dalton Highway and 1:250,000-scale mapping of the Wiseman and Chandalar Quadrangles (Brosgé and Reiser, 1964, 1971; DeYoung, unpublished mapping, 1978; Dillon and others, 1986). The calculated gravity and aeromagnetic signature of the cross section (sheet 2) corresponds closely to observed gravity and aeromagnetic profiles (Smith and Dillon, chap. 12). The section shows detailed geology of the southern end of generalized cross sections illustrated in figure 26 and on sheet 1.

The 'Geologic Overview' section below summarizes, from south to north, the regional geologic features exposed in each fault panel. The subsequent three sections focus on details of the terranes, structure, and metamorphism in the southern Brooks Range. The last section details the stratigraphy of some of the individual fault panels.

ACKNOWLEDGMENTS

This chapter is based primarily on cooperative geologic investigations by DGGS, the U.S. Geological Survey, and students and faculty of the University of Alaska Fairbanks, the University of California at Santa Barbara, California State University at Los Angeles, and Rice University at Houston. Although most of the geologic mapping by DGGS and the U.S. Geological Survey has been completed, many of the analytical studies are still in progress; therefore, the results presented here are preliminary.

The principal data are from eighteen 1:63,360-scale quadrangles in the Wiseman and Chandalar 1:250,000-scale quadrangles mapped by the author and others from DGGS since 1977. Much information came from exten-

[1]Deceased; formerly of DGGS, Fairbanks. This manuscript, although not completed at the time of the author's untimely death in July 1987, includes some of Dillon's most recent thinking on Brooks Range structure and stratigraphy. DGGS is pleased to publish this substantial contribution as Chapter 10 of this guidebook.

sive cooperative research in the Wiseman Quadrangle with members of the U.S. Geological Survey, particularly W.P. Brosgé, J.T. Dutro, Jr., T.E. Moore, and D.L. Jones. Fossils were identified by Dutro, Jones, A.G. Harris, K.J. Bird, N.J. Silberling, P. Quintero, and W.A. Oliver, Jr., of the U.S. Geological Survey Branch of Paleontology and Stratigraphy; by A.R. Palmer of the Geological Society of America; and by C.J. Smiley of the University of Idaho, among others. The fossils provide the age control for protoliths of the metasedimentary rocks in the Brooks Range near the Dalton Highway. Discussions of the geochemistry of the Wiseman Quadrangle by J.B. Cathrall, J.D. Hoffman, and Jack Antweiler of the U.S. Geological Survey have greatly improved my understanding of the economic geology of the area. I am also indebted to J.W. Cady and W.B. Hamilton (U.S. Geological Survey) for consultations on their geophysical and geologic interpretations of the area. Finally, numerous discussions with I.L. Tailleur (U.S. Geological Survey) have contributed to my understanding of the regional geology of the Brooks Range.

Many geologists from DGGS contributed to this paper. G.H. Pessel and C.G. Mull introduced me to the regional geology of the Brooks Range. J.D. Blum, D.N. Solie, L.C. Nicholson, J.A. Huber, K.K. Lamal, D.D. Adams, P.A. Adler, K.E. Adams, Mull, E.E. Harris, D.J. Gentry, and G.L. Myers participated in the mapping and petrologic studies of areas near the Dalton Highway. During 1984-87, students M.J. Kelly, Gentry, K.L. Crowder, and Lamal volunteered their services to DGGS to study the geology and petrology of the Brooks Range.

Most of what is known about the timing and conditions of emplacement of igneous rocks in the south-central Brooks Range comes from cooperative studies by the author and J.D. Blum with geologists at the above four universities. At the University of Alaska Fairbanks, D.L. Turner, D.D. Adams, R.J. Newberry, Nicholson, and Huber studied igneous rocks and related thermally metamorphosed rocks and vein deposits. G.R. Tilton, S.B. Mukasa, and J.A. Chen of the University of California provided the Nd-Sm and U-Pb analyses of zircons. T.E. Davis of California State University provided Rb-Sr isotopic analyses and interpretations. A.P. LeHuray and A. Zindler of Columbia University provided Nd-Sm isotopic analyses.

Structural and stratigraphic studies by H.G. Avé Lallemant, J.S. Oldow, R.R. Gottschalk, J.S. Phelps, F.E. Julian, G.M. Seidensticker, and others from Rice University are yielding new interpretations of the deformational history of the rocks near the Dalton Highway.

GEOLOGIC OVERVIEW

The rocks of the southern Brooks Range, northern Koyukuk basin, and Ruby geanticline in the Wiseman and Chandalar Quadrangles range in age from Proterozoic(?) to Cretaceous and in metamorphic grade from completely unmetamorphosed rocks to possibly polymetamorphic gneiss and schist of the amphibolite facies. (These rocks are depicted on fig. 139, sheet 2, or chap. 7.) The stratigraphy, structural style, and metamorphic grade are consistent within individual fault panels. Stratigraphy and metamorphic grade vary slightly between adjacent fault panels in the same lithotectonic terrane but vary greatly between panels in different terranes.

The Ruby terrane of the Ruby geanticline is composed of Proterozoic(?) and lower Paleozoic metasedimentary rocks with continental protoliths and Cretaceous granitic plutons and batholiths (Patton, chap. 4). Mid-Cretaceous granites intrude through the Ruby terrane into the Angayucham terrane and thrust system, stitching them together; they provide the upper limit for the timing of thrusting and amalgamation of the Ruby and Angayucham terranes. Upper Triassic and Lower Jurassic(?) radiolarian cherts in the Angayucham terrane provide a lower limit to the time of thrusting.

Fault panel 12 is the southernmost exposed panel in the map area; this panel is part of the northern Ruby geanticline (Ruby terrane) and structurally underlies rocks of fault panel 11. Fault panel 12 is composed of Proterozoic(?) to Paleozoic schist intruded by mid-Cretaceous granite plutons (Blum and others, chap. 11; Patton, chap. 4).

The Ruby and Arctic Alaska terranes may once have been contiguous, for they both have similar pre-Mesozoic continental metasedimentary and metaigneous rocks. However, there are distinct differences, particularly the thick, extensive Devonian carbonate rocks in the Arctic Alaska terrane; in the Ruby terrane, Devonian carbonate rocks are comparatively rare, thin, and discontinuous (Brosgé and others, 1973; W.W. Patton, Jr., oral commun., 1984). The Ruby terrane is widely intruded by mid-Cretaceous granitic plutons, which the Arctic Alaska terrane apparently lacks. Direct comparisons are difficult because exposures in the Ruby terrane are poor, and exploration and mapping has lagged behind that conducted in the well-exposed, economically well-endowed Brooks Range.

Rocks of the Mosquito terrane (fault panel 13) are exposed in a fenster 30 km east of the Dalton Highway between the South Fork and Malamute fault zones on the south side of the Brooks Range. Mosquito terrane rocks are not exposed along the Dalton Highway, where they are probably at a depth of about 5.5 mi (9 km), buried beneath rocks of fault panels 10 and 11 (sheet 2). Protolith ages for amphibolite-facies quartz schist, amphibolite, and granitic gneiss of fault panel 13 are thought to be the same as those of lithologically similar rocks in the Brooks Range and Ruby geanticline (fault panels 6, 8, 9, and 12). Because protoliths are similar to those of the Ruby and Arctic Alaska terrane, the Mosquito terrane is presumed to be an offset part of either the Arctic Alaska terrane or Ruby terrane within the Malamute-South Fork fault system.

The youngest and least metamorphosed rocks in the area are in the Angayucham terrane, located in the southern spurs of the Brooks Range and the northern Koyukuk basin (fault panels 10 and 11). A major fault zone, the Angayucham fault system, or Kobuk suture zone (fig. 26; sheet 1; Mull, 1982; Mull and others, 1987a) separates these rocks from older metamorphic

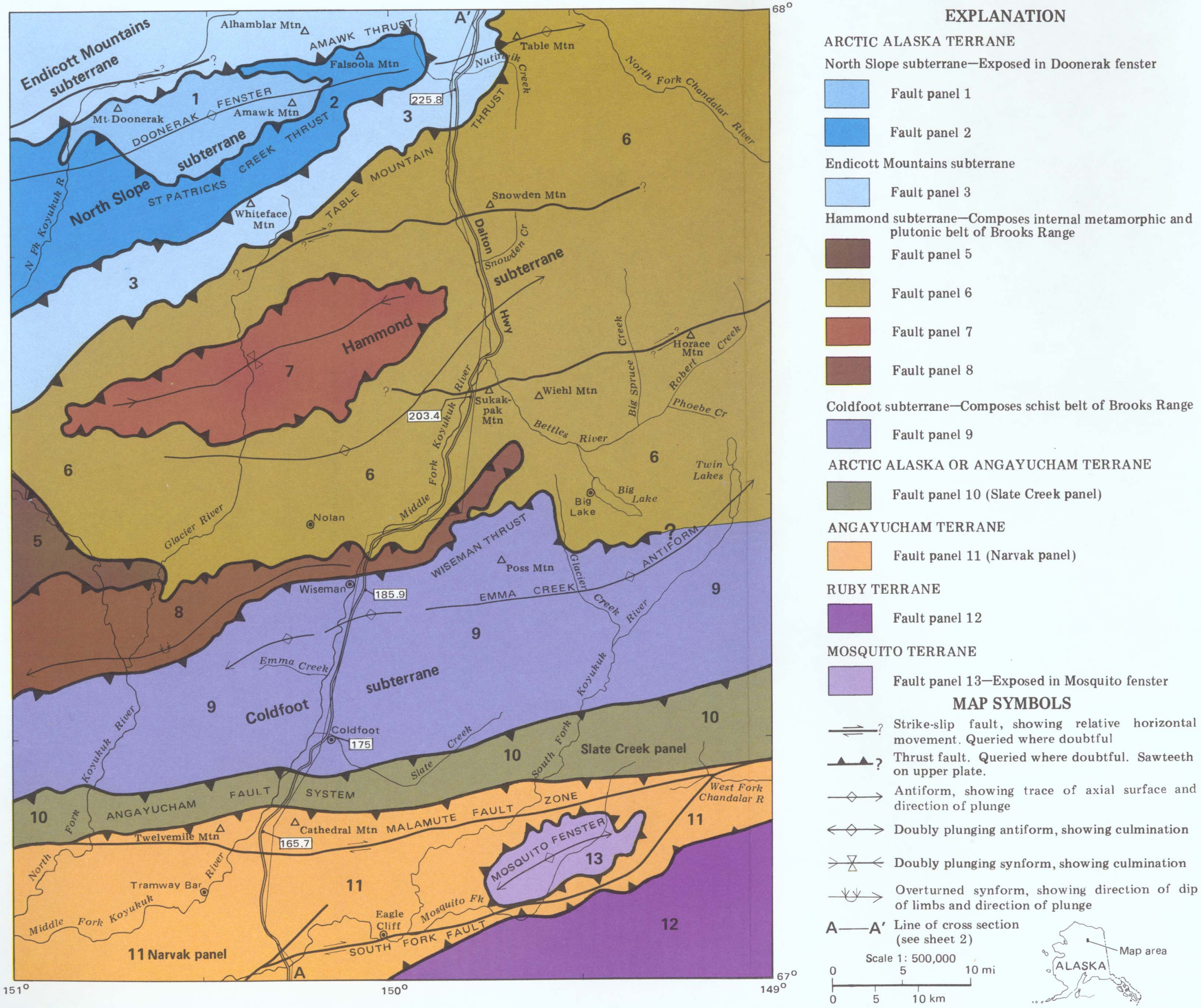

Figure 139. Map of fault panels and lithotectonic subterranes of northern Koyukuk basin and southern Brooks Range.

rocks to the north. At this boundary, the dominantly oceanic rocks of fault panels 10 and 11 are thrust over the southern edge of continental rocks of the Brooks Range.

The Angayucham terrane is composed principally of oceanic rocks that include diabase, pillow basalt, tuff, chert, graywacke, and serpentinized peridotite. It underlies a large part of northern Alaska to the south of the Brooks Range and is correlated with ophiolitic klippen in the northwestern Brooks Range, with the Christian Complex in the eastern Brooks Range, and with the Innoko and Tozitna terranes and Rampart Group of the southern Yukon and Koyukuk basins (Roeder and Mull, 1978; Jones and others, 1987). In most of these areas, geologic maps and supplementary geophysical data show that, regionally, the Angayucham terrane is a dismembered and structurally reversed ophiolite composed of three fault panels.

Upper Paleozoic to Mesozoic rocks of fault panels 10 and 11 are divided into four thrust slices that occur in the same sequence throughout the southern Brooks Range. The lower two slices, which constitute fault panel 10, are composed of metamorphosed phyllite and graywacke and are called the Slate Creek panel by Patton and Box (1989). They are overlain by fault panel 11, composed of a lower panel, the Narvak panel, which consists of pillow basalt and chert (Angayucham volcanic rocks) and an upper thrust panel, the Kanuti panel, which consists of metagabbro and ultramafic rocks (Patton and Box, 1989). Collectively, fault panels 10 and 11 are called the Angayucham terrane (sheet 2). Probable protoliths of these four panels were slope-rise sediments, which form the lower two panels, and oceanic basalt and chert and oceanic lower crust and mantle, which form the upper panel; together these lithologies typify oceanic crust.

During mid-Mesozoic plate convergence, the oceanic rocks of fault slices 10 and 11 were obducted as four regional thrust panels onto continental rocks of both the Ruby terrane in the Ruby geanticline and the Arctic Alaska terrane in the Brooks Range. The Angayucham thrust system and oceanic terrane are named for their occurrences in the Angayucham Mountains and Cosmos Hills, where the Angayucham volcanic rocks were recognized as a thrust plate by Fritts (1970). Thrusting of the Angayucham terrane onto the Arctic Alaska terrane caused imbricate thrusting, northward-verging folds, and tectonic-burial metamorphism within the lower plate in the Brooks Range.

The phyllite and graywacke panel (Slate Creek panel) was not included in the Angayucham terrane by Mull (1982), Hitzman and others (1982), Silberling and Jones (1984), or Mull and others (1987a), all of whom mapped the Angayucham terrane boundary as the distinct lithologic contact between mafic and ultramafic igneous rocks and the southern edge of the phyllite-and-graywacke belt. However, I include the phyllite-and-graywacke thrust panel in the Angayucham terrane because of metamorphic and lithologic similarities with the Angayucham-terrane metavolcanic rocks and dissimilarities from the schists of the Arctic Alaska terrane to the north. Similar relationships have been found along the south side of the central Brooks Range and the west side of the Ruby geanticline by Patton (oral commun., 1985). The phyllite and graywacke thrust panel is separated from the Arctic Alaska terrane by a mylonite zone at the north strand of the Angayucham thrust system.

In the Koyukuk basin, unmetamorphosed coarse marine and nonmarine conglomerate (molasse) of Albian age unconformably overlies the fault panels of the Angayucham terrane. Because clasts in the conglomerate record the progressive erosional stripping of the thrust panels of the Angayucham terrane during isostatic(?) rebound of the Brooks Range, this Albian conglomerate provides a minimum age of thrusting.

After thrusting and metamorphism and subsequent deposition of Albian molasse, rocks of the southcentral Brooks Range were folded around east-northeast-trending, doubly plunging axes. These folds formed during uplift following thrusting, during strike-slip faulting on the Malamute, South Fork, and related faults, or during both events.

The Arctic Alaska terrane composes most of the Brooks Range and extends from the north strand of the Angayucham thrust system north to the Arctic Ocean. The terrane is composed primarily of continental sedimentary and metasedimentary rocks of Proterozoic(?) through Mesozoic age with especially thick and varied Devonian strata. Silberling and Jones (1984) and Jones and others (1987) divided the Arctic Alaska terrane into several subterranes, each with regionally similar Paleozoic stratigraphy with distinct local variations. Minor modifications of their boundaries have been made to fit the geology shown in figure 139. Especially significant is the recurrence in most of the subterranes of the Hunt Fork Shale, whose distribution seems to preclude post-Devonian plate boundaries within the Arctic Alaska terrane.

The Arctic Alaska terrane is composed of four subterranes: Coldfoot, Hammond, Endicott Mountains, and North Slope subterranes. The Coldfoot subterrane, which is the structurally highest and southernmost of the four subterranes, includes Proterozoic(?) to Devonian polymetamorphic metasedimentary rocks of the Brooks Range schist belt (fig. 140); they are the most highly metamorphosed and probably the oldest rocks within the map area (fault panel 9). These Proterozoic(?) to lower Paleozoic rocks are intruded by Devonian(?) granitic and mixed felsic-mafic intrusive complexes and are overlain locally by Devonian bimodal volcanics. The rocks of the schist belt (Coldfoot subterrane) were thrust northward onto lower Paleozoic through Upper Devonian rocks of the Hammond subterrane of the Arctic Alaska terrane. The thrust faults are probably imbrications of the lower plate formed during emplacement of the Angayucham terrane.

Paleozoic greenschist-facies metasedimentary and metavolcanic rocks of the Hammond subterrane (fault panels 5 through 8) include Cambrian calc-schist and marble and Ordovician pelitic schist and graphitic marble. These lower Paleozoic rocks are unconformably overlain by Lower to Middle Devonian siliciclastic and volcaniclastic rocks that are in turn overlain in sequence by thick Devonian marble and limestone of the Skajit Limestone (fig. 141), upper Middle Devonian pelitic, calcareous, and siliceous clastic rocks of the Beaucoup Formation, and Upper Devonian Hunt Fork Shale. Prior to metamorphism, all these rocks were intruded by Devonian or Jurassic mafic plutons and by Devonian granitic and bimodal plutons.

Fault panel 3 is the Endicott Mountains subterrane of the Arctic Alaska terrane. The Endicott Mountains subterrane consists mainly of phyllitic, lower greenschist-facies Devonian felsic volcanic and volcaniclastic rocks interlayered with siliceous, calcareous, and pelitic metasediments that are unconformably overlain by the Upper Devonian Hunt Fork Shale. Undifferentiated nonfossiliferous lower Paleozoic rocks probably occur locally within fault panel 3 on the south side of the Doonerak fenster, but no fossils have been found to support this supposition. The Hunt Fork Shale is continuously exposed from the north side of the Doonerak fenster, where it forms part of a major thrust plate in the northern Brooks Range called the Endicott Mountains allochthon (Mull, 1982), around the east plunge of the fenster to its south side. The stratigraphy and structure of the Endicott Mountains allochthon and the Doonerak fenster are discussed in greater detail by Mull (1982; chaps. 5 and 6) and Mull and others (1987b; chap. 14).

Fault panels 1 and 2 encompass the North Slope subterrane of the Arctic Alaska terrane. These phyllitic lower greenschist- and pumpellyite-prehnite-facies lower Paleozoic to Triassic sedimentary and volcanic rocks are exposed in the Doonerak fenster and are stratigraphically and lithologically similar to the basement rocks of the North Slope. Structurally, these are the lowest rocks mapped in the area.

Parautochthonous lower Paleozoic rocks of the Doonerak fenster include black siliceous phyllite and quartzite with thin, fossil-bearing, Cambrian and Ordovician limestone interbeds and Ordovician mafic to intermediate volcanic and volcaniclastic rocks. The lower Paleozoic rocks are unconformably overlain by an upper Paleozoic to lower Mesozoic section with Lower Mississippian Kekiktuk Conglomerate at the base. The Kekiktuk is overlain conformably in sequence by Mississippian Kayak Shale, Mississippian to Lower Pennsylvanian Lisburne Group limestone, Permian Sadlerochit Group (Echooka Formation) sandstone, Triassic Shublik Formation limestone and calcareous shale, and Triassic Karen Creek Sandstone. This sequence is widespread in the northeastern Brooks Range and the subsurface of the Arctic Slope but occurs at only two localities south of the drainage divide of the Brooks Range (Mull and others, 1987b; chap. 14). The Doonerak fenster is a structural window into the stratigraphically distinct rocks of the North Slope subterrane, which are probably present beneath most of the northern Brooks Range and Arctic Slope.

Basement rocks of the North Slope subterrane in the Doonerak fenster are affected by imbricate thrust faulting and metamorphism typical of the southern Brooks Range. On the basis of studies of cleavage development in the Doonerak fenster and overlying terranes, there were probably two post-Triassic periods of metamorphism (Dillon, 1982; Julian and others, 1984). Geochronologic and stratigraphic data indicate that metamorphism took place before Albian time. Structural data are consistent with metamorphism during emplacement of the Angayucham terrane during Jurassic or Early Cretaceous time.

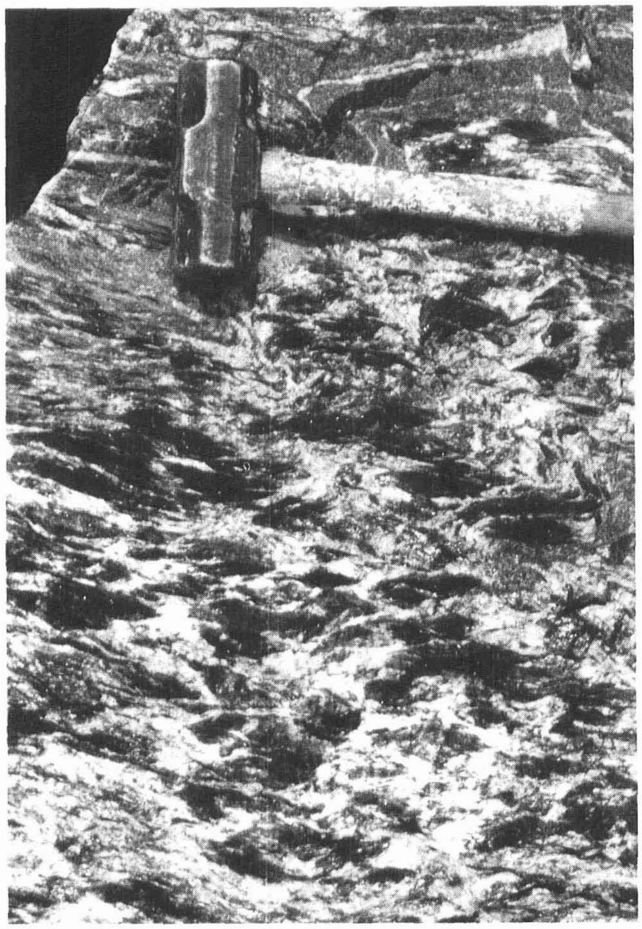

Figure 140. Knotty texture of muscovite-quartz schist of the Brooks Range schist belt, caused by two or three superimposed metamorphic fabrics.

Most of the thick sections of middle Paleozoic rocks of the Arctic Alaska terrane contain minor to abundant volcanic interlayers. These thick volcanic-rich, middle Paleozoic, possibly eugeoclinal sequences occur dominantly in the southern and central parts of the Wiseman and Chandalar Quadrangles south of the Doonerak fenster. Volcanic-poor, miogeoclinal or nonmarine middle Paleozoic sequences are restricted to the Endicott Mountains and North Slope subterranes. When restored to their prethrust position, the Devonian miogeoclinal sequences lie to the north of coeval eugeoclinal sequences. The hinge line between the miogeoclinal and eugeoclinal Devonian sequences may be marked by several converging Devonian and Early Mississippian unconformities.

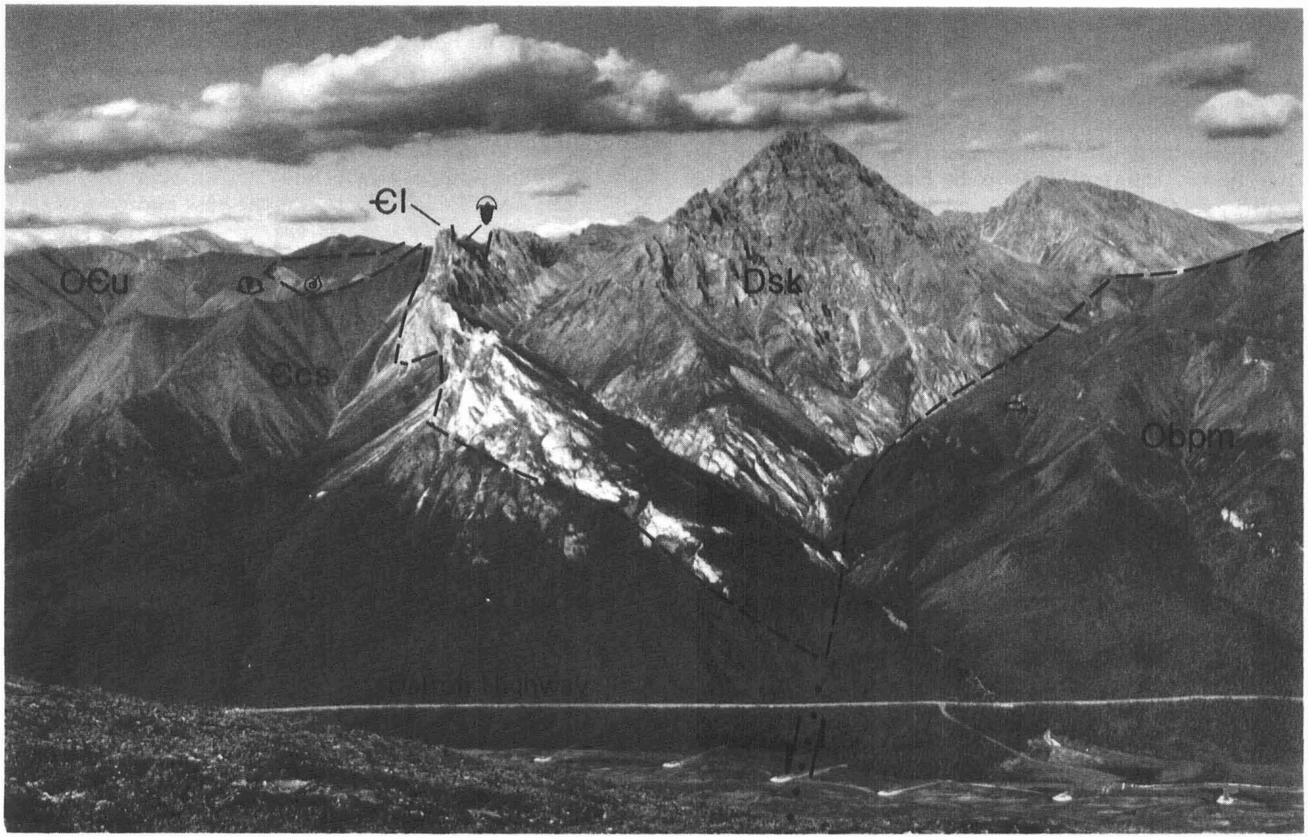

Figure 141. View eastward across ***Dietrich*** *River valley toward* ***Snowden*** *Mountain, near Mile 220.* ***Snowden*** *Mountain is composed of Devonian* ***Skajit*** *Limestone (Dsk) that overlies a sequence of Cambrian calc-schist* ***(€cs)*** *and trilobite-bearing limestone (€1) to the north and is faulted against Ordovician conodont-bearing black marble and graphitic phyllite (Obpm) to the south. Fossils symbols: , brachiopod;* ***O*** *, cephalopod; , trilobite; ,* ***cono-****dont.*

STRUCTURE

FAULTS AND FAULT PANELS

The lithotectonic terranes of the southcentral Brooks Range are cut by several large-displacement thrust faults and strike-slip faults. Although many of these faults have not been precisely located in the field, their approximate location is based on. stratigraphic, structural, and geophysical discordances.

There were two episodes of large-displacement faulting in the southcentral Brooks Range. During the first episode, the oceanic rocks of the Angayucham terrane in the Koyukuk basin were obducted onto the continental rocks of the Arctic Alaska terrane. At this time the lithotectonic terranes of northern Alaska were sutured together. This event produced regional metamorphism of the continental rocks that were overridden and pervasive thrust faulting in both the continental and oceanic plates. Most of the subterranes of the Arctic Alaska terrane are bounded by thrust faults formed during this Jurassic to Early Cretaceous event. The second event is represented by post-Early Cretaceous, high-angle, strike-slip faults that displace the thrust faults and bound some of the southern terranes.

THRUST FAULTS

Introduction

All rocks of the Arctic Alaska terrane in the southern Brooks Range have been significantly displaced by north-vergent thrusts. The Angayucham terrane was probably thrust northward onto the Arctic Alaska terrane during Late Jurassic and Neocomian time. Most thrust faults that formed in the Arctic Alaska and Angayucham terranes during this event dip to the south-southwest; however, many of the thrust faults are folded, and strikes and dips vary. North of the Brooks Range drainage divide, many of the major thrust faults are folded and dip to the north (fig. 26; sheets 1 and 2; Mull, 1982).

Thrust faults in the Arctic Alaska terrane

In vertical section (sheet 2), most fault panels in the Arctic Alaska terrane are bounded by thrust faults that separate horizontally elongate, south-dipping, diamond-shaped to lenticular rock bodies 10 to 20 km thick. To the north, the lenticular rock bodies are formed by individual thrust faults that cut upsection to the north through middle and upper Paleozoic metasedimentary strata. To the south, progressively older Proterozoic(?) basement and lower Paleozoic metasedimentary rocks are involved in thrusting. Thrust faults in the Arctic Alaska terrane are assumed to flatten to the south and join in a décollement at or above the base of the lithosphere of the Arctic Alaska plate.

Thrust fault zones in the Arctic Alaska terrane with slips of 10 to 100 mi separate sections of moderately contrasting stratigraphy. Although displaced, I think that there is enough stratigraphic similarity between rocks in adjacent fault panels of the Arctic Alaska terrane to limit displacement on the faults from ten to a few hundred miles. Consequently, the rocks of the Arctic Alaska terrane are considered parautochthonous. Large-scale, asymmetric isoclinal folds within the stratified rocks indicate northward vergence during thrusting. Most pre-existing structures and textures of the subducted continental rocks of the southern Brooks Range were obscured by deformation, cataclasis, and recrystallization during the Late Jurassic to Neocomian orogenic event.

A lower limit for the age of this convergence and thrusting is provided by genetically related thrust faults in the Angayucham terrane that cut lower Jurassic and older rocks. Thrust-related metamorphism affects Triassic and older rocks in the Arctic Alaska terrane. Lead-loss ages from zircons, Rb-Sr and Sm-Nd mineral and whole-rock isochrons, and older K-Ar cooling ages of 135 to 150 Ma (Late Jurassic to Early Cretaceous) date the closing of the high-temperature isotopic systems following metamorphism. The oldest K-Ar cooling ages related to thrusting are late Middle Jurassic (Patton, 1984). A large number of K-Ar cooling ages in the Brooks Range relate to postconvergence regional uplift and cooling from 100 to 140 Ma.

On many of the allochthons in the northcentral and western Brooks Range, widespread flysch and olistostromes of Neocomian age provide stratigraphic evidence for thrusting at that time, as does a reversal from southward- to northward-directed paleocurrents. The Neocomian flysch, derived from the overriding thrust sheet, was itself involved in the latest stages of thrusting. Coarse Albian conglomerates that unconformably overlie allochthonous rocks in the northern and southern foothills of the central Brooks Range provide a younger limit for thrusting in that area (Mull, 1982, 1985, chaps. 5 and 6; Mayfield and others, 1983). Pre-Albian thrusting is also indicated by Albian and younger sediments that unconformably overlie rocks and thrust faults of the Angayucham terrane along the southern margin of the Brooks Range (Dillon and Smiley, 1984). In addition, the Angayucham thrust system is intruded by 112-m.y.-old granites in the Ruby geanticline (Blum and others, chap. 11).

The best-known thrust fault in the Arctic Alaska terrane is the Amawk thrust (figs. 27, 79, and 139; sheet 2) (Dutro and others, 1976; Mull, 1982), which surrounds the Doonerak fenster and displaces rocks of the Endicott Mountains allochthon (fault panel 3) onto rocks of the North Slope subterrane (fault panels 1 and 2); it has at least 55 mi (90 km) of slip (Mull and others, chap. 14). Fault planes within this zone are marked by broken formations, phyllonite, and mylonite. Mapping by Dillon and others (1986) shows that the Amawk thrust zone is folded around the east and west ends of the Doonerak fenster, as are the rocks, metamorphic structures, and minor thrust faults in fault panels 1 through 3 and 6 (sheet 2). Rocks and structures exposed beneath the Amawk thrust are called the 'Doonerak fenster'; folded rocks and thrust faults of the combined upper and lower plates of the Amawk thrust zone are called the 'Doonerak antiform.'

The structurally lowest thrust fault exposed in the Arctic Alaska terrane near the Dalton Highway is the St. Patricks Creek thrust fault, which bounds fault panel 1 in the Doonerak fenster (sheet 2). The St. Patricks Creek thrust may merge with the Amawk thrust through the Blarney Creek thrust (Oldow and others, 1984), or it may be a separate, lower, imbricate structure. The St. Patricks Creek thrust clearly involves lower Paleozoic basement rocks in its upper and lower plates, as do strands of the Blarney Creek and Amawk thrust. Significantly, the St. Patricks Creek thrust appears to define a smaller fenster entirely within the Doonerak fenster, effectively making it a double-paned window! The St. Patricks Creek thrust and hypothetical lower thrust faults probably resulted from slip transmitted through the lower Paleozoic basement rocks to areas farther north (sheet 2). Therefore, rocks in the core of the Doonerak fenster are considered parautochthonous.

On the south side of the Doonerak fenster and structurally higher than the Amawk thrust, two other large-displacement thrust faults have been named within the Arctic Alaska terrane, as well as several unnamed thrust faults (sheet 2). The lowest of these faults, the Trembley Creek thrust, is near the base of fault panel 3 and places a thick section of Beaucoup Formation, Hunt Fork Shale, and felsic volcanic and intrusive rocks over a thin section of Hunt Fork Shale of the Endicott Mountains allochthon (fig. 78). The Trembley Creek thrust is also folded around the east plunge of the Doonerak antiform but cannot be identified on the north side of the fenster, where a thick section of Hunt Fork Shale is present at the base of the Endicott Mountains allochthon.

Above the Trembley Creek thrust, the Table Mountain thrust (fig. 80) displaces a thick section of Skajit Limestone and Whiteface Mountain volcanics at the base of fault panel 6 over the rocks of fault panel 3. The Table Mountain thrust is also folded around the eastern plunge of the Doonerak antiform but is not present on the north flank of the antiform.

Fault panel 7 in the Hammond subterrane may be a klippen because it is surrounded and apparently underlain by a fault zone whose location is based on local structural and stratigraphic discordances between underlying Devonian rocks of fault panel 6 and overlying rocks of unknown age of fault panel 7. Lithologically, the rocks in fault panel 7 are similar to both Ordovician and Devonian formations found in the area. If the rocks of fault panel 7 are Devonian in age, this fault may have relatively small displacement; if Ordovician in age, the fault moved at least 10 km. In either case, this thrust fault probably roots at the northern edge of the Coldfoot subterrane (fault panels 8 and 9).

The Coldfoot subterrane encompasses what is commonly called the 'schist belt' in the central Brooks Range. The northern boundary of the schist belt was first identified in the Survey Pass Quadrangle as a fault named the 'Walker Lake lineament' by Fritts and others (1971), who thought it was a major plate boundary. Later, Dillon and others (1980) suggested that the Walker Lake lineament is not a plate boundary because similar rocks are found on both sides of the fault in the western Wiseman Quadrangle. The northern boundary of the schist belt near the Dalton Highway is the Wiseman thrust (fig. 142), a regional fault zone that dips steeply to the south and brings higher grade, older(?) rocks of the Coldfoot subterrane (fault panel 9) onto lower and middle Paleozoic rocks of the internal metamorphic and plutonic belt, or Hammond subterrane (fault panels 4 through 8). Gottschalk (1987) and Gottschalk and Oldow (1988) have referred to the fault as the Minnie Creek thrust.

Thrust faults in the Angayucham terrane

The Angayucham thrust system has three important strands in the Wiseman and Chandalar Quadrangles. The northern strand of the Angayucham fault system separates the Slate Creek panel (fault panel 10) from the Arctic Alaska terrane (fig. 63). The fault is marked by a mylonite zone that separates high-grade schist belt polymetamorphic rocks of fault panel 9 from low-grade monometamorphic phyllites and graywackes of fault panel 10 to the south; it crosses the Dalton Highway at about Mile 172. Oldow and others (1987b) and Gottschalk and Oldow (1988) propose that this fault is a low-angle, extensional fault system. The middle strand, in the center of fault panel 10, is a zone of broken formations between the phyllite to the north and the graywacke to the south. The southern strand forms the southern boundary of fault panel 10 and separates the graywacke of the Slate Creek panel from mafic igneous rock and chert of the overlying Narvak panel of the Angayucham terrane. The southern strand crosses the Dalton Highway just north of Twelvemile and Cathedral Mountains near mile 167 (fig. 62). This major lithologic break is the boundary between the Angayucham and Arctic Alaska terranes as mapped by Silberling and Jones (1984) and Jones and others (1987) and as illustrated by Mull (1982) (fig. 26; sheets 1 and 2).

Twenty miles (30 km) east of the Dalton Highway, a window through the Angayucham fault system exposes rocks of the Mosquito terrane (fault panel 13). The steeply north-dipping northern boundary of fault panel 12 may also be part of the Angayucham fault system because it displaces mafic igneous rocks and phyllites of the Angayucham terrane over granitic rocks and migmatites of the northern Ruby terrane. The fault zone is marked by garnet amphibolite knockers of the Angayucham terrane and mylonite and mylonite gneiss formed cataclastically from granitic rocks of the Hodzana pluton in the Ruby terrane. Significantly, garnet-amphibolite knockers also occur in the Angayucham thrust zone at several locations along the northwest side of the Ruby geanticline (Patton, 1984). However, the contact between the Angayucham terrane and the Hodzana pluton has been further deformed by right slip on the South Fork fault zone or protoclasis along the border of the Hodzana pluton or both.

The Angayucham and correlative terranes with their associated thrust systems underlie most, if not all, of the Koyukuk basin and define the geanticlinal form of the Ruby terrane (Patton and others, 1977; Decker and Dillon, 1984; Jones and others, 1984). The thrust system separates the widespread upper Paleozoic and

Figure 142. View northwestward toward village of *Wiseman* and *Wiseman* thrust, northern boundary of schist belt in *Wiseman* Quadrangle. Proterozoic(?) to lower Paleozoic calcareous schist *(PzPcs)* and quartz-mica schist *(PzPqs)* are thrust northward *over* Devonian rocks of internal metamorphic and plutonic belt, which in this area consist dominantly of black phyllite and siltstone and chloritic sandstone and conglomerate of the *Beaucoup* Formation *(Db).*

lower Mesozoic oceanic rocks of the correlative Angayucham, Innoko, and Tozitna terranes and the Rampart and Christian Complexes (Jones and others, 1984a) from the continental rocks exposed in the mountain ranges of Arctic Alaska. The Angayucham terrane and fault system occur in klippen in the western Brooks Range and in the Christian Complex and also as a root zone along the southern side of the Brooks Range (Patton and others, 1977; Roeder and Mull, 1978). Thus, the Angayucham terrane and fault zone and their correlative rocks once covered about 240,000 km^2 of Arctic Alaska. Extensions into Siberia have been proposed by Churkin and Trexler (1980) and Box (1984).

The Angayucham fault system is the most important structural feature in northern Alaska. Estimates of displacement on the Angayucham fault system (excluding shortening in the upper and lower plates) vary according to the model of the origin of the oceanic rocks (Patton and others, 1977; Roeder and Mull, 1978; Mull, 1982; Mayfield and others, 1983; Box, 1985; Patton, 1984) but displacement must be at least >120 mi (200 km) and may possibly be several thousands of miles. Thrust imbrication in the upper and lower plates significantly increases the estimated amount of slip that occurred during plate convergence. Inverted structural stacking of thrust panels of the Angayucham terrane near the Dalton Highway places mantle rocks over crustal rocks over supracrustal rocks. This inverted structural stacking of the Angayucham terrane is also present regionally in northern Alaska and requires extensive shortening in the upper plate. Considering the extent and displacement of the Angayucham fault system, it must rank as one of the more important plate boundaries in the North American Cordillera. Miller (1987) and Oldow and others (1987b) suggest that the south flank of the Brooks Range and the Angayucham fault system have been modified by an episode of crustal extension and low-angle normal faulting that began in Albian time.

STRIKE-SLIP FAULTS

Introduction

High-angle, west-trending faults with substantial displacement are common in the southcentral Brooks Range. Many of these longitudinal faults are probably related to post-thrust folding. Other west-trending faults within the Arctic Alaska and Angayucham terranes have right-lateral separation of a few miles to tens of miles. Generally, cataclasis in the postmetamorphic high-angle faults formed narrow zones of brittlely deformed crushed rock and gouge. Locally, neomineralization may have occurred during strike-slip displacement.

Malamute and South Fork fault zones

The Malamute and South Fork fault zones form part of the northern and southern boundaries of fault panel 11 (Angayucham terrane), respectively. Right-lateral separations along the west-trending Malamute and South Fork faults displace the Angayucham terrane and thrust fault system, thus postdating its upper Jurassic to Neocomian emplacement. The Malamute-South Fork fault system also displaces Albian to Cenomanian sedimentary rocks locally. Stratigraphic evidence presented below suggests that the faults may have been active during deposition of the Albian conglomerates.

In the western Wiseman Quadrangle, an exposure of the main strand of the Malamute fault zone consists of a broad zone of gouge containing slickensided phacoidal lenses of various lithologic units that occur both to the north and to the south of the Malamute fault. In the same region, displaced strands of the Angayucham thrust system are displaced to the right 25 mi (40 km). Cretaceous nonmarine strata exposed within the area of figure 139 dip moderately northward and are apparently faulted against basalt and chert of the Angayucham terrane along the Malamute fault on the south side of Cathedral Mountain (sheet 2). Slickensided splays of the Malamute fault zone cut through Cretaceous strata exposed along the Middle Fork Koyukuk River, southwest of Coldfoot near Tramway Bar.

The South Fork fault is exposed in several outcrops 6 mi (10 km) east of the Dalton Highway at Eagle Cliff and along the Mosquito Fork Koyukuk River. At Eagle Cliff, northeast-trending splays cut through Cretaceous conglomerates and also bring slickensided and brecciated Cretaceous strata against deformed mafic volcanic rocks of the Angayucham terrane. The main strand crosses the Dalton Highway at the South Fork Koyukuk River (Mile 156.2). From there the South Fork fault apparently projects eastward along a prominent fault-line scarp on the south side of the Mosquito Fork Koyukuk River (Peter Coney, oral commun., 1984) through a melange zone exposed in the Mosquito Fork on the south side of the Mosquito terrane fenster.

The Malamute and South Fork fault zones pass on the north and south sides of the Mosquito fenster, respectively. The fenster resulted from erosion through a west-trending, double-plunging fold in the Angayucham fault system. This fold may result from transpression between the fault zones.

Right-lateral displacement of the Malamute-South Fork fault system is probably >55 mi (90 km). This estimate is based on the apparent displacement of Cretaceous granitic rocks and surrounding hornfelsed phyllites on the north side of the fault system on the East Fork Chandalar River (Brosge and Reiser, 1964) from hornfelsed phyllites within the fault system near the Dalton Highway 55 mi (90 km) farther west. These western phyllites occur as fault panels within the Malamute-South Fork fault system and provide only a minimum slip distance and piercing point between originally connected terranes. Substantially greater slip is required if the Cretaceous granite on the East Fork Chandalar River (the only Cretaceous granite in the Brooks Range) was displaced from the Jim River pluton, which crops out along the Dalton Highway. The Jim River pluton is the closest Cretaceous pluton intruded into the phyllites on the south side of the fault system (Blum and others, chap. 11).

The Malamute and South Fork fault systems project westward into faults in the Kobuk trench. They project eastward along the northern boundary between the Ruby geanticline and the Brooks Range to intersect the subsurface continuation of the northeast-trending Kaltag fault in the Yukon Flats. The Malamute-South Fork-Kobuk trench fault system is the northern limit for

Cretaceous granites for over 600 km in northern Alaska. In central Alaska and the northern Yukon Territory, the Tintina fault is the northern limit of Cretaceous granites for over 500 km. This similarity provides additional evidence for correlation of the Malamute-South Fork-Kobuk trench fault system with the Tintina fault. There is substantial evidence for mid-Cretaceous and older offset on the Tintina fault (Templeman-Kluit, 1976; Gabrielse, 1985) and for younger, post-Late Cretaceous movement on the Kaltag fault (Patton and Hoare, 1968). The timing and amount of displacement on the northeast-trending Kaltag fault permits the Malamute-South Fork-Kobuk trench fault system to have been offset from the Tintina fault by the Kaltag fault (Grantz, 1966; Patton, 1984).

Aeromagnetic evidence for amount of slip

Aeromagnetic evidence for large displacement on the Malamute-South Fork fault system presented by Decker and Dillon (1984) is illustrated in figure 143. Aeromagnetic data for the southern Brooks Range show five linear magnetic provinces, each corresponding to a distinct geologic terrane. From south to north, province 1 is a broad, low-gradient region underlain by mid-Cretaceous sedimentary rocks of the Koyukuk basin. Province 2 has a very high magnetic signature and high-amplitude magnetic anomalies and is underlain by mafic igneous rocks of the Angayucham terrane (fault panel 11). Province 3 is a narrow, low-gradient magnetic trough underlain by monometamorphic phyllite and metagraywacke (fault panel 10). Province 4 is a high-gradient region underlain by polymetamorphic rocks of the Brooks Range schist belt (fault panel 9). Province 5 is a broad, low-gradient region underlain by lower and middle Paleozoic basement rocks of the Arctic Alaska terrane (fault panels 1 through 8).

In the western Brooks Range, provinces 1 through 5 change trend from west to south and project onto the eastern Seward Peninsula, where similar rocks are found. Southeast of the Brooks Range across the Koyukuk

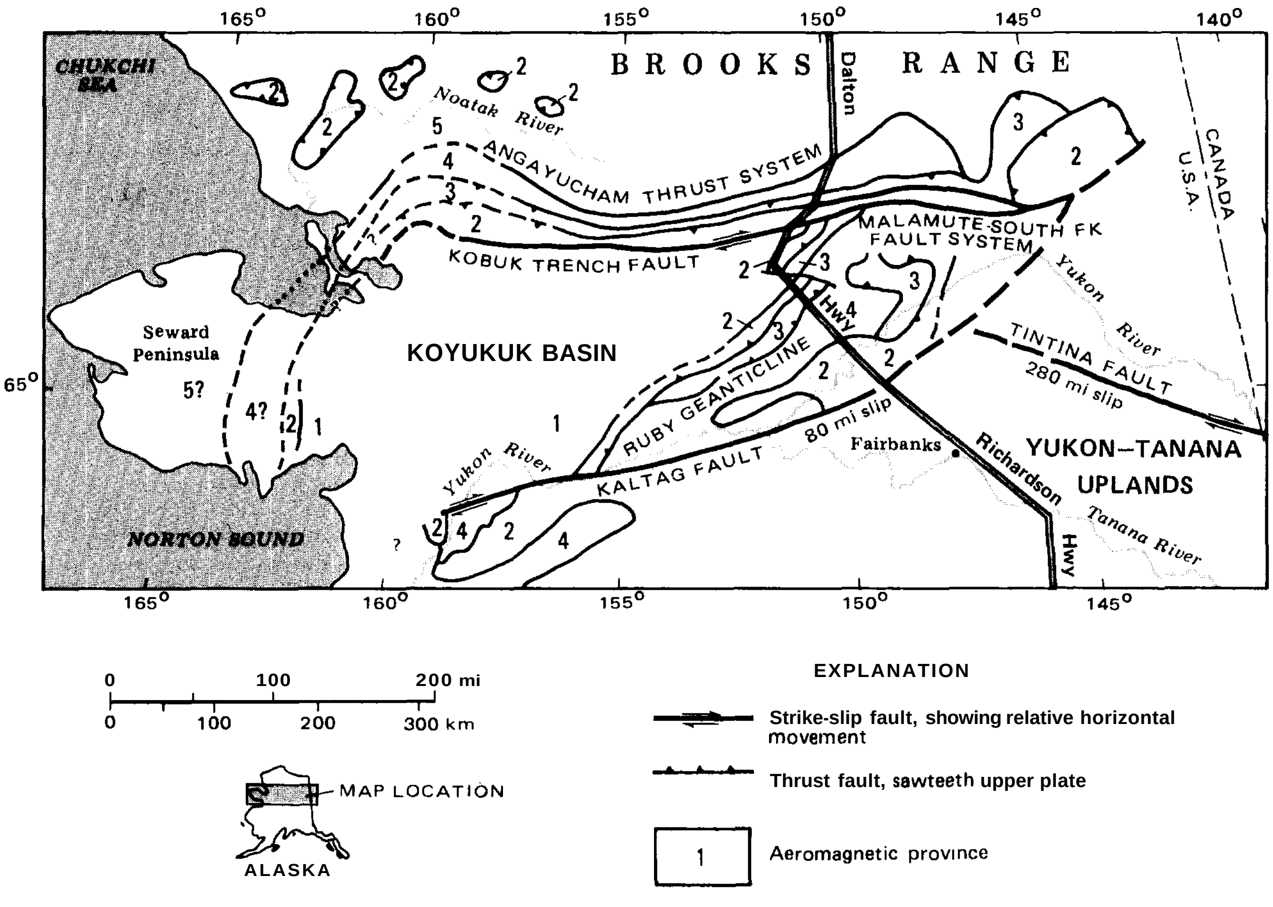

*Figure 143. Map showing aeromagnetic provinces and major strike-slip faults near Koyukuk basin and adjacent areas. Offset provinces are evidence for large-scale strike-slip movement on Kobuk-Malamute-South Fork fault system. Province descriptions: 1, area of low magnetic gradient underlain by Cretaceous sedimentary rocks in Koyukuk basin; **2,** area of high-amplitude anomalies underlain by pillow basalt of fault panel 11 (Naruak panel) of Angayucham terrane; **3,** narrow, low-gradient trough underlain by phyllite and metagraywacke of fault panel 10 (Slate Creek panel) of Arctic Alaska or Angayucham terrane; 4, high-gradient region underlain by metamorphic rocks of fault panel 9 (Brooks Range schist belt); 5, broad, low-gradient region underlain by lower and middle Paleozoic metasedimentary rocks of faults panels 1 through 8 of Arctic Alaska terrane. Map modified from Decker and Dillon (1984).*

basin on the northwest flank of the Ruby geanticline, mirror images of provinces 1 through 3 occur. They are repeated on the southeast flank of the Ruby geanticline because of folding about its northeast-trending axis. The rock units of these three Ruby-terrane provinces have been correlated with those in equivalently numbered magnetic terranes in the Brooks Range by Patton and others (1977), Jones and others (1984), and Dillon and others (1987). The basement rocks that structurally underlie the Ruby geanticline may be equivalent to provinces 4 and 5 of the Brooks Range. Provinces 2 through 5 continue eastward along the south flank of the Brooks Range beyond correlative provinces of the Ruby geanticline; thus, the original continuity of correlated provinces is disrupted and the axis of the Ruby geanticline is truncated. The Malamute-South Fork fault system occurs along a line of abrupt aeromagnetic discontinuity between the disrupted provinces where the axis is truncated. This important fault system displaces and postdates the upper Jurassic to Neocomian Angayucham fault system, which bounds provinces 2 and

High-angle, strike-slip(?) faults in the Arctic Alaska terrane

Three other west-trending, high-angle, strike-slip(?) fault zones cut rocks of the Arctic Alaska terrane. These faults do not bound fault panels but are important because they may have been formed during the same tectonic episode as the Malamute-South Fork fault system. The three faults all displace structures formed in the Arctic Alaska terrane during Jurassic to Neocomian obduction. The southern two faults have subhorizontal slickensides, and lithologic relationships across them seem to require at least a few miles of right slip. The magnitude and direction of slip on the northern fault are unknown.

The southernmost of the three faults crosses the Dalton Highway south of Sukakpak Mountain. This fault forms the brecciated and slickensided marble-schist contact on the south side of Sukakpak Mountain and contains an important stibnite-gold lode deposit described in chapter 13. The stibnitegold vein was emplaced during movement of the fault and probably formed during cooling after metamorphism related to emplacement of the Angayucham terrane. The high-angle fault trends eastward along the south side of Wiehl Mountain to Robert Creek. The fault truncates the south-dipping carbonate rocks of Sukakpak and Wiehl Mountains at the Dalton Highway; the carbonates do not crop out on the west side of the Middle Fork Koyukuk River-Dietrich River valley. Total slip on the fault is not known.

Similarly, the second west-trending, high-angle fault, called the 'Snowden Mountain fault,' forms the contact between the Devonian and older carbonates on Snowden Mountain and Ordovician pelitic schist to the south. The south-dipping carbonate belt is truncated by this fault at the highway and does not crop out on the west side of the Dietrich River valley.[2] Low-angle slickensides in the high-angle fault zone have mullions and chatter marks probably caused by right slip. This fault projects westward along the southern side of the Doonerak antiform toward right-separation faults in the Sheep Creek area of the central Wiseman Quadrangle 60 mi (100 km) southwest of Snowden Mountain. The axis of a large east-northeast-trending synform that runs the length of fault panel 7 may be displaced about 10 mi (16 km) to the right from a syncline in the Hunt Fork Shale near the headwaters of the Mathews River 5 mi (8 km) east of Snowden Mountain.

There is also sketchy evidence for a third high-angle fault that trends west-southwest along the northern side of the Doonerak antiform (Dillon and others, 1986), then bends westward a few miles north of Sheep Creek and continues through Wolverine Creek and onward across most of the Survey Pass Quadrangle. The magnitude and direction of slip have not been measured. Several miles of right separation is inferred from displaced formations along the fault zone in the western Wiseman Quadrangle. However, there is no evidence for a continuous high-angle fault along the north side of the eastern end of the Doonerak antiform. The fault could be one of several longitudinal faults that cut the antiform (Mull and others, chap. 14), but its large separation and apparent continuity well beyond the antiform are not consistent with this hypothesis. Small-scale structures along the northern flank of the antiform described by Avé Lallemant and others (1983) provide evidence for right slip that postdates thrust faulting.

NORTH-TRENDING FAULTS

North-trending normal faults have been mapped in several river valleys in the Wiseman Quadrangle. Only one normal fault has been identified within the area of figure 139. This fault, mapped beneath alluvium in the Middle Fork Koyukuk River valley, is inferred for two reasons: 1) Proterozoic or lower Paleozoic calcareous schist and marble intruded by Devonian(?) granite gneiss are widely exposed in the core of the antiform on the west side of the Dalton Highway and Middle Fork Koyukuk River valley near Emma Creek; however, neither the calcareous unit nor the granite crop out on the east side of the highway near Marion Creek; and 2) the structurally higher quartzose paragneiss that caps Emma Dome on the west side of the Dalton Highway underlies the entire schist belt to the east of the highway. Therefore, I suggest that a north-trending fault exists beneath the valley of the Middle Fork Koyukuk River. However, the axis of the antiform is not displaced laterally by this fault. Pure east-side-down dip-slip of at least 2 km is required to bury the calcareous schist and marble unit without significantly displacing the fold axis laterally. The position of the north-trending fault is indicated by aeromagnetic contours that swing abruptly from their typical east-west trend to an anomalous north trend in the Middle Fork Koyukuk River valley between Coldfoot and Wiseman. Faulting apparently postdates folding of the major rock units and thrust faults in the Coldfoot terrane.

[2] **Alternatively, the high-angle faults on the south sides of Sukakpak, Wiehl, and Snowden Mountains may be part of a regional folded thrust fault that forms the base of a regional allochthon composed dominantly of the Skajit Limestone (Devonian) and older carbonate rocks. A northeast-trending zone of faults on the northwest side of the Skajit Limestone north of Sukakpak Mountain could also be part of the folded thrust.–ED.**

FOLDS

Several generations of small- and large-scale folds deform the rocks of the Brooks Range. Premetamorphic and synmetamorphic folds are described briefly at the beginning of this section. The rest of this section focuses on broad, regional folds that warp major lithologic units, thrust faults, and metamorphic structures. These broad folds formed after metamorphism, perhaps during isostatic rebound following subduction or during strike-slip faulting or both. Erosion of these folds formed the Doonerak, Emma Creek, and Mosquito windows.

Premetamorphic and synmetamorphic, megascopic and mesoscopic, and isoclinal and intrafolial folds are ubiquitous in the southern Brooks Range. Most axes of small intrafolial folds parallel mineral lineations and their plunge depends on their location with respect to postmetamorphic folds. Mesoscopic, isoclinal, asymmetric, north-vergent folds with axes parallel to the regional east-northeast strike are common. These north-vergent folds deform the older post-Triassic(?) to pre-Albian cleavage (S_2) and have a younger axial-plane cleavage (S_3). Locally, these mesoscopic isoclinal folds are close to major thrust faults. From these observations, I inferred that the upper plates of the major thrust faults were probably moving northward while the younger cleavage was forming. The sense of vergence during formation of the S_2 cleavage has not yet been determined.

The broad postmetamorphic folds are typically east-northeast trending, symmetric to slightly asymmetric, and double plunging. The southernmost fold is the Mosquito antiform and fenster (fault panel 13), about 20 mi (30 km) east of the Dalton Highway in the northeastern corner of the Koyukuk basin. The Mosquito antiform deforms the Angayucham fault system. The westward downplunge continuation of this antiform beneath the Dalton Highway is clearly reflected in the gravity and aeromagnetic data illustrated in the cross section (sheet 2).

Shallowly plunging, upright to northward-vergent, overturned folds have affected the postorogenic Albian molasse deposits along the southern flank of the Brooks Range. These folds postdate the suturing of the Angayucham terrane to the Arctic Alaska terrane. Some of the folds occur along--and may be genetically related to--strands of the Malamute-South Fork fault system. Other folds in the western Wiseman Quadrangle seem best explained by post-Albian compression in the northern margin of the Koyukuk basin after both the thrust and strike-slip faulting events. Post-Albian compression is also indicated by low-angle faulting and local low-grade metamorphism in Albian molasse deposits farther west near the Cosmos Hills (Fritts and others, 1971; Hitzman and others, 1982). There, compression may be due to either strike-slip movement on the Kobuk trench fault system (Grantz, 1966) or unrelated compression of the northern margin of the Koyuk basin.

Metamorphic rocks of the schist belt in the Emma Creek drainage between Coldfoot and Wiseman are folded into a broad, west-southwest-plunging antiform. The fold axis of the Emma Creek antiform projects directly across the Middle Fork Koyukuk River valley and the Dalton Highway near Mile 180 (fig. 144). Eastward from Marion Creek the antiform projects into

Figure 144. View southeastward from Emma Dome toward Middle Fork Koyukuk Riuer valley. Calcareous schist and marble (PzPcs, PzPm) are exposed in core of antiform that trends east across valley and continues up Marion Creek, where a structurally higher schist (PzPqs) is exposed. Because the calcareous schist and marble (PzPcs, PzPm) are not exposed in Marion Creek, a fault (downdropped to the east) probably trends along the valley parallel to the Dalton Highway. Note eclogite locality of Gottschalk and others (1984). Remaininggeologic units. MzPzp, MzPzg, and MzPzv, Mesozoic to middle Paleozoic phyllite, graywacke, and mafic volcanic rocks of Angayucham terrane; Dhf, Upper Devonian Hunt Fork Shale.

and across the Chandalar Quadrangle (fig. 145). Proterozoic(?) or lower Paleozoic rocks intruded by Devonian granite form the core of this antiform. These older rocks are overlain along the north limb and east-plunging nose of the antiform by a north- and east-dipping stratigraphic sequence of lower Paleozoic metasedimentary rocks, Middle Devonian Skajit Limestone, and Upper Devonian Hunt Fork Shale.

A regional, northeast-plunging postmetamorphic antiform crosses the Dalton Highway near Mile 209, between Sukakpak Mountain and Dietrich Camp. The south limb is evident in the southeasterly dipping carbonate rocks that form Sukakpak and Wiehl Mountains. The dip reverses in the schist to the north. West of the Dalton Highway, the antiform is also visible in the metasedimentary rocks of fault panels 6 and 8 north of the Wiseman thrust (sheet 2). The north-dipping limb is paralleled to the north by a double-plunging, east-northeast-trending synform that cradles the klippe(?) of fault panel 7. The axis of the synform intersects the high-angle fault on the south side of Snowden Mountain.

North from Snowden Mountain to Nutirwik Creek, the strike and dip of major lithologic units and thrust faults change gradually from west-striking and south-dipping to north-striking and east-dipping, as the rocks are folded around the eastern plunge of the Doonerak antiform. This broad, doubly plunging fold trends east-northeast across the northern Wiseman and northwestern Chandalar Quadrangles for over 80 mi (130 km). It folds both the rocks and thrust faults of the Doonerak antiform (fault panels 1 and 2), including those in the upper plate of the Amawk thrust. The axis plunges east across the Dalton Highway north of Nutirwik Creek near Mile 230 and passes beneath Table Mountain.

METAMORPHISM

Data compiled from thin sections and outcrops from 1:63,360-scale quadrangles show that, in addition to thermal metamorphism due to intrusion, rocks of the Arctic Alaska terrane in the southern Brooks Range (fault panels 1 through 9) have been affected three times by regional metamorphism (M_1, M_2, and M_3, with corresponding cleavages S_1, S_2, and S_3). By comparison, only one regional metamorphic event has affected the rocks of the Angayucham terrane in the northern Koyukuk basin (fault panels 10 and 11), and cleavages produced in these rocks during this event parallel M3 in the Arctic Alaska terrane. Rocks in the Mosquito terrane (fault panel 13) have such strong final schistosity (S_3?) that evidence for earlier recrystallization is obscured. Few host rocks to late Mesozoic granites in the northeastern Ruby geanticline (fault panel 12) are present in the area we have studied in detail; those host rocks that do crop out were so thoroughly migmatized during intrusion that only the fact that they were metamorphosed prior to intrusion can be discerned (see Patton, chap, 4, and Patton and others (1977) for discussions of rocks in the Ruby geanticline).

Two cleavages (S_2 and S_3) and two metamorphic mineral assemblages are easily discernible in most outcrops and thin sections from the southern Arctic Alaska terrane. The metamorphic grades for both of these cleavages and mineral-forming events (M2 and M3) decrease northward. An older, pre-S_2 metamorphic event (M_1) is suspected on the basis of indirect evidence described below.

One or both of the two obvious metamorphic events that affected rocks in the southern Brooks Range (M_2 and M_3) and the discernible metamorphic events of the Angayucham, Mosquito, and Ruby terranes are attributed to increases in temperature, pressure, and rates of cataclasis due to tectonic crustal thickening during emplacement of the Angayucham terrane. The regional parallelism of the latest prominent cleavage within all areas and parallelism with the attitude of the Angayucham fault system provide evidence for recrystallization during thrusting. M_2 and M_3 metamorphism are nearly coaxial in the southern Brooks Range. Metamorphic grade of the M_2 and M_3 metamorphic events increases southward in the Arctic Alaska terrane toward the Angayucham fault system, probably because of increased tectonic loading by the ophiolite. The youngest cleavage (M_3) is axial planar to major north-vergent isoclinal folds that were probably formed during thrusting; mineral lineations trend north-south, parallel to the apparent thrust-transport direction.

The timing and P-T conditions of this thrust-related metamorphism in the Arctic Alaska terrane are discussed in Roeder and Mull (1978), Dillon and others (1980), Dillon (1982), Mull (1982), Dusel-Bacon and others (1989) and are summarized here. The M_3 metamorphism recrystallized Triassic and older rocks; the M_2 metamorphism recrystallized Mississippian and older rocks and may have affected Triassic rocks. Near the Dalton Highway, Albian sediments were not affected by either M_2 or M_3 metamorphism. The available data suggest that both M_2 and M3 were post-Triassic and pre-Albian.

Thrusting of oceanic rocks of the Angayucham terrane onto the continental basement of the older Brooks Range probably started in Late Jurassic time because some knockers in the Angayucham fault system produce Late Jurassic K-Ar (amphibole) cooling ages (Patton, 1984). In addition, graywacke involved in thrusting of the allochthons in the De Long Mountains contains Late Jurassic pelecypods (Mayfield and others, 1983). Thrusting continued through deposition of the Berriasian Okpikruak Formation and deformation of Valanginian to Hauterivian flysch in the northern Brooks Range (Mull, 1982, 1985). However, postmetamorphic uplift must have started close to this time because the oldest abundant K-Ar cooling ages in the southern Brooks Range are about 140 Ma (fig. 146). Furthermore, preliminary U-Pb metamorphic lead-loss ages for Devonian zircons from the Brooks Range and preliminary Rb-Sr and Nd-Sm mineral and whole-rock isochrons from the Wiseman Quadrangle yield metamorphic ages between 100 and 160 Ma (J.T. Dillon, DGGS, unpublished data, 1986). Therefore, the peak of metamorphism was probably during Late Jurassic to Neocomian time, about 120 to 160 Ma ago.

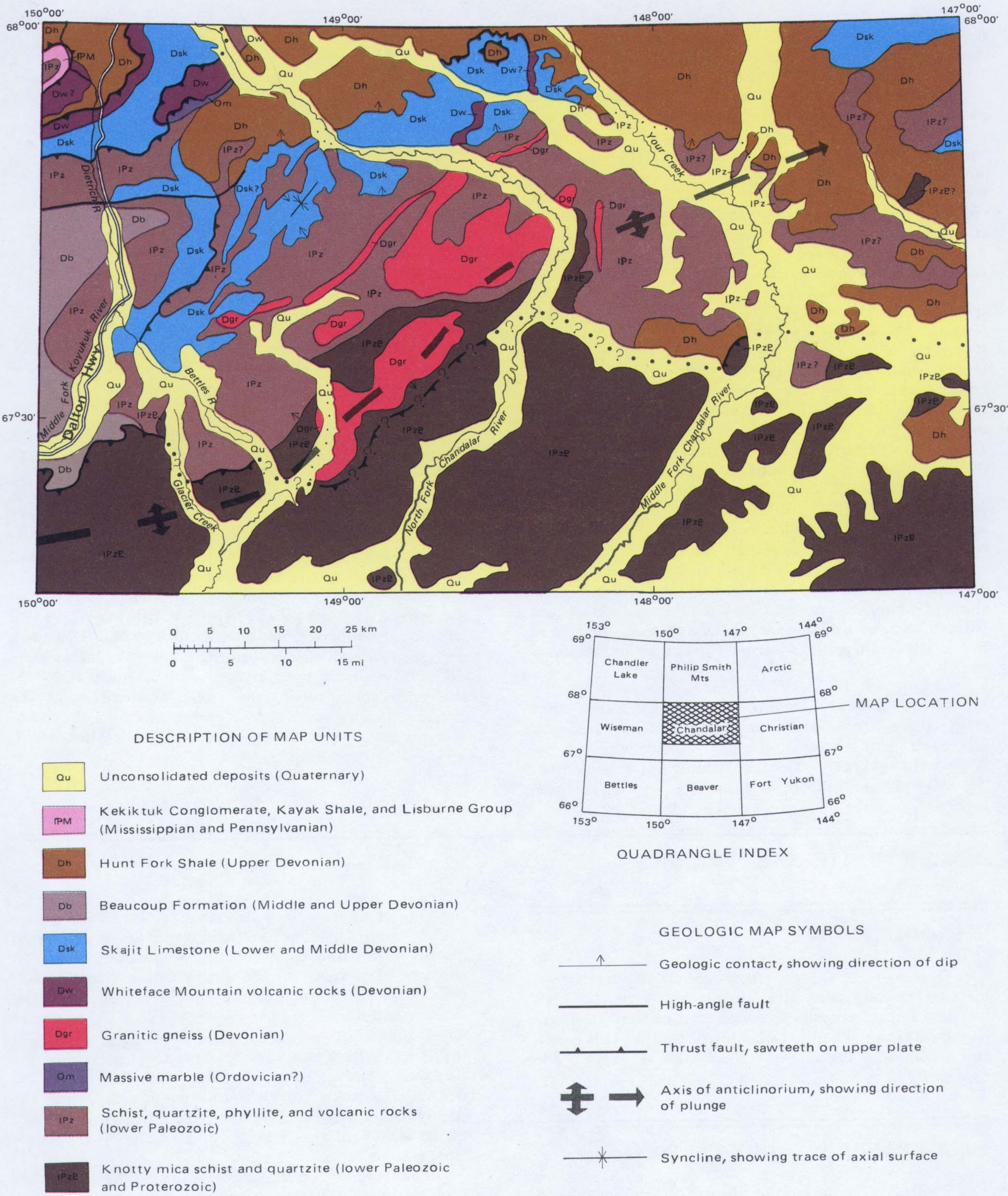

Figure 145. Generalized geologic map of northern Chandalar Quadrangle, showing eastward projection of Emma Creek antiform.

Regional evidence for deformation and metamorphism that predate Late Jurassic emplacement of the Angayucham terrane includes the following: 1) radiometrically dated Proterozoic(?) metamorphic rocks; 2) thermally overprinted, regionally metamorphosed rocks around Devonian plutons; 3) abrupt facies changes and unconformities in Devonian and lower Mississippian rocks; and 4) locally preserved older(?) metamorphic structures. However, I have not been able to locate specific pre-Jurassic faults or cleavage planes, and the nature of pre-Jurassic structural events must remain speculative. One or more pre-Devonian regional metamorphic events is required and they may have been accompanied by low-angle faulting.

Mineral assemblages associated with emplacement of the Angayucham terrane onto continental metamorphic terranes (Arctic Alaska and Ruby) that surround the Koyukuk basin include a facies series from garnet-amphibolite and blueschist to biotite-greenschist and epidote-amphibolite and finally to greenschist (Gilbert and others, 1977; Patton and others, 1977; Turner and others, 1979; Dusel-Bacon and others, 1989). Mineral assemblages from the Wiseman Quadrangle are similar to the facies series of the Alpine schist belt of New Zealand (Turner, 1981). Maximum temperatures >450 °C in the southern Brooks Range are inferred from the local presence of oligoclase. Most conodont alteration indexes from the Wiseman Quadrangle indicate temperatures of 350 to 400 °C (A.G. Harris, oral commun., 1984). Maximum pressure is controlled by the albite + chlorite = glaucophane transition; peak pressures were about 5 to 7 kbar with higher pressures and temperatures in the south, grading to lower pressures and temperatures in the north.

Metamorphic facies of the two earliest metamorphic events (M_1 and M_2) and possibly the youngest event (M_3) in the Arctic Alaska terrane are locally displaced by thrust faults formed during emplacement of the Angayucham terrane. The displacement is expected for the M_1 metamorphism because it is probably pre-Devonian and for the M_2 metamorphism because S_2 was isoclinally folded during thrust faulting. However, if M_3 metamorphism was the result of tectonic loading during thrusting, the truncation of M_3 mineral zones requires M_3 to be syntectonic.

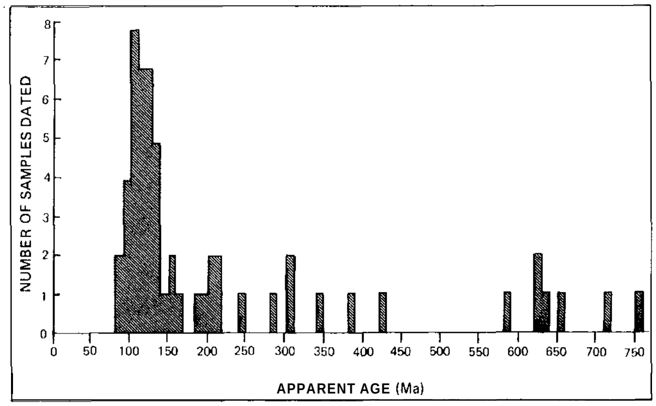

Figure 146. Histogram of ^{40}K-^{40}Ar ages of selected rock samples collected in Brooks Range. From Turner and others (1979).

The onset of rapid isostatic rebound and post-metamorphic cooling that followed thrust-related metamorphism can be estimated from isotopic ages and stratigraphic control. Most K-Ar cooling ages reported by Turner and others (1979) are <140 Ma (fig. 146), with a pronounced peak at 100 to 120 Ma. This peak corresponds with the deposition of thick Albian molasse deposits, on both the northern and southern flanks of the range, that record the erosional stripping of the Brooks Range thrust sheets and cooling of the polymetamorphic parautochthonous basement (Mull, 1982, 1985; Dillon and Smiley, 1984).

DESCRIPTION OF FAULT PANELS

INTRODUCTION

This section details, from south to north, the metasedimentary and metaigneous rocks in the fault panels of the southern Brooks Range and northern Koyukuk basin. The relationships of all the fault panels and terranes of the area are illustrated on sheet 2 and figure 139.

FAULT PANEL 12 (RUBY TERRANE)

Thermally metamorphosed rocks and migmatites were formed by the intrusion of the Mesozoic Hodzana (biotite-hornblende) granite pluton into metasedimentary country rocks that dominate the northern Ruby terrane (fault panel 12). The Hodzana pluton, the most widely exposed rock unit in the Ruby terrane in the area of figure 139, is described by Blum and others (1987; chap. 11). The northern edge of the Hodzana pluton is foliated and, locally, mylonitized along its contact with rocks of the Angayucham terrane (fault panel 11). Garnet-amphibolite knockers occur along the fault in several places near the Mosquito Fork Koyukuk River. Similar garnet-amphibolite knockers are found in the Angayucham thrust zone along the east side of the Koyukuk basin; Patton (1984) described them as tectonites formed early during emplacement of the Angayucham terrane. Therefore, the northern contact of the Hodzana pluton may be part of the Angayucham fault system.

The precise age of emplacement of the Hodzana pluton is not known. However, it is similar in lithology, Rb-Sr whole-rock isochron age, and nearly concordant K-Ar mineral-cooling ages to other Albian granitic plutons that intrude the Angayucham fault and the Ruby geanticline (Blum and others, 1987; chap. 11). Cataclastic rocks along the northern contact of the

Hodzana pluton can be related to the Angayucham fault system only if the pluton is pre-Albian. If the pluton is Albian, the fault is either a protoclastic margin to the Hodzana pluton or a strand of the South Fork fault that has been deeply eroded. In either case, the fault may be a reactivated strand of the Angayucham fault system.

FAULT PANEL 13 (MOSQUITO TERRANE)

Fault panel 13 is the Mosquito terrane of Wust and others (1984). Strongly lineated, amphibolite-facies garnet-biotite-muscovite-plagioclase-quartz schist, hornblende-plagioclase metabasite, and gneissic granitic plutons are exposed in a fenster through fault panels 10 and 11 (sheet 2). This assemblage is similar to old rocks found in the southern Arctic Alaska and central Ruby terranes.

FAULT PANELS 10 AND 11 (ANGAYUCHAM TERRANE)

INTRODUCTION

The Angayucham terrane includes Angayucham volcanic rocks (fault panel 11) and rocks in the phyllite-and-graywacke belt (fault panel 10) (fig. 147). These upper Paleozoic to lower Mesozoic rocks of the Angayucham terrane are unconformably overlain by Albian molasse deposits.

ANGAYUCHAM VOLCANIC ROCKS

Along the flanks of the Brooks Range and the Ruby geanticline, the Angayucham lithotectonic terrane is a stack of three major fault panels. Serpentinized ultramafic rocks and local layered gabbro of the Kanuti panel (Patton and Box, 1989) form the upper part of the Angayucham terrane; Angayucham volcanic rocks of the Narvak panel form the major exposures of the terrane (Roeder and Mull, 1978). These upper two panels constitute fault panel 11. Phyllite and graywacke of the Slate Creek panel constitute fault panel 10.

Ultramafic rocks are exposed only locally in the upper part of fault panel 11 near the Mosquito fenster. However, aeromagnetic highs in the northern Koyukuk basin over the Angayucham volcanic rocks may reflect ultramafic rocks with high magnetic susceptibilities at depth (sheet 2; Smith and Dillon, chap. 12). Gabbro and diabase are the dominant rock types in the upper part of fault panel 11 near fault panel 12. Basalt, locally with pillow textures, is the dominant rock type elsewhere in the lower part of fault panel 11. Diabase and gabbro occur as dikes and sills in the Angayucham volcanic rocks of fault panel 11 and as dikes or sills or tectonic slivers in the phyllite-and-graywacke of fault panel 10.

Metamorphosed and altered mafic igneous rock, chert, argillite, and limestone of the Angayucham volcanic rocks have regional metamorphic mineral assemblages of pumpellyite-prehnite facies, although greenschist and blueschist facies metabasite occur locally (Dusel-Bacon and others, 1989). Primary sedimentary and igneous textures are well preserved in all these

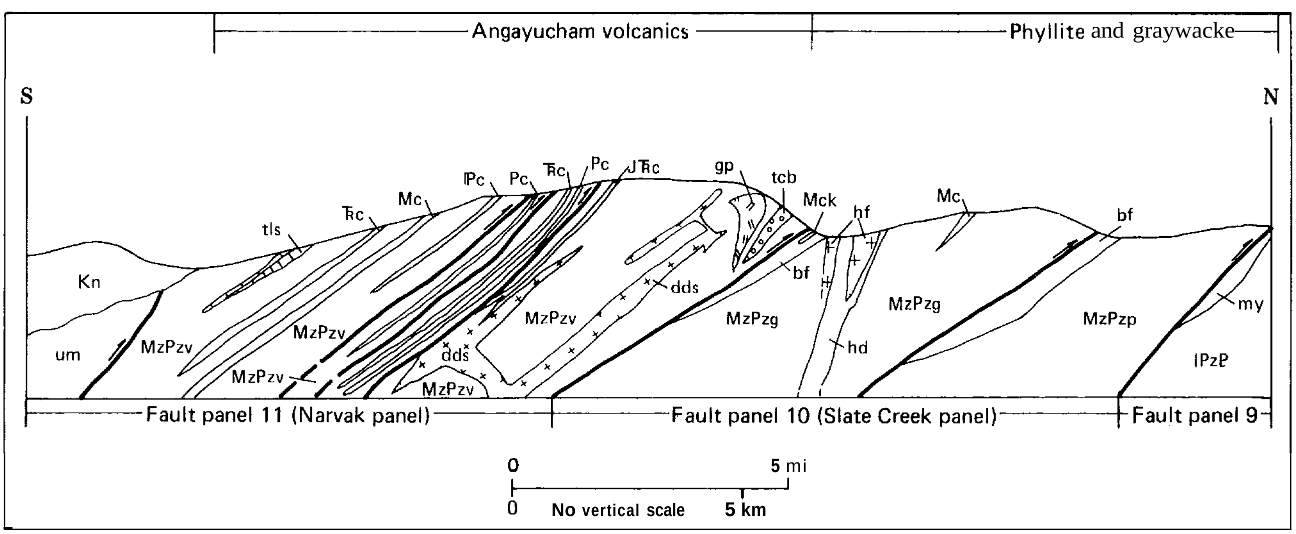

Figure 147. Diagrammatic north-south cross section of Angayucham terrane and southern end of Arctic Alaska terrane. Mixed chert ages indicate imbricate thrusting of long-lived oceanic crust. Geologic units: Fault panel 11 -- MzPzv, Mesozoic to Paleozoic volcanic rocks, dominantly pillow basalt; Mc, Mississippian chert; IPc, Pennsylvanian chert; Pc, Permian chert; Ƭc, Triassic chert; JƬc, Jurassic to Triassic chert; Kn, Cretaceous (Albian) nonmarine conglomerate; dds, diorite dikes and sills; gp, gabbro plug; tcb, tuffaceous conglomerate and breccia with Triassic chert clasts; tls, tuffaceous bioclastic limestone; um, ullramafic rocks. Fault panel 10 -- MzPzg, Mesozoic to Paleozoic graywacke; MzPzp, Mesozoic to Paleozoic phyllite; Mc, Mississippian chert; Mck, Mississippian chert knockers; bf, broken formation; hd, hornblende diorite (tectonic sliver?); hf, hornfels. Fault panel 9 -- lPzP, lower Paleozoic to Proterozoic(?) polymetamorphic rocks of schist belt; my, mylonite.

rocks. Conodont alteration indices from limestone interbeds in the Angayucham volcanic rocks indicate metamorphic temperatures of 160 to 300 °C (I.L. Tailleur, oral commun., 1985).

Pillow basalt, breccia, and tuff dominate the lower part of the Angayucham volcanic rocks. Abundant chert, rare limestone, and red and brown argillite interlayers also occur. Radiolarian chert and fossiliferous limestone interlayers range from Late Devonian to Early Jurassic(?) (D.L. Jones, oral commun., 1984), a period of almost 200 m.y. Repeated sequences of older chert layers overlying younger chert layers is best explained as being the result of tectonic imbrication by layer-parallel thrust faults (fig. 147).

Lenses of tuffaceous, bioclastic limestone interbedded with pillow breccia and tuff at Twelvemile Mountain (Mile 165.7) and at Heart Mountain in the western Wiseman Quadrangle contain fossils of mixed ages and provenances. Fossils within various limestone interbeds on Twelvemile Mountain include Visean to Permian foraminifera, endothyroids, echinoderms, ostracods, corals, bryozoans, brachiopods, and calcareous algae in oolitic, micritic grainstone (Bird, 1977; I.L. Tailleur, oral commun., 1985). Fossils within the lenticular limestone interbeds on Heart Mountain include Permian foraminifera and brachiopoda, crinoid and bryozoan debris, and Mississippian conodonts; adjacent chert layers contain Triassic radiolaria (Patton and Miller, 1973; Murchey and Harris, 1985; I.L. Tailleur, oral commun., 1985; Dillon and others, 1986). Mixed fossil ages of the limestones require a source area for carbonate detritus of mixed ages and provenances, such as a continental shelf or a seamount. Derivation of the bioclastic material from a shelf setting seems unlikely, as the siliceous clastic interlayers that would indicate proximity to a continent are very rare in the Angayucham volcanic rocks. The limestone was probably formed by bioclastic debris flows from a seamount in a volcanically active oceanic plateau during Mississippian, Permian, and Triassic time.

The age range of the volcanic rocks of the Angayucham terrane indicated by fossils from chert and limestone layers is as great as that underlying oceans today and requires either far-traveled or long-lived oceanic crust, or both. The Angayucham volcanics may represent a vast expanse of oceanic crust that was formed over a long period (Late Devonian through Early Jurassic) at a mid-ocean ridge and then obducted onto the Brooks Range. Alternatively, a smaller piece of Late Devonian to Early Mississippian oceanic crust carrying chert and volcanic rocks that had been deposited on it in an oceanic plateau setting may be represented in the upper part of the Angayucham terrane in fault panel 11.

PHYLLITE-AND-GRAYWACKE BELT

Fault panel 10 (Slate Creek panel of Patton and Box, 1989) consists of graywacke and subordinate phyllite in the southern part and phyllite and subordinate graywacke in the northern part (fig. 147; sheet 2). The area of predominantly graywacke is separated on the south from the Angayucham volcanic rocks by the south strand of the Angayucham fault system and on the north from the area of predominantly phyllite by the middle strand of the Angayucham fault system; both strands are marked by broken formations. The phyllite panel is more recrystallized and deformed than the graywacke panel but has only one prominent foliation. The phyllite structurally overlies polymetamorphic rocks of the schist belt that have two prominent foliations. The northern strand of the Angayucham fault system is marked by mylonite and separates the monometamorphic rocks of the Slate Creek panel from polymetamorphic rocks of the schist belt. Oldow and others (1987b) and Gottschalk and Oldow (1988) suggest that this fault is part of a late low-angle extensional fault system.

The phyllite-and-graywacke panels differ mainly in abundance of phyllite and degree of recrystallization. Both contain fine- to medium-grained brown metagraywacke, brown metasiltstone, and dark-gray phyllitic argillite, with relict clastic textures, and, locally, sole marks and well-preserved bedding, including graded bedding. The rocks are composed of subangular detrital grains of quartz (40 percent), albite (10 percent), and tourmaline and rocks fragments (35 percent) in a microcrystalline matrix (15 percent) of quartz, sericite, and chlorite. Rock fragments are 80 percent chert and 20 percent chloritic, volcanic fragments and argillite. Coarse lithic wacke composed of chert and volcanic fragments occurs as interlayers in the Angayucham volcanic rocks and in the graywackes near the graywacke-volcanic rock contact.

Mississippian radiolarian-chert interbeds and Triassic chert blocks occur in the graywacke (D.L. Jones and B.L. Murchey, written communs., 1984). Both the graywacke and phyllite panels were intruded by gabbro and diabase before metamorphism. Contacts of the Triassic chert blocks have not been seen and their significance with regard to the age of the surrounding strata is uncertain.

The metagraywacke and phyllite have well-preserved clastic textures and sedimentary structures, poorly developed greenschist-facies mineralogy, and one metamorphic fabric. Metamorphism is bracketed between the Mississippian age of intercalated chert and the Albian age of overlying unmetamorphosed conglomerate.

The structure and fabric of the graywacke and phyllite are similar to structural zone 2 metagraywacke described by Blake and others (1967). These low-grade monometamorphic metagraywackes contrast sharply with the tectonite fabric, upper greenschist- to albite-epidote-amphibolite-facies metamorphic mineralogy, and polymetamorphic structures of the Proterozoic(?) to Devonian schists in the structurally underlying Arctic Alaska terrane. Similar orientation and appearance of the latest cleavage in the polymetamorphic rocks (S_3) and in the cleavage in the overlying monometamorphic graywacke suggest synchronous metamorphism of rocks in both fault panels 9 and 10.

CRETACEOUS SEDIMENTARY ROCKS OF KOYUKUK BASIN

Near the Dalton Highway, unmetamorphosed Albian and Cenomanian coarse, near-shore, marine and nonmarine conglomerate units of the Koyukuk basin unconformably(?) overlie the Angayucham volcanic rocks (fault panel 11). Albian marine rocks in the basin were deposited mainly by turbidity currents and include facies transitional into the nonmarine rocks exposed on the north and east margins of the Koyukuk basin. The nonmarine rocks were deposited by braided and meandering streams that headed in both the Arctic Alaska and Ruby terranes. The respective south- and northwest-facing Albian shorelines of these continental terranes delineate the early form of the Koyukuk successor basin that was built on the lithotectonic terranes that were accreted during the Late Jurassic and Early Cretaceous (Neocomian).

Six rock units (three nonmarine, two marine, one transitional) have been distinguished in the Koyukuk basin strata on the basis of their clast types. The oldest nonmarine unit (fig. 148, Kic) is principally coarse conglomerate with minor sandstone and contains predominantly greenstone clasts with subordinate graywacke clasts. The middle nonmarine unit (fig. 148, Ksc) is mainly conglomerate with local interlayers of mudstone, carbonaceous shale, and coal; graywacke clasts predominate over greenstone and vein-quartz clasts. On the west flank of the Ruby geanticline, conglomerates (Ksc) contain abundant clasts of hypabyssal felsic to alkaline plutonic rocks, in addition to graywaclte, greenstone, and quartz clasts. The felsic igneous clasts probably came from 104 to 112 Ma plutons of the Ruby geanticline (for example, the Hodzana and Jim River plutons).

The upper nonmarine sandstone with subordinate conglomerate and local mudstone and coal interbeds (fig. 148, Kqc) contains clasts of vein quartz and graphitic schist. Measurements of clast imbrication, cross-bedding, and trough and channel axes in the Cretaceous nonmarine conglomerates of the Wiseman Quadrangle show that streams flowed to the south (fig. 148).

Sedimentary structures and fining-upward depositional cycles in the lower nonmarine units (Kic, Ksc) are typical of meandering-stream deposits. Bimodal clast

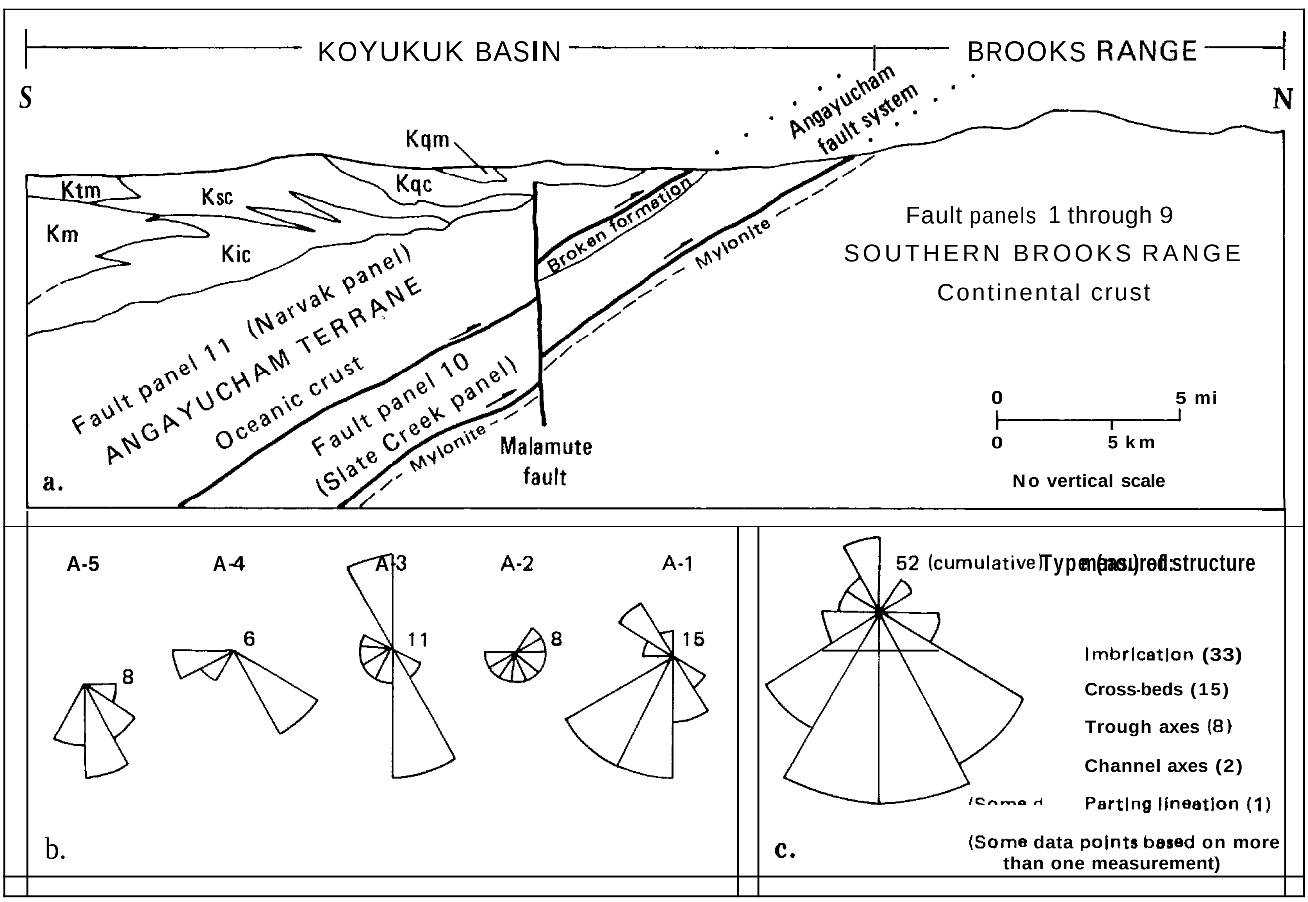

Figure 148. Summary diagrams for Cretaceous conglomerate and sandstone deposits in northern Koyukuk basin and southern Brooks Range: (a) Diagrammatic north-south cross section showing structural and stratigraphic relations. Geologic units: Kic, nonrnarine deposits containing clasts of igneous rock from fault panel 11 (Narvak panel); Ksc, nonmarine deposits containing clasts of metagraywacke from fault panel 10 (Slate Creek panel); Kqc, nonmarine deposits containing clasts of quartz from schist belt and internal metamorphic and plutonic belt; Kqm, marine interlayers in upper nonmarine sandslone Kqc; Km, marine deposits containing mostly clasts of graywacke; Ktm, transitional marine-to-nonmarine deposits containing mostly clasts of graywacke. (b) Rose diagrams showing paleocurrent trends in Wiseman 1:63,360-scale quadrangles. (c) Composite rose diagram summarizing paleocurrent trends (plotted in 30° increments) in Wiseman 1:63,360-scale quadrangles.

sizes, normal and reverse grading, abundant cross-bedding, and lack of depositional cycles in the upper nonmarine sandstone unit (Kqc) are typical of braided-stream deposits.

Marine interlayers (fig. 148, Kqm) have been identified in the upper nonmarine sandstone (Kqc) along the Middle Fork Koyukuk River near Chapman Creek (T.N. Nilsen, J.E. Decker, and J.M. Murphy, oral communs., 1985). Both the middle and upper nonmarine units (Ksc, Kqc) grade southward into marine rocks (fig. 148, Kqm, Km). This nonmarine-to-marine facies transition trends westward along the northern Koyukuk basin. Marine unit Km and coarser grained, transitional marine-to-nonmarine unit Ktm (fig. 148) consist of interbedded sandstone, conglomerate, siltstone, and shale in fining-upward sequences. Sedimentary structures indicate deposition by turbidity currents and other mass flows. Both units contain predominantly graywacke clasts in a matrix dominated by greenstone fragments.

Locally, marine unit Km contains Albian ammonites and pelecypods (Patton and Miller, 1973). Transitional unit Ktm contains medial Cretaceous foraminifera and ***Znoceramus*** prisms, both of which are typically found in shallow, marginal-marine environments (Paula Quintero, written commun., 1984). The marine unit (Km) locally contains Albian ammonites (Brosgé and Reiser, 1971). Fossil floras in nonmarine units Kqc and Ksc are Albian and Albian or Cenomanian (Dillon and Smiley, 1984). Patton (oral commun., 1984) has evidence that flora of the northern Koyukuk basin may be mainly Cenomanian in age.

Mafic clasts in the lowest nonmarine unit (Kic) are derived from the highest thrust panel in the Angayucham terrane (fault panel 11). Monometamorphic phyllite-and-graywacke clasts in the middle molasse unit (Ksc) were derived from the lower thrust panel of the Angayucham terrane (fault panel 10). Clasts in the highest unit (Kqc) are mainly polymetamorphic rocks and quartz veins derived from parautochthonous basement of the Brooks Range. Dillon and Smiley (1984) concluded that the sequence of clast types resulted from progressive erosional stripping of thrust sheets synchronous with uplift and cooling during final stages of the Brooks Range orogeny. A similar conclusion was reached by Patton and others (1977) and Box and others (1984) from the sandstone petrology of marine rocks (Km) of the Koyukuk basin. These conclusions are also compatible with the conclusions of Mull (1982, 1985), who pointed out that a progressive change in composition of Albian sediments on the north side of the Brooks Range suggests progressive stripping of allochthons. (See fig. 149.)

Stratigraphic, structural, and geochronologic evidence indicates concurrent fault movement and molasse deposition. Thick, coarse conglomerate layers with sedimentary structures typical of chaotic, mass-flow deposits predominate in nonmarine and transitional-marine mid-Cretaceous deposits of the northeastern Koyukuk basin (Kic, Ksc, Kqc). The mode of deposition of these conglomerates suggests steep, possibly faulted escarpments in a nearby source area. The east-west-trending, marine to nonmarine transition zone in the northern Koyukuk basin is extremely narrow (<15 mi; 24 km) compared to the broad (∿100 mi; 160 km), synchronous, nonmarine-marine transition in mid-Cretaceous rocks of the Arctic Slope. The narrowness of the facies transition zone, the parallel trends of the facies transitions with strands of the Malamute-South Fork fault system, and the coarse mass-flow deposits

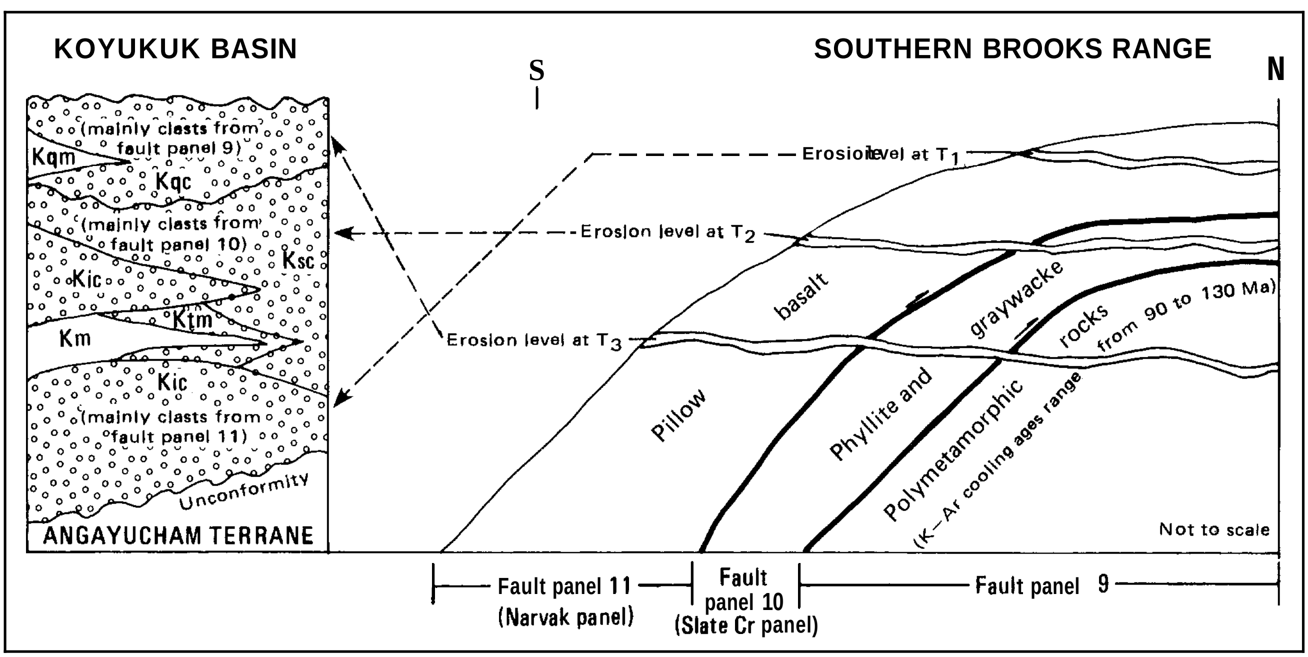

Figure 149. Diagrammatic columnar section and north-south cross section illustrating clast assemblages in Koyukuk basin and erosion levels in southern Brooks Range during mid-Cretaceous time. See figure 148 for explanation of geologic units.

provide evidence for synchronous displacement, erosion, and deposition along the southern margin of the Brooks Range.

Further evidence for synchronous uplift and deposition comes from metamorphic mineral ages. The K-Ar histogram (fig. 146) illustrates the age distribution of K-Ar dates in the central Brooks Range (Turner and others, 1979). Apparenlly, most ages older than about 150 Ma result from partial resetting of older isotopic systems and from inherited argon; those ages younger than about 150 Ma date metamorphic cooling of polymetamorphic rocks of the southern Brooks Range during Late Jurassic and Early Cretaceous time. Closer analysis of the data of Turner and others (1979) and of new K-Ar data (J.T. Dillon, DGGS, unpublished data) shows that the older metamorphic cooling ages (120 to 150 Ma) systematically come from the southernmost edge of the range. Both the structural model for metamorphism (northward thrusting of a southward-thickening slab of oceanic crust) and the metamorphic mineral zones (highest grades and pressures of metamorphism in the southernmost polymetamorphic rocks) seem to indicate a deeper tectonic burial for the polymetamorphic rocks of the southernmost Brooks Range than those farther north. The combined evidence for deeper burial and older cooling ages for these rocks requires relatively rapid uplift and erosion for the southernmost part of the Arctic Alaska terrane. This evidence supports the conclusion of Mull (1982, 1985), who suggested that the schist belt and internal metamorphic and plutonic belt of the Brooks Range were uplifted during the Albian, probably due to isostatic rebound. The suggestions of Miller (1987) and Oldow and others (1987b) that the southern flank of the Brooks Range was affected by an episode of extensional faulting is also compatible with the concept of isostatic rebound during the Albian.

FAULT PANELS 8 AND 9 (COLDFOOT SUBTERRANE)

INTRODUCTION

Rocks in fault panels 8 and 9, the southernmost rocks in the Arctic Alaska terrane, constitute the Brooks Range schist belt. Called the Coldfoot subterrane (Silberling and others, 1984; Jones and others, 1987), these fault panels include the most metamorphosed and probably the oldest rocks in that subterrane (fig. 150).[3] The siliceous metasedimentary rocks may be Proterozoic; lower Paleozoic(?) calcareous protoliths form the structurally lowest unit. Both the siliceous and calcareous rocks must be older than the Middle Devonian Ambler volcanic rocks that overlie them and may include Proterozoic, lower Paleozoic, and Devonian rocks. The older rocks and the Ambler volcanic rocks are intruded by Devonian(?) granitic and bimodal plutons and are overlain unconformably by siliceous and pelitic clastic rocks of the Upper Devonian Hunt Fork Shale.

PROTEROZOIC(?) AND LOWER PALEOZOIC ROCKS

Paragneiss and banded-and-knotty schist, which has probably been metamorphosed three times, form the pre-Devonian(?) basement in fault panels 6, 8, and 9. Protoliths in the most widespread unit (fig. 150, PzEsq, PzEqs) are interlayered quartzite and carbonaceous argillaceous sediment in layers a few centimeters thick. Regionally, these siliceous units structurally or stratigraphically overlie calcareous units (fig. 150, PzPcs, PzPm). Calcareous rocks near the contact with calcareous schist seem to be stratigraphically interlayered or tectonically mixed with the siliceous rocks. All of the rocks near the contact are isoclinally folded because of metamorphism and thrusting.

Both the siliceous and calcareous rocks contain interlayers of mafic metavolcanic rocks and are intruded by dikes and sills of metadiabase and gabbro and by plutonic complexes that include rocks of felsic and mafic composition. The crystallization age of these igneous rocks is uncertain; preliminary isotopic ages of the intrusive rocks range from Proterozoic(?) to Jurassic. Dated Early or Middle Devonian granitic plutons intrude the Proterozoic(?) or lower Paleozoic siliceous (PzEqs, PzEsq) and calcareous (PzPcs, PzEm) rocks in fault panels 6 and 8; lithologically correlative Devonian(?) felsic and mafic plutons intrude the siliceous and calcareous rocks in fault panel 9 near Emma Creek.

DEVONIAN VOLCANIC ROCKS

Devonian volcanic rocks, including an unnamed chloritic quartzite and the Ambler volcanic rocks, discordantly overlie the Proterozoic(?) or lower Paleozoic banded-and-knotty schist (PzEqs, PzEsq) of the schist belt in the western part of the Wiseman Quadrangle (Dillon and others, 1986). Discordance is inferred from the simpler metamorphic history of the Devonian rocks when compared with the Proterozoic(?) and lower Paleozoic rocks of units PzEqs and PzEsq. The Ambler volcanic rocks correlate with the Ambler sequence of Hitzman and others (1982) but are not exposed in fault panel 8 near the Dalton Highway.

The oldest Devonian units in the schist belt near the Dalton Highway are chloritic quartzite (fig. 150, Dsc) and calcareous chlorite-quartz schist that together may

[3]In this discussion, Dillon has combined the description of Proterozoic(?) and lower Paleozoic paragneiss and banded-and-knotty schist of fault panel 9, which most workers consider the Brooks Range schist belt (Coldfoot subterrane of Jones and others, 1987). with a relatively narrow band of less metamorphosed Devonian rock of fault panel 8, which is part of the internal metamorphic and plutonic belt (Mull, chap. 5) (Hammond subterrane of Jones and others, 1987). Dillon's treatment differs somewhat from the terms used in figure 139 and sheet 2, in which the schist belt is restricted to fault panel 9. Dillon combined the discussion of fault panels 8 and 9 because he felt that the knotty mica schist is stratigraphically overlain by the less metamorphosed Devonian rock. Although a stratigraphically continuous section of the schist belt and overlying less metamorphosed rock is not exposed in the vicinity of the Dalton Highway, this relationship is exposed in the western part of the Wiseman Quadrangle to the west and in the central Chandalar Quadrangle to the east.--ED.

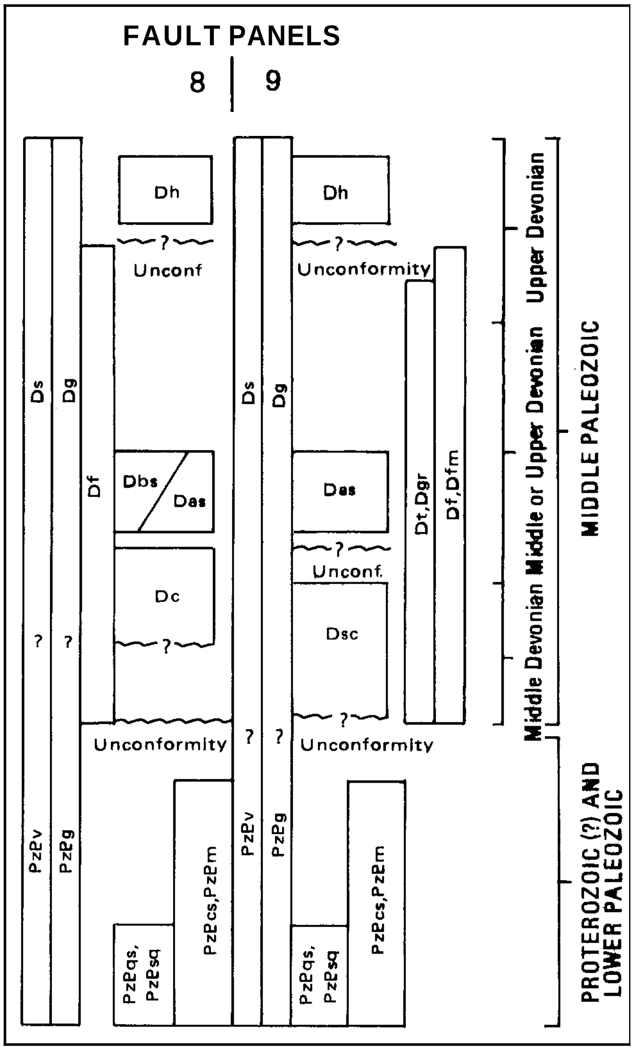

Figure 150. Correlation diagram for rock units in fault panels 8 and 9.

represent a basal sandstone; these rocks are exposed as a narrow band at the extreme southern edge of the schist belt, where they underlie fault panel 10 and the north strand of the Angayucham fault system. The Ambler volcanic rocks (fig. 150, Das, Df) and the Beaucoup Formation (fig. 150, Dbs) (Dutro and others, 1979) interfinger locally and overlie the chlorite-quartz schist. The Ambler volcanic rocks are distinguished from the Beaucoup Formation by the presence of abundant felsic volcanic interlayers and the more heterogeneous lithology of the Ambler rocks. In fault panels 8 and 9, the Beaucoup Formation is represented only by black and dark-green graphitic and chloritic phyllite with lenses of brown and black dolomitic marble and interlayers of greenschist. The Beaucoup Formation is lithologically much more varied farther north in fault panels 3 and 6.

In the central and western Wiseman Quadrangle, west of the Dalton Highway, the Ambler volcanic rocks are composed of graphitic quartz schist, phyllite, and fine-grained, black, pyritic quartzite with interbeds of marble, calcareous schist, and mafic and felsic volcanic rocks. Because of their economic significance, felsic and associated mafic metavolcanic rocks are the distinguishing lithologies in the Ambler volcanic rocks. The felsic metavolcanic rocks include light-colored felsic flows, airfall tuff, and blastoporphyritic intrusive rocks that contain albite + white mica ± biotite ± garnet. Light-colored, fine-grained calcareous felsic schist with intermixed calc-schist and talc schist may represent metamorphosed submarine tuff. Metabasite and principally blocky, dark epidote-chlorite-albite-actinolite greenschists make up 50 percent of the metavolcanic interlayers within the Ambler volcanic rocks.

At one locality within the Wiseman Quadrangle, Devonian conodonts are found in platy gray to black, quartz-bearing marble interlayers within the Ambler volcanic rocks. Farther west, in the Survey Pass and Ambler River Quadrangles, Devonian megafossils also occur in marble layers in the Ambler volcanic rocks (Hitzman and others, 1982).

Zircons from the felsic volcanic rocks and the granitic plutons have been dated as Devonian (Dillon and others, 1980; Newberry and others, 1986; Dillon, unpublished data). Because both the volcanic rocks and plutons are Lower or Middle Devonian, they could be genetically related. Therefore, Ambler volcanic rocks and associated ore deposits may occur near Devonian granitic plutons.

Economically, the Ambler volcanic rocks constitute one of the most important lithologic units in Alaska. At least 14 major volcanogenic Cu-Zn-Ag massive-sulfide prospects have been found in the Ambler mining district in the Survey Pass and Ambler River Quadrangles of the southern Brooks Range, 180 mi (290 km) west of the Dalton Highway (Hitzman and others, 1982). Aggregate reserves reported for three of the deposits are valued at $8 billion. The richest deposit (Arctic Camp) contains 35 million tons of ore that averages 4 percent copper, truly a world-class deposit. The ore in most deposits is closely associated with felsic volcanic rocks.

The Ambler volcanic rocks and the distinctive Proterozoic(?) and lower Paleozoic banded-and-knotty schist are exposed discontinuously for nearly 270 mi (450 km) across the southern Brooks Range schist belt from the Kiana Hills in the Baird Mountains Quadrangle through the Ambler mining district to the central Wiseman Quadrangle (Hawley, 1982). In the eastern Wiseman Quadrangle, exposures of the Ambler volcanic rocks are rare, but east of the Dalton Highway in the Chandalar Quadrangle, the Coldfoot and Hammond terranes may contain exposures of the Ambler volcanic rocks, particularly near Devonian granites.

DEVONLAN HUNT FORK SHALE

The youngest rocks in fault panels 8 and 9 are dark, graphitic, pelitic quartzose schist with locally preserved graded bedding. In the western Wiseman Quadrangle, these rocks discordantly overlie the Ambler volcanic rocks and are lithologically and stratigraphically correlated with the Hunt Fork Shale (Dillon and others, 1986). These rocks locally overlie mica schist at the southern edge of the schist belt east of the Dalton Highway and are on strike with chloritic quartzite that overlies schist west of the highway.

Correlation of the Hunt Fork metasediments, the Devonian felsic plutons, and volcanic rocks of the schist belt (fault panels 8 and 9) with similar Paleozoic rocks of fault panels 3 and 5 through 7 to the north provides evidence that fault panels 8 and 9 have not been displaced large distances from fault panels 3 and 5 through 7 since the Devonian.

METAMORPHISM

Two episodes of metamorphism (M_2 and M_3) affected protoliths of all Devonian and Proterozoic(?) or lower Paleozoic rocks in the schist belt. The third and oldest metamorphic event (MI)--the existence of which is controversial--affected only the protoliths of the Proterozoic(?) or lower Paleozoic banded-and-knotty schist of fault panel 9, which constitutes the schist belt as it is thought of by most workers (Jones and others, 1987; Mull and others, 1987a).

The youngest metamorphism (M_3) is represented by a semipenetrative, schistose axial-plane cleavage (S_3) and probably formed between Triassic and Neocomian times. The M_2 metamorphism, evident from a penetrative schistosity (S_2) that parallels lithologic layering, is presently bracketed within the same time interval. Schistosity S_2 is partially to mostly transposed by semipenetrative schistosity S3. Schistosities S_2 and S3 were both developed in the greenschist to albite-epidote-amphibolite facies. Metamorphism M_1 may have formed the gneissic bands and caused pre-M_2 metamorphic dewatering in the Proterozoic(?) or lower Paleozoic banded paragneiss schists. Partial transposition of S_1 during the S_2 event cut these bands into rods. Where S_3 cleavage cuts the rods at oblique angles, rods of quartzite and graphitic schist are disrupted into distinctive black and white lenses and rods that result in the distinctive knotty structure of these units (fig. 140).

Several regional features of the geology of the southern Brooks Range support pre-Late Devonian metamorphism of the Proterozoic(?) or lower Paleozoic banded-and-knotty country-rock schist of the southern Brooks Range schist belt: 1) lack of knotty structure and presence of relict sedimentary and volcanic structures in the unconformably(?) overlying Devonian metamorphic rocks in the Wiseman, Survey Pass, Ambler River, Baird Mountains, and Shungnak Quadrangles; 2) intrusion of banded and knotty schists by Devonian plutons in the Chandalar, Wiseman, Survey Pass, Ambler River, and Shungnak Quadrangles (Dillon, unpublished data; Dillon and others, 1980, 1987); 3) intrusion of Proterozoic(?) plutons into banded schists in the western Wiseman Quadrangle; 4) apparent pre-M_2 metamorphic dewatering of the banded-and-knotty schist; and 5) Proterozoic(?) K-Ar ages obtained from metamorphic minerals in the Baird Mountains, Ambler River, and Wiseman Quadrangles (Turner and others 1979; Mayfield and others, 1983; R.B. Forbes, oral commun., 1985; Dillon, unpublished data).

Metabasite bodies in the banded-and-knotty schists locally contain relict blueschist- (Dillon and others, 1981a, b) and eclogite(?)- (Gottschalk and others, 1984) facies mineral assemblages. Textural relationships of these two high-pressure mineral assemblages are complex; although both high-pressure assemblages predate at least the retrograde part of the M_3 metamorphism, they may have been formed during different events. There is considerable disagreement about the age and significance of high-pressure mineral assemblages in metabasites in northern Alaska (Turner and others, 1979; Dusel-Bacon and others, 1989; Forbes, oral commun., 1985). The proximity of most blue amphibole-bearing assemblages in northern Alaska to the Angayucham terrane leaves little doubt that most are related to emplacement of that terrane (Patton and others, 1977; Dusel-Bacon and others, 1989). However, the petrology and geochronology of some occurrences are best explained by local preservation of Upper Proterozoic blueschist-facies mineral assemblages (Gilbert and others, 1977; Turner and others, 1979; Forbes, oral commun., 1985). The ability of blue amphibole to persist as a metastable phase is controversial.

In fault panels 3 through 9, at least three generations of quartz veins are distinguished by crosscutting relationships between the veins and structures formed during M_2 and M_3 metamorphic events. Metamorphic structures are postdated by the youngest veins and predated by the oldest veins. Intermediate-age veins cut S_2 cleavage but are cut by S_3 cleavage. Preliminary analyses of fluid inclusions in quartz veins from the Wiseman and Chandalar Quadrangles show that the youngest veins were emplaced during cooler and more hydrous conditions than existed during emplacement of the older two generations of veins (K.K. Lamal, oral commun., 1985).

FAULT PANELS 4 AND 5 (HAMMOND SUBTERRANE)

Fault panels 4 and 5 encompass parts of the Hammond subterrane in the western Wiseman Quadrangle, where these panels were originally differentiated. The Devonian stratigraphy of fault panels 4 and 5 are generally similar to those of fault panel 6 but are not described here because only a small portion of fault panel 5 and none of fault panel 4 are present near the Dalton Highway.

FAULT PANEL 6 (HAMMOND SUBTERRANE)

METAMORPHISM

The evidence for M_1 metamorphism is not found in fault panel 6 of the Hammond subterrane, except near Twin Lakes in the Chandalar Quadrangle. But the lower Paleozoic through Devonian rocks of fault panel 6 are cut by cleavages S_3 and S_2, which also affect all the rocks in fault panels 8 and 9. Both M_2 and M_3 metamorphic-mineral assemblages decrease in grade northward from high-greenschist facies to low-greenschist facies in the broad region underlain by fault panels 5 through 7. The S_2 cleavage is represented everywhere by a penetrative, commonly layer-parallel schistosity. S3 cleavage in the south is a regionally developed, millimeter-spaced, semipenetrative axial-plane schistosity. It gives way to a locally developed, centimeter-spaced, semipenetrative phyllitic cleavage in the northern Hammond subterrane. However, locally, near major thrust faults in the north, such as the Amawk thrust, S_3 cleavage can be highly developed. All the rocks are affected by a locally developed crenulation cleavage.

PROTEROZOIC(?) AND LOWER PALEOZOIC ROCKS

The banded-and-knotty schist is exposed in the Hammond subterrane only near Twin Lakes in the central Chandalar Quadrangle, where it is probably the structurally lowest and oldest unit in fault panel 6 (fig. 145, lPzP). The thrust-fault boundary between the Coldfoot subterrane and the Hammond subterrane has not been mapped in detail east of the Dalton Highway, and its location is uncertain. However, the knotty mica schist is intruded at Twin Lakes and Phoebe Creek by Devonian granitic plutons that pin it to lower Paleozoic rocks indigenous to fault panel 6. These relationships require the presence of the banded-and-knotty schist at depth in other areas of fault panel 6. In addition, in our attempt to model the geologic cross section along the Dalton Highway (sheet 2), high-density, high-magnetic-susceptibility knotty mica schist was required at depth in fault panel 6 to obtain a good fit (Smith and Dillon, chap. 12).

LOWER PALEOZOIC ROCKS

Introduction

Recent paleontologic studies in the Chandalar D-6 Quadrangle near the Dalton Highway show that lower Paleozoic rocks (fig. 151) are widespread in fault panel 6 (Palmer and others, 1984; A.G. Harris and J.T. Dutro, Jr., oral communs., 1985; Dillon and others, 1987). Previously, early Paleozoic fossils in the Wiseman and Chandalar Quadrangles were limited to the Dooneralc fenster (fault panels 1 and 2) (Dutro and others, 1976, 1984a; C.G. Mull and A.G. Harris, written communs., 1985).

Geology

The oldest fossils found in fault panel 6 were discovered at the northern base of the Snowden Mountain carbonate massif by D.N. Solie (DGGS) (figs. 141 and 152). The fossils consist of early Middle Cambrian trilobites and phosphatic brachiopods of Siberian affinities in thin brown limestone beds (fig. 151, Єl) interlayered with dolomitic and graphitic phyllite and chlorite-quartz grit, sandstone, and conglomerate (figs. 74 and 76, Єcs) (Palmer and others, 1984). Trilobite debris occurs in a layer <20 cm below the base of the marble in Snowden Mountain. Dolomite at the base of the massive dolomitic marble contains fragments of poorly preserved conodont species that range from late Late Cambrian to Permian. A few miles to the east, Devonian megafossils occur near the base of the dolomite. Exposures of the contact between the trilobite-bearing lower Cambrian limestone and overlying post-Late Cambrian (Devonian?) massive dolomite in the cliff face are excellent. There is no evidence for faulting at the contact, so the contact is stratigraphic, even though most of the Middle and Late Cambrian epochs are missing between the trilobite-bearing rocks and the overlying marble. These relationships require an unconformable contact (A.G. Harris, oral commun., 1985).

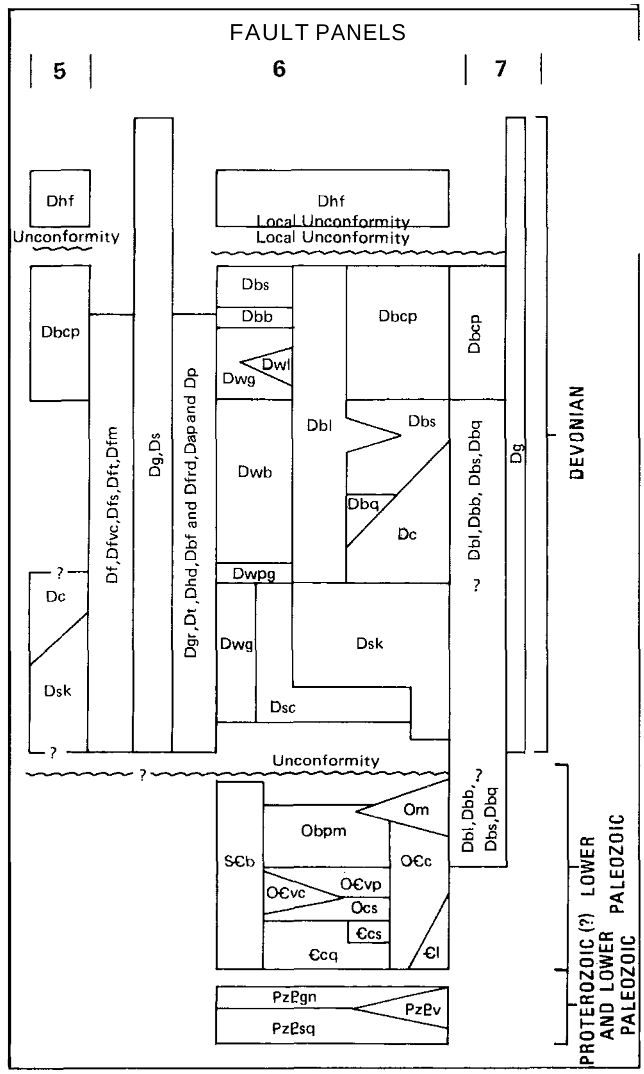

Figure 151. Correlation diagram for rock units in fault panels 5, 6, and 7.

Two and one-half miles (4 km) along strike to the northwest of the trilobite locality, cliff-forming massive marble of the Skajit Limestone contains Devonian megafossils. There, Skajit Limestone directly overlies graphitic phyllite with crinoidal limestone interlayers that contain Middle Ordovician conodonts (J.T. Dutro, Jr., oral commun., 1983; A.G. Harris, written commun., 1985). Data from these localities indicate that Devonian marble unconformably overlies rocks of lower Middle Cambrian through Middle Ordovician age.

Rocks stratigraphically(?) below and on strike with rocks at the trilobite locality include interlayered tan- to

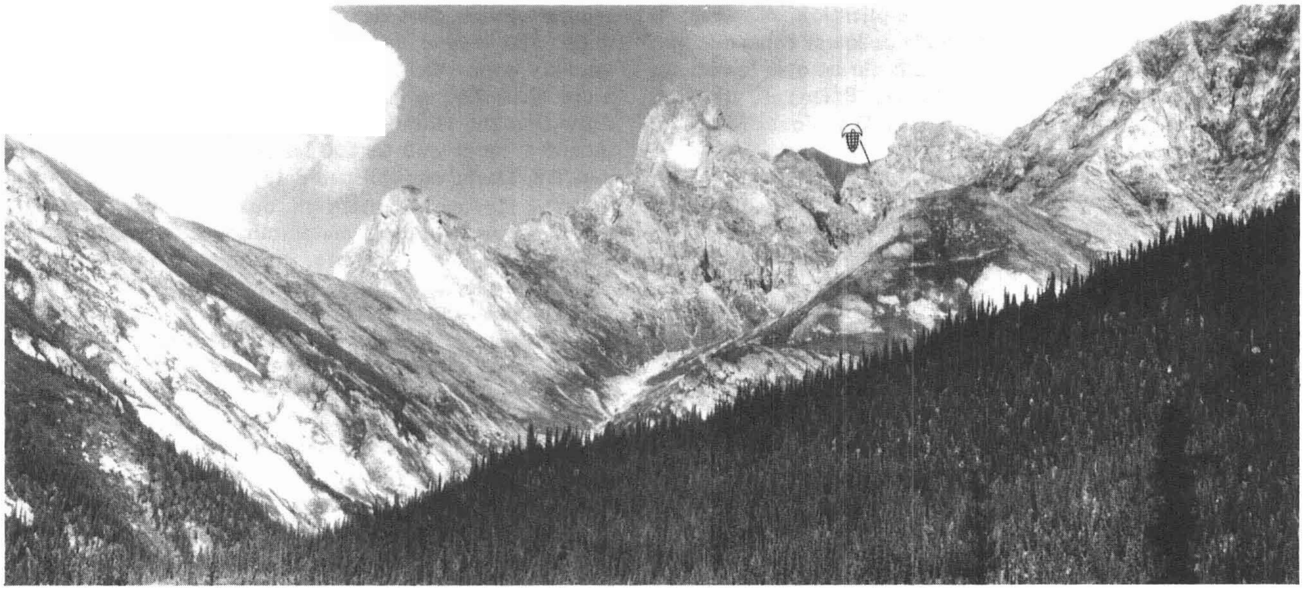

***Figure 152.** View eastward from near Mile 216 of limestone pinnacles of mostly Devonian Skajit formation on north side of Snowden Mountain. Cambrian trilobites were recovered at base of cliff east of saddle at right center. Photograph by C.G. Mull, July 1987.*

orange-weathering muscovite-chlorite-quartz schist and quartzite, chloritic calcareous schist, dolomitic phyllite, calcareous shale-chip siltstone, sandstone, and granule conglomerate, sandy marble, and gray, green, and purple phyllite (fig. 151, Ocs, Gcq, **O€vp**). On strike across the **Dietrich** River valley south of Big Jim Creek, rocks of similar lithology include calcareous shale-chip siltstone and sandstone (fig. 151, **O€** c). This unit grades with decreasing grain size into gray, black, purple, and green phyllite (OC vp); unit **O€**c grades with increasing grain size into polymictic granule to boulder conglomerate with clasts of intermediate volcanic rock, quartz, dolomite, and marble set in a calcareous, schistose matrix (fig. 151, **O€vc**).

Middle Ordovician conodonts were found at several localities immediately to the north and south of **Snowden** Mountain in the Chandalar D-6 Quadrangle in black to gray, crinoidal marble interlayers in thinly interbedded, black, carbonaceous, calcareous, and pelitic phyllite, brown calcareous phyllite, and brown dolomitic **mudstone** (fig. 151, Om and Obpm) (A.G. Harris. written commun., 1985). Four conodont localities are on the south side of **Snowden** Mountain, where **fossiliferous** Devonian quartzite, black phyllite, and marble overlie the **conodont-bearing** Ordovician black phyllite and marble (Brosgé and Reiser, 1964). Three Ordovician conodont localities are in similar rocks exposed 7 km north of **Snowden** Mountain in upper Nutirwik Creek. Megafossils from thick marble interlayers within this section are also Middle Ordovician (J.T. Dutro, Jr., oral commun., 1985). Orange dolomite and thick gray marble that contain Late Ordovician to Silurian **megafossils** are thrust onto Middle Ordovician and Devonian rocks of Whiteface Mountain volcanics in this area.

Near the Big Spruce, Sheep Creek, and Twin Lakes areas of the central Chandalar Quadrangle, chloritic and calcareous schist (Ocs) is overlain by black phyllite and marble **(Obpm)**. These rocks are assigned an early Paleozoic age because they overlie Proterozoic(?) or lower Paleozoic banded-and-knotty schist (fig. 151, **PzPsq)** and are overlain by thick marbles of the Skajit Limestone (fig. 151, Dsk). In this area, Early or Middle Devonian granitic plutons intruded the banded-and-knotty schist and lower Paleozoic(?) rocks and then the overlying Skajit (sheet 2).

Local correlations

The lower Paleozoic rocks of fault panel 6 are similar in age, lithology, and fauna to the pre-Mississip**pian** rocks of the North Slope subterrane in the Doonerak fenster (fault panels **1** and 2). Both areas have early Middle Cambrian trilobites of Siberian affinities (Palmer and others, **1984),** Middle Ordovician black phyllite and marble (A.G. Harris, oral commun., **1985),** and Cambrian or Ordovician intermediate volcaniclastic rocks **(O€vc)** (Brosgé and Reiser, 1971; Mull and others, chap. 14). The lower Paleozoic rocks of the two areas are probably related and are thought to once have been continuous beneath most of the area north of fault panel 9 (sheet 2). They were apparently deeply eroded and then unconformably overlain by Middle Devonian calcareous and quartzose sediments in the Hammond and Endicott Mountains subterranes and by Mississippian sediments farther north in the North Slope **sub**terrane.

Two additional lithologic correlations of the lower Paleozoic rocks are plausible. First, calcareous lower Paleozoic(?) rocks southeast of fault panel 6 are **ap**parently in contact with lithologically similar rocks of fault panel 9 **(PzPcs, PzPm)**. The boundary between fault panels 6 and 9 has not been traced farther east because lithogically similar calcareous rocks occur on both sides of it, which makes the contact difficult to map. Because of the lithologic similarity between **PzPcs** and **PzPm,** the calcareous rocks of fault panel 6 may be lower Paleozoic. An important implication of this correlation is that the Proterozoic(?) or lower Paleozoic banded-and-knotty schist was probably thrust over lower Paleozoic(?) calcareous schist (PzPcs, PzPm) in the Emma Creek antiform as suggested by D.L. Jones (oral commun., 1983, 1984).

A second plausible correlation involves pink-weathering calcareous, black, silty phyllite with inter-layers of black-brown crystalline dolomite that crops out west of the Dalton Highway in the southwestern part of fault panel 6. These rocks have been tentatively assigned to unit Dbs (fig. 151; sheet 2). However, no fossils have been found in them and they may be lower Paleozoic because they are lithologically similar to lower Paleozoic rocks of the Doonerak fenster and the Snowden Mountain area (Dillon and others, 1986).

Regional correlations

The lower Paleozoic rocks of fault panels 1, 2, and 6 may be correlated with coeval strata elsewhere in the Brooks Range. In the Romanzof Mountains, 180 mi (290 km) northwest of the Dalton Highway, and the Baird Mountains, 260 mi (420 km) west of the highway, Proterozoic(?) or lower Paleozoic basement rocks occur beneath middle Paleozoic and younger rocks. Structural, stratigraphic, and petrologic evidence exist for lower Paleozoic rocks in the schist belt surrounding the Chandalar pluton 20 mi (32 km) east of the Dalton Highway, the Arrigetch and Igikpak plutons 100 to 140 mi (160 to 225 km) to the west, the Cosmos Hills window 180 mi (290 km) to the west, and the Hub Mountain (Mount Angayukaqsraq) area, 260 mi (420 km) to the west (Dillon and others, 1980; Hitzman and others, 1982). The lower Paleozoic rocks are unconformably overlain by Devonian to Mississippian metasediments and locally stratigraphically overlie older(?), higher grade metamorphic rocks. These high-grade rocks are represented by knotty mica schist in the schist belt and dated Proterozoic(?) plutons and para-gneiss in the Ernie Lake area, 80 (130 km) west of the Dalton Highway, and in the Hub Mountain area. Similarities in the stratigraphic settings and lithologies, fossil ages, and fossil affinities permit correlation of the lower Paleozoic rocks along the east-West extent of the Brooks Range (Dillon, 1985; A.G. Harris, oral commun., 1985).

DEVONIAN PLUTONS AND AMBLER VOLCANIC ROCKS

Strongly foliated Early or Middle Devonian granite and granodiorite plutons intruded into Paleozoic carbonates and calcareous shales are exposed along the Dalton Highway and to the east in the Chandalar Quadrangle (fig. 145, Dgr). The Devonian felsic plutonic rocks in the Chandalar Quadrangle are grouped into two major categories by Newberry and others (1986): 1) the Horace Mountain plutons, and 2) the Baby Creek batholith. The description of these Devonian plutons is modified from Newberry and others (1986).

The Horace Mountain plutons are relatively small (0.5 to 10 mi^2; 1 to 15 km^2) masses that appear to represent cupolas at the roof of a larger, shallowly eroded pluton. The Baby Creek batholith includes four major and several minor deeply eroded granite masses, possibly connected at depth, with a total surface area of more than 140 mi^2 (400 km^2). Rb-Sr and U-Pb ages show that the Horace Mountain and Baby Creek plutons intruded about 380 to 400 Ma (J.T. Dillon, G.R. Tilton, and T.E. Davis, unpublished data).

The Horace Mountain plutons are largely composed of silica-oversaturated, metaluminous, porphyritic biotite ± hornblende granite, hornblende-biotite granodiorite porphyry, leucogranite, and porphyritic muscovite(?)- biotite granite. Porphyritic phases occur in the smaller masses and contain hornblende and feldspar pseudomorphs and quartz phenocrysts in a very fine-grained groundmass. Larger masses are porphyritic in the interior, with blastoporphyritic K-feldspar in a (1 to 2 mm) groundmass. Primary accessory minerals are generally limited to sphene, magnetite, zircon, and apatite.

Zones of quartz-sericite-pyrite-chalcopyrite ± chlorite alteration affect about 25 percent of the rocks of the Horace Mountain plutons at Horace Mountain and near Robert and Big Spruce Creeks. In addition, moderately to weakly developed sericitic alteration without significant copper-porphyry mineralization occurs in most exposures of the Horace Mountain plutons. Epidote-chlorite-calcite propylitic alteration is present locally in Horace Mountain plutons. Pb-Zn-Cu skarns are abundant along the northwestern side of the Horace Mountain plutons.

Rocks of the Baby Creek batholith are dominantly silica-oversaturated, peraluminous, biotite- and muscovite-bearing granites. In the field, these peraluminous granites seem to grade into the metaluminous Horace Mountain granites. The predominant granitic rock of the Baby Creek batholith is very similar in composition, mineralogy, and texture to the Arrigetch-Igigpak and Okpilak S-type granites of the Brooks Range. Slightly porphyritic phases are most common. White mica and apatite are the most common accessory minerals; garnet, pyrite, zircon, and magnetite are present, and hornblende is noticeably absent. Coarse-grained white micas of primary(?) and secondary origins occur throughout the Baby Creek batholith. Tin anomalies are common in some areas around the batholith.

Field relations between the two suites of Devonian plutonic rocks in the Chandalar mining district are not clear because the two suites are mostly present in separate masses. Distinct contacts between phases have not been found. However, the Horace Mountain pluton crops out within 1.2 m (2 km) of the Baby Creek batholith near Phoebe Creek. There the two suites have a broad overlap of phases and textures. These relationships are most easily explained if the suites are consanguineous. Field relations are complicated because granitic members and leucocratic phases of the two suites are identical in hand specimen and are distinguishable only through petrography and whole-rock chemical analysis. Moreover, both have strong foliations that obscure original textures and structures. Measured isotopic and geochemical signatures overlap. Genetic relations of the two suites are also unclear; they could either represent one magma modified by contamination, differentation, or some other process, or two different magma types.

Newberry and others (1986) conclude that, chemically and mineralogically, the Horace Mountain plutons and the Baby Creek batholith represent I- to S-type granitic suites (Chappell and White, 1974), respectively, and that most other major Brooks Range granitic plutons are S type. They distinguished the S-type granites as two-mica granites that contain 1 to 3 percent corundum, but no hornblende, and have $K_2O/K_2O + Na_2O$ values between 0.4 and 0.7 and $Al_2O_3/Na_2O + K_2O + CaO$ values between 1 and 1.8. White and others (1986) show that these criteria prove only that the granites are mildly peraluminous and that peraluminous granites can be derived from either S- or I-type melting. Consequently, the genetic implications of Chappel and White's (1974) classification scheme may not apply to the Chandalar granites. Although Newberry and others (1986) concluded that the suites represent different magmas, I favor the hypothesis that the two suites represent different levels of erosion into variably contaminated cogenetic plutons.

Geologic and geochronologic data indicate that the foliated Chandalar plutons are of Early and Middle Devonian age; the data also permit only a slight difference in age between the Horace Mountain plutons and the Baby Creek batholith. Pb-alpha dating of zircon from the Baby Creek area yielded an age of 380 to 400 Ma (Brosgé and Reiser, 1964); zircon Pb-Pb dating yields ages of 370 to 400 Ma. Projected on a concordia diagram, the U-Pb data yielded an upper intercept (magmatic) age of 402 ± 8 Ma for the Horace Mountain plutons, 382 ± 20 Ma for the Baby Creek batholith, and 390 ± 15 Ma for all the Chandalar plutons. Rb-Sr studies of the Baby Creek batholith yield an age of 380 ± 12 Ma and an initial Sr ratio of 0.707, although there is unresolved scatter (T.E. Davis, written commun., 1985). The isotopic data neither reject nor support any small differences in age between the two plutonic suites that may be proposed to account for observed petrologic differences. K-Ar determinations on muscovite and biotite from the Chandalar plutons yield ages of 100 to 125 Ma (Brosgé and Reiser, 1964). These K-Ar ages presumably date cooling from mid-Mesozoic M3 metamorphism.

Preliminary Rb-Sr whole-rock data indicate that the mafic component of the Devonian Ambler volcanic rocks has low initial Sr values, indicating a mantle origin similar to that proposed for I-type granitic rocks by Newberry and others (1986). The formation of the mafic and felsic extrusives of the bimodal Ambler volcanics may be related to the apparent synchronous intrusion of I- and S-type granites. In this hypothesis, the mafic volcanic and I-type granitic components are mantle melts, whereas the felsic volcanic and S-type granitic components are fused parts of the continental crust.

Devonian bimodal metaplutonic and metavolcanic rocks lie in parallel, west-trending belts in the southern Brooks Range. Distribution of the plutonic and volcanic rocks overlaps in volcanic centers found south of the Doonerak fenster in the Arctic Alaska terrane in the Wiseman, Chandalar, Coleen, and Survey Pass Quadrangles. The Devonian age is interpreted from isotopic analyses of U and Pb of over 40 zircon fractions from these felsic metaigneous units. Considering concordia plots, Pb-Pb ages from the discordant zircon fractions, and fossil ages derived from marbles intercalated in the volcanic sequences, Dillon and Tilton (1985) find the same age range from 360 to 410 Ma in volcanic rocks as was found in the plutons. This range is attributed to variations in crystallization ages and analytical error and to the U-Pb systematics of the Brooks Range zircons. Overlapping age and distribution of the Devonian plutonic and volcanic rocks provide evidence for cogenesis. Devonian granites of Arctic Alaska are correlated with Devonian granitic batholithic rocks of the North American Cordillera.

Lower intercepts on U-Pb concordia diagrams for these zircons of the volcanic and plutonic rocks range from 105 to 150 Ma and bracket the end of lead loss that results from metamorphism. The age of this metamorphic event corresponds to the Late Jurassic and Neocomian emplacement of the Angayucham terrane.

DEVONIAN VOLCANIC AND SEDIMENTARY ROCKS

Devonian rocks in fault panel 6 include three main units (fig. 151): 1) a lower unit of Middle Devonian coarse, siliceous, clastic metasediments (Dsc, Dwg) overlain in part by marble and dolomite of the Skajit Limestone (Dsk) and felsic metavolcanic rocks of Whiteface Mountain volcanic rocks with associated purple and green phyllite and tuff (Df, Dbf, Dwpg); 2) a middle unit of finegrained Middle or Upper Devonian graphitic, calcareous, and pelitic metasediment (Dbs, Dbl, Dwg) conformably overlain by metasandstone and conglomerate (Dbcp); and 3) an upper unit of unconformable overlying Upper Devonian Hunt Fork Shale (Dhf).

The Middle Devonian siliceous clastic rocks (Dsc) lie unconformably(?) on lower Paleozoic(?) rocks and grade laterally and upward into the Skajit Limestone (Dsk) and Whiteface Mountain volcanic rocks (Dwg). The siliceous clastic rocks (Dsc) are green- to buff-weathering, partly calcareous chloritic metasiltstone, sandstone, and quartz conglomerate with some carbonate clasts. They contain local interlayers of limestone; gray, green, and purple phyllite; chlorite-quartz grit; and felsic volcaniclastic(?) rocks.

The Whiteface Mountain volcanic rocks are laterally equivalent to parts of the Beaucoup Formation, the Skajit Limestone, and the Middle Devonian siliceous clastic rocks. The Whiteface Mountain volcanic rocks are composed mainly of a lower black, chloritic phyllite unit with thin lenses of brown, finely crystalline dolomite (Dwb) and an upper unit of buff-weathering, green-gray calcareous phyllite, graywacke, and conglomerate with limestone interlayers. Distinguishing lithologies include felsic volcanic and volcaniclastic interlayers (Df, Dfm) and maroon, purple, and green phyllite and tuff (Dwpg). In most places, the Whiteface Mountain volcanics form relatively subdued slopes with few good exposures (fig. 77).

The exact stratigraphic position of the Whiteface Mountain volcanic rocks is not known. In some places, it occur beneath the Skajit Limestone and may include undifferentiated Ordovician or Lower Devonian volcaniclastic rocks. Elsewhere, the Whiteface Mountain volcanic rocks seem to grade laterally into the Beaucoup and Skajit formations. These volcanic rocks are tentatively considered to be coeval with all Middle and lower Upper Devonian rock units.

The Devonian Whiteface Mountain volcanic rocks are similar in lithology, age, and stratigraphic position to the mineral-rich Devonian Ambler volcanic rocks and probably formed during the same magmatic event. Felsic volcanic rocks seem to have been erupted throughout Middle Devonian time from volcanic centers near the Mathews River, Koyuktuvuk and Nutirwik Creeks, and Horace and Table Mountains. Coeval strata to the south are eroded and are covered by the Hunt Fork Shale to the east and north; they are exposed only to the west and southwest. Westward from Nutirwik Creek and the Mathews River in the western Chandalar Quadrangle, Middle Devonian units of the Whiteface Mountain volcanic rocks grade laterally into coarse, siliceous, clastic metasediments (Dsc) at about the Dietrich River.

In the same area and along the same southwestern trend, the Skajit Limestone grades laterally southwestward into coarse, siliceous, clastic rock within a short distance. The facies change occurs in the Dietrich River valley. Massive, gray marble over 1,600 ft (500 m) thick forms spectacular cliffs on the east side of the Dalton Highway between Big Jim Creek and Nutirwik Creek (fig. 77). The marble is interbedded with siliceous clastic and volcaniclastic rocks that thicken to the southwest. Across the Dalton Highway on the west side of the Dietrich River valley and north of Big Jim Creek, the Skajit Limestone forms two layers, each less than 150 ft (50 m) thick (fig. 76). It pinches out completely about 20 km to the southwest and does not reappear for over 20 mi (30 km) to the west, where it again thickens rapidly.

North to south facies changes in the Devonian metasedimentary rocks (particularly the Skajit Limestone) are more common in the Brooks Range than east-west facies changes. The northern limit of the Skajit Limestone is a nearly linear east-trending facies change into siliceous clastic rocks of the Beaucoup Formation. The Skajit Limestone probably was deposited in an intertidal environment. Limestone reefs in the Ambler volcanic rocks and the Beaucoup Formation to the south and north, respectively, indicate deposition at slightly greater depths within the euphotic zone. Shallow fault-bounded Devonian basins are inferred to the north and south of the Skajit platform, where the Beaucoup and Ambler formations were deposited. The linear northern and southern facies limits of the Skajit were probably controlled by Devonian growth faults.

In the Wiseman Quadrangle and western Chandalar Quadrangles, the Skajit Limestone consists mainly of massive gray marble (figs. 69, 77, 141, 152, and 153)

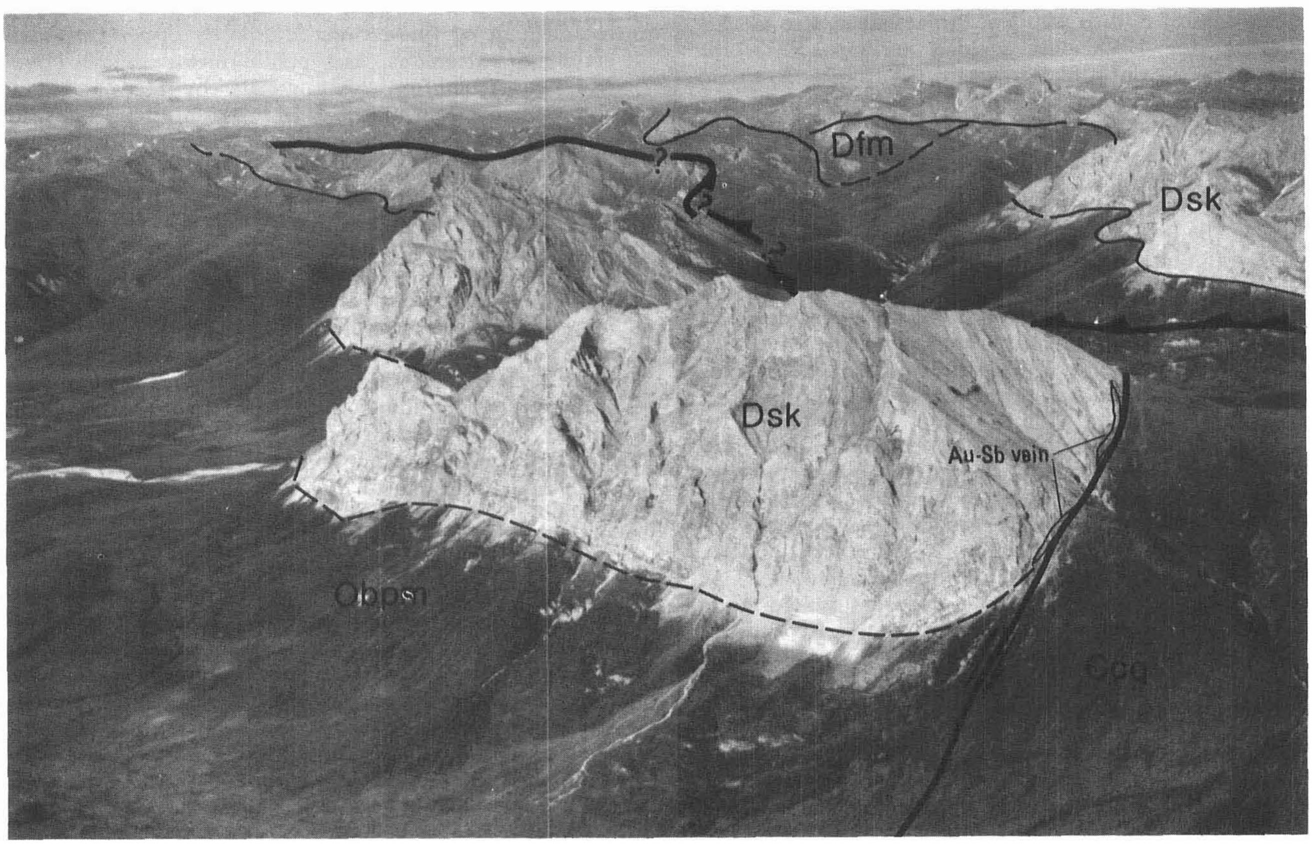

Figure 153. View northeastward toward Sukakpak Mountain (in foreground), which is composed of a synclinal fold of Devonian and older Skajit Limestone (Dsk) that overlies Paleozoic(?) rocks on the north (Obpm) and is in fault contact with lower Paleozoic rocks (Ꞓcq) on the south. Gold-bearing quartz-stibnite vein is present in fault zone on south side of Sukakpak Mountain. In distance, bimodal igneous rocks (Dfm) of Devonian(?) age are present at base of Skajit. Dillon Mountain is visible beyond Sukakpak Mountain at left center. Photograph by C.G. Mull, August 1972.

with layers and replacement bodies of finely crystalline gray dolomite. The Skajit Limestone includes pink-weathering dolomitic phyllite and locally important subunits and interlayers of carbonate conglomerate, quartzite, metabasite, and graphitic and calcareous schist.

Although fossils are rare in the Skajit Limestone, the formation is dated by Middle and early Late Devonian fossils in its type section in the Wiseman Quadrangle. The Skajit Limestone in the Chandalar Quadrangle is Middle Devonian (Brosgé and Reiser, 1964). Elsewhere in the Brooks Range, the Skajit Limestone has been correlated lithologically mainly with Devonian carbonate rocks but locally with Silurian and Ordovician carbonate rocks. These older carbonate units should be assigned to different formations.

Sedimentary structures including algal laminations, fenestral oolites, conglomerate, cut and fill, and dolomite breccia (Henning, 1982) in the marble and dolomite of the Skajit Limestone in the Wiseman Quadrangle indicate deposition in an intertidal environment. The regional paucity of fossils in the Skajit Limestone also supports deposition in an intertidal, hypersaline environment because shell-bearing faunas are rare in analagous modern environments. The fossil-poor nature of the Skajit Limestone cannot be explained entirely by recrystallization. Megafossils are much more common in thin, biothermal marble layers in the Ambler and Whiteface Mountain volcanic rocks, in the Beaucoup Formation, and in the Hunt Fork Shale than in nearby similarly metamorphosed thick marbles of the Skajit Limestone. Furthermore, the regional preservation of small-scale sedimentary structures in the Skajit Limestone suggests--despite metamorphic recrystallization--that fossils should have been preserved if they were originally present.

The Beaucoup Formation in fault panel 6 is thick and lithologically heterogeneous. It consists of Middle and Upper Devonian carbonaceous, siliceous, calcareous metasedimentary rocks and rare felsic metavolcanic rocks. The formation grades laterally and downward into the Whiteface Mountain volcanic rocks and Skajit Limestone. In the souihern part of the Wiseman Quadrangle, the Beaucoup Formation (Dutro and others, 1979) is laterally equivalent to the Ambler volcanic rocks and regionally lies unconformably on lower Paleozoic rocks.

Basal Beaucoup Formation conglomerate has been found unconformably overlying older rocks in some places. Where the Beaucoup Formation grades downward into the Skajit Limestone (Dsk) or laterally into the black phyllite of Whiteface Mountain volcanic rocks (Dwb), it is composed predominantly of black limestone interlayered with black calcareous and pelitic phyllite, black metasiltstone, and thin, dark-gray marble and brown dolomite (Dbs, Dbq, Dbl). The upper part of the Beaucoup Formation is composed mainly of siliceous clastic rocks, including calcareous chlorite-muscovite-quartz metasandstone, grit, and conglomerate (Dbcp), with local interlayers of black phyllite and limestone (Dbs, Dbb, Dbl), purple and green phyllite, and felsic volcanic rocks.

In many places, in absence of paleontologic data, the Beaucoup Formation is difficult to distinguish from the overlying Upper Devonian (Frasnian) Hunt Fork Shale. Hunt Fork Shale is composed predominantly of brown-weathering lithic graywacke and black phyllite with interlayers of fossiliferous limestone and green-weathering quartz sandstone. Bedding and graded bedding are locally well preserved.

The Hunt Fork Shale grades downward into the Beaucoup Formation in some areas but unconformably overlies upper Middle Devonian and older rocks in other areas. Phyllite and semischist of the Hunt Fork Shale conformably overlie the Beaucoup Formation but unconformably overlie the Skajit Limestone and Ordovician or Silurian marble in the Hammond subterrane. Basal conglomerate, discordant contacts, and missing sections at other locations in the Wiseman Quadrangle provide evidence for a regional unconformity beneath the Upper Devonian Hunt Fork Shale.

FAULT PANEL 7 (HAMMOND SUBTERRANE)

The rocks in fault panel 7 are not fossiliferous and their age is unknown. Lithologies within fault panel 7 are quartzite and black, calcareous phyllite with thin gray-marble interlayers (fig. 151, Dbb), black quartzite (fig. 151, Dbq), and black to gray platy limestone (fig. 151, Dbl). These rocks are similar to the fossiliferous carbonaceous phyllite and limestone in the Devonian (fig. 151, Dbs) and Ordovician units (fig. 151, Obpm, Om) of fault panel 6.

The rocks in fault panel 7 are bounded by structural contacts and overlie Middle to Upper Devonian rocks. If fault panel 7 is composed of Devonian rocks, displacement on the bounding faults could be small; if composed of Ordovician rocks, as seems likely, displacement on the faults must be much greater.

FAULT PANEL 3 (ENDICOTT MOUNTAINS SUBTERRANE)

Fault panel 3 is called the Endicott Mountains subterrane (Silberling and Jones, 1984) and the Endicott Mountains allochthon (Mull, 1982; Mull and others, 1987a; chap. 14); it structurally overlies the Doonerak fenster above the Amawk thrust. The Endicott Mountains allochthon encompasses Devonian to Triassic rocks that are exposed mainly between the Doonerak fenster and the north front of the range. However, near the Dalton Highway the Endicott Mountains allochthon is rooted on the south side of the Doonerak fenster beneath thrust plates of lower Paleozoic rocks of the Hammond subterrane.

Devonian rocks in the fault panel consist of the Whiteface Mountain volcanic rocks overlain by various facies of the Beaucoup Formation, which are overlain in turn by the Hunt Fork Shale (fig. 154). These formations were described under fault panel 6. (See also Mull and others, chap. 14.)

The Whiteface Mountain volcanic rocks occur only along the south and east sides of the Doonerak fenster. They are composed of calcareous phyllite, siltstone, and graywacke with limestone interlayers, and intrusions and interlayers of felsic volcanic rocks. Undifferentiated lower Paleozoic rocks may be mapped together with the Whiteface Mountain volcanic rocks (W.P. Brosgé, oral commun., 1984).

The lower Upper Devonian Beaucoup Formation occurs at the top of the Whiteface Mountain volcanic rocks and below the phyllites of the Hunt Fork Shale. The Beaucoup Formation and the Hunt Fork Shale occur on the north, east, and south sides of the Doonerak fenster. The Beaucoup Formation consists of interlayered calcareous chlorite-muscovite-quartz metasandstone, grit, and conglomerate; quartzite; purple and green phyllite; fossiliferous black limestone and phyllite; and local felsic volcanic rocks.

Upper Devonian phyllites of the Hunt Fork Shale are interlayered, brown-weathering graphitic phyllite and metagraywacke with relict bedding, graded bedding, flame structures, and rip-up clasts. On the north side of the fenster, the Hunt Fork Shale grades up into the Upper Devonian to Lower Mississippian Kanayut Conglomerate.

Regionally, the Kanayut Conglomerate is overlain stratigraphically by a sequence of Upper Paleozoic and Lower Mesozoic rock units, including the Mississippian Kayak Shale, the Mississippian to Lower Pennsylvanian Lisburne Group, the Permian Echooka Formation of the Sadlerochit Group and Permian Siksikpuk Formation, and, finally, the Triassic to Jurassic Otuk Formation. Mississippian and younger rocks on the Endicott Mountains allochthon are present only north of the Doonerak fenster, in synclines in the northern Endicott Mountains, and along the northern mountain front. These units have not been recognized on the south side of the fenster. They are described in greater detail by Mull and others (chap. 14).

Stratigraphic differences in the Endicott Mountains allochthon on opposite sides of the fenster are explained by the tendency of thrust faults to cut upsection and involve younger rocks to the north. The Amawk thrust cuts upsection and gradually cuts out the Whiteface Mountain volcanic rocks as they pass from the south side around and over the east end of the fenster through outcrops along the Koyuktuvuk Creek and klippen near Falsoola Mountain (sheet 2). Similarly, the Trembley Creek thrust and Table Mountain thrust that lie above the Endicott Mountains allochthon apparently cut out the Kanayut Conglomerate and the upper Paleozoic rocks along the south side of the Doonerak fenster but pass above them on the north side of the fenster. Only the Hunt Fork Shale and the Beaucoup Formation appear to be continuous in the Endicott Mountains allochthon from the south to the north side of the Doonerak antiform.

FAULT PANELS 1 AND 2 (NORTH SLOPE SUBTERRANE)

Rocks in fault panels 1 and 2 (fig. 154) are exposed in the core of the Doonerak fenster and constitute the North Slope subterrane of Silberling and Jones (1984) and Jones and others (1987). They are stratigraphically similar to the basement of the Arctic Slope and northeastern Brooks Range and are described in detail by Mull and others (chap. 14).

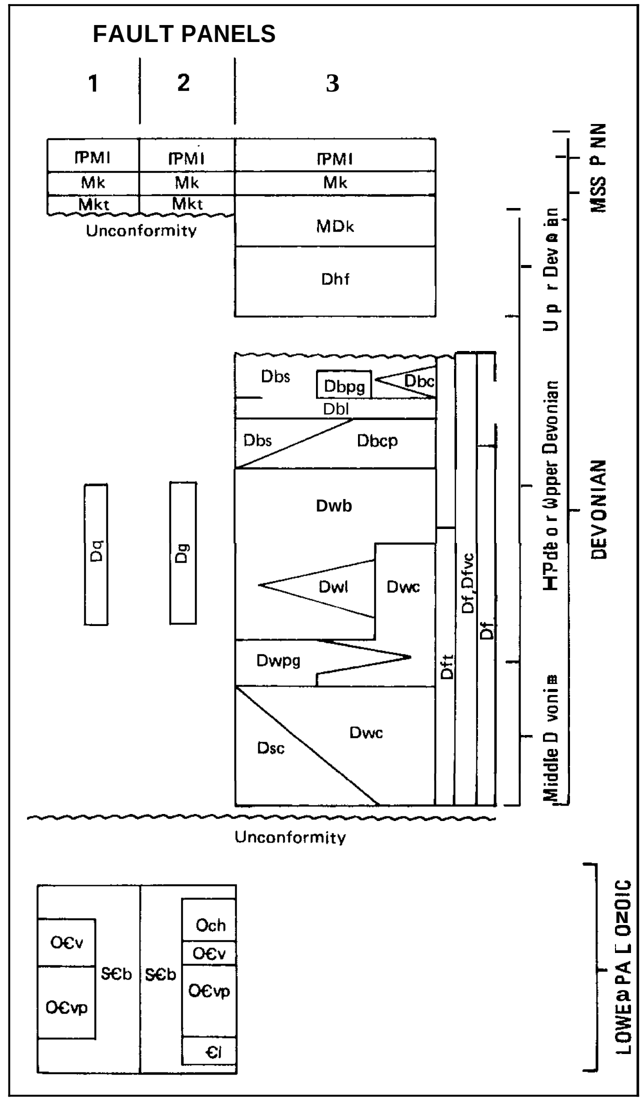

Figure 154. Correlation diagram for rock units in fault panels 1, 2, and 3.

The lower Paleozoic basement consists of dark, siliceous calcareous phyllite with interlayered mafic to intermediate volcanic rocks (S€b). Fossil-bearing interlayers in the phyllite include limestone with early Middle Cambrian trilobites (€l) (Dutro and others, 1984a); chert interlayers with Ordovician conodonts (D.L. Jones and A.G. Harris, oral commun., 1984); and black phyllite with Silurian graptolites (Michael Churkin, oral commun., 1983). The mafic to intermediate volcanic rocks (O€v) and related volcaniclastic rocks, conglomerate, and purple and green phyllite (OCvp) were dated by K-Ar whole-rock and Ar^{39}-Ar^{40} incremental-heating methods and yielded Late Cambrian to Middle Ordovician ages of 465 to 520 Ma (Dutro and others, 1976). These rocks have been informally named the Apoon assemblage (Oldow and others, 1987d). Lower Paleozoic rocks of similar age and lithology are found near Snowden Mountain in fault panel 6 (fig. 151, €l, O€vc, O€vp, S€b) and in the

Romanzof Mountains of the northeastern Brooks Range, 180 mi (290 km) northeast of the Doonerak fenster.

Devonian rocks have not been recognized in the Doonerak fenster. They were either not deposited in the Doonerak area or were deposited and then stripped by erosion before deposition of the Mississippian Kekiktuk Conglomerate or have yet to be recognized (Moore, 1984). Locally, in the western part of the fenster, Dillon and others (1986) mapped Devonian(?) epiclastic-volcaniclastic rocks that are lithologically similar to those in the structurally overlying Devonian Whiteface Mountain volcanic rocks and Beaucoup Formation of fault panel 3. Thin, tuffaceous shales have also been found in the Mississippian Kekiktuk conglomerate.

A sequence of Mississippian to Triassic rocks unconformably overlies the lower Paleozoic rocks within the fenster. This sequence includes the Mississippian Kekiktuk Conglomerate and Kayak Shale, the Mississippian to Lower Pennsylvanian Lisburne Group, the Permian Echooka Formation of the Sadlerochit Group, and the Triassic Shublik Formation and Karen Creek Sandstone (Dutro and others, 1976; Mull, 1982; Mull and others, chap. 14). This stratigraphic sequence is otherwise confined to the Arctic Slope and northeastern Brooks Range. It differs from stratigraphic sequences in the Endicott Mountains allochthon (subterrane) to the north in that 15 to 50 ft (5 to 20 m) of Kekiktuk Conglomerate exposed in the core of the Doonerak fenster is replaced by more than 7,500 ft (2,500 m) of the nonmarine to marine Kanayut Conglomerate and Noatak Sandstone and a comparable thickness of Hunt Fork Shale on the allochthon. Furthermore, the Permian and Triassic rocks of the allochthon are more siliceous (cherty) and apparently more distal than those in the fenster. Mull and others (chap. 14) show that the Endicott Mountains allochthon is more similar stratigraphically to rocks of the southcentral Brooks Range and has been telescoped a minimum of 55 mi (88 km) over the parautochthonous rocks in the core of the fenster. They suggested that movement along the Amawk thrust may have been two to three times the minimum distance calculated.

Despite these differences in lithology, thickness, and stratigraphy, the upper Paleozoic rocks of the Doonerak fenster, Arctic Slope, and northeastern Brooks Range are similar enough to those of the Endicott Mountains allochthon that all of these rocks must represent parts of an originally continuous sedimentary basin. Therefore, the Amawk thrust, though an important thrust fault, does not represent a terrane boundary. Moreover, post-Mississippian terrane boundaries are also precluded in the Hammond, Endicott Mountains, and North Slope subterranes.

Within the Doonerak fenster along the Amawk thrust (Mull, 1982) and Blarney Creek thrust (Julian and others, 1984), the lower Paleozoic basement rocks are locally imbricated onto the Mississippian rocks. Deeper within the fenster, the St. Patricks Creek thrust cuts mainly through lower Paleozoic rocks. Involvement of lower Paleozoic basement in thrust faults within the North Slope subterrane in the Doonerak fenster (fault panels 1 and 2) and in the Hammond subterrane (fault panel 6) suggests that the lower Paleozoic basement rocks in the Doonerak fenster are parautochthonous.

Low-grade M_3 and M_2 metamorphism affected all rocks in the North Slope subterrane in the Doonerak fenster (Dillon, 1982; Julian and others, 1984; Dusel-Bacon and others, 1989). Cambrian through Lower Triassic rocks within fault panel 2 were recrystallized twice during the M_3 and M_2 phases of metamorphism. Greenschist-facies recrystallization during the M_2 metamorphism produced a pervasive metamorphic foliation. Deformation during the M_3 metamorphism produced kink folds of S_2 cleavage and formed phyllitic axial-plane cleavage S3. These structures affect both the upper and lower plate rocks of the Amawk thrust zone (Julian and others, 1984). The fact that these structures clearly affect Mississippian rocks and appear to affect Lower Triassic rocks in fault panel 2 provides a lower age limit of Lower Triassic(?) for M_2 and M_3 metamorphism in the Arctic Alaska terrane.

Similar structures are interpreted as evidence for the presence of M_2 and M_3 metamorphism in Mississippian and Triassic rocks of fault panel 1. However, the metamorphic grade of fault panel 1 is that of pumpellyite-prehnite facies, so it is difficult to precisely correlate the barely visible, locally developed cleavage found in Triassic rocks of fault panel 1 with easily seen, regionally developed cleavage in higher grade Devonian and Mississippian rocks of fault panels 2 and 3, respectively.

The St. Patricks Creek thrust fault forms the boundary between fault panels 1 and 2 and apparently displaces greenschist-facies rocks of fault panel 2 onto pumpellyite-prehnite facies rocks of fault panel 1. The truncated metamorphic facies have two important implications: 1) the St. Patricks Creek thrust fault must have at least 6 mi (10 km) of slip to account for the truncation; and 2) the thrusting must have occurred after M_2 metamorphism.

CHAPTER 11.

REGIONAL SIGNIFICANCE OF THE JIM RIVER AND HODZANA PLUTONS

By J.D. Blum,[1] J.T. Dillon,[2] and A.E. Blum[3]

INTRODUCTION

The Jim River and Hodzana plutons (figs. 58 and 155) are two of many Lower Cretaceous plutons that intruded the Ruby geanticline. The Jim River pluton also intruded the narrow band of Devonian to Jurassic oceanic crustal rocks of the Angayucham terrane that separates the Precambrian to lower Paleozoic metasediments of the Ruby geanticline from the Cretaceous sediments and volcanic-arc rocks of the Yukon-Koyukuk basin. The Hodzana pluton intruded both the Angayucham terrane and metasedimentary rocks of the Ruby geanticline about 110 Ma, thus stitching the terranes together.

Reconnaissance studies of other Cretaceous plutons in northcentral Alaska have shown distinct compositional and isotopic differences between the plutonic rocks that intruded the Ruby geanticline and those that intruded the Yukon-Koyukuk basin (Arth and others, 1984; Miller, 1984; Puchner, 1984). Plutons of the Ruby geanticline are 104 to 111 m.y. old, are composed of biotite and two-mica granite, have initial $^{87}Sr/^{86}Sr$ values of 0.706 to 0.730 and $\delta^{18}O$ values >8.5. Plutons of the western Yukon-Koyukuk basin are 99 to 110 m.y. old, are composed of hornblende-pyroxene monzonite, syenite, quartz monzonite, and associated subsilicic alkaline phases, and have no reported isotope values. Plutons of the eastern Yukon-Koyukuk basin are 80 to 84 m.y. old, are composed of hornblende-biotite tonalite, granodiorite, and granite, and have initial $^{87}Sr/^{86}Sr$ values of 0.7038 to 0.7047 and $\delta^{18}O$ values <8.5.

Arth and others (1984) and Miller (1984) concluded that significant amounts of continental crust were either the source of or were incorporated into the magmas of plutons that intruded the Ruby geanticline. They also concluded that plutons which intruded the eastern Yukon-Koyukuk basin are characteristic of island-arc or convergent continental-margin magmatism and that plutons which intruded the western Yukon-Koyulcuk basin are enigmatic in origin. An important question is whether compositional differences between the Lower Cretaceous plutons of the Ruby geanticline and those of the western Yukon-Koyukuk basin are due to magma generation by different processes or result from similar magmas that assimilated different types and amounts of crustal rock during intrusion.

RESULTS OF STUDY

A petrologic and isotopic investigation of the Jim River and Hodzana plutons was initiated to test the importance of wall rock assimilation on the composition and isotopic signature of the Lower Cretaceous plutons. Preliminary data are reported below; details are reported by Blum and others (1987). Modal volume and field data show that most of the Jim River pluton is composed of biotite-amphibole-pyroxene syenite and monzonite, with a central core of biotite-amphibole granite; the northwestern Hodzana pluton is composed of biotite-amphibole monzodiorite and granite. Twenty K-Ar age determinations from biotite and amphibole of the Jim River and Hodzana plutons range from 103 to 111 Ma and average 106 ± 6 Ma. A three-point whole-rock Rb-Sr isochron for the Jim River pluton yields an age of 111 Ma and an initial $^{87}Sr/^{86}Sr$ value of 0.7079. A five-

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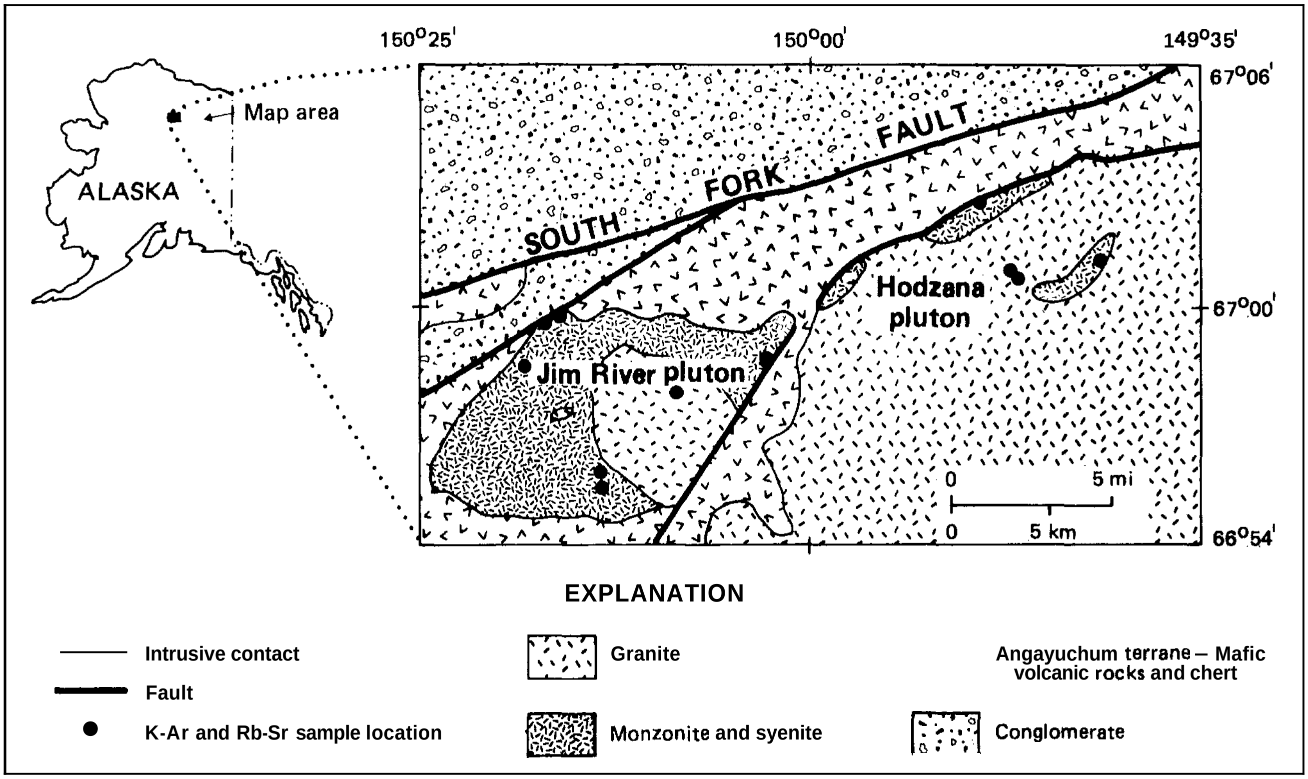

Figure 155. Generalized geologic map of the Jim River-Hodzana plutons area. Map modified from Blum and others (1987).

point whole-rock Rb-Sr isochron for the Hodzana pluton yields an age of 108 Ma and an initial $^{87}Sr/^{86}Sr$ value of 0.7078. Three whole-rock $\delta^{18}O$ values for the Jim River and Hodzana plutons range from 6.75 to 7.30.

These data indicate that the Jim River pluton and northwestern Hodzana pluton have different compositions, lower initial $^{87}Sr/^{86}Sr$ values, and lower $\delta^{18}O$ values than most of the other Lower Cretaceous plutons that intruded the Ruby geanticline. They are more similar, compositionally and isotopically, to the plutons that intruded the Yukon-Koyukuk basin. The unique syenitic to monzonitic composition of the Jim River pluton may represent a more primary mantle-derived magma that underwent minor contamination where it intruded oceanic rocks of the Angayucham terrane and severe contamination where it intruded continental rocks of the Ruby geanticline. This may help to explain why mostly syenitic to monzonitic magmas are found in the Angayucham terrane, mostly monzodioritic magmas are found in the transition zone, and mostly granites are found in the Ruby geanticline.

Contact relationships at the present depth of exposure suggest that wall rock assimilation was severe where the Hodzana pluton intruded the Ruby geanticline and minimal where the Hodzana and Jim River plutons intruded the Angayucham terrane. Wall rock composition and the amount of assimilation appear to be controlling factors in the composition and isotopic signature of the plutons. These plutons may record a transition in magma composition at the terrane boundary between the Angayucham terrane and the Ruby geanticline.

CHAPTER 12.

GRAVITY-AND-MAGNETIC MODEL OF THE DALTON HIGHWAY CORRIDOR THROUGH THE BROOKS RANGE

By D.T. Smith[1] and J.T. Dillon[2]

INTRODUCTION

Rocks in the Wiseman and Chandalar Quadrangles of the central Brooks Range have been metamorphosed twice and have undergone several deformations. These rocks are cut by at least two distinct types of faults: 1) southerly dipping thrust faults that imbricate oceanic rocks of Angayucham terrane onto continental rocks of the Brooks Range; and 2) high-angle faults with displacements of 300 ft to several miles that offset the thrust faults. No well logs, seismic lines, or exploratory holes are available to determine subsurface geology; however, detailed 1:63,360-scale geologic maps, gravity measurements, and aeromagnetic maps of the Dalton Highway corridor have recently been completed. In an attempt to define parameters of the subsurface structure, a two-dimensional gravity-and-magnetic model was made of the Dalton Highway corridor from the South Fork Koyukuk River to Chandalar Shelf (see sheet 2). Details of the model are given by Smith (1986).

METHODOLOGY

Two-dimensional gravity-and-magnetic models were made by drawing polygons that represent rock bodies of similar physical properties on the geologic cross section. The polygons, assumed to be horizontally infinite perpendicular to the cross section, were assigned density and magnetic-intensity values. These values were entered into a computer program that computed the gravity-and-magnetic profile of the model at desired points along the cross section by a numerical technique of Talwani and others (1959).

Where computed values compare favorably to corresponding measured values, the model is acceptable but not necessarily unique. Where computed values do not compare favorably, the density or magnetic value assigned to a polygon(s) is incorrect, or the shape of the polygon(s) is incorrect, or the geologic structure does not meet the two-dimensional requirement. The model was refined by adjusting the densities or geometries, or both, of the polygon(s) until a satisfactory fit between the observed and calculated values was obtained. Polygon adjustments were constrained by known geology, physical rock properties, and plausible changes in geology and rock properties at depth.

Because of the nonunique nature of gravity and magnetic solutions, more than one model can usually be found to explain a given set of anomalies. For this reason, the close correspondence of the gravity or magnetic profile that we calculated with those that are observed prove only that our geologic cross section is plausible.

GEOPHYSICAL DATA

The observed gravity values were obtained from the U.S. Geological Survey and are based on measurements made by the National Geodetic Survey at roughly 1-mi (1.5 km) intervals along the Dalton Highway in the Brooks Range. Bouguer gravity values were calculated by assuming a density of 2.67 g/cm^3; the values were not terrain corrected.

The effect of not applying a terrain correction can be estimated by comparison with data from a transect of a topographically and geologically similar valley in the Wiseman Quadrangle, west of this study area. Here, the terrain correction values have a general, near-linear

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increase northward up the valley (fig. 156), in contrast to a much greater linear decrease of Bouguer values, and thus would not severely affect the near-surface geologic interpretation. When local variations of up to 5 mgals were superimposed on the trend of terrain corrections, they correlated with abrupt changes in topography. These areas might affect the measured gravity, but no attempt was made to remove this effect.

Density values assigned to the gravity model above 9 mi (14.5 km) (table 4) were based primarily on published values for rocks in the southwestern part of the Wiseman Quadrangle (Hackett and Dillon, 1982), from measured values for rocks collected along the Dalton Highway, and in part on values given in the 'Handbook of Physical Constants' (Daly and others, 1966).

The observed magnetic profile was taken from the aeromagnetic maps of the Wiseman and Chandalar Quadrangles (Decker and Dillon, 1982). The profile was digitized along the cross-section line at points that correspond to the projected gravity stations. A 50,200-gamma regional field with a north-dipping slope of 0.75 gamma/mi was removed from the mapped values.

Magnetic intensities were assigned to the bodies of the model based on physical rock-property data collected by Lindsley and others (1966), Hackett and Dillon (1982), and from measured physical properties of rocks collected along the Dalton Highway (see table 4). Magnetic values given as susceptibilities were converted to magnetic intensities by the relationship

$$I = J + kf,$$

where

I is the total intensity of magnetization in gammas,

J is the natural remanent magnetization in gammas,

k is the volume susceptibility in cgs units,

and

F is the earth's total magnetic field in gammas.

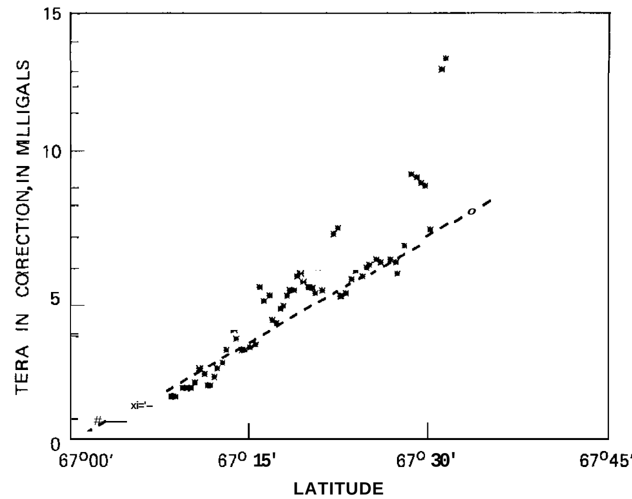

Figure 156. Graph of terrain-correction values vs. latitude for gravity stations west of study area. Stations trend roughly north-south up valley (similar to gravity stations used in this study). Dashed line represents general trend of values; deviations from dashed line are areas where topography steepens abruptly.

The natural remanent magnetization was not known for most rocks in the study area, but the few measured values are small and inconsistent in direction and thus taken as equal to zero. A 0.5 oersted value was used for the magnetic field (U.S. Naval Oceanographic Offiae, 1966).

GRAVITY-AND-MAGNETIC MODELS

Figure 157 and sheet 2 are based on a geologic cross section of the Dalton Highway from 67° to 68° N. latitude. Descriptions of the rock units modeled in this study are provided in chapter 10 and in the explanation of map units for the strip maps in the road log (fig. 82). The location of the cross section was chosen to coincide as closely as possible to gravity stations established along the Dalton Highway by the National Geodetic Survey.

Gravity stations were projected east-west onto the cross section in areas where the two diverged. We used the computer program HYPERMAG (version 1.4) and an interactive two-dimensional gravity-and-magnetic modeling program to compute the gravitational and magnetic effects of the bodies depicted in the geologic cross section (Saltus and Blakely, 1983).

The gravity model presented here is actually two models superimposed: 1) a model of geologic structure above 9 mi (14.5 km), and 2) a model of a crustal root below 20 mi (33 km). The crustal root was included to remove the regional gravity gradient. The root's geometry was manipulated until it fit the regional gradient.

Three assumptions were made when the two models were superimposed. First, the regional gradient was assumed to originate primarily from a -0.42 g/cm^3 density contrast at the crust-mantle interface (lower crust 2.92 g/cm^3, upper mantle 3.34 g/cm^3). This density contrast is typical of continents (Dziewonski and others, 1975). Second, the rock between the base of the modeled geology and the top of the crustal root was assumed to have uniform density and distribution and was also assumed not to contribute to the anomaly pattern. Third, a base density of 2.8 g/cm^3 was assumed (density contrasts were used in modeling).

The parameters of the modeled root agree well with values calculated by others for the Brooks Range. The base of the modeled root reaches a maximum depth of 24 mi (39 km) near the northern edge of the cross section and thins to a depth of 20 mi (33 km) about 10 mi (15 km) south of the southern edge of the cross section. This agrees with an approximate 25-mi (40 km) depth indicated by arrival times for the Brooks Range (Estabrook, 1985), and the 24-mi (39 km) depth esti-

***Table 4.** Density and magnetic-intensity values used to compute two-dimensional gravity-and-magnetic model of Dalton Highway corridor through Brooks Range*

Unit no.	Unit name	Density (g/cm^3)	Magnetic intensity (cgs)
1	Hodzana quartz monzonite plutons	2.725	0.0
2	Cretaceous molasse of the Koyukuk province	2.60	7.5 E-06
3	Angayucham ultramafic	3.01	5.0 E-04
4	Angayucham gabbro	2.93	7.0 E-04
5	Angayucham cherts and basalt	2.88	7.0 E-04
6	Angayucham cherts and phyllite	2.72	5.0 E-06
7	Angayucham graywacke	2.71	5.0 E-06
8	Amphibolite of the Mosquito terrane	2.88	0.0
9	Hunt Fork(?)	2.74	5.0 E-06
10	Knotty mica schist (basement rocks)	2.815 to 2.78	1.0 E-04 to 7.5 E-05
11	Eclogite	2.80	7.5 E-05
12	Calc-schist and marble	2.73	7.5 E-04
13	Devonian granites	2.66	7.5 E-04
14	Hornfels	2.80	7.5 E-04
15	Skajit and Middle Devonian siliceous claatic rocks	2.70 to 2.75	2.0 E-05 to 0.0
16	Sub-Skajit siliceous clastic rocks	2.71 to 2.76	5.0 E-05 to 7.5 E-06
17	Beaucoup Formation	2.70	2.5 E-04
18	Beaucoup and Skajit equivalent	2.70	1.0 E-05
19	Doonerak uPz	2.66 to 2.90	2.5 E-05 to 3.0 E-06
20	Doonerak lPz	2.68	1.5 E-05
21	Hunt Fork	2.65 to 2.68	2.6 E-05 to 7.0 E-06

mated by Woollard and others (1960) on the basis of gravity measurements. By comparison, a 20.6-mi (33 km) crustal thickness at Fairbanks was calculated by Woollard and others (1960) based on gravity data, and a 20-mi (32 km) depth was estimated by Hansen and others (1968) from a seismic-refraction profile. Both of these estimates are similar to the 20.6-mi (33 km) minimum depth of the modeled root.

The polygons of the magnetic model are identical to those of the gravity model, with the exception of the polygon that represents a granitic pluton at Marion Creek. This exception takes into account geologic differences on opposite sides of an important sub-parallel, high-angle fault between the gravity stations and the area of the geologic cross section and magnetic-profile line.

RESULTS

The computed gravity and magnetic profiles are a fairly good match to the observed profiles (fig. 157) and thus support the geologic interpretation given in the cross section. Areas that do not match the observed profiles are discussed below.

One obvious mismatch of the gravity profile occurs at the southern edge of the schist belt. In this area, a fault parallels the cross section, and thus the body does not fulfill the two-dimensional requirement of the modeling program. Moreover, an abrupt northward increase in elevation at the Angayucham fault system probably decreases observed gravity values, because the values have not been terrain corrected.

Two other problem areas are present in the northern part of the section. Here, the observed gravity profile has steps that are not mimicked by the computed gravity. These steps might represent unmapped faults or could be related to topography.

On the computed magnetic profile, the southernmost magnetic high between the South Fork and Malamute fault zone is overestimated by the model. The overestimate is not as much as it appears in the section line, because the measured aeromagnetic data directly to the east and west of the profile reach values comparable to the calculated value. Another important mismatch is present in the area of Minnie and Gold Creeks, between the Wiseman thrust and Sukakpak Mountain, where the magnetic signatures have similar shapes but the calculated profile significantly underestimates the observed profile. Perhaps the magnetic intensity for rocks of the Devonian and older Skajit Limestone and siliceous clastic unit in this area is much higher than that measured in other areas from which the susceptibility measurements were extrapolated. Alternatively, Proterozoic or lower Paleozoic rocks may be closer to the surface than shown in the cross section. This alternative would be supported by the recent discovery farther north in the Chandalar Quadrangle of early Paleozoic fossils in rocks previously thought to be Devonian (Dillon, chap. 10).

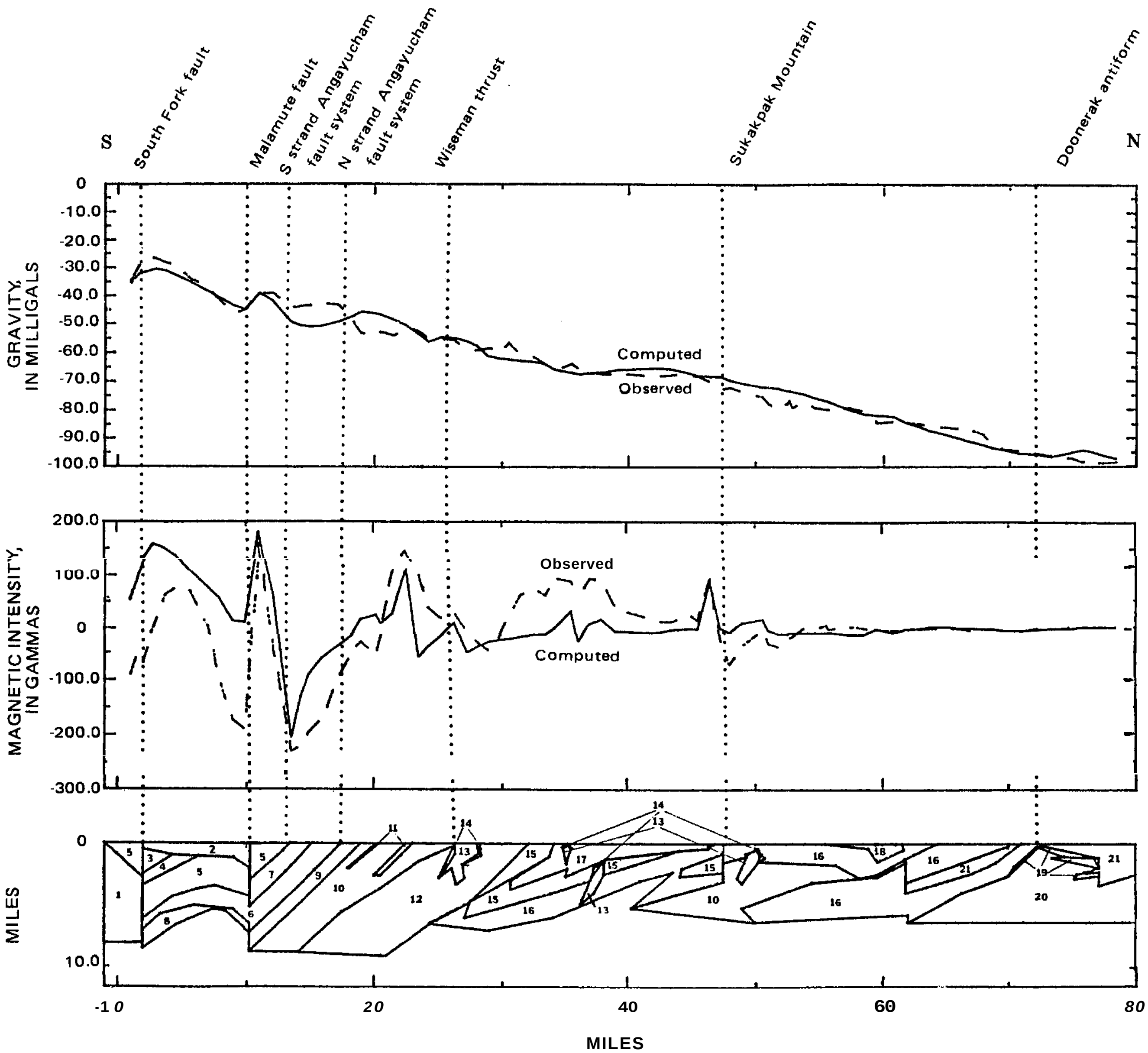

Figure 157. Simplified north-south cross section along Dalton Highway used in gravity-and-magnetic modeling. See figure 139 for line of section and table 4 for description of units.

A sharp dipole anomaly occurs over an unnamed ridge just south of Sukakpak Mountain. This anomaly was modeled as partially originating from a small pluton and related hornfels that have a high magnetic intensity. The anomaly may also be partially caused by local high relief. North of Sukakpak Mountain the magnetic anomalies flatten out and little can be learned from them.

CONCLUSIONS

The two-dimensional models show, to a first approximation, that the geologic cross section accounts for the observed gravity-and-magnetic values and that the depth to the root increases from about 20.6 mi (33 km) in the south to 24.2 mi (39 km) in the north. The models are being improved by inclusion of terrain corrections with the simple Bouguer gravity values used as the observed gravity in this study. We also propose to use two-and-one-half-dimensional modeling where necessary. This is a modeling technique similar to the two-dimensional modeling, except that the polygons have a finite extent perpendicular to the plane of the cross section.

CHAPTER 13.

GOLD DEPOSITS IN THE UPPER KOYUKUK AND CHANDALAR MINING DISTRICTS

By J.T. Dillon,[1] K.K. Lamal,[2] and J.A. Huber[3]

ABSTRACT

In the upper Koyukuk and Chandalar mining districts, placer-gold deposits surround areas of auriferous veins. Because of such spatial and mineralogical relations, we believe these placer deposits were derived from erosion of nearby lode deposits. The lode deposits are multistage epithermal quartz veins, containing stibnite, arsenopyrite, scorodite, and gold in a quartz-calcite gangue, with minor metacinnabar, molybdenite, and galena. The veins were deposited by carbon-dioxide-rich fluids boiling(?) at temperatures of 200 to 300 °C in dilatant faults. Structural evidence within the veins points to intermittent fault displacement during vein emplacement: veins and faults cut Late Jurassic to Early Cretaceous regional metamorphic structures in the country rock, and regional geologic relationships are consistent with displacement during postmetamorphic cooling.

Placer- and lode-gold deposits form a west-trending belt about 12 mi (20 km) wide and 90 mi (150 km) long, indicating a narrow west-trending source. Gold in these deposits has been either remobilized from supracrustal rocks during metamorphic dewatering or has come from deep-crust or mantle rocks.

A mantle source seems less likely than a supracrustal source because emanations from the mantle would be expected to spread out in the lower crust and produce a more diffuse pattern. Furthermore, the deep crust beneath the upper Koyukuk and Chandalar mining districts was probably depleted of its labile components during the Devonian and thus seems an unlikely potential source.

Gold remobilization from auriferous supracrustal sources in lower Paleozoic rocks during metamorphism, or from cupolas of Devonian felsic and mafic plutons, seems more likely. Because the gold deposits are spatially associated with areas where roofs of Devonian plutons occur at depths of a few miles or less, remobilization from the cupolas of plutons is most likely. Whatever the source, the gold has been remobilized during subsequent metamorphic events.

INTRODUCTION

Lode-gold and associated placer-gold deposits of the upper Koyukuk and Chandalar mining districts (fig. 158) have produced more than 320,000 oz of gold since 1893. This report discusses the spatial and geochemical relations between the lode-gold and placer-gold deposits in the Chandalar and upper Koyukuk mining districts and introduces hypotheses on the original gold source.

Geologic maps of the Chandalar and Wiseman Quadrangles (Brosgé and Reiser, 1964, 1970) were used to prepare this report. Information on mineral occurrences in the Chandalar and upper Koyukuk mining districts by Reed (1938), Grybeck (1977a,b), DeYoung (1978), Cobb (1981), Dillon (1982), and Ashworth (1983) was also incorporated. Brosgé and Reiser (1970) published detailed geochemical studies of quartz-stibnite veins in the Wiseman area and concluded that antimony-rich gold veins were the source of local placer-gold deposits there. The timing of metamorphism and formation of fluid inclusions in metamorphic rocks is discussed in Turner and others (1979) and Dillon and others (1980).

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REGIONAL GEOLOGY

The Chandalar and upper Koyukuk mining districts are underlain mainly by Cambrian through Devonian calcareous, carbonaceous, and siliceous clastic metasedimentary rocks locally interlayered with Devonian mafic and felsic metavolcanic rocks. These rocks were intruded by Devonian plutons, regionally metamorphosed twice during Mesozoic time under greenschist to lower amphibolite facies conditions, and later uplifted and faulted. The Paleozoic sedimentary and volcanic rocks were deposited on a basement of Proterozoic(?) and lower Paleozoic metasedimentary rocks.

LITHOLOGIES

Upper Proterozoic(?) to lower Paleozoic, Cambrian, Ordovician, and Silurian basement rocks (fig. 158, **PzPcs, PzPqs, O€sm, S€b**) in the Chandalar and **Wiseman** Quadrangles are exposed in the Doonerak and **Snowden** Mountain areas and are probably exposed near Bettles River, at Ernie Lake, and within the schist belt in the southern Brooks Range.

Proterozoic(?) or lower Paleozoic basement rocks (fig. 158, **PzPcs, PzPqs**) include polymetamorphic quartzose clastic rock, carbonaceous shale, calcareous sandstone, and mafic igneous rock. Pre-Devonian metamorphism probably produced the oldest amphibolite-facies mineral assemblages in these rocks. Lower Paleozoic, Devonian and, locally, Mississippian metasedimentary rocks **unconformably(?)** overlie the Proterozoic(?) and lower **Paleozoic** basement rocks in most parts of the study area (**Dillon** and others, 1980).

Near the Dalton Highway from Robert Creek to the Hammond River, Cambrian rocks are chloritic-quartz schist, chloritic calcareous schist, and local marble; Ordovician rocks in the same area are graphitic phyllite and schist, with thin interlayers of black crinoidal marble. Black carbonaceous, siliceous phyllites of Cambrian, Ordovician, and Silurian ages, local Cambrian marble, and Ordovician intermediate volcanic rocks crop out in the Doonerak fenster north of the map area.

Protoliths of polymetamorphic Devonian rocks are limestone, sandstone, siltstone, and conglomerate, with interbeds. dikes, and sills of felsic and mafic metavolcanic rocks. **The** oldest Devonian unit is unnamed and consists of coarse-grained siliceous metaclastic rocks (Dsc). Massive marbles of the Middle Devonian and older Skajit Limestone (Dsk) overlie the older metaclastic unit (**Brosgé** and Reiser, 1964). Interbedded with the massive marbles are many layers of graphitic, calcareous schists and metamorphosed felsic and mafic volcanic rocks. These interlayers predominate where the Skajit Limestone changes laterally (northward) and upsection into the Middle and Upper Devonian **Beaucoup** Formation (Db) of mixed carbonate and clastic rocks. These units are all overlain by siliceous clastic rocks of the Upper Devonian Hunt Fork Shale (Dh).

Felsic plutons (Dgr) that grade from meta-aplite to augen gneiss intrude the Proterozoic(?) and lower Paleozoic rocks. Coarse-grained augen gneiss grades outward into finegrained, hornblende-bearing **grano**diorite gneiss that intrudes hornfelsed and altered country rock. Injection migmatite and high-grade hornfels (Dt) are common along the contacts of the most deeply eroded plutons. Metamorphosed hornblende granodiorite, aplite dikes, copper and **molyb**denite porphyries, copper-zinc-silver skarn, and other altered rocks are most commonly found around the cupolas of partially unroofed plutons. The hornblende granodiorite may be comagmatic with the augen-bearing quartz monzonite or may represent a separate magma of approximately the same age. Similarly, metagabbro plutons that are abundant near some of the felsic plutons may be genetically related to them.

Zircons from several of the large plutons yield preliminary U-Pb ages of 390 $\pm$ 20 Ma. Data from preliminary whole-rock Rb-Sr analyses of several plutons form a poorly defined **isochron** that yields an initial Sr-isotope ratio between 0.705 to 0.712 and an apparent age between 320 to 385 Ma.

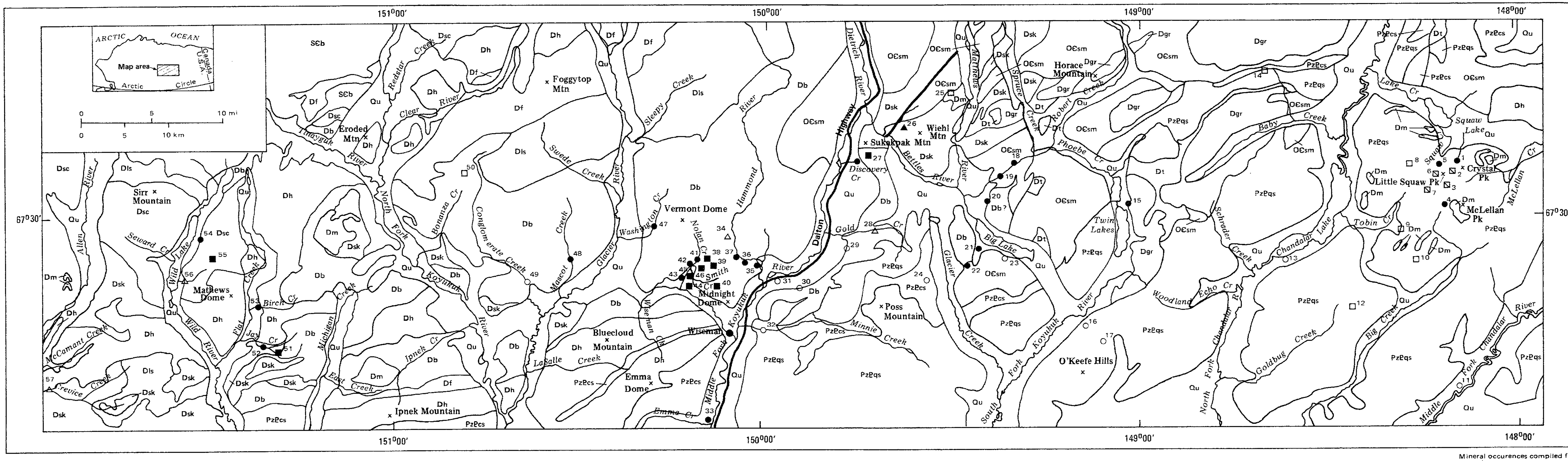
Felsic and mafic volcanic rocks are locally **inter**layered with Devonian limestone. These volcanic rocks are extrusive equivalents of the granitic and mafic plutons which they surround and are particularly significant because they are stratigraphic equivalents of the copper-rich felsic Ambler volcanic rocks. The volcanics have been traced across the Brooks Range into this area from the Ambler mining district, where they are the host rock for world-class massive-sulfide deposits (Hitzman and others, 1982).

Quaternary alluvium, colluvium, and glacial deposits unconformably overlie the polymetamorphic rocks (Hamilton, **1978b, 1979b)** and contain the placer gold of the upper Koyukuk and Chandalar mining districts.

METAMORPHISM AND STRUCTURE

The oldest metamorphic episode may be recorded by the pre-Mississippian cleavage of the Proterozoic(?) and lower Paleozoic basement rocks. Later, the Devonian, Mississippian, Permian, and upper Triassic rocks of the central **Wiseman** Quadrangle were metamorphosed twice (**M_2** and **M_3**) (**Dillon**, chap. 10). Metamorphic grade for **M_2** and **M_3** metamorphism increases to the south. **M_2** metamorphism produced a penetrative cleavage defined by **albite-epidote-amphibo**lite facies minerals; **M_3** metamorphism produced a semipenetrative cleavage defined by upper greenschist facies minerals. **M_3** metamorphism occurred under 5 to 6 kb pressure at 350 to 450 $^{\circ}$C. Because **M_3** metamorphism occurred during Late Jurassic to Early Cretaceous thrusting (**Dillon** and others, **1980**), M_2 metamorphism must have occurred between Late Triassic and Early Cretaceous time.

Many ages of faulting have been recognized within the study area, but only two are mentioned here. Late Jurassic or Early Cretaceous thrust faults were first



Mineral occurences compiled from Brosgé (1970), Cobb (1981), and Dillon (unpublished data, 1982).

MAP SYMBOLS

- Contact
- High-angle fault
- Gold-bearing quartz vein
- Productive quartzgold vein
- Quartz-stibnite-gold vein
- Stibnite occurrence
- Gold-placer prospect
- Productive gold placer
- A Gold placer containing fragments of stibnite-quartz vein

DESCRIPTION OF MAP UNITS

- **Qu** Quaternary deposits, undifferentiated
- **Dh** Upper Devonian Hunt Fork Shale—Carbonaceous phyllite and metagraywacke
- **Dls** Middle or Upper Devonian(?) black rocks—Black, partly calcareous phyllite; carbonaceous, micaceous limestone; and black quartzite
- **Db** Middle and Upper Devonian Beaucoup Formation—Chloritic siltstone, grit and conglomerate, and schist and graywacke. Unit is locally calcareous and carbonaceous
- **Dsk** Devonian Skajit Limestone—Limestone, dolomite, and marble
- **Dsc** Devonian siliceous clastic rocks—Chloritic quartzite, quartz phyllite, grit, and conglomerate
- **Df** Devonian felsic schist—Felsic metavolcanic and associated metasedimentary rocks
- **Dt** Devonian tactite—Calc-silicate hornfels and skarn
- **Dgr** Devonian granitic gneiss—Muscovite-biotite quartz monzonite and aplitic granite
- **Dm** Devonian metabasite—Basic metavolcanic and metaplutonic bodies, many of which have not been mapped. Locally may be Jurassic in age
- **S€b** Cambrian to Silurian black quartz phyllite—Quartzose siltstone and areillite
- **O€sm** Cambrian(?) and Ordovician schist and calcareous schist—Feldspathic chlorite-quartz schist, calcareous schist, and marble
- **PzPqs** Proterozoic(?) or lower Paleozoic schist and marble--Calcareous schist and marble, locally interlayered with graphitic schist and marble
- **PzPcs** Proterozoic(?) or lower Paleozoic schist—Calcareous schist and marble, locally interlayered with graphitic schist and muscovite-quartz schist

Figure 158. Generalized geologic map of upper Koyukuk and Chandalar mining districts, showing distribution of lode- and placer-gold deposits.

recognized by Brosgé and Reiser (1964, 1971). These thrust faults are typically shallow-dipping zones of mylonitic, phacoidal cataclastic rock, broken formation, and melange. Postmetamorphic (post-Early Cretaceous) faults are generally steeply dipping zones of gouge with slickensided, brecciated walls that cut older metamorphic structures. Many are intruded by quartz and calcite veins and, in at least three locations, the quartz veins contain gold. Most of the gold-bearing veins, therefore, postdate Early Cretaceous metamorphism.

GOLD DEPOSITS

Spatial, mineralogic, and geochemical similarities between lode- and placer-gold deposits provide evidence that the placer gold was derived from the nearby lode deposits. The larger auriferous lode and placer deposits of the upper Koyukuk and Chandalar mining districts are described below and shown on figure 158.

PLACER DEPOSITS

Placer-gold deposits have produced most of the gold found in the upper Koyukuk and Chandalar mining districts. These deposits are usually confined to Quaternary alluvium that overlies bedrock in deeply cut, nonglaciated valleys. Productive placer deposits are common around known lode-gold deposits of the Little Squaw Peak (fig. 158, locs. 1, 4, and 5) and Nolan Creek areas (fig. 158, locs. 35 through 37 and 41 through 43). Gold placers are also found close to smaller lode-gold deposits at Sukakpak Mountain (fig. 158, loc. 27), Jay Creek (fig. 158, loc. 52), and near Wild Lake (fig. 158, loc. 54). Reed (1938) provides a more complete description of the placer deposits.

Rounded, flattened nuggets and rough, coarse, or wire gold are found in the placer deposits. Some nuggets with relict crystal shapes and attached pieces of quartz weigh up to 40 oz; these are particularly common close to gold-bearing quartz veins. Some of the gold concentrates at Big Lake and Gold, Vermont, and Crevice Creeks contain angular pieces of gold-bearing stibnite-quartz vein material that must have been derived directly from the lode deposit (fig. 158, locs. 22, 28, 34, and 57; Cobb, 1981; Dennis Stacey and Pelham Jackson, personal commun., 1982). Other minerals found in placer concentrates, such as arsenopyrite, pyrite, stibnite, scheelite, galena, cinnabar, and copper nuggets, are typically associated with gold-bearing quartz veins. Anomalously high concentrations of gold, antimony, arsenic, silver, lead, copper, and mercury are also found in soil samples near gold-bearing quartz veins. Most placer deposits contain some extremely rounded, flattened gold, which probably came from a more distant source.

LODE DEPOSITS

Gold-bearing quartz veins of the upper Koyukuk and Chandalar mining districts have been mined since the turn of the century. The mineralogy, chemistry, and age of these lode deposits are described here with emphasis on the evidence that links them to the productive placer-gold deposits and distinguishes them from barren quartz veins.

Lode-gold deposits occur in an east-west-trending belt from Wild Lake to Squaw Lake (fig. 158). In general, gold-bearing quartz veins are 4 to 16 in. (10 to 40 cm) thick and pinch and swell along strike for distances up to 0.5 mi (1 km). Many of the veins occur along steeply dipping, postmetamorphic faults and portions are brecciated, displaced, slickensided, and boudinaged by subsequent movement along the fault zones; other portions are recemented by late-stage gangue. These relationships indicate that the veins formed during displacement.

The gold-bearing veins are composed of quartz-calcite gangue, with gold, stibnite, arsenopyrite, and scorodite as the principal ore minerals and pyrite, tetrahedrite, cinnabar, molybdenite, galena, and sphalerite as accessory minerals. The gold-bearing quartz veins are especially enriched in antimony and arsenic but also have anomalously high amounts of molybdenum, mercury, silver, lead, and zinc. Soils and placer deposits derived from these veins are also enriched in these elements (Brosgé and Reiser, 1970, table 2). Most auriferous veins are located between Nolan Creek (fig. 158, locs. 39, 40, and 46) and Little Squaw Creek (fig. 158, locs. 2, 3, 6, 7, and 8) and are surrounded by productive placer-gold deposits.

Antimony is the dominant metal in vein deposits of the upper Koyukuk mining district, whereas arsenic is the dominant metal in vein deposits of the Chandalar mining district. Typical auriferous veins that occur in the Nolan-Wiseman area of the upper Koyukuk mining district are extremely rich in stibnite, with arsenic-bearing accessory minerals. One such vein, studied in detail, is present on Sukakpak Mountain (fig. 158, loc. 27), which is on the axis of a large northeast-trending synform overturned to the northwest. Quartz-stibnite veins can be found along the south limb of the synform, along the west-trending, faulted contact between the Skajit Limestone (Dsk) and underlying mica-quartz schist (Db), and along faults near the contact within the Skajit and the schist. The veins strike parallel to the contact and dip steeply 65° to 80° SE.

At Sukakpak Mountain, there were two stages of vein emplacement. The older veins are composed of barren, massive, fine-grained, milky-white quartz, with minor copper sulfides. They intruded the wall rock along joint sets that trend N. 60° W. Fluid-inclusion data from the barren quartz veins give an uncorrected trapping temperature of 199 to 293 °C and a composition of 6.5 percent by weight sodium chloride. Barren quartz veins were later fractured by right-lateral, strike-slip faults.

The second stage of vein emplacement is represented by injection of the mineralizing fluid into openings along the fault zones and subsequent crystallization of quartz, stibnite, and gold. Quartz occurs both as ribbons and as crystals that fill cavities; stibnite occurs as brecciated and rehealed crystals or large euhedral crystals up to 8 in. (20 cm) long; and gold occurs as crystals or wires in stibnite and second-stage quartz. Fluid-inclusion data indicate that the mineralizing fluid crystallized from boiling(?) carbon-dioxide-rich fluids at temperatures from 212 to 254 °C.

The veins are continuously exposed for 435 ft (135 m) between 3,500 and 3,700 ft elevation and have an average thickness of 2.5 ft (0.75 m). Thirteen channel samples from this exposure yielded an average grade of 17.4 percent stibnite and 0.44 oz/ton gold. There is no appreciable alteration of the wall rock or of earlier vein material, which indicates thermal equilibrium between the wall rock and vein material.

Gold-quartz veins in the Chandalar mining district are rich in arsenopyrite and scorodite and contain minor amounts of stibnite. Veins at the Little Squaw gold mines in the Chandalar mining district are described in detail by Ashworth (1983). In general, the veins in the Little Squaw area are found along northwest-trending, high-angle normal faults that cut across regional schistosity. In order of decreasing abundance, accessory minerals are arsenopyrite, galena, sphalerite, stibnite and pyrite. Gold occurs as flakes and wires in quartz, as small veinlets, and along sulfide borders in association with galena, arsenopyrite, and more rarely stibnite or sphalerite. At least two generations of quartz are recognized in each vein. The earlier phase, referred to as 'barren quartz,' is massive (up to 3 ft [1 m] wide), white, coarsely crystalline (crystals 8 to 12 in. [20 to 30 cm] long), ,and contains <5 percent sulfides. Studies on heating and freezing indicate barren quartz crystallizes from carbon-dioxide-rich, boiling(?) fluids at temperatures of 250 to 310 °C.

Subsequent movement along the high-angle faults opened space along the dilatant zone that permitted crystallization of main-stage mineralization. The main-stage quartz occurs discontinuously next to the barren quartz and is finer grained and sometimes vuggy. Sulfides are more abundant in this generation of quartz. Fluids from inclusions in this quartz were also carbon-dioxide-rich and probably boiled at temperatures of 250 to 295 °C during crystallization into main-stage quartz.

As with the Sukakpak Mountain vein system, there is no apparent wall-rock alteration in the Little Squaw vein system, again indicating an equilibrium between vein material and wall rock. Both the Little Squaw and Sukakpak veins were crystallized in fault zones from carbon-dioxide-rich, boiling(?) fluids at 200 to 300 °C in a two-stage process that involved crystallization of barren quartz followed by crystallization of sulfide-bearing, auriferous vein quartz. Both are postmetamorphic, epithermal veins emplaced along active fault zones.

PRIMARY SOURCES OF GOLD

Because gold deposits in the upper Koyukuk and Chandalar mining districts form a nearly east-west-trending belt from Little Squaw Lake to Wild Lake, the primary source of the gold should be an east-west-trending geologic feature. Moreover, because the gold-bearing veins are younger than all metamorphic supracrustal rocks exposed in the area, the gold has been either remobilized from a preexisting source within the supracrustal rocks or was a primary emanation of the mantle or lower crust.

Primary mantle emanations rich in gold, arsenic, antimony, molybdenum, and mercury seem unlikely. Furthermore, supercritical fluids that diffuse from the mantle through the crust, which is about 35 km thick in the southern Brooks Range, should be more widely distributed.

Abundant evidence for Neocomian (120 to 145 Ma) metamorphism is found in the supracrustal rocks. Moreover, the temperatures and pressures in the deep crustal rocks were probably higher during metamorphism and stayed higher longer than in the supercrustal rocks. This is consistent with Albian (100 to 115 Ma) emplacement of auriferous veins remobilized from a crustal source, especially a deep one. The lithology of deep crustal rocks in the study area is unknown, but they probably have a continental origin. The hypothesized deep crustal source for the gold does not explain, however, the restricted distribution of the gold deposits. Neocomian metamorphism affected the entire southern Brooks Range, but rich gold deposits are found in relatively limited districts. Further, the Devonian granitic plutons with minimum-melting-point granitic compositions, high initial Sr-isotope ratios, and inherited zircons were apparently derived by anatexis of the deep crust (Dillon and others, 1980). Consequently, the deep crust in this area should have been depleted in labile elements before Neocomian time. Nevertheless, a restricted, probably granitic, source in the deep crust cannot be eliminated because the present constituents of the deep crust have not been determined.

There are three potential sources of gold within the supracrustal rocks: 1) older, remobilized' gold deposits; 2) granitic and mafic plutonic rocks; and 3) greenstones. Paleoplacers might be enriched in elements now associated with the lode deposits. Within the western Wiseman Quadrangle, there is a fair correlation between exposures of potential paleoplacer deposits (quartz-rich metaconglomerates that underlie and overlie the Skajit Limestone [Dsk]) and the occurrence of present-day placers at Wild Lake and Crevice, Jay, Birch, and Nolan Creeks (fig. 158, locs. 41, 42, 43, 52, 53, 54, 56, and 57; Brosgé and Reiser, 1970). The metaconglomerates apparently do not continue eastward to the Chandalar area from the productive Nolan Creek area; thus the paleoplacer hypothesis seems an incomplete explanation for the belt of gold deposits.

Gold also may have been remobilized during metamorphic dewatering of the lower Paleozoic rocks. There is geochemical evidence for copper-lead and silver-arsenic-antimony-mercury-molybdenum-gold occurrences in the lower Paleozoic rocks of the Doonerak area (Cathrall and others, 1984). However, the geochemical signature of the Doonerak rocks does not appear to extend into other areas of lower Paleozoic rocks near the gold deposits.

Ashworth (1983) suggests that mafic igneous rocks are a likely gold source in the Little Squaw quartz-vein gold deposits. Greenschists occur near both the Little Squaw and the Sukakpak Mountain vein systems. Studies of vein deposits in medium-grade metamorphic belts that contain significant amounts of greenstone show that vein constituents could have been supplied by greenstone, such as at Otago, New Zealand (Henly and others, 1976); the Yellowknife gold deposit, Northwest Territories, Canada (Boyle, 1955, 1959); and the Ranagiri gold fields, Andra Predesh, India (Ghosh and others, 1970). In each deposit, gold mobilization and emplacement occurred during cooling following metamorphism and can be explained by a metamorphic-secretion theory. Although greenschists occur near known gold-bearing quartz veins throughout the Wiseman and Chandalar Quadrangles, they are widespread, occurring in many areas where lode- and placer-gold deposits are absent.

Field relations show that many of the granitic plutons within the study area are slightly younger than the nearby mafic plutons. This suggests that both granitic and mafic plutons may have been potential sources for gold.

The gold source best corresponds, geologically, to a belt of shallowly eroded-and-buried Devonian granitic plutons (Brosgé and Reiser, 1964; Dillon and others, 1980). No lode deposits are found where plutons are deeply eroded between Chandalar and Big Lakes, but deposits commonly occur near cupolas and hornfelsed roofs. Anomalously high concentrations of antimony, arsenic, molybdenum, and copper in stream-sediment samples near cupolas of granitic plutons in the Chandalar Quadrangle (Marsh and others, 1978a,b, 1979; Adams and Dillon, 1989) support this geochemical association.

The granitic plutons provide the expected diffuse, east-west-trending source of gold in the shallow crust, whereas the overlying cupolas provide a depositional locus for hydrothermal fluids generated during late magmatic crystallization. By this hypothesis, metamorphic remobilization caused the gold to move only a small distance from the site of magmatic deposition. Similar geometric relations occur in the Shungnak mining district in the western Brooks Range. Moreover, the elements gold, antimony, arsenic, molybdenum, and copper are typically associated with felsic magmas and tend to be concentrated during the late stages of crystallization in the cupola as hydrothermal emanations.

CONCLUSIONS

The most likely prospects for gold deposits are in the roofs and cupolas of Devonian granitic and mafic plutons mapped by Brosgé and Reiser (1964, 1971) and Dillon and others (1981a,b, 1986). Recent geologic mapping in the Chandalar C-5 and C-6 Quadrangles shows a northward and westward distribution of cupolas and provides evidence for a genetic link between the mafic and granitic plutons. A primary mantle source for the gold seems unlikely, but a deep crustal source is difficult to eliminate except on the basis of the restricted distribution of the gold deposits.

Primary granitic sources for the gold are expected to occur where plutons are shallowly overlain by country rock, in areas of hornfelsed rocks injected by hornblende granodiorite and aplite dikes and sills, and beneath the felsic volcanic rocks in the Wiseman C-1, B-1, and Chandalar C-5, C-6, D-5, and D-6 Quadrangles. Potential for locating new gold deposits derived from primary granitic sources is highest along and north of Bettles River and Robert Creek in the Chandalar Quadrangle and westward from Sukakpak Mountain to the Nolan area in the Wiseman mining district (fig. 158). The gold potential is high because there is convincing geophysical evidence for shallowly buried plutonic cupolas in these areas (Cady, 1978). The potential for low-grade, Carlin-type gold deposits surrounding volcano-plutonic complexes near Nutirwik Creek, about 25 mi north of Sukakpak Mountain, needs further study.

CHAPTER 14.

STRATIGRAPHY AND STRUCTURE OF THE DOONERAK FENSTER AND ENDICOTT MOUNTAINS ALLOCHTHON, CENTRAL BROOKS RANGE[1]

By *C.G.* Mull,[2] K.E. Adams,[3] and J.T. Dillon[4]

ABSTRACT

The Doonerak fenster (figs. 26 and 159) is the most significant locality for understanding the regional structure and tectonic history of the central Brooks Range. It is a major northeast-trending antiform bounded by the Endicott Mountains allochthon, which to the north forms the entire northern Endicott Mountains of the central Brooks Range. The fenster contains a distinctive section of Lower Mississippian to Upper Triassic rocks that unconformably overlies Cambrian to Ordovician volcanic and metamorphic rocks (fig. 160). Rocks in the fenster are overlain by the Amawk thrust, the sole fault of the north-vergent Endicott Mountains allochthon. The allochthon consists of a thick interval of Upper Devonian strata overlain by a relatively thin interval of Mississippian to Lower Cretaceous rocks (fig. 160). Distinctive lithologic characteristics that distinguish the sequence on the allochthon from coeval rocks in the fenster include the following: 1) siliceous Permian and Triassic strata on the allochthon, in contrast to nonsiliceous coeval strata in the fenster; 2) maroon-and-green shale with associated barite and siderite nodules in the Permian sequence on the allochthon, in contrast to correlative dark-gray to black shale in the fenster; 3) thick Upper Devonian strata conformably underlying Mississippian rocks on the allochthon, in contrast to a regional unconformity beneath a thin basal Mississippian quartzite and conglomerate horizon in the fenster. Conversely, rocks in the fenster have characteristics similar to autochthonous rocks in the northeastern Brooks Range and Arctic Slope subsurface.

Restoration of the Endicott Mountains allochthon to a position south of the Doonerak fenster results in a logical reconstruction of the Mississippian through Triassic depositional basin. A minimum of 55 mi (88 km) of tectonic overlap of the Endicott Mountains allochthon over the Doonerak fenster is indicated by regional relations; actual shortening is at least an order of magnitude greater. Major thrusting apparently occurred during Early Cretaceous time and was followed by folding in Late Cretaceous and Tertiary time. The Doonerak antiform was probably uplifted during the later stage of folding.

INTRODUCTION

The Doonerak fenster can be found about 35 mi (55 km) south of the mountain front in the Endicott Mountains of the central Brooks Range. It is a large antiformal feature that can be traced for at least 70 mi (110 km) southwestward from the Dalton Highway in the area of the Dietrich River. The most significant and obvious geologic relations, however, are confined to the eastern end of the fenster, east of the North Fork Koyukuk River (fig. 159); this is the area in which the most detailed geologic studies have been concentrated.

The Endicott Mountains allochthon is best exposed north of the Doonerak fenster. Most of the major rock units on the allochthon are accessible from the Dalton Highway from a few miles south of the Brooks Range crest at Atigun Pass northward to the Brooks Range mountain front at Galbraith Lake.

In the discussion that follows, most of the geographic names in the vicinity of the Doonerak fenster may be found on figure 159 or on U.S. Geological Survey Wiseman and Chandalar 1:250,000-scale quadrangle maps.

[1]The original version of this paper appears in the volume, 'Alaskan North Slope Geology' (Tailleur and Weimer, 1987).

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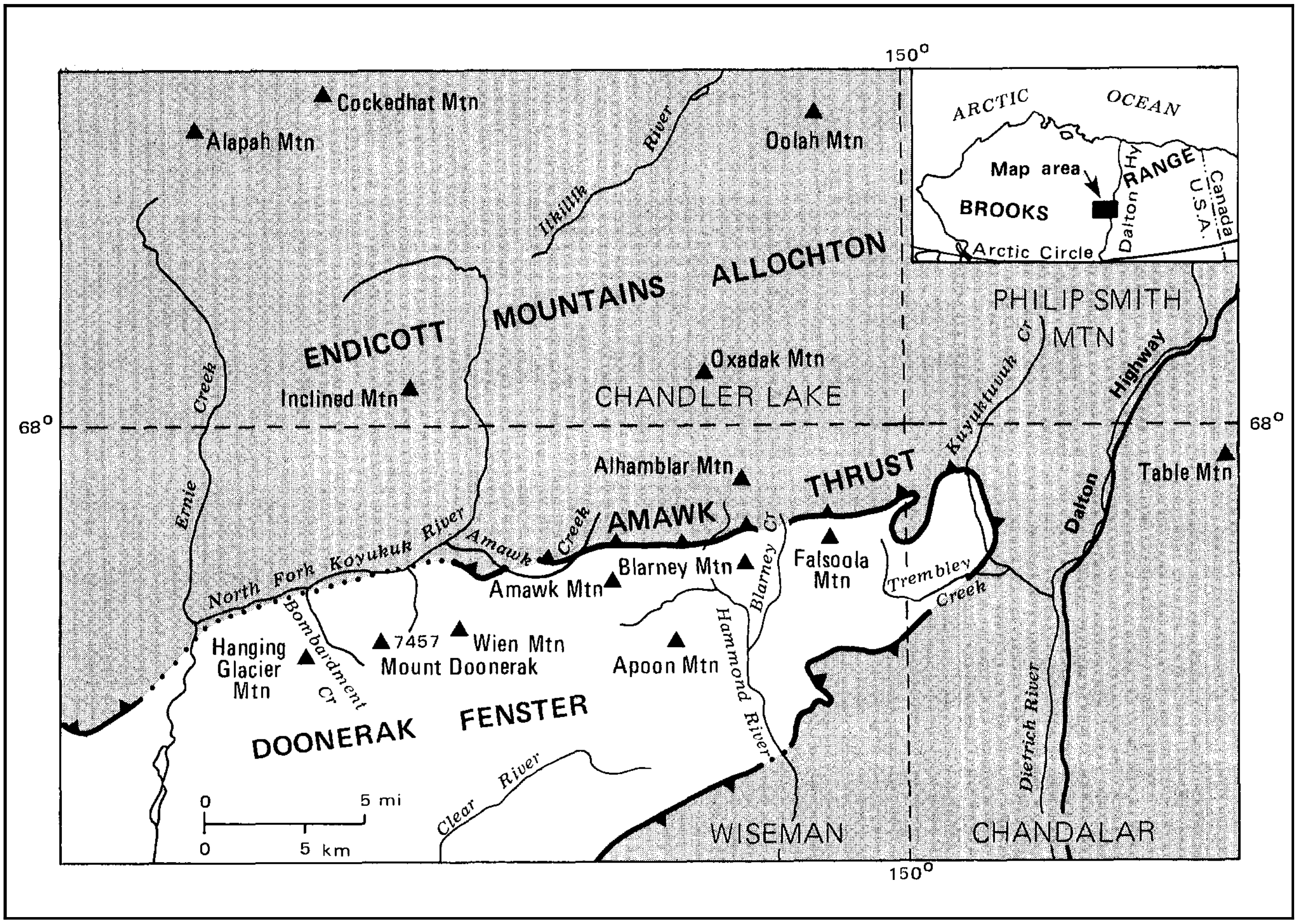

Figure 159. Index map of eastern end of Doonerak fenster.

PREVIOUS INVESTIGATIONS AND SUMMARY OF GEOLOGIC RELATIONS

The Mount Doonerak area was first mapped at a recdnnaissance scale of 1:250,000 as part of U.S. Geological Survey investigations of the Wiseman and Chandalar Quadrangles (Brosgé and Reiser, 1960, 1964, 1971). In these investigations, a long belt of north-dipping Mississippian rocks, consisting of Lisburne Group carbonates and Kayak Shale, was mapped from the North Fork Koyukuk River on the west to the Dietrich River on the east (fig. 159). A thin quartzite and conglomerate unit that underlies the Kayak Shale and unconformably overlies lower Paleozoic slate, phyllite, and mafic volcanic rocks was assigned to the Upper Devonian Kanayut Conglomerate. Brosgé and Reiser also mapped a major north-dipping thrust fault that overlies the Lisburne Group. They noted that the base of the overlying thrust plate consists of several thousand feet of black shale and phyllite of the Upper Devonian Hunt Fork Shale and grades upward into Kanayut Conglomerate (fig. 160).

In 1971 Mull made a reconnaissance trip into the area mapped by Brosge and Reiser and recognized that the quartzite and conglomerate unit beneath the Kayak Shale is similar to the Mississippian Kekiktuk Conglomerate in the northeastern Brooks Range and Arctic Slope subsurface (Mull, unpublished field notes, 1971). In those areas, Upper Devonian sediments are absent, so that Kekiktuk unconformably overlies older Paleozoic rocks (fig. 160). A more significant discovery was made at a locality on Amawk Creek (fig. 161), 5 mi (8 km) northeast of Mount Doonerak, where Mull (unpublished field notes, 1971) collected the trace fossil ***Zoophycos*** and several brachiopods from a fine-grained calcareous sandstone that overlies the Lisburne Group. The sandstone lithology and the presence of ***Zoophycos*** were noted as characteristic of the Permian Echooka Formation of the Sadlerochit Group, which is widespread in the northeastern Brooks Range. Furthermore, the brachiopods from the calcareous sandstone were identified as Permian by J.T. Dutro, Jr. (written commun., 1971, 1973), who confirmed the close similarity of the sandstone to the Echooka. The resemblance of strata in the Doonerak area to strata in the northeastern Brooks Range was later reinforced by discovery of another section of the Echooka Formation and a section of

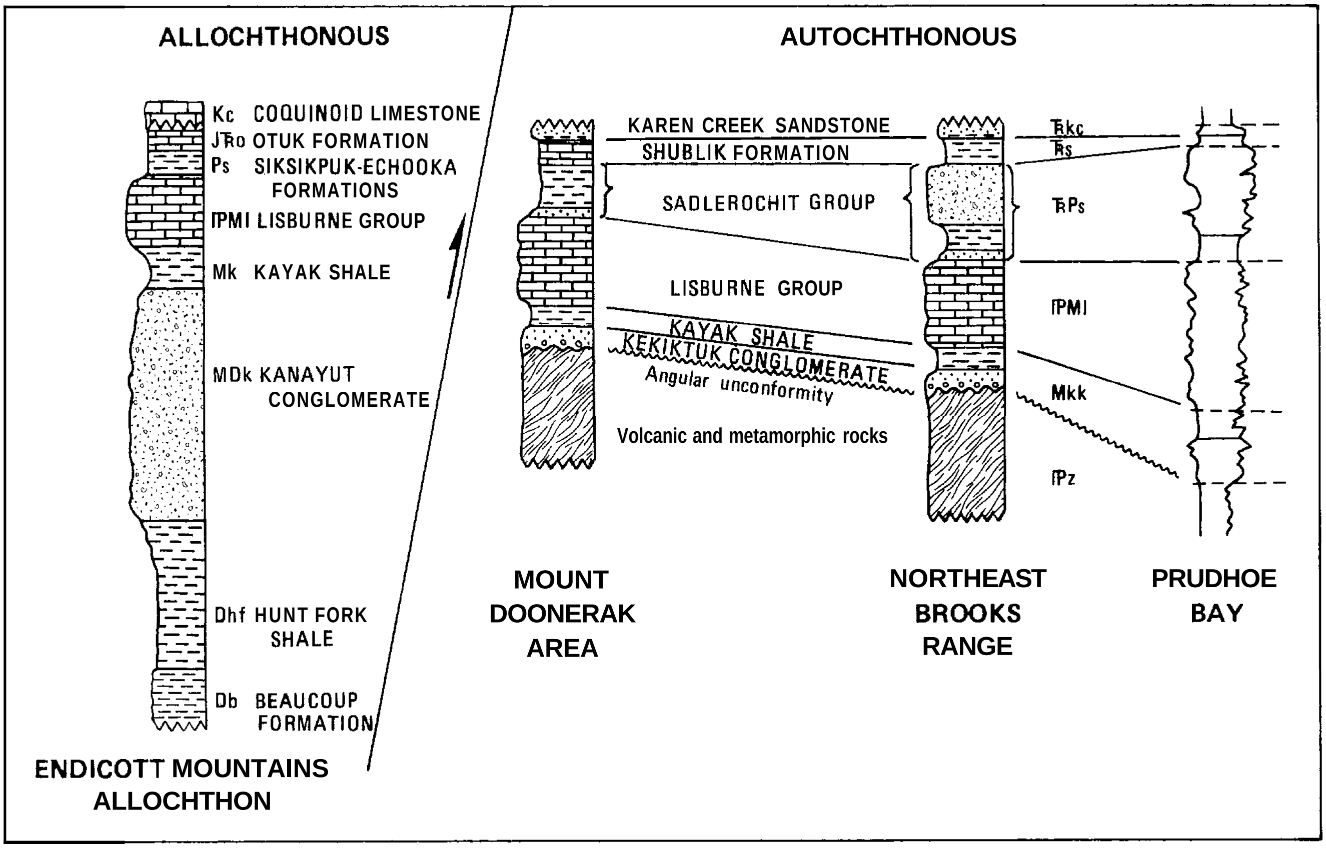

Figure 160. Generalized columnar sections of allochthonous and autochthonous rock units in northern Alaska.

earthy, phosphatic limestone and shale of the Triassic Shublik Formation at Bombardment Creek, 3 mi (5 km) northwest of Mount Doonerak. Both the Shublik and Echooka Formations in the Doonerak area were noted as markedly different from coeval equivalents that overlie the Lisburne Group along the Endicott Mountains front in the Tiglukpuk Creek area, 40 mi (65 km) to the northwest. The close similarity of the Mississippian to Triassic stratigraphy in the Doonerak area to that in the northeastern Brooks Range and subsurface of the Arctic Slope, and the presence of an overlying folded thrust sheet of markedly different lithology, suggest an analogy to the western Brooks Range, where stacked and folded allochthons of coeval, but widely diverse, facies indicate major crustal shortening.

These stratigraphic and structural observations suggest that all or most of the northern Endicott Mountains are allochthonous and that the Doonerak fenster is a window that exposes an autochthonous or parautochthonous stratigraphic section typical of the northeastern Brooks Range and Arctic Slope subsurface (fig. 160; sheet 1). Other implications are that the type localities of the Hunt Fork Shale, Kanayut Conglomerate, Kayak Shale, Alapah and Wachsmuth Limestones of the Lisburne Group, Siksikpuk Formation, and Otuk Formation are all allochthonous and have been thrust a minimum of 35 mi (56 km) northward over autochthonous or parautochthonous rocks. In addition, the absence of Upper Devonian rocks in the fenster implies that the Doonerak area may have been part of the Late Devonian orogenic belt from which the clastic wedge of the Kanayut Conglomerate and Hunt Fork Shale was derived (Mull and Tailleur, 1977; Newman and others, 1979). The source of the Kanayut Conglomerate had previously been inferred to be as much as 200 mi (320 km) farther north beneath the Arctic Slope or Beaufort Sea (Brosgé and Tailleur, 1970; Brosge and Dutro, 1973).

As a result of these discoveries, more detailed studies comparing the stratigraphy of the Doonerak fenster with the stratigraphy of the northeastern Brooks Range were carried out by the U.S. Geological Survey. Dutro and others (1976) suggested an alternative structural interpretation to the north-vergent folded-thrust model. They suggested that two separate thrust faults bound the flanks of the Doonerak antiform: one on the south side of the antiform emplaced by north-vergent thrusting, and a second on the north side of the antiform emplaced by south-vergent thrusting. This interpretation considerably minimizes the magnitude of crustal shortening. In a concurrent study, Armstrong and others (1976) investigated the Carboniferous section in the fenster.

West of the Doonerak area in the Schwatka Mountains, Mull and others (1976) and Mull and Tailleur (1977) reported the presence of another window that exposes a stratigraphic sequence similar to that of the Doonerak fenster and northeastern Brooks Range. The window was interpreted to be overlain by north-vergent folded thrust sheets of contrasting lithologies.

No further studies of the Mount Doonerak area were reported until Mull (1982) noted that a siltstone at the top of the Shuhlik Formation at Bombardment Creek correlated with the Karen Creek Sandstone of the northeastern Brooks Range. In addition, a pervasive, south-dipping slaty cleavage in the Sadlerochit Group was cited as compatible with north-vergent thrusting of the allochthon, which was named the Endicott Mountains allochthon; the sole fault was named the Amawk thrust (fig. 161). Dillon (1982) compared the cleavage of Devonian rocks of the allochthon with that of lower Paleozoic, Mississippian, ana Triassic rocks in the fenster. He concluded that rocks above and below the Amawk thrust have the same two sets of cleavage and that they were probably affected hy the same post-Triassic deformations.

During the 1981-83 field seasons, the Doonerak fenster was studied as a part of the U.S. Geological Survey's Alaska Mineral Resource Appraisal Program. This was a cooperative project that involved field parties from both the U.S. Geological Survey, led by W.P. Brosgé, J.T. Dutro, and J.B. Cathrall, and DGGS, led by J.T. Dillon. Preliminary reports by Dillon and others (1983, 1986) showed the Doonerak structure as a fenster surrounded by the Amawk thrust. Dutro and others (1984) and Palmer and others (1984) discovered Cambrian fossils within the fenster, and Cathrall and others (1984) showed that the Doonerak fenster was geochemically unique. Dillon (1985) mapped Devonian rocks of the Endicott Mountains allochthon around the eastern end of the Doonerak fenster to its southern limb and found that the Devonian rocks are overlain by lower and middle Paleozoic rocks of a higher allochthon.

The following field season, Mull and Adams (1984) traced the Amawk thrust eastward along the north flank of the fenster from the North Fork Koyukuk River to Falsoola Mountain (fig. 159) and mapped the stratigraphy and structure adjacent to the thrust. Kekiktuk Conglomerate was mapped unconformably overlying older rocks at numerous localities. Imbricate slices of Lisburne Group and Kayak Shale were also mapped, particularly in the eastern end of the window. Several east-west-trending, vertical to steeply south-dipping reverse faults were mapped; some of these faults cut the Amawk thrust. In general, a marked increase in structural complexity was noted from west to east.

Controversy concerning the structural interpretation of this area developed as a result of studies by researchers from Rice University. Julian and others (1984) and Oldow and others (1984), mapping in the eastern end of the Doonerak antiform, confirmed north-vergent thrusting of the Endicott Mountains allochthon. They did not observe Kekiktuk Conglomerate, however, and therefore suggested that the contact of the Mississippian rocks with the underlying lower Paleozoic rocks is not an unconformity as previously thought but is a thrust instead, which they named the Blarney Creek thrust. They proposed that the Doonerak fenster is part of a duplex structure, in which the Lisburne Group and Kayak Shale were detached from the underlying lower Paleozoic rocks along the Blarney Creek thrust and then moved a considerable distance northward. In view of this interpretation, they considered lithologic correlation of rocks in the fenster with rocks of the Arctic Slope to be tenuous. In addition, they reported that structural fabrics above and below

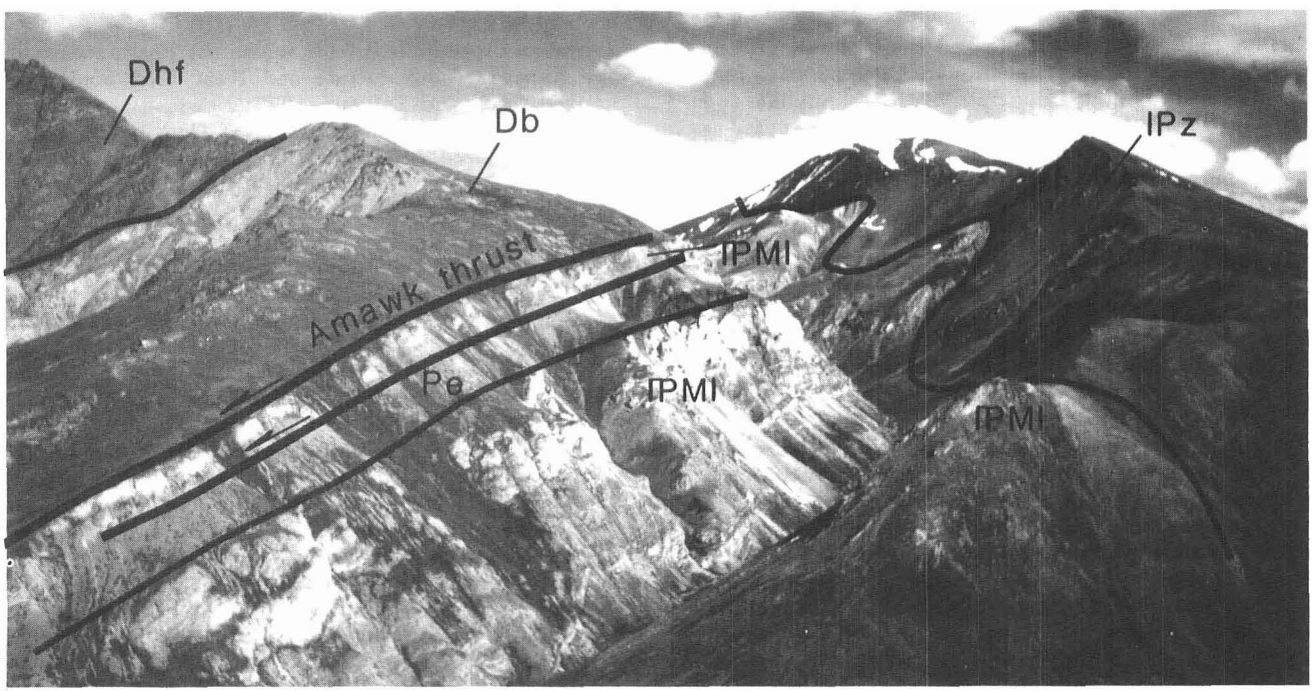

Figure 161. View eastward up Amawk Creek toward Amawk Mountain. Amawk thrust dips north and overlies imbricated Lisburne Group limestone and thin Echooka *Formation* of Sadlerochit Group on north wall of Amawk Creek canyon. Geologic units: *lPz*, lower Paleozoic *argillite* and metavolcanic rocks; Db, *Beaucoup* Formation; Dhf, Hunt Fork Shale; *lPMl*, Lisburne Group; Pe, Echooka Formation *of* Sadlerochit Group.

the base of the Mississippian are identical and showed no sign of a pre-Mississippian structural event. Phelps (1986), however, reported evidence of Devonian tectonism in the area. Seidensticker and others (1985) and Seidensticker (1986) later recognized the presence of the Kekiktuk Conglomerate at the base of the Carboniferous sequence in the window and suggested that the Blarney Creek thrust followed an originally disconformable contact between Carboniferous and lower Paleozoic rocks and acted as the upper detachment surface of a duplex structure. Phelps and others (1985) proposed that the late stage of high-angle faulting in the window is related to west-northwest thrusting, or a separate phase of deformation that involved significant east-west compression.

The following sections of this report summarize the major stratigraphic and structural relations of the Doonerak fenster. The summary is based both on published data and on our detailed geologic mapping in the eastern end of the fenster. Descriptions of specific important localities in the fenster have been presented by Mull and others (1987b).

STRATIGRAPHY OF THE DOONERAK FENSTER

The generalized stratigraphy of the core of the Doonerak fenster is illustrated in figure 160 and summarized below. Various parts of the section have been described by Armstrong and others (1976), Dutro and others (1976), and Mull (1982).

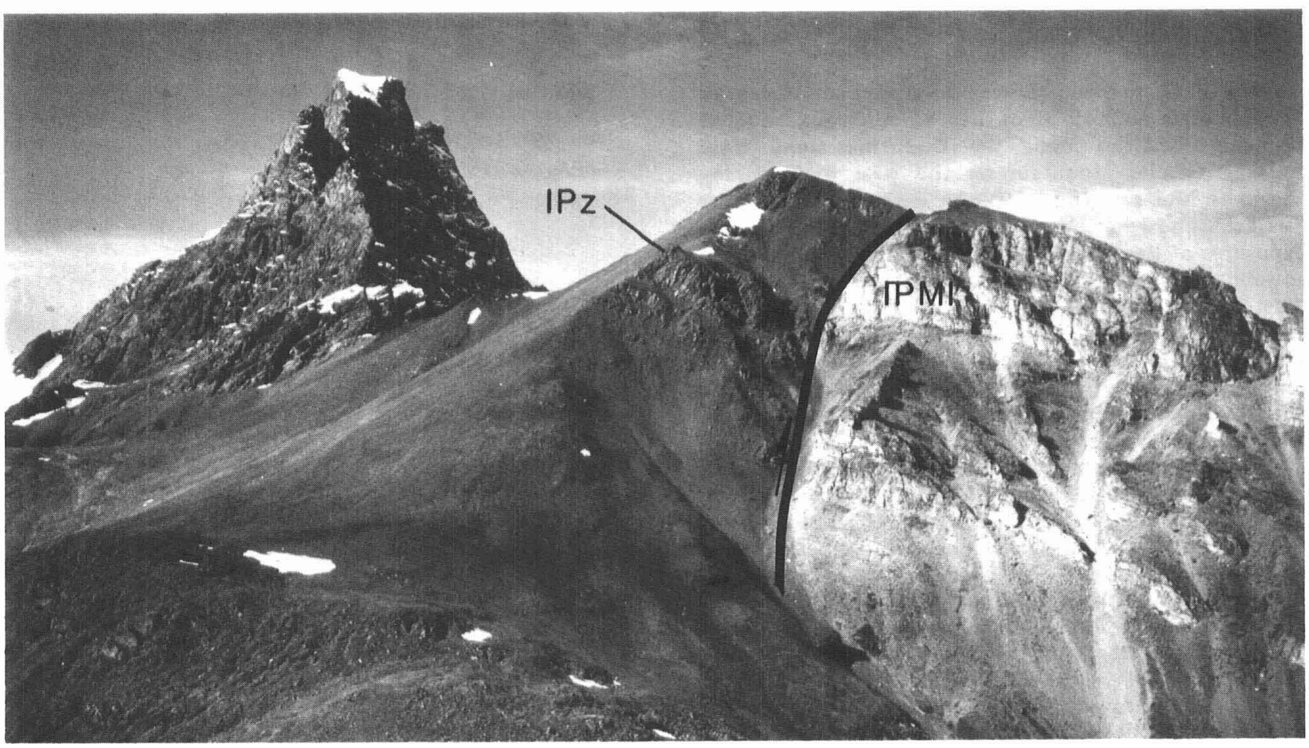

***Figure 162.** View westward toward high-angle fault along north face of Mount Doonerak (elevation 7,457 ft; 2,300 m). Throw is $>$1,000 ft (320 m). Geologic units: lPz, lower Paleozoic argillite and metavolcanic rocks; lPMl, Lisburne Group.*

LOWER PALEOZOIC ROCKS

The core of the Doonerak fenster is composed of two major suites of lower Paleozoic rocks: 1) dominantly basaltic and andesitic volcanic rocks, and 2) phyllitic and schistose rocks (Dutro and others, 1976). This sequence has been mapped from west of the North Fork Koyukuk River to near Trembley Creek at the eastern end of the Doonerak fenster (fig. 159). The massive, resistant volcanic rocks form most of the Doonerak massif (fig. 162) and adjacent high peaks of Hanging Glacier, Wien, and Apoon Mountains (fig. 159). The volcanic rocks are also present at lower elevations on Amawk and Falsoola Mountains (fig. 159). Most of the volcanic complex consists of black and green to green-gray tuff, breccia, agglomerate, and porphyritic amygdaloidal flows, cut by dikes of gabbro, diabase, and diorite. In some places, the volcanic rocks appear to be deeply weathered beneath the overlying unconformity.

The rest of the core of the fenster is composed dominantly of black to dark-gray argillite, phyllite, and thin-bedded siltstone. These rocks appear to be interbedded with the igneous complex, although at least one major thrust fault, the St. Patricks Creek thrust, has been mapped within the complex (Dillon and others, 1986). Julian (1986) reported four south-dipping, fault-bounded units within the lower Paleozoic complex.

Trilobites and brachiouods from rare limestone beds interbedded with the phyllite date the sequence as Middle Cambrian (Dutro and others, 1984); mafic dikes that cut the volcanic rocks yielded K-Ar ages that range from 465 to 520 Ma (Late Cambrian to Early Ordovician). The lower Paleozoic rocks have been weakly metamorphosed to prehnite-pumpellyite and lowest greenschist facies (Dutro and others, 1976).

CARBONIFEROUS ROCKS

KEKIKTUK CONGLOMERATE

We have mapped Lower Mississippian Kekiktuk Conglomerate at the base of the Carboniferous sequence at many localities along the Doonerak trend from Hanging Glacier Mountain on the west to Trembley Creek on the east (fig. 159). An entire section of the Kekiktuk, however, has not been measured, because exposures of the formation are generally poor and often covered with blocks of carbonate talus from the overlying Lisburne Group. The formation is relatively thin, about 130 ft (40 m) thick, and grades upward into soft black phyllitic shale of the Kayak. In a few places, the Kekiktuk is absent, probably because of nondeposition on high areas along the irregularly eroded top of the lower Paleozoic rocks.

The most conspicuous lithology of the Kekiktuk Conglomerate is a horizon up to 65 ft (20 m) thick consisting of light-gray quartzite (fig. 163) and a basal chert- and quartz-pebble conglomerate. Individual quartzite and conglomerate beds are up to 3 ft (1 m) thick and contain horizontal and trough cross-stratification and small- to large-scale channels. In places, yellow-green phyllite and yellow-brown siltstone and sandstone are interbedded with the quartzite. Overlying the quartzite-and-conglomerate horizon is an interval of predominantly black, phyllitic shale with minor amounts of schistose quartzite and silty argillite that grades upward into Kayak Shale.

Locally, the quartzite-and-conglomerate unit is underlain at a sharp contact by a lower interval of purple, green, and gray phyllite. This interval, in turn, is underlain by a basal unit of yellow-green to lightgreen, schistose stretched-pebble conglomerate, 20 ft (6 m) thick, that contains clasts of chert, quartzite, and green to dark-gray phyllite. The basal unit unconformably overlies black lower Paleozoic phyllite at a sharp irregular contact.

The variation in thickness and lithology of the Kekiktuk Conglomerate may be the result of deposition on the irregular topography of the underlying lower Paleozoic volcanic and phyllitic rocks. In paleohigh areas, resistant volcanic rocks are unconformably overlain by the massive white quartzite-and-conglomerate unit; in other areas of still greater relief, the basal quartzite and conglomerate is absent. Significantly, the schistose stretched-pebble conglomerate appears to be present only in areas underlain by the black phyllite.

KAYAK SHALE

The Kekiktuk Conglomerate is overlain by the Kayak Shale (figs. 160 and 163), a relatively incompetent unit that consists dominantly of soft, black, phyllitic shale. The upper part of the black shale is interbedded with thin, yellow-brown-weathering silty limestone beds that contain abundant crinoidal debris and scattered brachiowods: these beds are diagnostic of the Kayak Shale. The lower part of the Kayak contains interbeds of hard quartzitic siltstone. Because of its nonresistant nature, the Kayak is characterized by drag folds, fault slivering, and other structural complexities. No completely exposed sections are known, but the formation is at least 320 ft (97 m) thick.

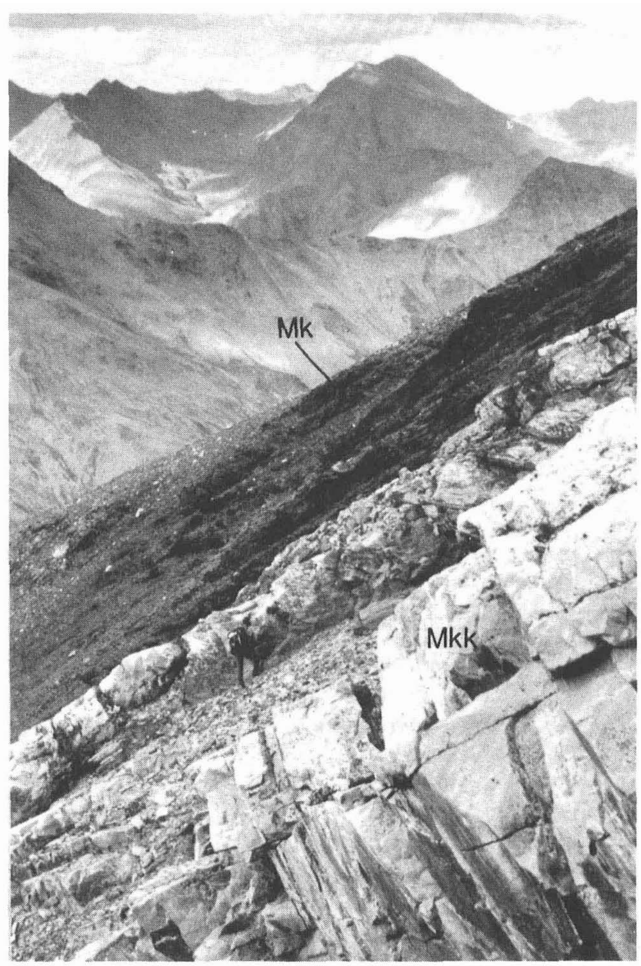

Figure 163. North-dipping, light-gray quartzite of *Kekiktuk* Conglomerate (Mkk) and overlying Kayak Shale (Mk) on north side of Amawk Mountain. Note geologist for scale.

LISBURNE GROUP

Lisburne Group carbonate rocks are the most prominent stratigraphic unit in the Doonerak fenster. The Lisburne, about 1,000 ft (300 m) thick, forms massive, light-gray-weathering cliffs of limestone and dolomite (fig. 164). A composite section of the Lisburne Group in the Doonerak fenster has been described by Armstrong and others (1976) and Armstrong and Mamet (1978). They show that in the Doonerak fenster the Lisburne consists dominantly of bryozoan-echinoderm wackestone and packstone that indicate deposition on an open-marine shelf to open platform. Part of the section consists of thin-bedded, gray to dark-gray argillaceous wackestone and packstone that contain a rich coral fauna. The argillaceous material and darker color suggest a foreslope environment of deposition. Also found in some horizons are gray dolomite and dark-gray to black nodular chert.

On the basis of foraminifera dating, the age of the Lisburne Group in the Doonerak fenster ranges from Late Mississippian (Meramec) to Early Pennsylvanian (Morrowan) (Mamet zones 11 to 20) (Armstrong and others, 1976). Conodonts from the top 1 ft (0.3 m) of the Lisburne Group at Bombardment Creek were dated by A.G. Harris (written commun., 1985) as early Early Pennsylvanian (early Morrowan).

From west to east the Lisburne in the fenster becomes more complexly deformed. Between the North Fork Koyukuk River and Bombardment Creek (fig. 159), a single, unbroken stratigraphic sequence of the Lisburne forms prominent north-dipping dipslopes cut by deep canyons. From Bombardment Creek east to Amawk Mountain (fig. 159), northdipping Lisburne Group and Kayak Shale are overlain by a thrust sliver of Lisburne and Kayak. East of Amawk Mountain, sheets of Lisburne and Kayak define the crest and north flank of the Doonerak antiform. Farther east on Falsoola Mountain, between Blarney and Kuyuktuvuk Creeks (fig. 159), isoclinal folds and several thrust repetitions of the Lisburne are folded across the crest of the antiform (fig. 164). At the easternmost end of the window, imbricated and disrupted Kayak Shale and Lisburne Group rocks (fig. 165) wrap from the north flank of the antiform around its eastern plunge along Kuyuktuvuk Creek to the south flank of the antiform along Trembley Creek (fig. 159).

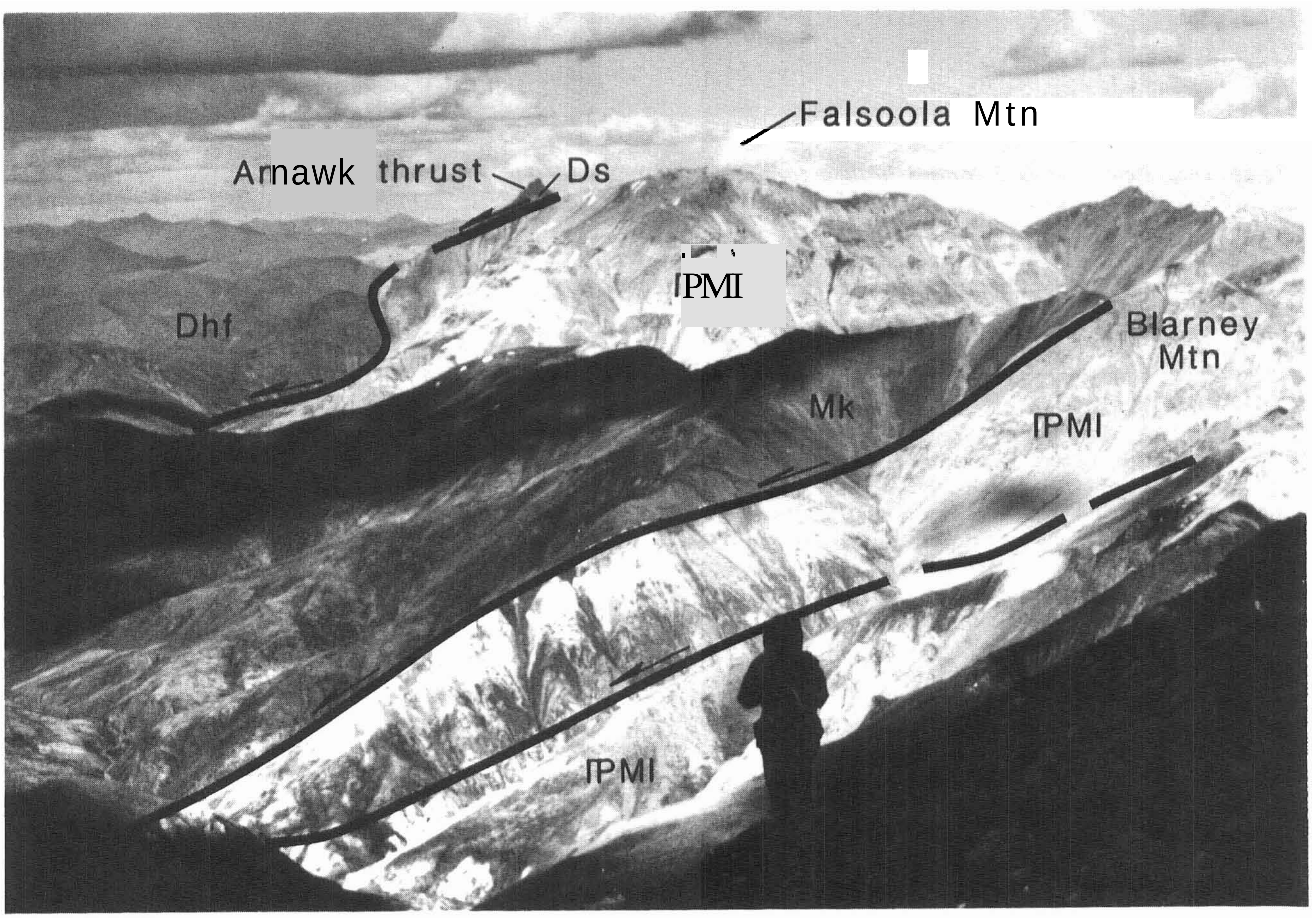

Figure 164. View eastward from Amawk Mountain along north flank of Doonerak fenster to Blarney and Falsoola Mountains. Compare greatly thickened and folded, regionally flat-lying Lisburne Group on Falsoola Mountain with thin sheets of Lisburne on Blarney Mountain. Geologic units: Ds, sandstone and conglomerate of Endicott Mountains allochthon; Dhf, Hunt Fork Shale; Mk, Kekiktuk Conglomerate and Kayak Shale; IPMl, Lisburne Group.

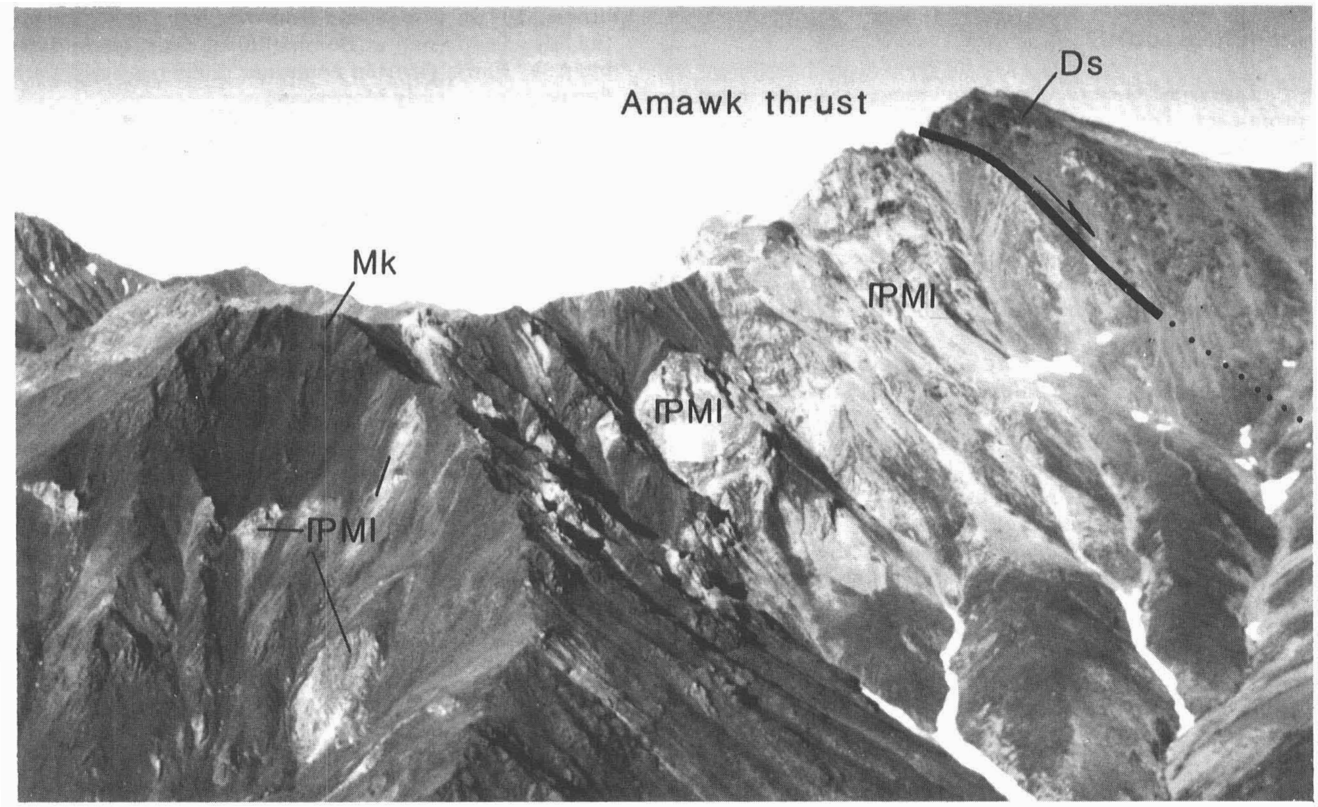

Figure 165. View westward toward broken formation composed of blocks of Lisburne Group limestone (IPMl) and Kayak Shale (Mk) at eastern end of Falsoola Mountain. Devonian clastic rocks (Ds) overlie Amawk thrust at upper right.

PERMIAN AND TRIASSIC ROCKS

SADLEROCHIT GROUP

The Lisburne Group is disconformably overlain by Permian strata of the Sadlerochit Group (fig. 160). These beds are best exposed at the mouth of Bombardment Creek and in a cliffside outcrop on the north side of Amawk Creek (fig. 161, Pe); the unit also occurs as a series of conspicuous yellow-brown-weathering dipslopes that overlie the Lisburne along the valley of the North Fork Koyukuk River. No other exposures of Sadlerochit Group are present east of Amawk Creek.

The Sadlerochit Group consists of two units: 1) a lower unit of pyritic, yellow-brown-weathering, very fine-grained calcareous sandstone and siltstone 190 ft (58 m) thick; and 2) an upper unit of black, phyllitic, silty shale 215 ft (66 m) thick. The contact between the two units is sharp and may be a disconformity. Beds in the lower interval are 2 to 20 in. (5 to 50 cm) thick, wispy laminated, and occasionally fine upwards. Most beds in the upper interval are masked by a pervasive slaty cleavage, but the few beds that are preserved range from 12 to 25 in. (30 to 60 cm) thick. A sharp contact separates the upper unit from the overlying calcareous shale of the Shublik Formation. The contact, exposed only at Bombardment Creek, appears to be a disconformity but is marked by calcite veins, which may indicate bedding-plane faulting.

The trace fossil *Zoophycos* and a brachiopod fauna including *Anidanthus* sp., *Martinia* sp., *Orulgania?* sp., *Spiriferella* sp., and *Attenuatella* sp. (J.T. Dutro, Jr., written commun., 1971, 1972, 1984, 1985) are present in the lower calcareous siltstone unit. The brachiopods date the lower unit as Early Permian (Wolfcampian) and indicate a correlation with the Joe Creek Member of the Echooka Formation of the Sadlerochit Group. The Echooka Formation is widespread in the northeastern Brooks Range and contains abundant *Zoophycos* which in northern Alaska is considered diagnostic for that formation (Detterman and others, 1975). The black, phyllitic shale unit that overlies the calcareous siltstone unit is undated and its regional correlation is uncertain; however, it may be a distal equivalent of the Ikiakpaurak Member of the Echooka Formation.

SHUBLIK FORMATION AND KAREN CREEK SANDSTONE

In the Doonerak fenster, the Shublik Formation and Karen Creek Sandstone are exposed near the mouth of Bombardment Creek only (fig. 166). In this area, the Shublik Formation is over 140 ft (37 m) thick; however, the section is folded and faulted, which prevented us from measuring the true thickness.

Figure 166. North-dipping Shublik Formation *(TRs)* and Karen Creek Sandstone *(TRk)* at mouth of Bombardment Creek, northwest of Mount Doonerak.

The Shublik at Bombardment Creek consists of dark-gray to black limestone, and sooty, earthy, calcareous shale that contains abundant phosphate nodules. The lithology is typical of the Shublik Formation in the northeastern Brooks Range and subsurface of the Arctic Slope. Some beds contain numerous impressions of the thin, flat pelecypods Daonella sp., Halobia sp., and Monotis sp. These fossils date the Shublik Formation as late Middle Triassic (Ladinian) to Late Triassic (late Carnian to early Norian) (N.J. Silberling, written commun., 1985). The Shublik grades upward into a 6 ft (2 m) thick interval of dense, dark-gray to black siltstone and very fine- grained sandstone (Mull, 1982). This dense horizon has the same lithology and stratigraphic position as the Karen Creek Sandstone described by Reed (1968) in the northeastern Brooks Range.

STRATIGRAPHY OF THE ENDICOTT MOUNTAINS ALLOCHTHON

STRATIGRAPHY NORTH OF THE DOONERAK FENSTER

The stratigraphic sequence of the northern Endicott Mountains allochthon is composed dominantly of rocks that range in age from Late Devonian to Early Cretaceous (Neocomian) (fig. 160). Major rock units include the Upper Devonian Beaucoup Formation (Dutro and others, 1979); the Upper Devonian Hunt Fork Shale and Upper Devonian and Lower Mississippian Kanayut Conglomerate (Nilsen and Moore, 1984a); the Lower Mississippian Kayak Shale and Mississippian to Lower Pennsylvanian Lisburne Group (Armstrong and others, 1970); the Permian Siksikpuk (Patton, 1957; Siok, 1985) and Echooka Formations (Detterman and others, 1975); the Triassic and Lower Jurassic Otuk Formation (Mull and others, 1982; Bodnar, 1984); and a Cretaceous (Neocomian) coquinoid limestone (Jones and Grantz, 1964; Bodnar, 1984). Only the basal units of the allochthon--the Beaucoup Formation, the Hunt Fork Shale, and the Kanayut Conglomerate--are present in the immediate Doonerak area (figs. 40 and 161). Carboniferous and younger rocks are present only in a few infolded synclines within the mountains and along the northern Endicott Mountains front.

Three major characteristics readily distinguish the allochthonous sequence of the Endicott Mountains allochthon from the autochthonous sequence in the

Doonerak fenster. First, thick Upper Devonian clastic rocks of the Beaucoup Formation, Hunt Fork Shale, and Kanayut Conglomerate are present on the allochthon, whereas no Upper Devonian rocks are present in the autochthonous or parautochthonous complex of the fenster. Second, siliceous beds and variegated maroon-and-green shale associated with barite and siderite nodules are present in the Permian Siksikpuk Formation on the allochthon but are absent in the coeval Sadlerochit Group of the Doonerak fenster. Third, siliceous beds occur in the Triassic part of the Otuk Formation on the allochthon but are absent in the Triassic Shublik Formation of the fenster.

Armstrong and others (1976), however, pointed out the similarity in lithology, petrology, and age between the Lisburne Group of the Doonerak fenster and the northeastern Brooks Range, and the slightly older Lisburne on the allochthons of the central and western Brooks Range. On the basis of foram and conodont dating, they reported that the upper part of the Lisburne in the fenster is as young as Early Pennsylvanian (Morrowan) and that the Lisburne on the Endicott Mountains allochthon is no younger than Late Mississippian (Chesterian). Conodonts no older than Morrowan, however, have recently been recovered from the top of the Lisburne at a number of localities on the Endicott Mountains allochthon (A.G. Harris, written commun. 1985, 1986); thus, the contrast between the age of the top of the Lisburne in the Doonerak fenster and on the Endicott Mountains allochthon may not be as marked as previously thought. Armstrong and others (1976) also demonstrated that the base of the Lisburne in the Doonerak fenster is early Late Mississippian (Meramecian) and younger than the base of the Lisburne on the Endicott Mountains allochthon, which is Early Mississippian (Osagean).

STRATIGRAPHY SOUTH OF THE DOONERAK FENSTER

The stratigraphic sequence that overlies the Hunt Fork Shale on the north side of the Doonerak fenster (in ascending order, Kanayut Conglomerate, Kayak Shale, Lisburne Group, Siksikpuk Formation, and Otuk Formation; fig. 160) has not been recognized on the Endicott Mountains allochthon south of the fenster. Continuity of the allochthon on opposite flanks of the fenster, however, can be demonstrated by the Hunt Fork Shale and Beaucoup Formation, which can be mapped from the north around the eastern plunge of the antiform to the south side of the fenster.

In addition to the Hunt Fork Shale and Beaucoup Formation, the allochthon south of the fenster includes the Middle to Upper Devonian Whiteface Mountain volcanic rocks (fig. 27; Dillon and others, 1986; Dillon, chap. 10). The Whiteface Mountain volcanics are composed of felsic volcanic rocks, limestone, and calcareous phyllite, siltstone, and graywacke, and may include undifferentiated lower Paleozoic rocks. The Beaucoup Formation (Dutro and others, 1979) is present at the top of the Whiteface Mountain volcanics and below the Hunt Fork Shale; it consists of quartzite, felsic volcanic rocks, purple-and-green phyllite, black phyllite and fossiliferous, black limestone, and interlayered calcareous chlorite-muscovite-quartz metasandstone, grit, and conglomerate. The Hunt Fork Shale consists of interlayered brown-weathering graphitic phyllite and metagraywacke that contain relict sedimentary features, including graded bedding, flame structures, and rip-up clasts.

The differences in stratigraphy of the Endicott Mountains allochthon on opposite sides of the fenster can be explained by the tendency of thrust faults to cut upsection in the direction of movement; in the case of the Brooks Range, this is to the north. Thus, from the south side to the north side of the Doonerak antiform, the Amawk thrust gradually cuts out the Middle Devonian Whiteface Mountain volcanic rocks as it passes around and over the eastern end of the antiform through outcrops along Kuyuktuvuk Creek and klippen near Falsoola Mountain (fig. 159). Similarly, a thrust fault above the Endicott Mountains allochthon apparently cuts out the Kanayut Conglomerate and other upper Paleozoic and lower Mesozoic rocks on the allochthon along the south side of the fenster but daylights and passes above them on the north side.

STRUCTURE OF THE DOONERAK FENSTER

THRUST FAULTS

The Doonerak fenster is well defined by the Amawk thrust (Mull, 1982), the sole thrust at the base of the Endicott Mountains allochthon (figs. 40, 161, and 167). The thrust has regional north dip along the entire north flank of the fenster (Mull, 1982; Mull and Adams, 1984; Dillon and others, 1986). It can be mapped around the eastern plunge of the antiform along Kuyuktuvuk Creek and then westward up Trembley Creek (fig. 27).

In many places along the northern side of the eastern Doonerak fenster, the approximate location of the thrust is marked by the contact between the light-gray-weathering cliffs of the Lisburne Group carbonates and the slope-forming horizon of the dark-gray-weathering phyllite and schistose quartzite and conglomerate of the Beaucoup Formation (fig. 161). In most places, though, the Beaucoup Formation is poorly exposed and the thrust fault is in a broad covered interval up to several hundred feet thick. Isolated exposures of the Beaucoup(?) schistose quartzite and conglomerate in this covered interval are commonly intensely shattered and filled with quartz veins. The Beaucoup is overlain by several thousand feet of Hunt Fork Shale that in many places are the lowest rocks exposed above the Amawk thrust.

Near the junction of Bombardment Creek and the North Fork Koyukuk River (fig. 159), the Amawk thrust is present within a relatively narrow covered

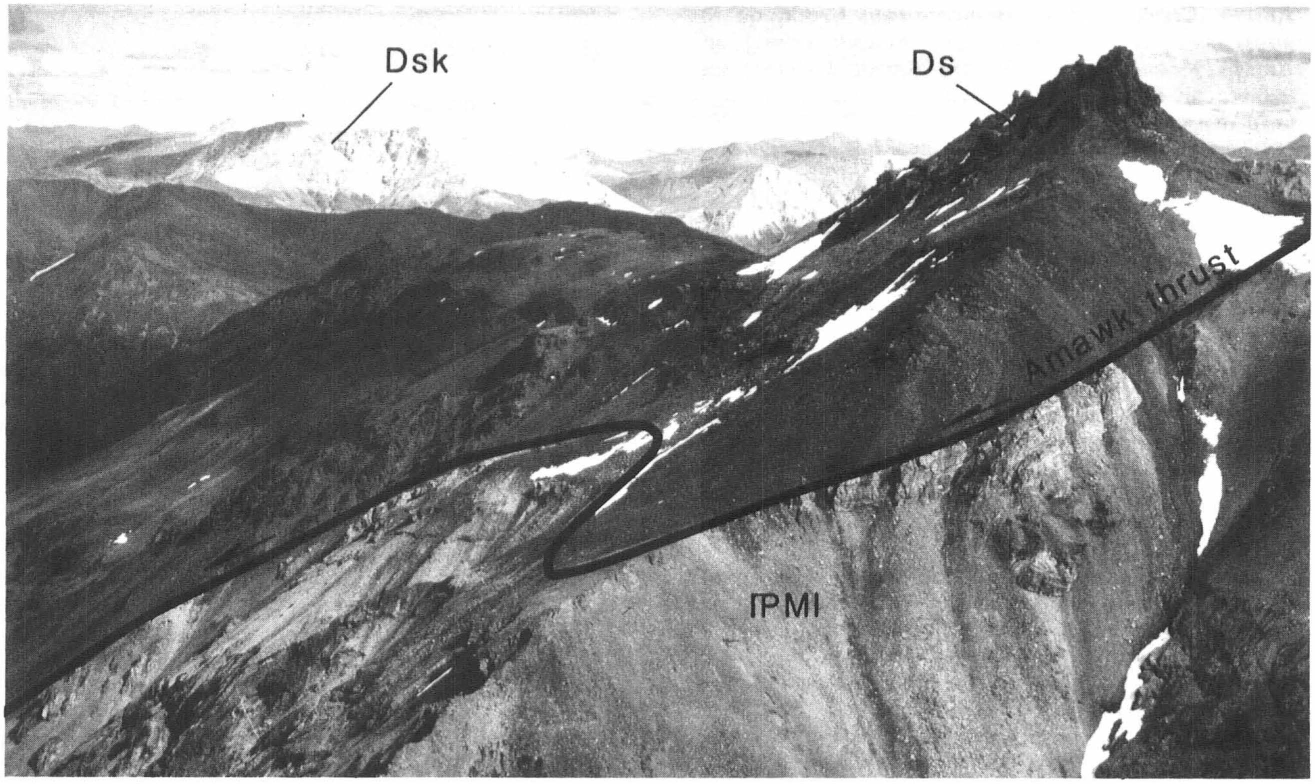

Figure 167. View eastward toward north-dipping Amawk thrust on Falsoola Mountain. Schistose sandstone and conglomerate and interbedded micaceous phyllite (Ds) at base of Endicott Mountains allochthon overlie limestone of Lisburne Group (lPMl). Light-gray cliffs in distance are composed of Devonian and older Skajit Limestone (Dsk) on Table Mountain allochthon, which overlies Endicott Mountains allochthon down eastern plunge of Doonerak antiform.

interval that separates the Upper Triassic Karen Creek Sandstone from intensely deformed lower Paleozoic phyllitic schist and schistose quartzite. To the east, on a small north-dipping flatiron west of the mouth of Amawk Creek (fig. 159), light-yellow-gray-weathering phyllite and schistose quartzite overlie a dipslope of the Lisburne Group; at this location, the thrust plane dips 35° N. From Amawk Creek east to Blarney Creek (fig. 159), the thrust is generally in a broad covered interval above the Lisburne. On Falsoola Mountain, near the eastern end of the fenster, a dipslope of the Lisburne Group underlies 100 ft (30 m) of Devonian schistose sandstone and conglomerate and interbedded micaceous phyllite; the thrust plane at this locality dips 20° N. (fig. 167). In Trembley Creek valley on the south side of the eastern end of the fenster, south-dipping Lisburne Group is overlain by a cataclastic interval of sheared phyllite that contains numerous sheared ovoid masses of gray dolomite that range from ½ in. to over 3 ft (1 cm to 1 m) long.

In most places on the north side of the fenster, rocks exposed at the base of the Endicott Mountains allochthon are predominantly shale and phyllite of the Beaucoup Formation or Hunt Fork Shale. At Falsoola Mountain, however, the base of the allochthon is composed of several hundred feet of resistant dark-gray to dark-red-brown, schistose quartzitic sandstone; minor schistose quartz- and chert-pebble conglomerate; and interbedded gray, green-gray, and yellow-gray micaceous phyllite (fig. 167). These rocks may be a basal coarse clastic unit of the Hunt Fork Shale or they may be part of the Kanayut Conglomerate.

The variability in rock types that overlie the Amawk thrust is probably a result of unmapped thrust faulting and isoclinal folding. Large-scale deformation in the lower part of the allochthon is difficult to assess because of the generally nonresistant nature of the Hunt Fork Shale and Beaucoup Formation. However, intense isoclinal folding is visible in the gorge of the upper North Fork Koyukuk River and on a few well-exposed cliff faces northeast of Alhamblar Mountain and near Inclined Mountain (fig. 44). This type of deformation is probably characteristic of the lower part of the Endicott Mountains allochthon on the entire north side of the Doonerak fenster.

In the core of the fenster, our mapping of the Carboniferous to Triassic rocks indicates a marked increase in structural complexity toward the east. Between Hanging Glacier Mountain and Bombardment Creek (fig. 159), the Carboniferous to Triassic section is an unbroken, north-dipping stratigraphic sequence that unconformably overlies the lower Paleozoic volcanic and argillite complex of the core. Both Permian and Triassic rocks are present in this area but are not found east of

Amawk Creek. East of Bombardment Creek, on the north side of Wien Mountain, Amawk Creek, and Amawk Mountain (fig. 159), the Kayak, Lisburne, and Sadlerochit are imbricated. On Blarney Mountain at the head of the Hammond River, Kekiktuk Conglomerate is also imbricated in a thrust sheet of the Kayak that overlies at least two imbricate sheets of the Lisburne (fig. 164). Beneath the Lisburne in this area, Julian and others (1984) and Oldow and others (1984) mapped the Blarney Creek thrust between the Kayak and the Kekiktuk or lower Paleozoic rocks. Still farther east, on Falsoola Mountain, imbricated sheets of the Lisburne Group are isoclinally folded. At the eastern end of the fenster near Kuyuktuvuk Creek, disjunct blocks of Lisburne are floating in a highly deformed shale matrix (fig. 165). The structure is so chaotic that the sequence can be described as a broken formation. The marked increase in overall structural complexity toward the eastern end of the fenster suggests that deeper levels of the Amawk thrust are exposed to the east (W.K. Wallace, oral commun., 1987).

In addition to the structural complexities in the Mississippian and younger rocks of the fenster, a major thrust fault, the St. Patricks Creek thrust, has been mapped within the lower Paleozoic rocks in the core of the antiform (Dillon and others, 1986). Moreover, Julian and others (1984) and Oldow and others (1984) suggested that the Doonerak fenster is part of a duplex structure and that the rocks here referred to as autochthonous and parautochthonous are grossly allochthonous. The Blarney Creek thrust, St. Patricks Creek thrust, and other unnamed faults within the core of the antiform are presumably part of the duplex system.

DIRECTION OF THRUST EMPLACEMENT

Regional geology and tectonic reconstructions favor major north-vergent thrusting in the formation of the Brooks Range orogenic belt (Mull, 1982; Mayfield and others, 1983). However, Dutro and others (1976), from studies of the Doonerak area, and Churkin and others (1979), from studies of the western Brooks Range, suggested the possibility of a major south-vergent thrusting event in the formation of the Brooks Range orogenic belt. Crane (1980), Dutro (1980), Mayfield (1980), Metz (1980), Mull (1980), and Nelson (1980) cited a wide variety of data that questioned the interpretation of Churkin and others. In the Doonerak fenster, a pervasive slaty cleavage in the Echooka Formation consistently dips south relative to bedding and is compatible with north-vergent emplacement of the Endicott Mountains allochthon (Mull, 1982). Similar south-dipping cleavage is present in the Lisburne Group. On the basis of detailed structural studies, Julian and others (1984) and Oldow and others (1984) also suggested north-vergent thrusting.

HIGH-ANGLE FAULTS

A number of east-west-trending, vertical to steeply south-dipping, high-angle reverse faults are mapped at the eastern end of the Doonerak fenster. They have an estimated throw of as much as 1,500 ft (450 m). These faults cut the north-dipping carbonate rocks of the Lisburne Group at a number of localities (fig. 162). At Falsoola Mountain, near the eastern end of the fenster, high-angle faults cut both the isoclinally folded Lisburne Group and the Amawk thrust. This relation indicates that high-angle faulting postdated thrust emplacement of the Endicott Mountains allochthon (fig. 168) and suggests that the high-angle faulting is unrelated to the formation of the duplex.

AGE OF DEFORMATION

Dating the deformation in the area of the Doonerak fenster is based on inferences from regional stratigraphic relations, primarily on the north side of the Brooks Range. Within the fenster itself, stratigraphic evidence shows only that the Endicott Mountains allochthon was emplaced after deposition of the Late Triassic Shublik Formation and Karen Creek Sandstone. On the north flank of the Endicott Mountains, however, thrust sheets containing thick turbidites of the Okpikruak Formation have been dated as early Neocomian (Berriasian), based on the pelecypod *Buchia okensis* (Jones and Grantz, 1964). The thickness of the turbidites and presence of olistostromes (Mull, 1982, 1985; Crane, 1987) suggest a major orogeny at the beginning of the Cretaceous. The deformation is more precisely dated by a coquinoid limestone horizon at the top of the Endicott Mountains allochthon (fig. 160); the horizon contains the late early Neocomian (Valanginian) pelecypod *Buchia sublaevis* and is overlain at a thrust contact by the Okpikruak Formation. In the northern foothills, thrust sheets of the Endicott Mountains allochthon and other allochthons are unconformably overlain by gently deformed post-orogenic sediments of the Albian Fortress Mountain Formation (Mull, 1982, 1985; Mayfield and others, 1983). These relations suggest that the Endicott Mountains allochthon was emplaced between the early Neocomian and the Albian.

Regional studies suggest that the orogenic belt of the central Brooks Range was formed by at least two stages of deformation: 1) an initial stage of intense crustal shortening in the Early Cretaceous (pre-Albian), and 2) a stage of substantial vertical uplift in the core of the range, probably during Early (Albian) to Late Cretaceous (Mull, 1982, 1985; Mayfield and others, 1983). Similar relations are evident on the south side of the range (Dillon and Smiley, 1984). High-angle faulting, uplift, and folding of the Doonerak anticline probably occurred during the later stage of vertical uplift.

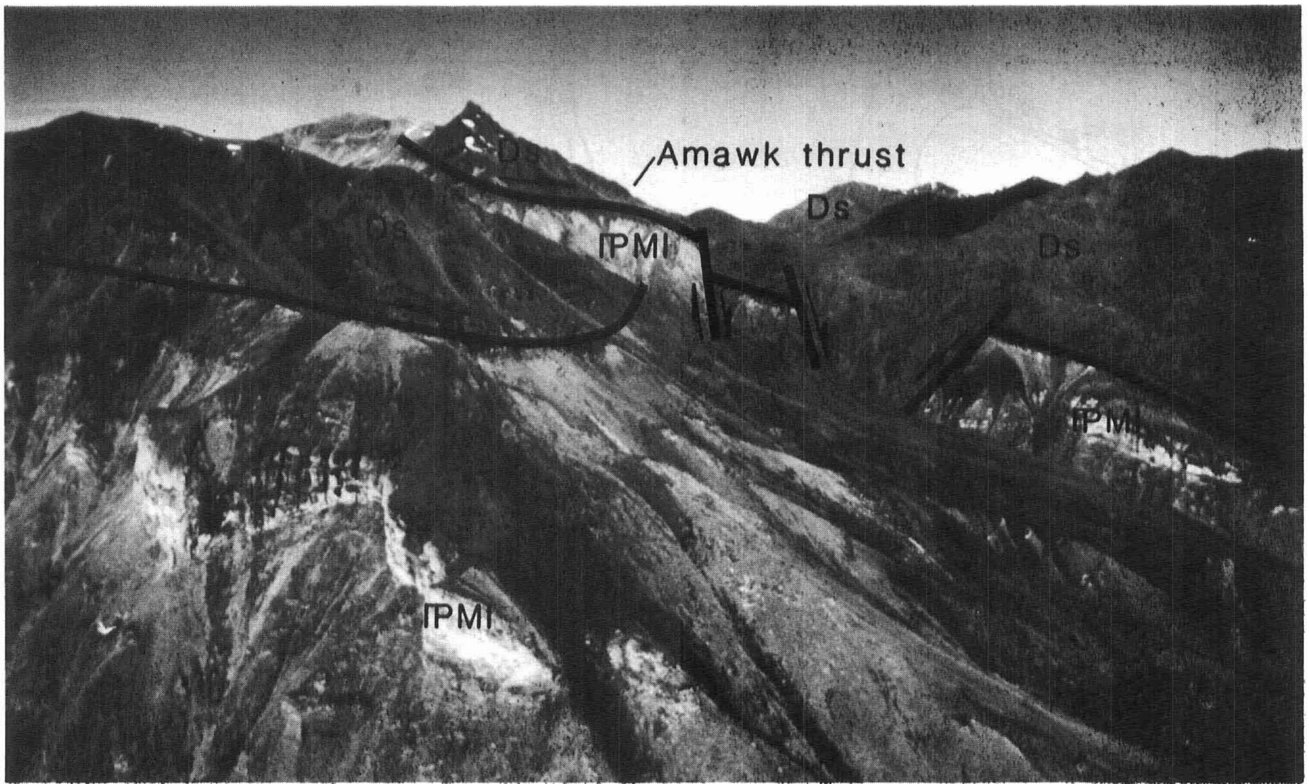

Figure 168. View westward toward Amawk thrust cut by high-angle faults on north side of Falsoola Mountain. Geologic units: Ds, Upper Devonian rocks; IPMl, Lisburne Group.

PALINSPASTIC RESTORATION OF THE ENDICOTT MOUNTAINS ALLOCHTHON

Restoration of the Endicott Mountains allochthon to a position south of the Doonerak fenster (fig. 169) results in a logical reconstruction of the Mississippian through Triassic depositional basin. In reconstructing the depositional basin, the stratigraphy of the Echooka and Siksikpuk Formations (Permian) and the Shublik (Triassic) and Otuk (Triassic and Jurassic) Formations is particularly useful because these formations show marked differences in lithology across the basin. In order to compare and contrast the Permian and Triassic stratigraphy of the Endicott Mountains allochthon with that in the Doonerak window, a number of stratigraphic sections have been measured (Adams, 1989). A preliminary evaluation of the stratigraphic data shows a systematic change in the formations from northeast to southwest but only when rocks on the allochthon are moved south of the fenster (Adams and Mull, 1985).

To illustrate this overall trend of the Permian and Triassic rocks, details of three of the stratigraphic sections--an autochthonous or parautochthonous section at Bombardment Creek in the Doonerak fenster and two allochthonous sections, one in the Atigun Gorge-Cobblestone Creek area and one farther to the southwest at Gray Mountain--are tabulated in figure 170 and discussed below.

Regional stratigraphic data (Brosgé and Dutro, 1973) show that the margin of the Permian basin was to the north and east near the present Arctic Coast and that within the basin there is a general decrease in grain size and clastic input to the south and west. This trend is illustrated by a pyritic, calcareous siltstone to very fine-grained sandstone at the base of the Permian sequence. In the northeastern Brooks Range, this siltstone interval is part of the basal member of the Echooka Formation (Detterman, 1975); on the Endicott Mountains allochthon it is an unnamed basal horizon in the Siksikpuk Formation (Siok, 1985). When the Endicott Mountains allochthon is restored south of the Doonerak fenster, this unit systematically thins to the south and west, until, in the Gray Mountain area, it is only a few feet thick.

With the decrease in clastic input to the south and west, there is a corresponding increase in shale and siliceous mudstone content. Maroon-and-green shale and siltstone with associated barite and siderite nodules dominate the allochthonous Siksikpuk Formation (Siok, 1985). In the restored position, these shales grade laterally to the northeast into dark-gray to black shale and siltstone found in the Doonerak area and northeastern Brooks Range. Similarly, a siliceous mudstone

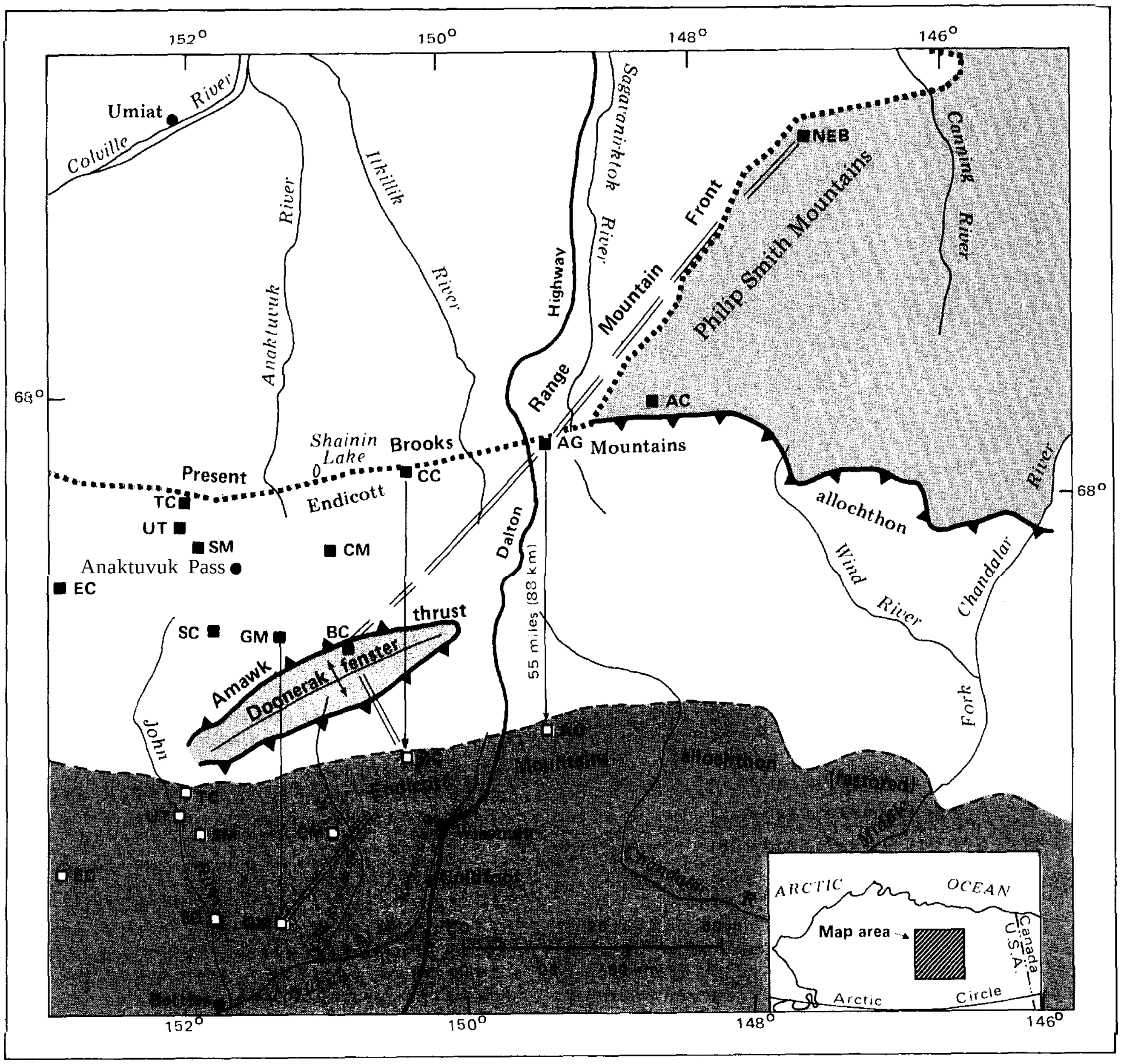

Figure 169. Palinspastic map of Endicott Mountains allochthon, based on preliminary evaluation of stratigraphic and structural data. Map shows possible northernmost location of allochthon before thrusting (shaded area at bottom); actual location is an undetermined distance farther south. Filled squares mark present location of Permian or Triassic sections; open squares mark restored position.

unit within the Siksikpuk Formation on the Endicott Mountains allochthon pinches out to the north and east: the unit is present at Gray Mountain, but there are only a few siliceous beds in the Atigun Gorge-Cobblestone Creek area and none in the Doonerak fenster.

The general increase in siliceous content to the south and west is also documented in the Triassic strata (Bodnar, 1984) when the allochthonous rocks are restored south of the Doonerak area. In the Gray Mountain area, the Triassic-Jurassic Otuk Formation on the allochthon is characterized by black shale and silicified mudstone and limestone. The silica content decreases to the northeast so that in the Doonerak fenster and northeastern Brooks Range the coeval Shublik Formation is characterized by soft, earthy, sooty limestone and shale containing abundant phosphate nodules.

A systematic change from northeast to southwest is also noted in the Upper Triassic Karen Creek Sandstone, which overlies the Shublik Formation in the northeastern Brooks Range and in the Doonerak fenster and is a member of the Otuk Formation on the northeastern Endicott Mountains allochthon (Bodnar, 1984). In the northeastern Brooks Range, the Karen Creek Sandstone is over 100 ft (36 m) thick, but it thins southwestward

Stratigraphic criteria		SW Allochthonous: Gray Mountain	Allochthonous: Atigun Gorge Cobblestone Creek	Autochthonous: Mount Doonerak area	Autochthonous: Northeast Brooks Range NE	
Karen Creek Sandstone		Absent	25 cm thick	3 m thick	30 m thick	
Limestone member	Otuk Formation	25 m thick	15 m thick	Absent	Absent	Shublik Formation
Chert member	Otuk Formation	3.5 m thick	Present	Absent	Absent	Shublik Formation
Sooty shale	Otuk Formation	Absent	Present	Abundant	Abundant	Shublik Formation
Phosphate nodules	Otuk Formation	Absent	Present	Abundant	Abundant	Shublik Formation
Siliceous mudstone	Siksikpuk Formation	24 m thick	Minor	Absent	Present	Echooka Formation
Maroon and green shale	Siksikpuk Formation	44 m thick	Minor	Absent	Absent	Echooka Formation
Barite and siderite	Siksikpuk Formation	Minor	Minor	Trace	Absent	Echooka Formation
Calcareous, pyritic siltstone		1.7 m thick	13 m thick	54 m thick	60 m thick	Echooka Formation
Pre-Lisburne stratigraphic units		Kayak Shale Kanayut Conglomerate Hunt Fork Shale	Kayak Shale Kanayut Conglomerate Hunt Fork Shale	Kayak Shale Kekiktuk Conglomerate lower Paleozoic volcanic and metamorphic rocks	Kayak Shale Kekiktuk Conglomerate lower Paleozoic volcanic and metamorphic rocks	

Figure 170. Stratigraphic criteria used to restore Mississippian through Triassic depositional basin in central Brooks Range to prethrusted position.

to <6 ft (2 m) thick at Bombardment Creek in the Doonerak fenster and to <3 ft (1 m) thick on the eastern end of the Endicott Mountains allochthon.

Restoration of the Endicott Mountains allochthon to a position south of the Doonerak fenster also results in a logical progression of Upper Devonian and Mississippian stratigraphic units. A thick section of Hunt Fork Shale (Upper Devonian) and Kanayut Conglomerate (Upper Devonian and Lower Mississippian) regionally underlies the Kayak Shale (Lower Mississippian) and Lisburne Group (Lower Mississippian and Lower Pennsylvanian) on the Endicott Mountains allochthon. In the Doonerak fenster and northeastern Brooks Range, however, the Hunt Fork and Kanayut are absent and thin Lower Mississippian Kekiktuk Conglomerate underlies Kayak Shale and Lisburne Group. The Kekiktuk, in turn, is unconformably underlain by lower Paleozoic metamorphic and volcanic rocks that are not present on the Endicott Mountains allochthon.

This preliminary analysis of the Endicott Mountains allochthon and its relation to the Doonerak fenster indicates a minimum of 55 mi (80 km) of northward displacement for the Endicott Mountains allochthon. This figure represents the minimum overlap measured from the northern limit of allochthonous rocks at the present mountain front to the approximate northernmost point, south of the Doonerak fenster, from which the allochthon could have been derived. No allowance has been made for the northern limit of the Endicott Mountains allochthon in the subsurface north of the mountain front or for shortening by imbrication and folding within the allochthon. Further, the minimum restoration is unrealistically low because the lower Paleozoic volcanic and metamorphic rocks of the core of the Doonerak fenster are not a likely source for the abundant quartz and chert found in the Kanayut Conglomerate on the Endicott Mountains allochthon. Consequently, the net crustal shortening represented by the Endicott Mountains allochthon is probably at least an order of magnitude greater than the minimum distance of 55 mi (88 km) shown on the palinspastic map. Available geologic mapping is not sufficiently detailed to permit construction of balanced cross sections across the area.

CHAPTER 15.

SEDIMENTOLOGY OF THE KANAYUT CONGLOMERATE

By T.E. Moore,[1] T.H. Nilsen,[2] and W.P. Brosgé[3]

ABSTRACT

The allochthonous Upper Devonian and Lower Mississippian(?) Kanayut Conglomerate is one of the coarsest, thickest, and most laterally extensive fluvial sequences in North America. The formation consists, in ascending order, of the Ear Peak Member (conglomerate, sandstone, and shale), the Shainin Lake Member (conglomerate and sandstone), and the Stuver Member (conglomerate, sandstone, and shale). Thickness, paleocurrent, and maximum clast-size data show that the compositionally mature Kanayut was shed from a northern highland toward the southwest into a basin that was open to the south. At the maximum extent of progradation of the Kanayut depositional system, braid-plain deposits (the Shainin Lake Member) accumulated in proximal areas and meandering-stream deposits (Ear Peak and Stuver Members) in relatively more distal locations. These deposits prograded over Upper Devonian prodelta and shallow-marine deposits of the Hunt Fork Shale and Noatak Sandstone and were transgressed in the Early Mississippian by the shallow-marine basal sandstone member of the Kayak Shale.

Near the Trans-Alaska Pipeline, the Kanayut Conglomerate crops out in two major west-trending and north-vergent thrust plates. The northern thrust plate exposes the thickest section known of Kanayut Conglomerate (8,600 ft; 2,624 m) and consists of very thick sections of the Ear Peak and Stu'ver Members and a relatively thin section of the Shainin Lake Member. These thicknesses may have resulted from the ponding of meandering-stream deposits between the major entry points of sediment into the Kanayut basin at Shainin Lake and in the northeastern Brooks Range. The southern thrust plate exposes a thinner 2,300-ft-thick (700 m) sequence of the Kanayut that cannot easily be divided into the various members. This section generally consists of finer grained deposits containing thinner fluvial cycles that resulted from deposition by smaller and more basinward distributary channels.

INTRODUCTION

The Kanayut Conglomerate and associated units of the Upper Devonian and Mississippian Endicott Group form one of the most extensive fluvial-deltaic complexes in North America. These units crop out along the crest of the Brooks Range in northern Alaska over an east-west distance of 600 mi (950 km) and a north-south distance of about 40 mi (65 km) (fig. 171). The Kanayut Conglomerate is very resistant and forms steep mountains and ridge crests, especially in the area traversed by the Trans-Alaska Pipeline (fig. 172).

The original depositional basin of the Kanayut Conglomerate and associated units was wider than the width of the present exposure belt because the rocks are stacked in generally east-striking and north-vergent allochthonous and imbricated thrust sheets. The Kanayut Conglomerate and associated units were deposited as a very large, coarse-grained delta that prograded southwestward during the Late Devonian and subsequently retreated during the Late Devonian and Early Mississippian (Nilsen and others, 1981a).

Our stratigraphic and sedimentologic data include 26 measured sections from the Kanayut Conglomerate that were presented in a series of U.S. Geological Survey open-file reports (Nilsen and others, 1980a, b, 1981b, 1982; Nilsen and Moore, 1982a). We have also previously summarized the distribution, depositional facies, regional sedimentologic variations, and tectonic significance of these units (Nilsen, 1981; Nilsen and Moore, 1982b; Moore and Nilsen, 1984; Brosgé and others, 1988). The purpose of this report is to briefly review the major features described in our earlier publications and to discuss the stratigraphic and sedimentologic relations of the Kanayut Conglomerate and associated units near the Trans-Alaska Pipeline and Dalton Highway.

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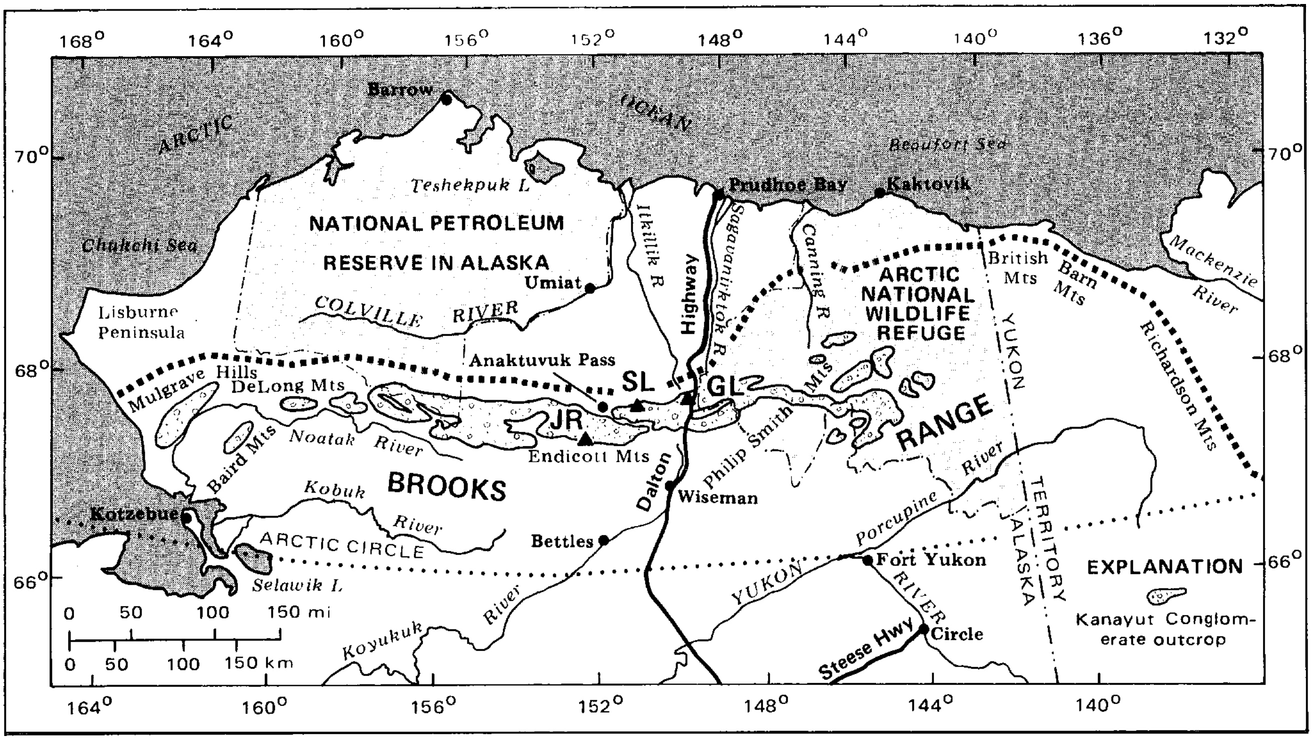


Figure 171. Index map of northern Alaska showing distribution of Kanayut Conglomerate. Dotted line marks northern edge of Brooks Range. Columnar sections were measured at Galbraith Lake (GL) and John River (JR); type sections were measured near Shainin Lake (SL).

GEOLOGIC SETTING AND PREVIOUS WORK

The middle and upper Paleozoic rocks of the Brooks Range can be divided into three principal sequences: 1) platform carbonate rocks of the Silurian to Upper Devonian Baird Group; 2) conglomerate, sandstone, and shale of the Upper Devonian to Upper Mississippian Endicott Group; and 3) platform carbonate rocks of the Mississippian to Pennsylvanian Lisburne Group (Tailleur and others, 1967). Tailleur and others (1967) considered the Endicott Group to represent a major Late Devonian offlap-onlap cycle, a sequence of events that was caused by the outbuilding of a large clastic wedge from a northern source area (Donovan and Tailleur, 1975). On the basis of regional stratigraphic relations, Nilsen (1981) divided the Endicott Group into autochthonous and allochthonous sequences (fig. 173). The allochthonous sequence was presumably thrust northward on top of the autochthonous sequence during Mesozoic orogenesis (Mull and others, 1976; Mull and Tailleur, 1977), although the exact palinspastic relation between the sequences has not yet been established (Nilsen, 1981).

The autochthonous sequence of the Endicott Group rests depositionally on deformed lower Paleozoic rocks and consists of the nonmarine Kekiktuk Conglomerate and overlying marine Kayak Shale. These units are overlain by the marine Mississippian and Pennsylvanian Lisburne Group and Permian and Triassic Sadlerochit Group. The autochthonous Endicott Group occurs mainly in the eastern Brooks Range and in the subsurface of the Arctic Slope; it is also present south of the outcrop belt of the allochthonous sequence of the Endicott Group, most notably in the Doonerak antiform (chap. 14).

The allochthonous sequence is restricted to structurally higher areas along the crest of the Brooks Range and consists, in ascending order, of the marine Hunt Fork Shale, marine Noatak Sandstone, nonmarine Kanayut Conglomerate, and marine Kayak Shale (Bowsher and Dutro, 1957; Chapman and others, 1964; Porter, 1966b). Dutro and others (1979) reported that the marine Beaucoup Formation, Frasnian (Late Devonian) in age, is locally conformably preserved at the base of the allochthonous sequence. The allochthonous sequence of the Endicott Group is conformably overlain by the Mississippian and Pennsylvanian Lisburne Group and overlying Permian and Triassic Etivluk Group. The allochthonous sequence is distinguished from the autochthonous sequence of the Endicott Group by its substantially greater thickness, faulted base, structural position, and age. In the remainder of this report, we will focus on the stratigraphy and sedimentology of the Kanayut Conglomerate and other associated units of the allochthonous sequence.

The Kanayut Conglomerate was described by Bowsher and Dutro (1957) in the Shainin Lake area and by Porter (1966b) in the Anaktuvuk Pass area; it was

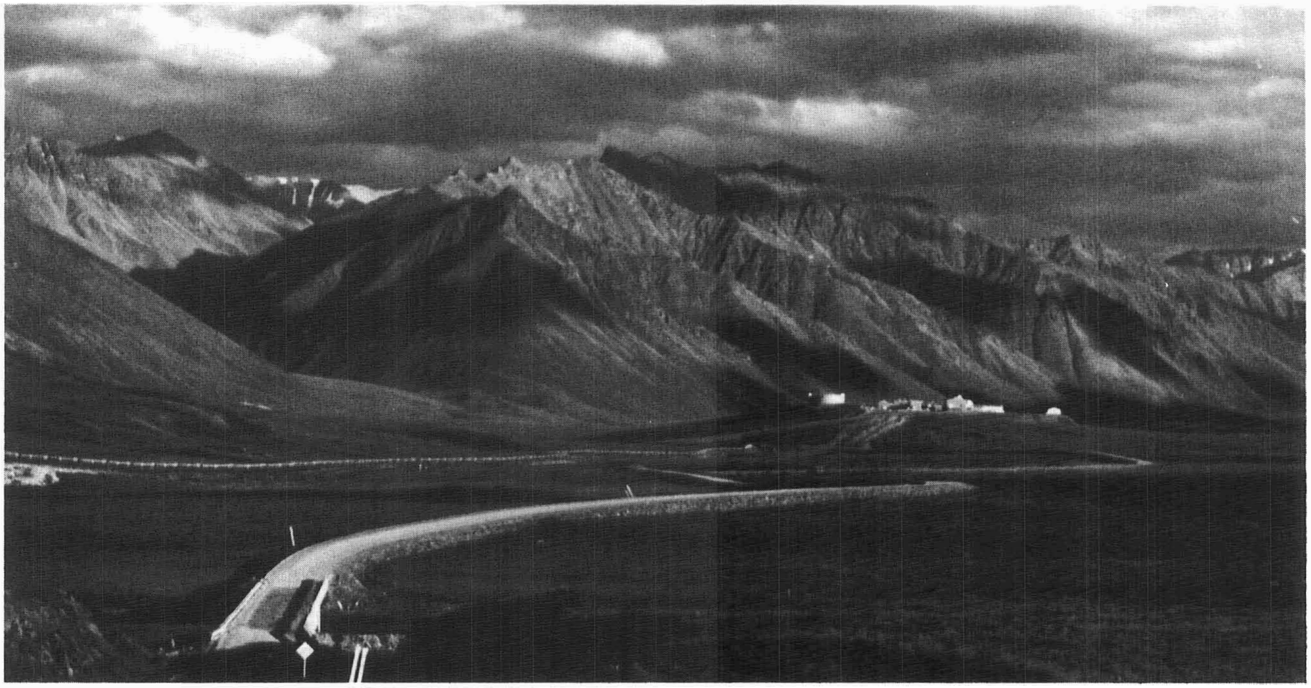


Figure 172. View southward toward peaks of Kanayut Conglomerate east of Atigun River, eastern Endicott Mountains. Peaks are up to 7,600 ft (2,300 m) high; vertical relief is 4,000 to 5,000 ft (1,200 to 1,500 m). Pump Station 4 is on low hill at right center. Photograph by C.G. Mull, August 1984.

mapped in considerable detail in some areas by Brosgé and Reiser (1962, 1964, 1965, 1971) and Brosgé and others (1976, 1979a, b). As a result of early stratigraphic work, the Kanayut Conglomerate was subdivided into three informal members and one formal member, in ascending order: the basal sandstone member (marine), the lower shale member (fluvial), the middle conglomerate member (fluvial), and the Stuver Member (fluvial). Brosgé and others (1988) reassigned the basal sandstone member of the Kanayut Conglomerate to the Noatak Sandstone, leaving the Kanayut Conglomerate as an almost entirely nonmarine formation consisting of three members. Nilsen and Moore (1984a) revised the stratigraphic nomenclature of the Kanayut Conglomerate and described type sections for its members in the Shainin Lake area. They defined and formally named the lower member the Ear Peak Member (conglomerate, sandstone, and shale) and the middle member the Shainin Lake Member (conglomerate and sandstone) to accompany the previously named Stuver Member (conglomerate, sandstone, and shale) (figs. 174 and 175).

Although brachiopods of latest Devonian age (J.T. Dutro, Jr., written commun., 1982) are present in the uppermost beds of the Kanayut Conglomerate about 9 mi (15 km) south of Anaktuvuk Pass, and plant fossils generally indicate a Late Devonian age, some plant fossils from the upper beds suggest a possible early Mississippian age (S.H. Mamay, written commun., 1980, 1982). We therefore consider the Kanayut Conglomerate to be of Late Devonian and Early Mississippian(?) age. Brachiopods from the underlying Hunt Fork Shale and Noatak Sandstone are of Frasnian and Famennian

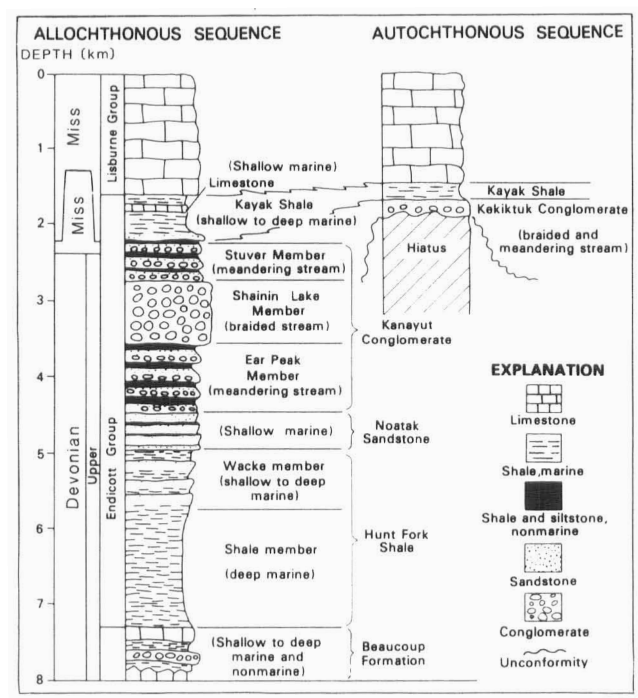


Figure 173. Simplified columnar sections show stratigraphic nomenclature and inferred depositional environments of allochthonous and autochthonous sequences of Endicott Group in central and northeastern Brooks Range. Modified from Nilsen and Moore (1984a).

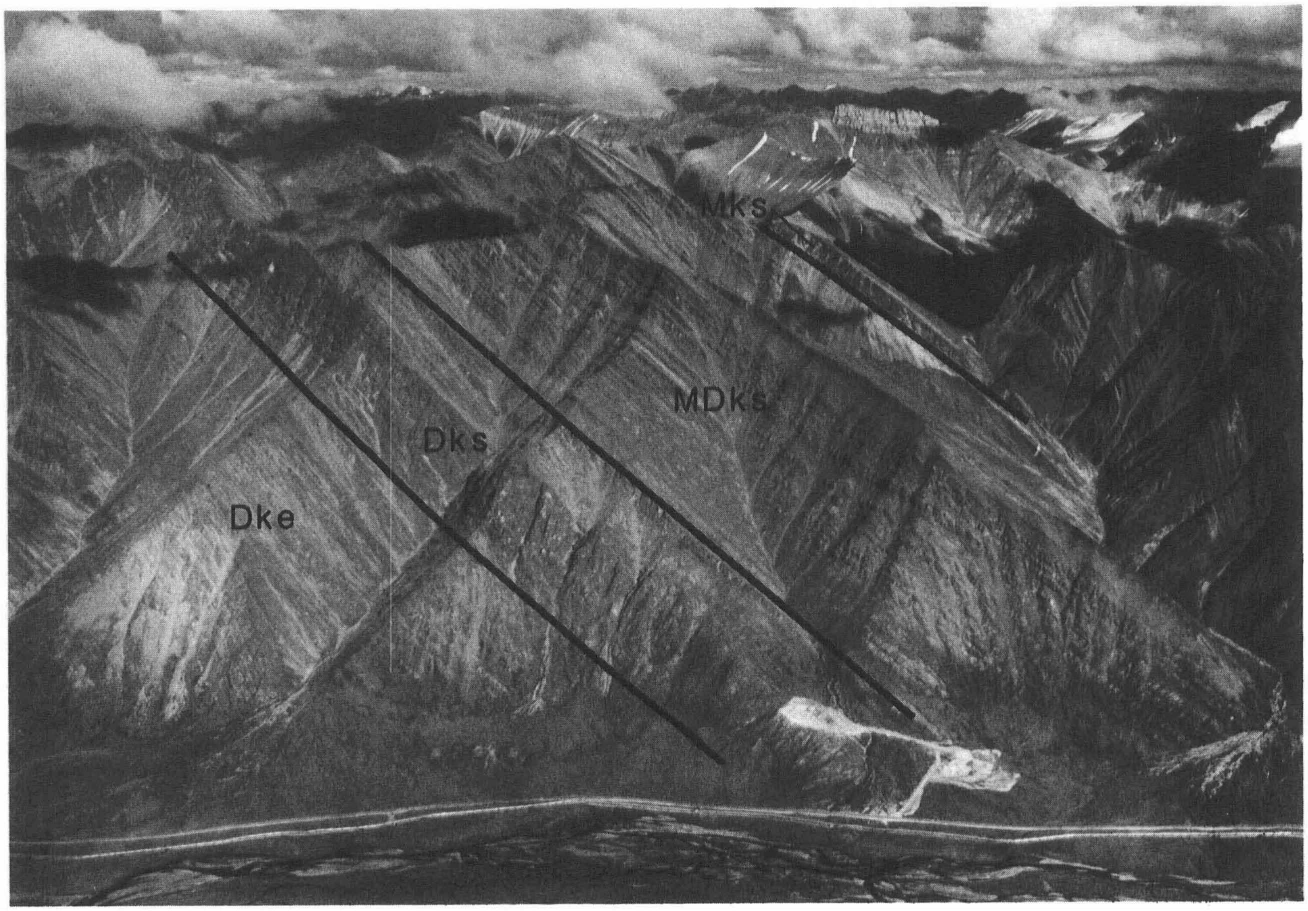


Figure 174. View eastward across Atigun River valley showing all three members of Kanayut Conglomerate and base of Kayak Shale. Geologic units: Dke, Ear Peak Member; Dks, Shainin Lake Member; MDks, Stuver Member; Mks, basal sandstone member of Kayak Shale; Mk, Kayak Shale. Photograph by C.G. Mull, August 1987.

(Late Devonian) age and those from the overlying Kayak Shale are of Kinderhookian and Osagean (Early Mississippian) age (Nilsen and others, 1980b).

The Kanayut Conglomerate is compositionally very mature, containing about 82 percent chert, 15 percent vein quartz, and 3 percent quartzite clasts. Point-count estimates from sandstone thin sections indicate a framework composition of 30 to 70 percent quartz, 10 to 50 percent chert, lesser amounts of argillite, quartzite, granitic and gneissic rock fragments, and minor amounts of feldspar, biotite, mica, tourmaline, and quartz-mica tectonite rock fragments. The composition appears to vary little laterally or vertically. Although the source area must have contained abundant siliceous sedimentary rocks and vein quartz, its original composition is volumetrically difficult to ascertain because of the apparent removal of most labile detritus by chemical weathering in the source area and by attrition during sediment transport.

SEDIMENTARY FACIES

HUNT FORK SHALE

Shale deposited in low-energy and probably deep-marine settings (at least below wave base) forms most of the lower shale .member of the Hunt Fork Shale (fig. 173) The lower member is as thick as 2,300 ft (700 m) in the Philip Smith Mountains Quadrangle (Brosgé and others, 1979a) and contains thin, graded siltstone beds that increase in abundance upward. The overlying wacke member of the Hunt Fork Shale (Dutro and others, 1977) is composed of shale and shaly siltstone interbedded with fine- to medium-grained sandstone that contains abundant brachiopod fossils. The wacke member is as thick as 2,300 ft (700 m) in the Philip Smith Mountains Quadrangle (Brosgé and others, 1979a) and is thought to have been deposited in marine-slope, outer shelf, and channel-mouth-bar environments (fig. 175).

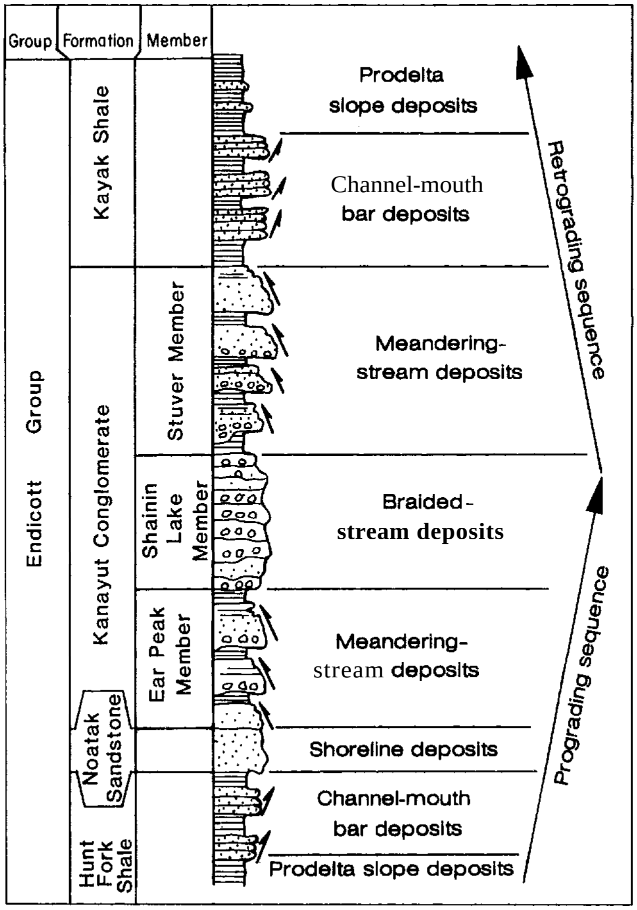


Figure 175. Diagrammatic columnar section of *allochthonous* sequence of Endicott Group, central Brooks Range. Entire sequence is about *14,000 ft (4,500 m)* thick. Arrows inclined to right indicate coarsening-upward sequences; arrows inclined to left indicate fining-upward sequences.

NOATAK SANDSTONE

The Noatak Sandstone (Dutro, 1952, **1953a,** b), previously mapped as the basal sandstone member of the Kanayut Congfomerate in the central and eastern Brooks Range (Brosge and others, **1979a,** b), consists of calcareous sandstone interbedded with black shale. The Noatak is locally conglomeratic and contains marine megafossils in many places. It is as thick as 3,200 ft (1,000 m) in the western Brooks Range (Tailleur and others, 1967) and 2,000 ft (600 m) in the Philip Smith Mountains Quadrangle (**Brosgé** and others, **1979a)** but is absent along the northern margin of much of the central and eastern Brooks Range. The Noatak Sandstone was probably deposited in a nearshore, inner shelf setting (fig. 175).

KANAYUT CONGLOMERATE

EAR PEAK MEMBER AND STUVER MEMBER

The Ear Peak Member and Stuver Member of the Kanayut Conglomerate consist of a series of **thinning-** and fining-upward cycles of conglomerate, sandstone, and shale. Because these cycles and associated sedimentary structures are similar to those described from many modern meandering rivers, we provisionally interpret the members to have been deposited by meandering streams on a flood plain (fig. 175).

Beds at the base of the cycles typically truncate underlying shale or paleosols and consist of massive or crudely plane-parallel conglomerate or conglomeratic sandstone that contain abundant rip-up **clasts** of shale, siltstone, and paleosol material.

Overlying the basal conglomeratic beds are parallel, stratified beds of sandstone that are, in turn, overlain by trough cross-stratified beds of sandstone. Trough amplitudes gradually decrease upward as grain size decreases. These deposits represent channel fill by sand waves that migrated downchannel as the stream gradually migrated laterally.

The lower coarse-grained part of each cycle typically ranges from 6 ft (2 m) to as much as 100 ft (30 m) thick and is characterized by long, inclined surfaces that cut across the vertical sequence (epsilon **cross-stratification).** These surfaces, which are present at a number of outcrops, are thought to be the original, inclined surfaces of point bars. They are most visible in **sandstone-**rich cycles and are typically 16 to 50 ft (5 to 15 m) high, 80 to 250 ft (25 to 75 m) long, and 3 to 10 ft (1 to 3 m) thick.

The upper part of each cycle consists of thinly bedded, current-ripple-marked, fine-grained sandstone with thin shale interbeds. The ripple-marked sandstone contains abundant mica, clay, and carbonaceous material. Climbing ripples are locally common in these deposits, as well as plant fossils and root impressions. These thin beds of sandstone are interpreted as levees deposited on the inner parts of meander loops by overbanking processes during flood stages.

The uppermost part of each cycle consists of interchannel and flood-plain shale and siltstone that are 3 ft (1 m) to as much as 200 ft (60 m) thick. The shale varies from red brown to black, probably depending on the amount of exposure to the atmosphere. The red shale was probably deposited chiefly on higher ground of the flood plain; the black shale, in lower swampy areas. Many cycles contain red shale directly over the sandy-levee facies, succeeded upward by black shale. Both red and black shale contain abundant fossil-plant debris, much of it in situ. Mudcracks, raindrop imprints, and features that might represent burrows, but are more likely root casts, are common. The Ear Peak Member locally contains very thick sections of red-brown shale, particularly in the eastern Brooks Range. These deposits may represent large flood-plain areas traversed by few river channels.

The fine-grained upper part of the cycles commonly contains highly oxidized yellow, orange, and red **hori-**

zons that we interpret as paleosols. These layers are generally massive and argillitic and locally contain abundant pyrite. They range from 4 in. (10 cm) to as much as 6% ft (200 cm) thick and are laterally continuous for at least 160 ft (50 m). They are generally developed in levee deposits that consist of interbedded sandstone and siltstone 3 to 10 ft (1 to 3 m) thick and are commonly situated at the top of the coarsening-upward cycles. Where present, the paleosols are commonly mottled, noncalcareous, and appear to overprint or destroy other sedimentary features, such as plant fossils and ripple marks.

Some of the sandstone cycles within the meandering-stream facies do not exhibit fining- and thinning-upward trends. These bodies of sandstone are locally channelized and may form symmetrical vertical cycles that characteristically contain abundant rip-up clasts and fragments of levee, interchannel, and flood-plain material. The bodies may be crevasse-splay deposits formed where levees had been breached during large floods.

SHAININ LAKE MEMBER

The Shainin Lake Member consists of interbedded conglomerate and sandstone that have been deposited by braided streams (fig. 175). Characteristic of these deposits are fining-upward couplets of conglomerate and sandstone, in which the base of each conglomerate bed truncates the underlying sandstone. In some sections, conglomerate rests on conglomerate to form amalgamated beds; the sandstone is absent as a result of either nondeposition or erosion.

The conglomerate-sandstone couplets are thought to represent deposition of various kinds of bars within a braided-stream complex. Sandstone was deposited on the flanks, tops, and downstream edges of gravel bars as thin, but wide, lens-shaped bodies characterized by parallel stratification, low-angle trough cross-stratification, or very low angle, inclined, tabular cross-stratification. The sandstone probably accumulated during the waning stages of floods and on the protected downstream margins of bars.

The largest conglomerate clasts are found in the Shainin Lake Member. The conglomerate is typically well imbricated, clast supported, and contains a sandstone or pebbly sandstone matrix. Long axes of pebbles are oriented parallel to flow and have proven to be useful paleocurrent indicators for the Shainin Lake Member. Paleosols, levee deposits, shale, and siltstone are rare and generally <3 ft (1 m) thick. Where present, the paleosols are typically marked by discontinuous, red oxidized layers a few inches thick at the top of conglomerate and sandstone beds.

KAYAK SHALE

In its type area near Shainin Lake, the Kayak Shale is about 1,000 ft (300 m) thick. It rests conformably on nonmarine deposits of the Stuver Member and has been subdivided into five members, in ascending 'order: 1) fine-grained basal sandstone (130 ft; 40 m); 2) lower black shale (600 ft; 180 m); 3) argillaceous limestone (80 ft; 24 m); 4) upper black shale (130 ft; 40 m); and 5) red limestone (16 ft; 5 m) (Bowsher and Dutro, 1957). The three lower members can be traced along the entire Brooks Range, despite marked changes in total thickness from thrust faulting. In the southern and eastern Brooks Range, the Kayak Shale is generally <500 ft (150 m) thick and consists mostly of black shale.

The basal sandstone member typically consists of thinly cross-stratified and ripple-marked, fine-grained quartzose sandstone with abundant burrows; it represents nearshore deposition, probably in tidal sand flats. Paleocurrent directions from the basal sandstone are highly variable and indicate flow toward the southwest, southeast, and northeast (Nilsen and others, 1980a). The overlying lower black shale member contains some thin, graded beds of fine-grained sandstone that appear to be either turbidites or vertical accumulations of storm-generated sediment overflow. The black shale member was probably deposited on a prodelta slope or outer shelf; the argillaceous limestone member consists of massive, fossiliferous debris flows of argillaceous limestone (Nilsen and Moore, 1982a). In general, the Kayak Shale was-deposited in progressively deeper water, except for the uppermost part in the central and eastern Brooks Range, where it shoals into platform carbonate rocks of the Lisburne Group.

REGIONAL SEDIMENTOLOGY AND PALEOGEOGRAPHY

The thickness of the Kanayut Conglomerate varies from east to west and also from thrust plate to thrust plate. It appears to be thickest near Galbraith Lake, about 42 mi (70 km) east of Shainin Lake, where a composite section 8,600 ft (2,624 m) thick was measured. The Kanayut Conglomerate gradually thins westward and southward to thicknesses of about 1,000 ft (300 m) at Siavlat Mountain near Howard Pass and in the Mulgrave Hills area of the western Brooks Range (Moore and Nilsen, 1984). The most marked thinning takes place in the coarse-grained Shainin Lake Member. This member is 1,725 ft (526 m) thick in its type section near Shainin Lake but cannot be distinguished in the Westernmost and southernmost outcrops of the Kanayut Conglomerate (Nilsen and Moore, 1984b).

The Kanayut Conglomerate is generally coarsest in the Shainin Lake region and in the northeastern Brooks Range and fines toward the west and south. This is illustrated by a contour map of maximum clast sizes measured in the Kanayut Conglomerate (fig. 176). Well-defined maxima are located in the Shainin Lake area, where clasts are as large as 9 in. (23 cm) in length, and in the easternmost outcrop area, where clasts are as

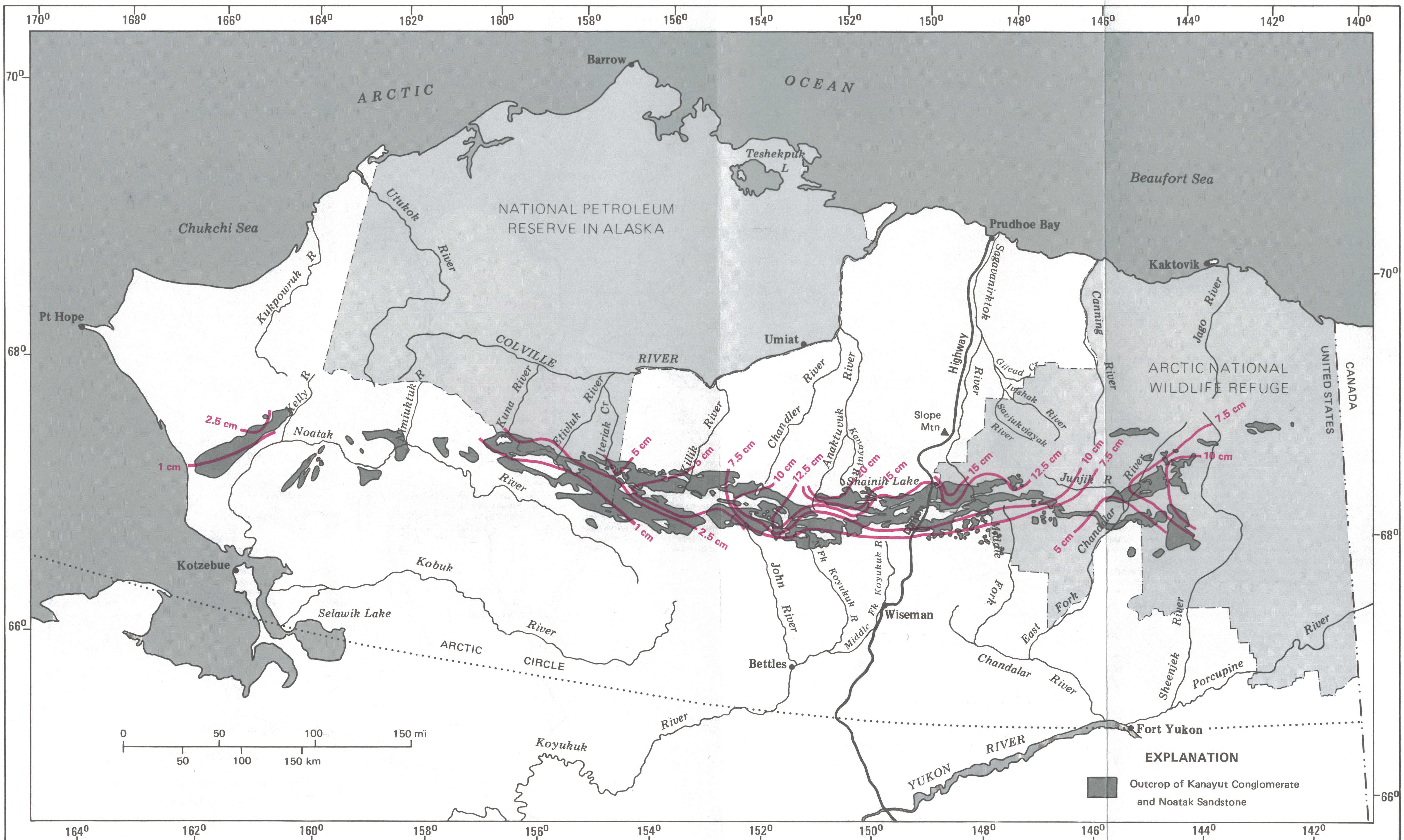


Figure 176. Contour map showing distribution of maximum clast sizes (cm) in Kanayut Conglomerate on basis of 218 measurements. Contour interval is variable.

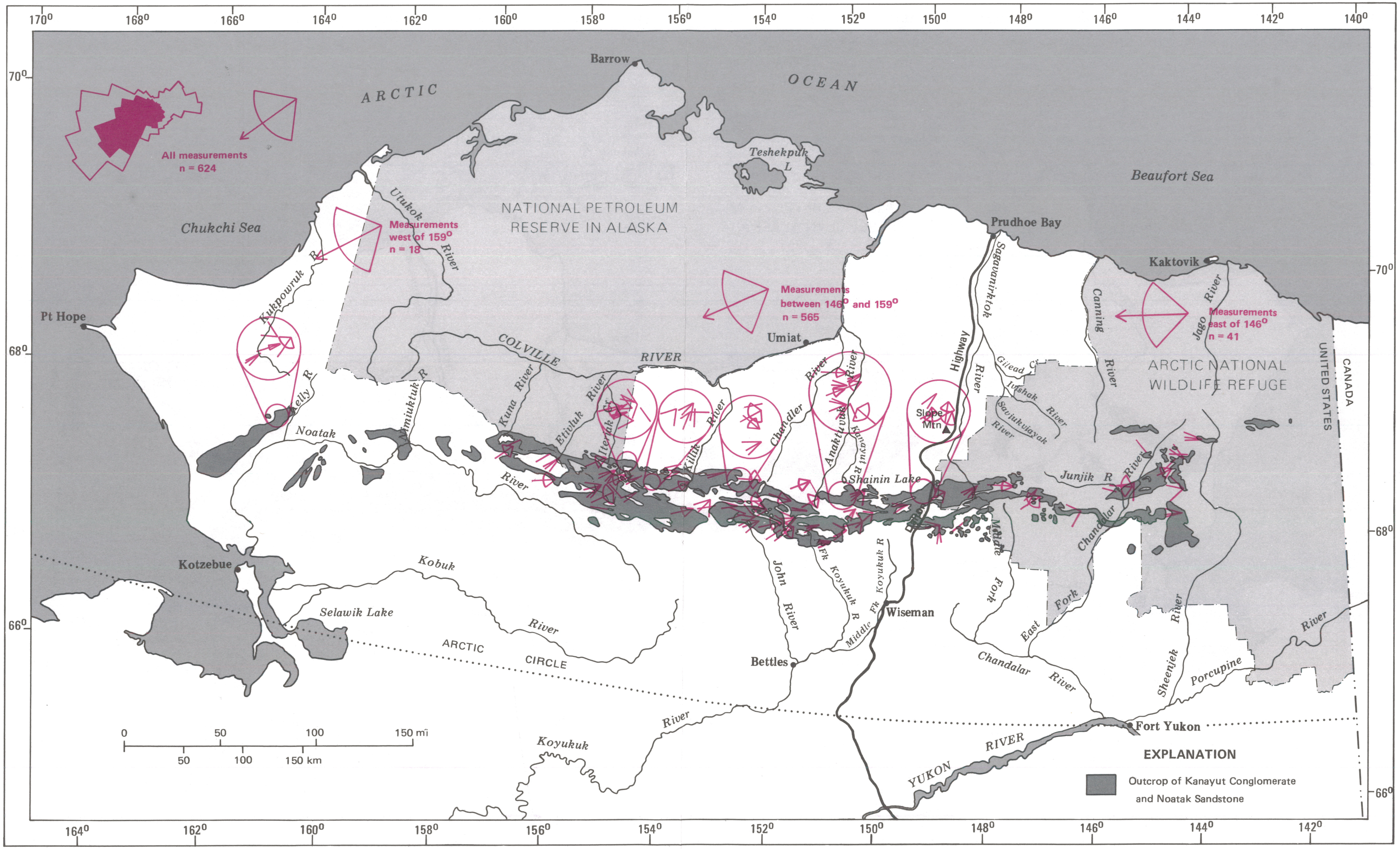


Figure 177. Paleocurrent map of Kanayut Conglomerate. Rose diagram, vector mean, and standard deviation at top summarize data from 624 measurements. Open area on rose diagram indicates measurements that gave sense of paleocurrent direction only; filled area indicates measurements that gave more precise paleocurrent direction. Sedimentary features measured include 285 trough cross-strata, 157 imbrication and long-axis orientation, 82 tabular cross-strata, 66 primary current lineations, 16 current-ripple marks, 8 flute marks, 4 pebble trains, 4 erosional scours, 2 surfaces with aligned plant fragments, 1 channel-margin orientation, and 1 fluid-escape fold.

large as 4 in. (10 cm) in length. Other clast-size maxima are present west of the headwaters of the Junjik River, near the Sagavanirktok River, east of Iteriak Creek, and south of the Kuna River. The shape of the contour lines around the clast-size maxima indicates that sediment transport was generally toward the south and west and away from the northern edge of the outcrop belt of the Kanayut Conglomerate. Clast-size data suggest that coarse-grained detritus entered the Kanayut basin from two major trunk systems, one located at the eastern end of the basin and the other near Shainin Lake. A number of smaller systems along the northern margin of the basin may have also contributed sediment to the basin.

Paleocurrent data consistently indicate westward or southwestward sediment transport and are thus compatible with the clast-size data (fig. 177). The azimuthal-vector mean and standard deviation of 624 measurements from the Kanayut Conglomerate is $242^\circ + 47^\circ$ and range from $172^\circ \pm 28^\circ$ to $304^\circ \pm 30^\circ$ for locations with more than four measurements. In general, the paleocurrent data indicate that sediment transport was somewhat more southerly ($229^\circ \pm 46^\circ$) for exposures in the western Brooks Range than for exposures in the central ($240^\circ \pm 46^\circ$) and eastern ($269^\circ \pm 48^\circ$) Brooks Range. There is little variation in sediment transport direction among the three fluvial members of the Kanayut Conglomerate; transport directions in the underlying and overlying marine units are much more variable (Nilsen and others, 1980a).

We have interpreted the finer grained Ear Peak and Stuver Members as deposits of meandering streams and the Shainin Lake Member as a deposit of braided streams. There are, however, regional variations in the fluvial character of the Kanayut Conglomerate. These include northeast to southwest decreases in both the thickness of meandering-stream cycles and the amount of erosional truncation of underlying units by the basal strata of individual fluvial cycles (Moore and Nilsen, 1984). In addition, the ratios of cumulate thicknesses of conglomerate to sandstone and sandstone to shale in the Kanayut Conglomerate decrease from northeast to southwest (Moore and Nilsen, 1984). These features indicate that streams were somewhat smaller and more distributary in southern and western exposures. Braided-stream deposits, contained mainly in the Shainin Lake Member, decrease in abundance toward the southwest. Moreover, significant amounts of marine strata are intercalated with fluvial strata in the westernmost exposures of the Kanayut Conglomerate but are rare in northeastern exposures.

The above data indicate that the Kanayut Conglomerate and associated marine units represent a major deltaic complex of Late Devonian and Early Mississippian age that prograded toward the south and west from northern and eastern highlands (fig. 178). At least two major trunk river systems fed detritus to the Kanayut delta and deposited thick braided-stream sequences (the Shainin Lake Member) in proximal areas during the time of maximum progradation. The proximal part of the delta probably consisted of a broad conglomeratic flood plain that graded southwestward into braided streams surrounded by fine-grained flood plains. More distal parts of the delta consisted of meandering streams and extensive fine-grained flood plains that graded laterally into muddy interdistributary bays and sandy shoreface deposits at the margin of the subaerial part of the delta. These rivers were probably smaller, sandier, and more ephemeral than those of the upper delta plain. At the interface between the lower delta plain and the delta front, tidal and estuarine conditions probably extended into the distributary channels. Offshore, the delta consisted of sandy channel-mouth and offshore bars that graded seaward into muddy prodelta slope sediments that accumulated mainly by vertical fallout from turbid flood- and storm-related ocean currents.

The fluvial part of the delta eventually subsided below sea level and was transgressed in Early Mississippian time by marine sediments of the Kayak Shale. The basal sandstone member of the Kayak Shale contains flaser bedding, herringbone cross-stratification, reactivation surfaces, and abundant bioturbation, all suggestive of an intertidal environment (Nilsen and others, 1980b, 1981b, 1982). Eventually, during Late Mississippian and Pennsylvanian time, a carbonate platform formed over the former delta.

Distribution of clast sizes, paleocurrents, and sedimentary facies, together with systematic variation in thickness, consistently indicates that the Kanayut Conglomerate was derived from a south-facing mountainous belt along the northern and eastern margins of the basin. Comparison of the maximum clast-size data with those of other ancient and modern river systems suggests that the coarse clastic material of the Kanayut Conglomerate in the Shainin Lake area must have been deposited close to the margin of the basin. The edge of the basin in that area, however, has either been removed by faulting or is not exposed. The data argue strongly against a southern highland, such as the Doonerak antiform, that might have shed sediment northward into the Kanayut basin. Instead, available evidence argues for a Devonian basin that was open to the south, and later, during the Mesozoic, underwent large-scale northward thrusting of a minimum of 55 mi (88 km) (Mull and others, chap. 14).

MEASURED SECTIONS OF THE KANAYUT CONGLOMERATE NEAR THE TRANS-ALASKA PIPELINE

Along the Trans-Alaska Pipeline between Atigun Pass and Galbraith Lake, the Kanayut Conglomerate crops out in a series of south-dipping thrust sheets (fig. 179). These thrust sheets imbricate and duplicate the three members of the Kanayut Conglomerate, the underlying Hunt Fork Shale and Noatak Sandstone, and the overlying Kayak Shale. The thrusts root in a major regional detachment surface in the Hunt Fork Shale (Moore, 1984). Upward, the thrust plates contain successively southerly, more distal facies of the Kanayut Conglomerate.

On the basis of regional structural and sedimentologic relations, the Kanayut Conglomerate can be divided into two major thrust sequences near the Trans-

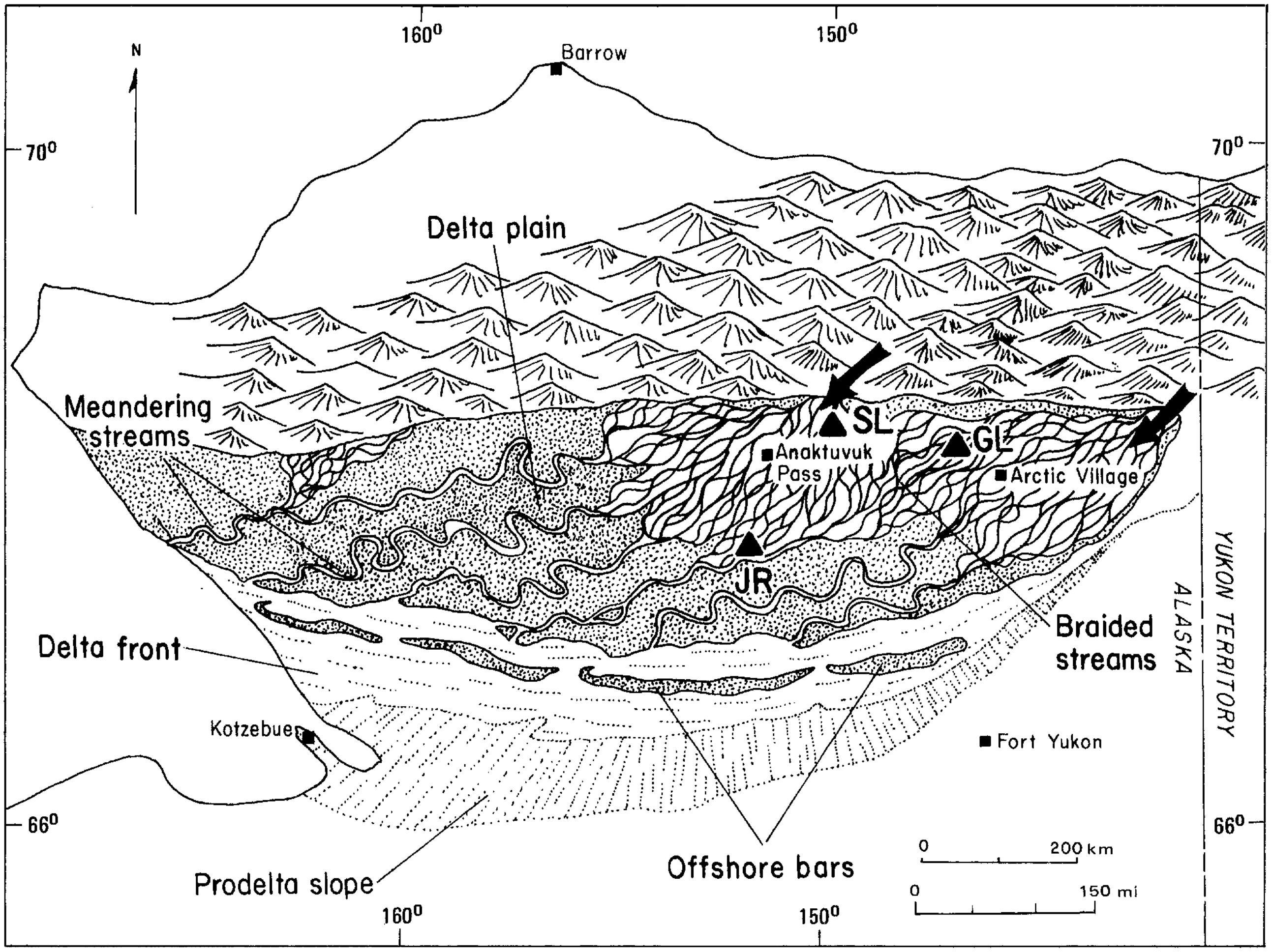


Figure 178. Paleogeographic map of Kanayut Conglomerate and associated marine units showing approximale extent of delta. Columnar sections were measured at Galbraith Lake (GL) and John River (JR); type sections were measured near Shainin Lake (SL).

Alaska Pipeline. The northern sequence is well exposed along the Atigun River valley south of Galbraith Lake and north of the junction of the east and west forks of the Atigun River (fig. 179). In this area, the Kanayut Conglomerate reaches its maximum thickness of 8,600 ft (2,624 m). A composite section of the entire thickness, previously reported by Nilsen and others (1982), is described below. The sections were all measured on high ridges on both sides of the Atigun River valley about 9 mi (15 km) south of Galbraith Lake near Mile 260 on the Dalton Highway. The northern sequence is characterized by very thick sections of both the Ear Peak and Stuver Members and a relatively thin section of the Shainin Lake Member. The unusually thick finer grained deposits in this area may have resulted from deposition midway between the entry points of the two major trunk systems into the basin. This position, located along the northern depositional axis of the basin, was away from the major sources of coarse-grained sediment and associated braid-plain deposits.

The second, more southerly, thrust sequence of Kanayut Conglomerate crops out near Atigun Pass (Mile 244) above a thick, structurally complex sequence of Hunt Fork Shale and Noatak Sandstone. The Kanayut Conglomerate in this thrust sequence generally consists of medium- to thick-bedded sandstone. conglomerate, and shale. Because the Ear Peak and Shainin Lake Members are difficult to differentiate in this area, and the Stuver Member is not recognizable, the Kanayut Conglomerate is shown as undivided in figure 179.

We have not measured a section of Kanayut Conglomerate in the Atigun Pass area, but we have measured a complete section of Kanayut about 70 mi (110 km) to the west on the John River (fig. 171). This section, previously presented by Nilsen and others (1982), is located in the same structural position and is probably part of the same thrust sequence as that at Atigun Pass. Although the Kanayut Conglomerate cannot be easily divided into its three members for mapping purposes, we have attempted to do so on the basis of the John River section.

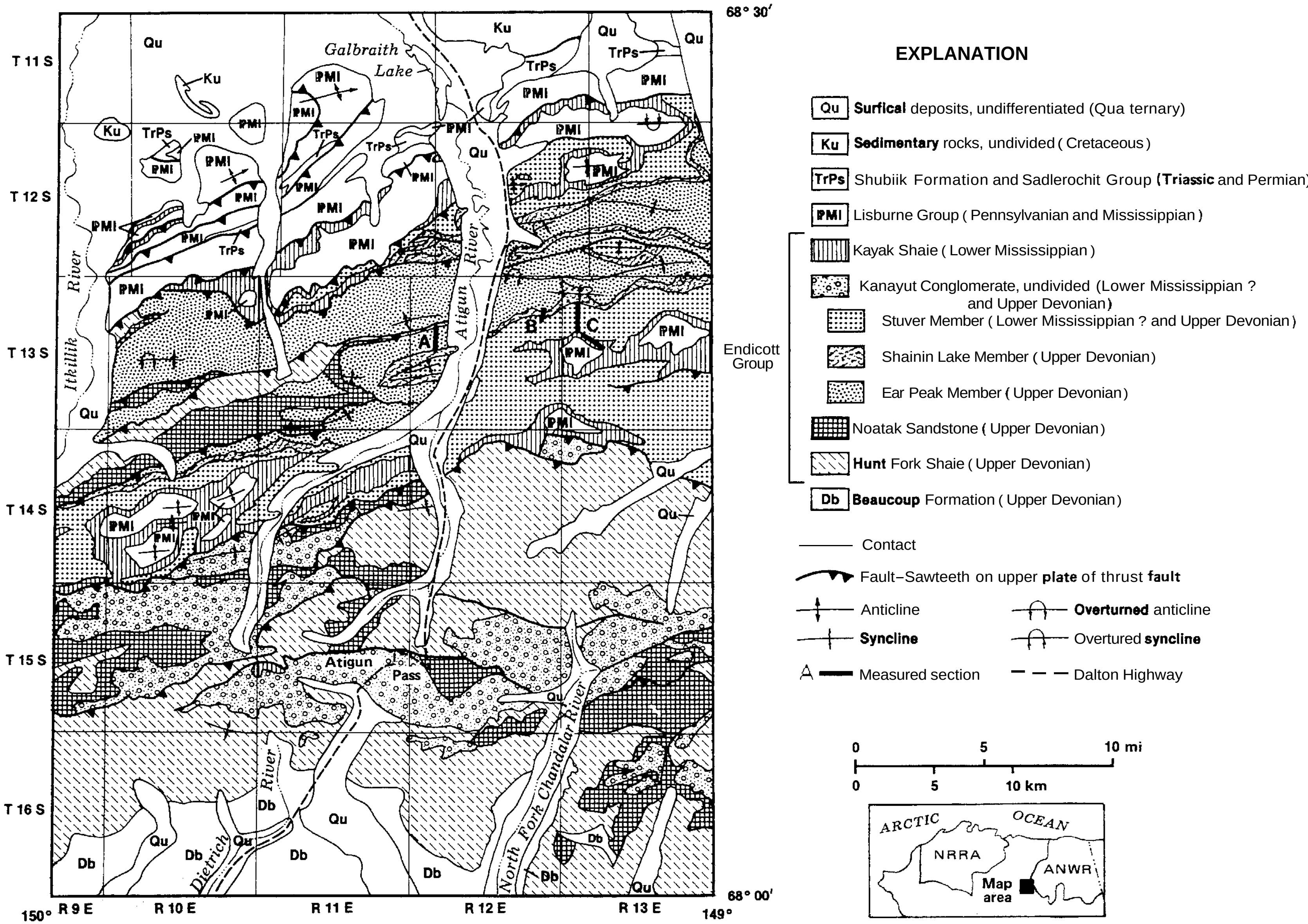


Figure 179. Simplified geologic map of Kanayut Conglomerate and associated units near Dalton Highway and Trans-Alaska Pipeline along Atigun River, northcentral Endicott Mountains. Geology modified from Brosgé and others (1979a). Columnar sections described in text: A, Ear Peak Member (fig. 180); B, Shainin Lake Member (fig. 183); C, Stuver Member (fig 185).

ATIGUN RIVER VALLEY AREA

EAR PEAK MEMBER

A section that includes the upper part of the Hunt Fork Shale and the overlying Ear Peak Member of the Kanayut Conglomerate was measured along the west side of the Atigun River valley, 9 mi (15 km) south of Galbraith Lake (NW¼, sec. 18, T. 13 S., R. 12 E.), Philip Smith Mountains Quadrangle (fig. 180). The lower part of the section was measured along a small south-flowing tributary creek; the upper part was measured on the shoulder of the ridge south of the creek. The complete measured section is 4,632 ft (1,412 m) thick and can be viewed from near Mile 260 on the Dalton Highway.

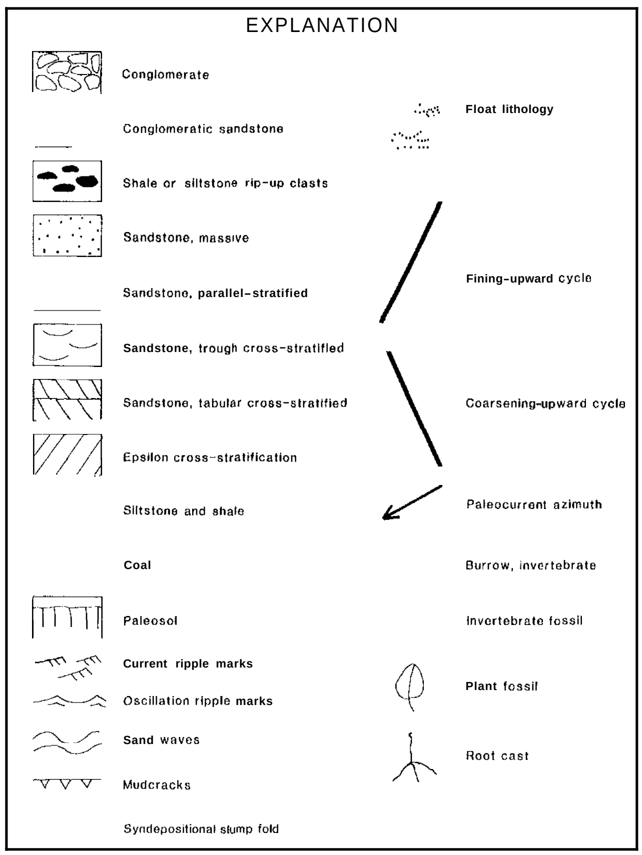
Because the lowermost part of the Kanayut Conglomerate is fine grained and interfingers with the Hunt Fork Shale at this locality, neither we nor Brosge and others (1979a) recognized the Noatak Sandstone here. We placed the contact between the Hunt Fork Shale and the Ear Peak Member of the Kanayut Conglomerate at the lowest prominent couplet of conglomerate and coarse-grained sandstone (830 ft; 253 m), thus delimiting a section of the Ear Peak Member that is 3,806 ft (1,160 m) thick.

The lower 827 ft (252 m) of the measured section consists of the Hunt Fork Shale. At the base of the Hunt Fork are thiekening- and coarsening-upward cycles of shale, siltstone, and fine- to medium-grained sandstone that contain scattered fragments of plant fossils. The fine-grained parts of the cycles are gray to black, ripple marked, and contain abundant marine Scolithus-type burrows. The coarse-grained upper parts of the cycles are trough cross-stratified, parallel stratified, or massive. These cycles have been deposited as marine bars in a delta-front to delta-slope setting.

Within the Hunt Fork Shale, fining- and thinning-upward fluvial cycles intercalated with *Scolithus*-bearing marine strata are present between 329 and 827 ft (100 and 252 m) above the base of the section. The fluvial cycles average about 33 ft (10 m) thick and are best developed in the interval between 492 and 755 ft (150 m to 230 m). They typically consist of fine- to medium-grained sandstone that fines upward into gray and black ripple-marked siltstone and shale. The base of each cycle is erosive. The sandstone is trough cross-stratified and contains abundant rip-up clasts, plant fragments, and, rarely, pebbles as large as 4/5 in. (2 cm). Large inclined surfaces (epsilon cross-stratification) are well developed. Marine strata that are present at the tops of some of the fining-upward cycles consist of burrowed, thinly bedded siltstone and fine- to very fine grained sandstone that are parallel stratified, ripple marked, and lenticular. The marine deposits decrease in thickness and abundance upward toward the base of the overlying Ear Peak Member of the Kanayut Conglomerate. Burrowed intervals of marine(?) origin, however, are present 1,935 ft (590 m) above the base of the section in the lower part of the Ear Peak Member.

The lowest part of the Ear Peak Member is marked by two prominent fining-upward cycles at 830 and 859 ft (253 and 262 m) that contain conglomeratic sandstone at their bases. The cycles near the base of the member, between 830 and 1,329 ft (253 and 405 m), typically contain fine- to medium-grained, trough cross-stratified sandstone at their base and are overlain by parallel-stratified, very fine grained sandstone, siltstone and shale. The fine-grained upper part of some cycles is thin and contains Scolithus-type burrows, which indicate a marine origin. Two cycles between 984 and 1,247 ft (300 and 380 m) contain thick accumulations of interstratified ripple-marked, fine-grained sandstone and shale that lack burrows and may represent either fluvial-flood-plain or freshwater-lake deposits. The interbedded marine and fluvial strata near the contact between the Hunt Fork Shale and the Kanayut Conglomerate record a gradual progradation of coarse-grained delta-plain meandering streams over fine-grained delta-front sediments. The complex interfingering probably reflects vertical and lateral accretion of fluvial-meandering-channel, interdistributary-bay, brackish-swamp, low-energy-shoreline, and freshwater-lake deposits.

The middle and upper parts of the Ear Peak Member, between 1,329 and 4,593 ft (405 and 1,400 m), consist of thinning- and fining-upward cycles inferred to be meandering-stream deposits (fig. 181). These cycles contain coarser grained deposits than those at the base of the member and thicker basal sandstone and conglomerate units, which exhibit multiple scour horizons and local trough cross-stratification. The coarse basal



[To be used with figures 180, 183, 185, and 188]

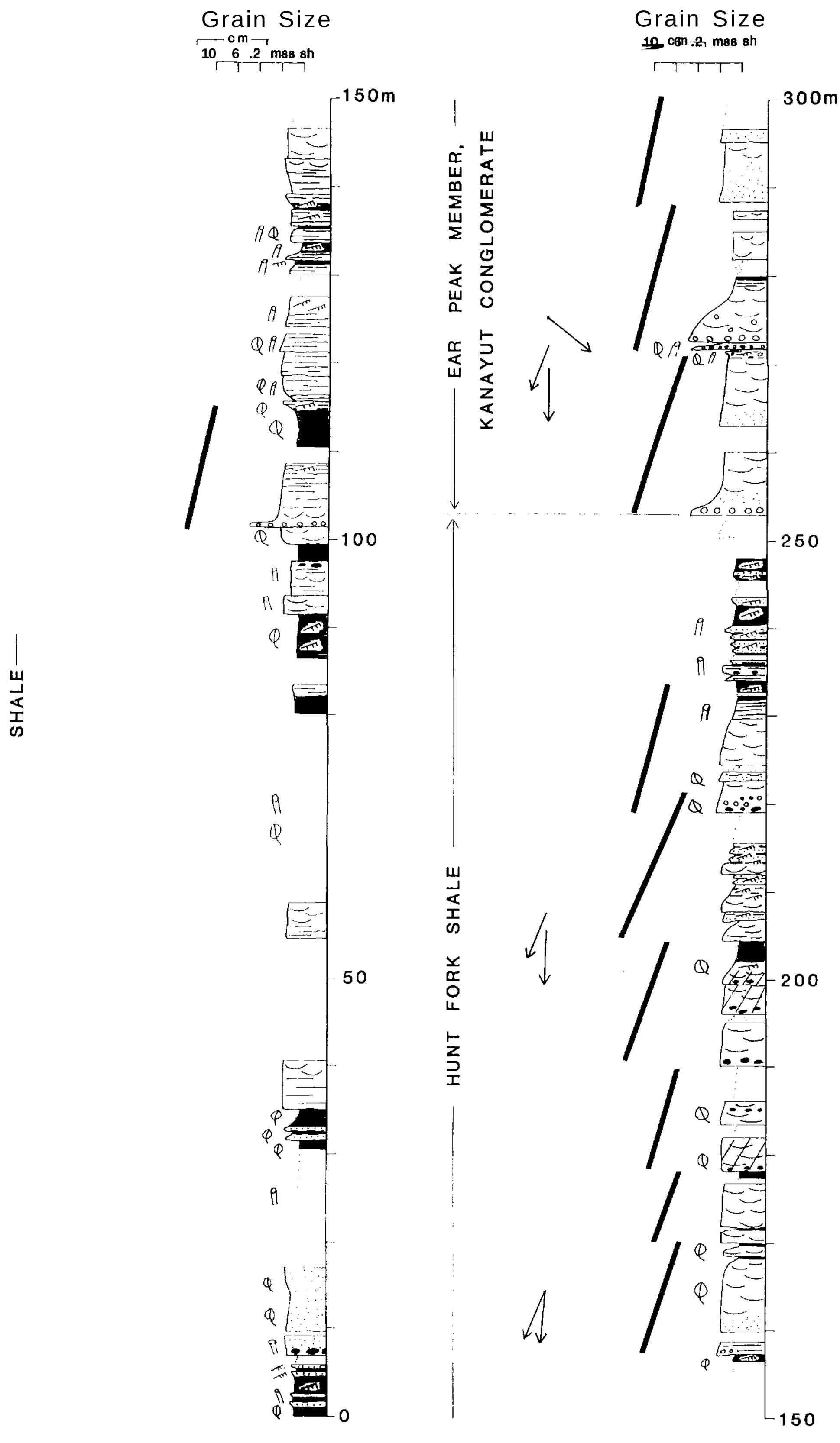


Figure 180. Columnar section of uppermost part of Hunt Fork Shale and overlying Ear Peak Member of Kanayut Conglomerate measured on west side of Atigun River valley. See figure 179 for location of section.

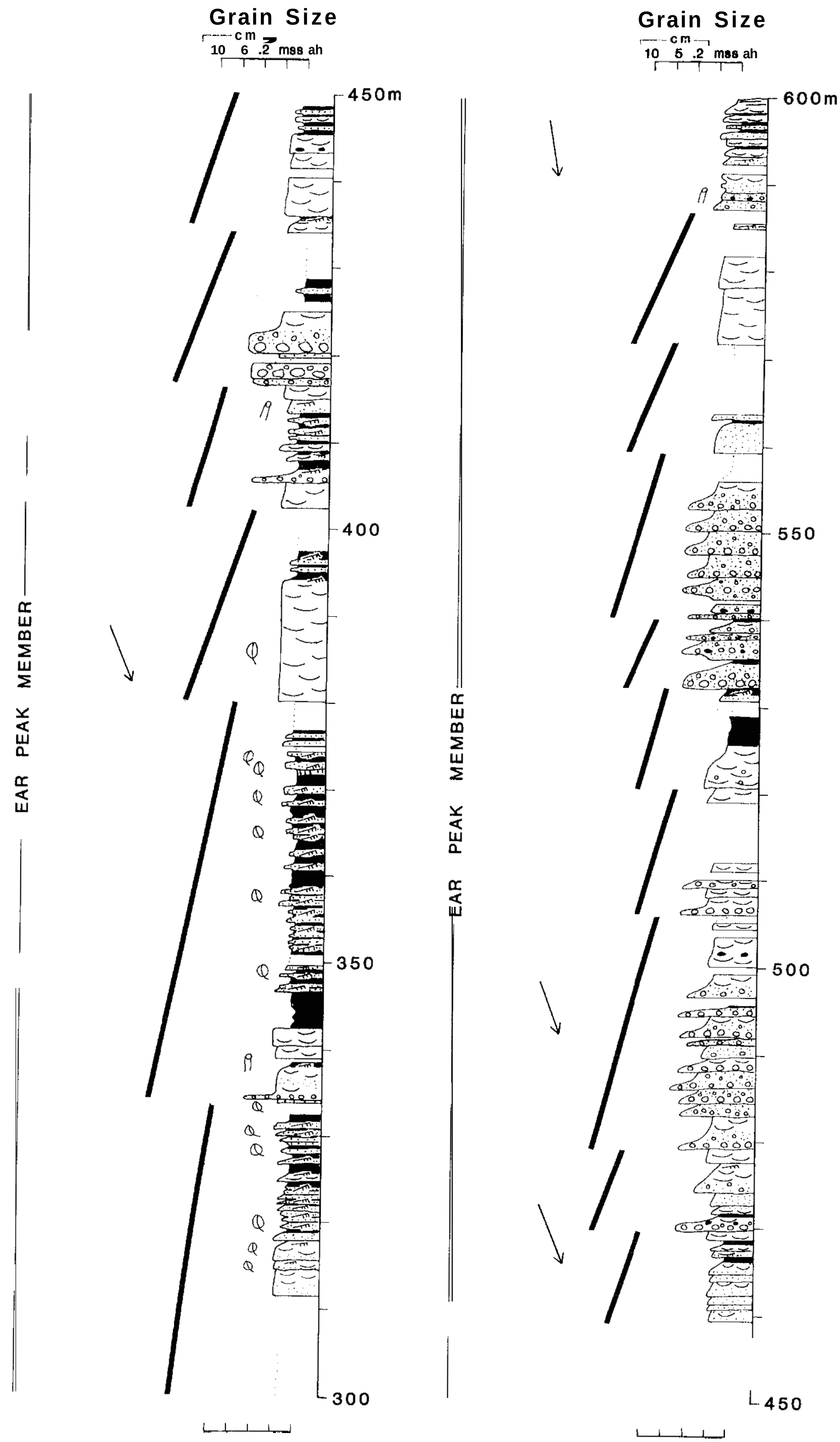


Figure 180. Continued.

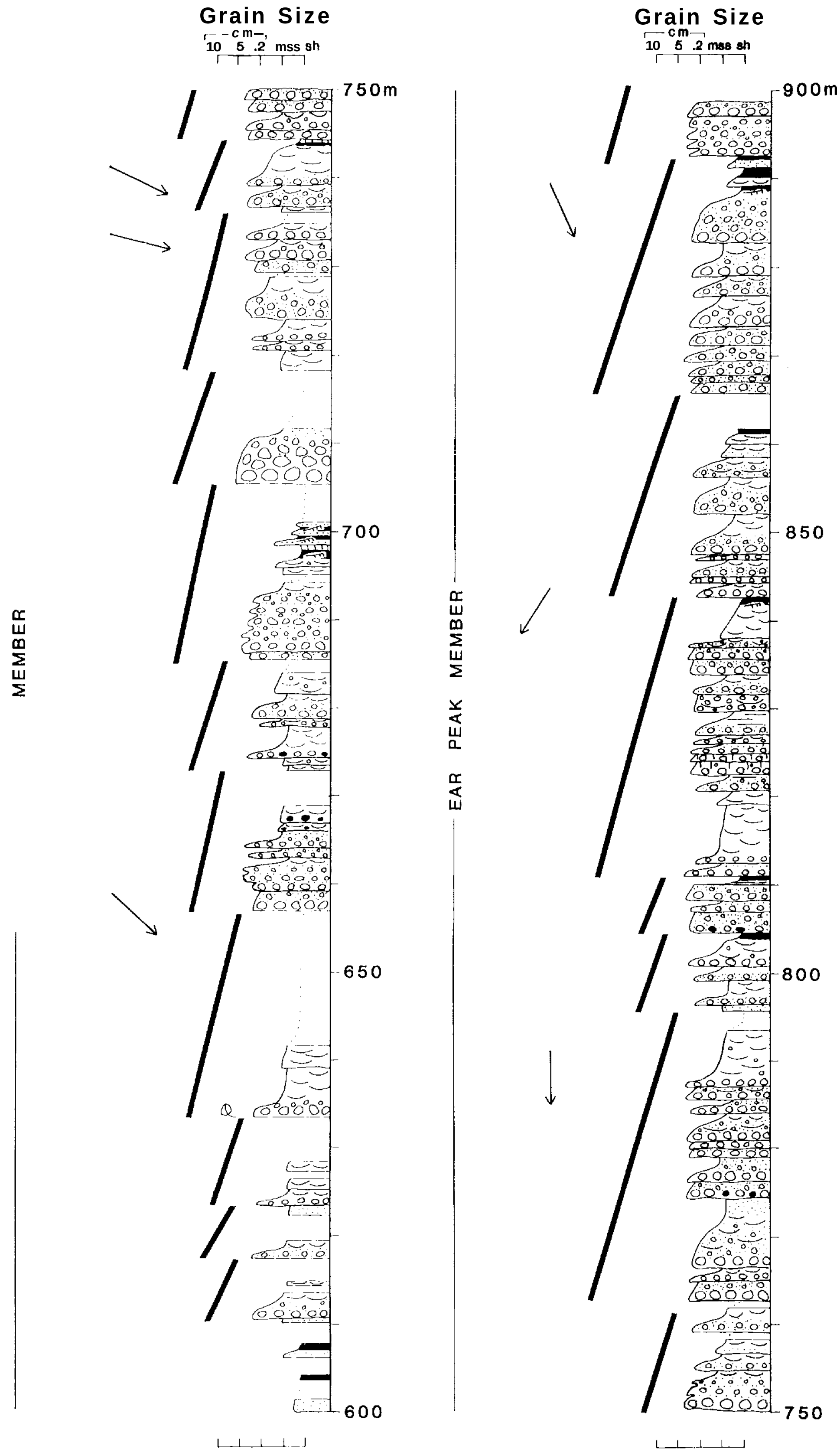


Figure 180. Continued.

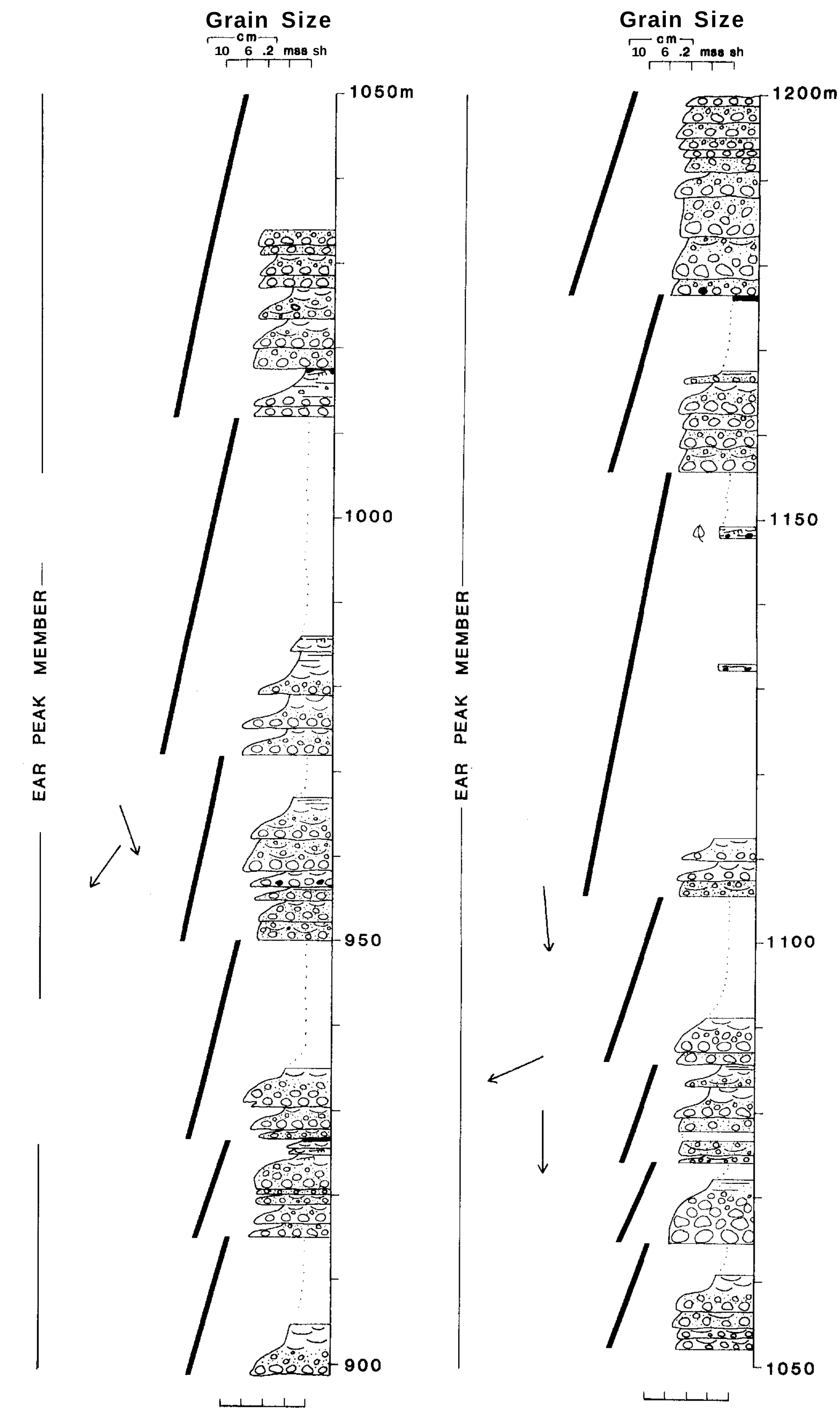


Figure 180. Continued.

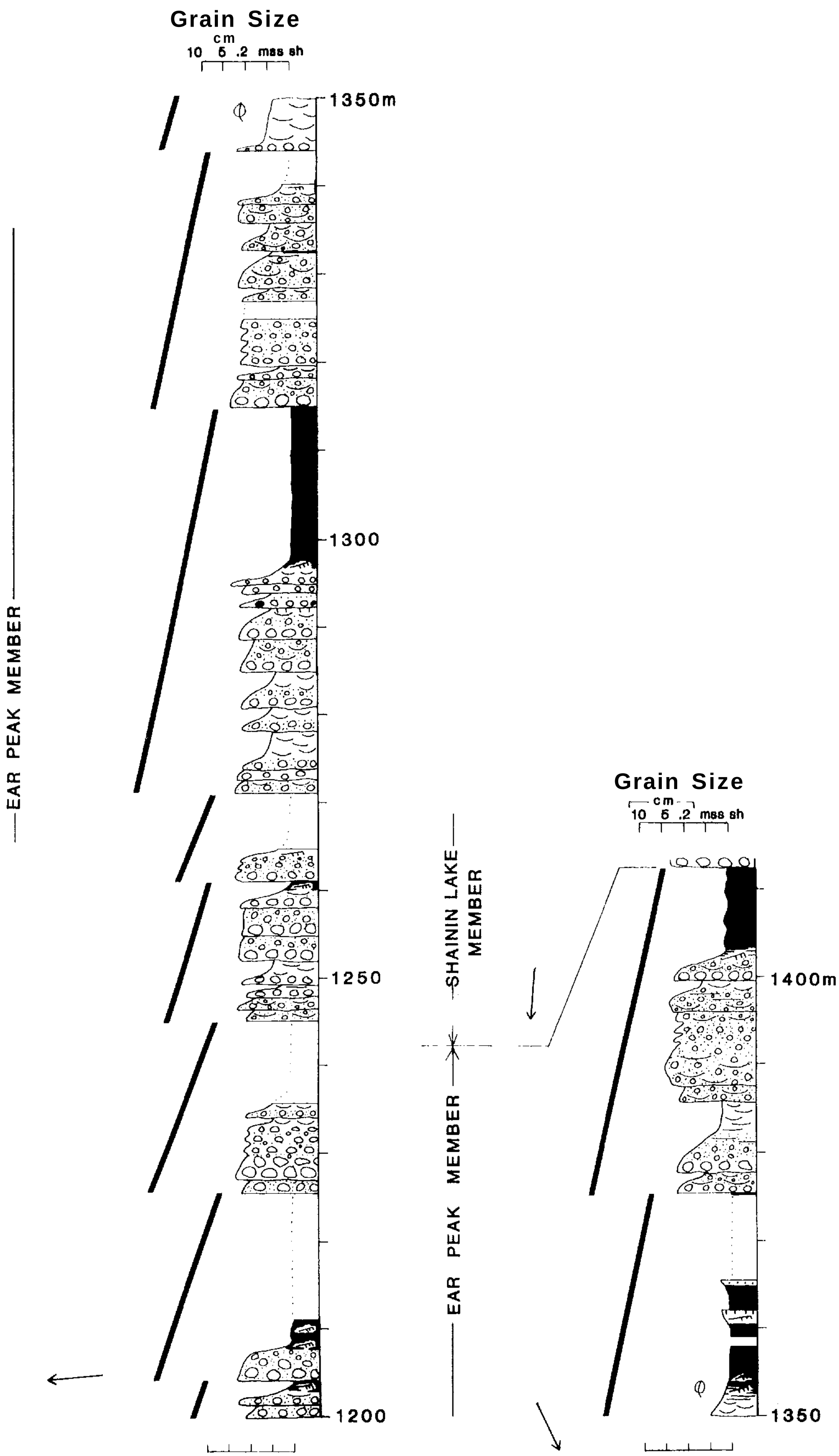


Figure 180. Continued.

facies of each cycle probably represents channel and lower point-bar deposits. The basal deposits are typically overlain by trough cross-stratified, fine- to medium-grained sandstone, which is capped by red, maroon, or black siltstone and shale. These fine-grained sediments probably accumulated on the upper surface of point bars, levees, and flood plains. Shale and siltstone intervals are thin or nonexistent in the middle part of the Ear Peak Member between 2,362 and 2,920 ft (720 and 890 m) but increase in thickness upward toward the top of the section. Near the top, the flood-plain deposits are as thick as 66 ft (20 m) and commonly rest directly on conglomeratic strata without intervening sandstone strata of substantial thickness. One well-exposed flood-plain sequence near the top, between 4,430 and 4,479 ft (1,350 and 1,365 m), contains two cycles of black shale, thin ripple-marked sandstone, and a paleosol. These thin, coarsening-upward cycles are 7 to 10 ft (2 to 3 m) thick and probably record outbuilding of levees into adjacent swampy lowlands or lakes on the flood plains.

The contact with the overlying Shainin Lake Member at the top of the section is abrupt and marked by the disappearance of shale and the appearance of massive sandstone and conglomeratic strata (fig. 182). This contact reflects the influx of coarser grained sediment and the transition to deposition by braided streams.

Cycles are generally of intermediate thickness (66 to 98 ft; 20 to 30 m) near the base of the Ear Peak Member, thinnest (33 to 66 ft; 10 to 20 m) in the middle part, and thickest (148 ft; 45 m) near the top. Clast size increases regularly upward. In the Hunt Fork Shale and in the lowest part of the Ear Peak Member, conglomerate is rare and maximum clast size is 4/5 in. (2 cm); in the uppermost part of the Ear Peak Member, maximum clast size is 2 in. (5 cm). Two pebble counts from the Ear Peak Member show that clast composition is typical of the Kanayut Conglomerate, averaging 85 percent chert and 15 percent vein quartz. Quartzite is absent, and red chert composes two percent of the conglomerate clasts.

Paleocurrent measurements from the section are consistent in orientation and indicate southward sediment transport. There is little difference in transport direction between the Hunt Fork Shale and the Ear Peak Member of the Kanayut Conglomerate. The mean and standard deviation of eight measurements from the Hunt Fork Shale are $186^{\circ} \pm 21^{\circ}$; mean and standard deviation of 17 measurements from the Ear Peak Member are $171^{\circ} \pm 38^{\circ}$.

SHAININ LAKE MEMBER

A complete section of the Shainin Lake Member of the Kanayut Conglomerate was measured on a steep, east-facing valley wall along a small northeast-trending creek about 9 mi (15 km) southeast of Galbraith Lake (NE¼, sec. 12, T. 13 S., R. 12 E.), Philip Smith Mountains Quadrangle (fig. 183). The Shainin Lake Member at this location is 509 ft (155 m) thick and is conformably underlain and overlain by the Ear Peak and Stuver Members, respectively. The contacts between the members are abrupt, and the members can be differentiated by the absence of shale in the Shainin Lake Member. The basal 39 ft (12 m) of the Stuver Member (1 cycle) were also measured at this location.

The measured section consists almost exclusively of interstratified conglomerate, conglomeratic sandstone, and coarse-grained sandstone. Fine-grained sandstone, siltstone, and shale are present only in the basal part of the Stuver Member in the uppermost part of the section. Conglomerate beds are as thick as 30 ft (7.3 m), are lenticular, and contain clasts as large as 4 in. (10 cm) diam. Sandstone beds are as thick as 5 ft (1.4 m), are commonly conglomeratic, and contain pebbles as large as 1 in. (3 cm) diam.

The conglomerate and sandstone beds form fining-upward couplets, in which the conglomeratic base of each couplet is channeled into the underlying sandstone or conglomeratic sandstone (fig. 184). The conglomerate is clast-supported; typically massive to crudely parallel stratified; and has a well-defined fabric, characterized by imbricated pebbles and the long axes of clasts orientated parallel to flow. The matrix consists of sandstone and finer conglomeratic sandstone. Beds of sandstone are generally parallel to trough cross-stratified but are locally massive. They form thin, flat lenses in outcrops that are truncated by beds of conglomerate. In the interval between 328 and 492 ft (100 and 150 m), the conglomerate beds are commonly overlain by a scoured surface and another conglomerate bed, which forms a sequence of amalgamated conglomerate. This interval contains the coarsest conglomerate. Although partly covered, two fining- and thinning-upward cycles can be recognized toward the top of the section. The overlying conglomerate is scoured into the fine-grained deposits and locally contains well-developed flute casts at its base.

The lower 509 ft (155 m) of the section was probably deposited by braided streams. Fining-upward conglomerate-sandstone couplets within this lower interval were probably formed by lateral and vertical accretion of migrating longitudinal bars. Migrating dunes and sand waves superimposed on larger bars resulted in cross-bedded sandstone. The transition to the thicker, fining-upward cycles of the Stuver Member may mark a change to deposition by a meandering-stream system. The coarse basal parts of these thicker cycles are interpreted to be river-channel and point-bar deposits, and the fine upper parts to be upper point-bar, levee, and flood-plain deposits.

Two pebble counts of the Shainin Lake Member indicate an average composition of 88 percent chert, 11 percent vein quartz, and 1 percent quartzite. Red chert clasts are very abundant and form about 40 percent of the total number of clasts.

Four paleocurrent measurements of trough cross-strata axes, clast imbrication, and clast long axes indicate that sediment transport was consistently southward during the deposition of the Shainin Lake Member. However, flute casts in the uppermost part of the section are oriented toward the west, suggesting a change to more westerly sediment transport during deposition of the basal strata of the Stuver Member. The mean and standard deviation for all seven paleocurrent measurements from this section are $188^{\circ} \pm 43^{\circ}$.

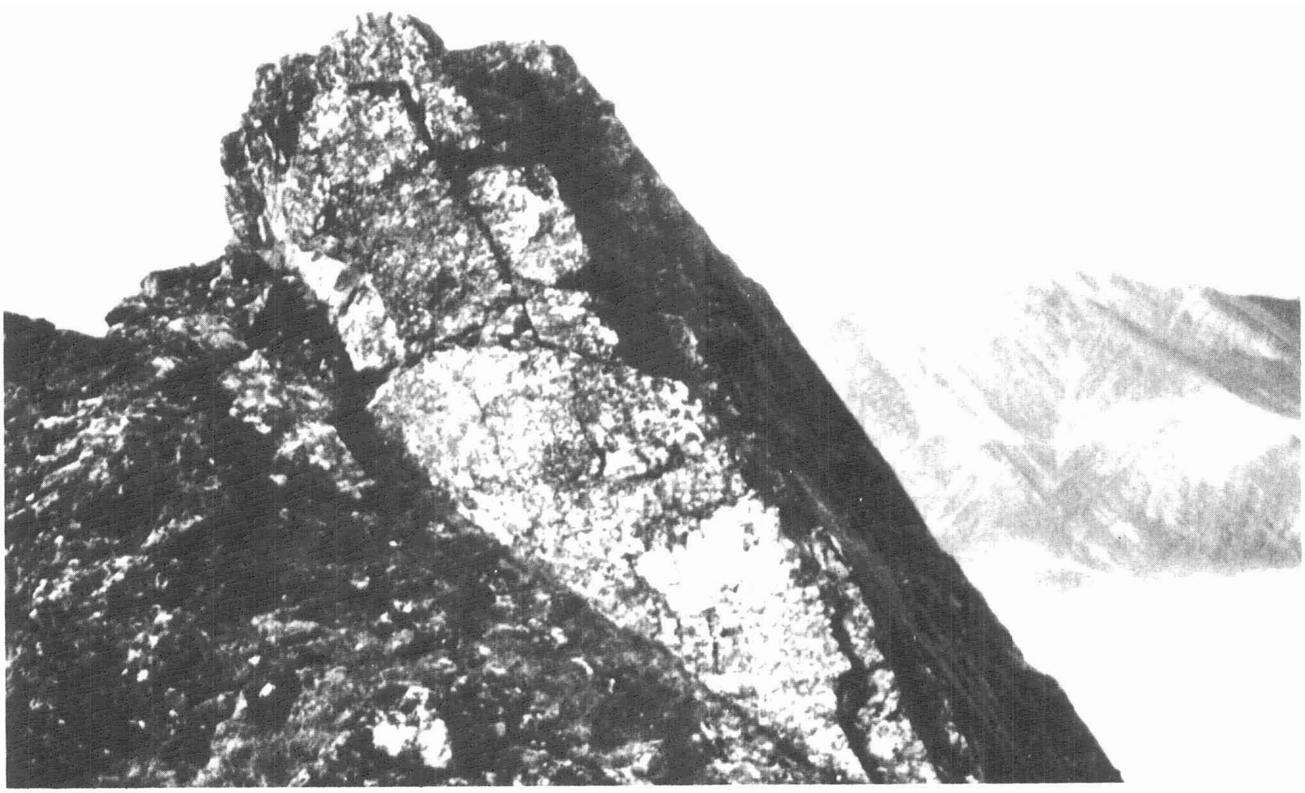


Figure 181. Typical fining-upward cycle in middle part of Ear Peak Member of Kanayut Conglomerate (2,300 ft [710 m] above base of section), west side of Atigun River. Thich conglomerate beds in lower part of cycle (at center) cut into poorly exposed shale (at left). Beds shown are about 19 ft (6 m) thick.

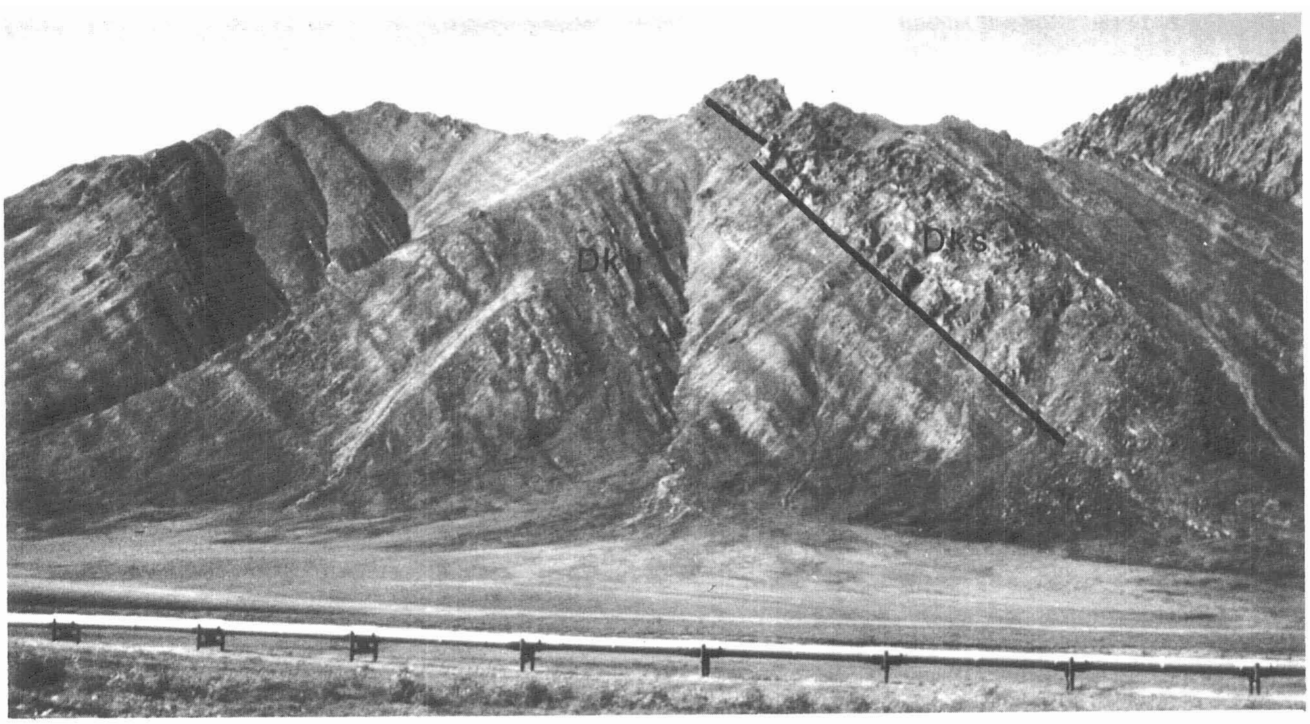


Figure 182. View of Ear Peak (Dke) and Shainin Lake (Dks) Members of Kanayut Conglomerate on west side of Atigun River valley. Note sharp contact between two members, marked by appearance of massive, resistant conglomerate and sandstone beds.

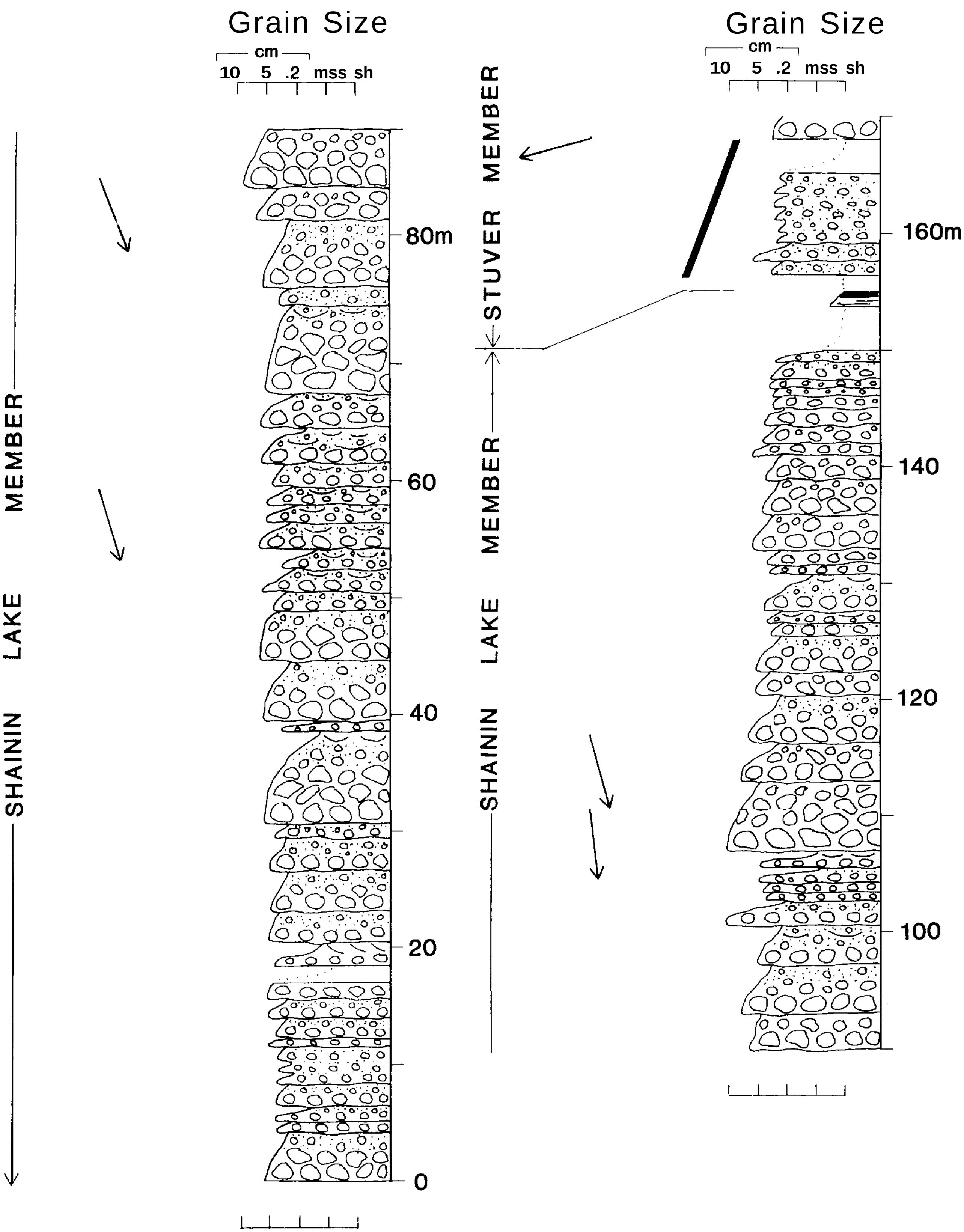


Figure 183. Columnar section of Shainin Lake Member of Kanayut Conglomerate measured on east side of Atigun River. See figure 179 for location of section and figure 180 for explanation of symbols.

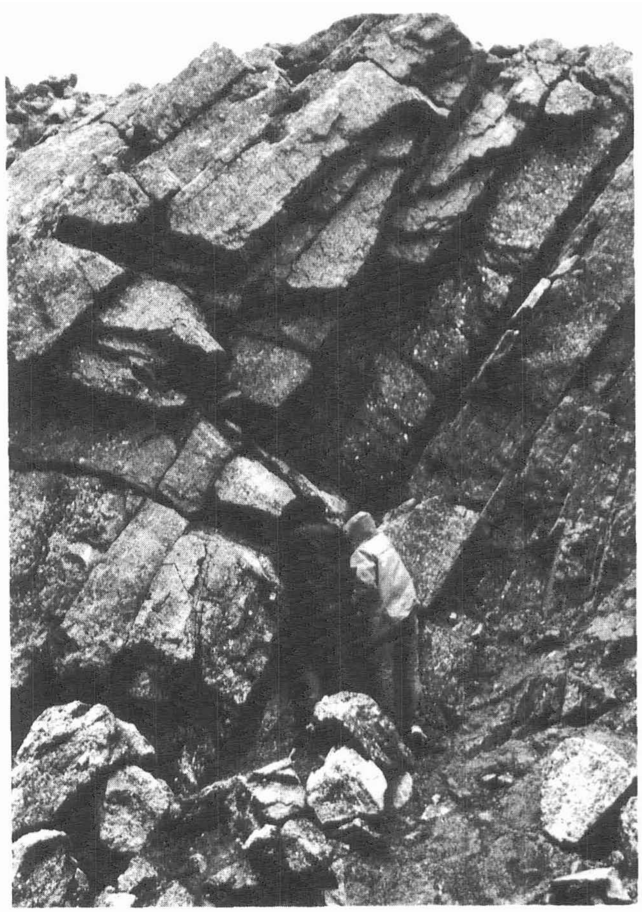


Figure 184. Geologists examine massive conglomerate beds in Shainin Lake Member of Kanayut Conglomerate in quarry at Mile 260.8 on Dalton Highway. Conglomerate beds fine upward and are locally channeled into underlying beds. Photograph by C.G. Mull, August 1987.

STUVER MEMBER

A thick section (4,295 ft; 1,309 m) of the Stuver Member of the Kanayut Conglomerate (fig. 185) was measured along the crest of a north-trending ridge (fig. 186) about 3.7 mi (6 km) east of Atigun River valley (secs. 3, 10, and 11, T. 13 S., R. 13 E.), Philip Smith Mountains Quadrangle. Less than one percent of this section is covered, and contacts with the underlying Shainin Lake Member and overlying Kayak Shale are well exposed.

The contact with the Shainin Lake Member is marked by the disappearance of massive conglomerate beds and the first appearance of distinct fining-upward cycles that contain 3- to 6-ft-thick (1 to 2 m) shale intervals. At the base of the cycles are beds of conglomerate or conglomeratic sandstone that have a maximum clast size of 2.75 in. (7 cm), as compared with a maximum clast size of 3.3 in. (8.5 cm) in the uppermost part of the Shainin Lake Member. These generally massive beds have an erosional contact with underlying strata and may contain rip-up clasts as large as 3.3 ft (1 m). The coarse-grained beds grade upward into trough cross-stratified and current-ripple-marked sandstone that is overlain by extensive sequences of interbedded shale, siltstone, and very fine grained sandstone as thick as 180 ft (55 m).

The fining-upward cycles and the overall fining-upward pattern of the Stuver Member indicate meandering-stream deposits and gradual retreat of the Kanayut delta. The conglomerate and conglomeratic sandstone at the base of the cycles represent channel-lag deposits; the overlying trough cross-stratified and ripple-marked sandstone represents channel-fill deposits; and the thick sequences of shale, siltstone, and fine-grained sandstone in the upper part of the cycles represent flood-plain, crevasse-splay, and levee deposits.

The fine-grained flood-plain deposits form distinct units within the measured section. Within an overall fining-upward cycle, flood-plain deposits contain several types of coarsening-upward sequences that average 3 ft (1 m) thick. Some of those fining-upward cycles consist of bundles of fine- to coarse-grained sandstone, which appear to be crevasse-splay deposits (fig. 187); whereas others consist mostly of siltstone, very fine grained sandstone, and paleosols, which appear to be levee deposits. Sandstone within the fine-grained deposits is rippled, laminated, massive, and typically laterally discontinuous. The flood-plain deposits are red, orange, green, brown, or black, probably depending on the degree of oxidation of the sediments. Plant fragments, root traces, and some rip-up clasts are present in these deposits, and mudcracks are developed in many horizons.

Near the base of the Kayak Shale, the Stuver Member contains evidence of alternating marine and fluvial environments. Couplets of gray, very fine to medium-grained sandstone, capped by oscillation ripple marks, probably represent limited marine wave activity. The contact with the overlying shallow-marine basal sandstone member of the Kayak Shale is marked by the appearance of thoroughly bioturbated intervals and abundant oscillation ripple marks. The presence of distinct fining-upward cycles, interpreted to be fluvial in origin, permits subdivision of the 210-ft-thick (64 m) basal sandstone member into three separate fluvial and marine cycles (not shown on fig. 185). The overlying black shale is platy and contains a few thin sandstone interbeds that may be turbidites. Within the lowermost bed of the overlying limestone are crinoidal debris, corals, and brachiopods.

Conglomerate clasts from a sandstone bed at 4,200 ft (1,280 m) above the base of the section, about 98 ft (30 m) below the base of the Kayak Shale, consist of 82 percent chert, 17 percent vein quartz, and 1 percent quartzite. Conglomerate clasts in the basal sandstone member of the Kayak Shale consist of 83 percent chert, 15 percent vein quartz, and 2 percent quartzite in one area and 83 percent chert and 17 percent vein quartz in a second area. Clasts range from 2.75 in. (7 cm) at the base of the Stuver Member to 0.4 in. (1 cm) near the contact with the basal sandstone member of the Kayak Shale.

Forty-six paleocurrent measurements from the Stuver Member at this location have a vector mean and standard deviation of $203^{\circ} \pm 45^{\circ}$. One measurement of clast imbrication and long-axis orientation taken from the Kayak Shale yielded a sediment-transport direction of 170°.

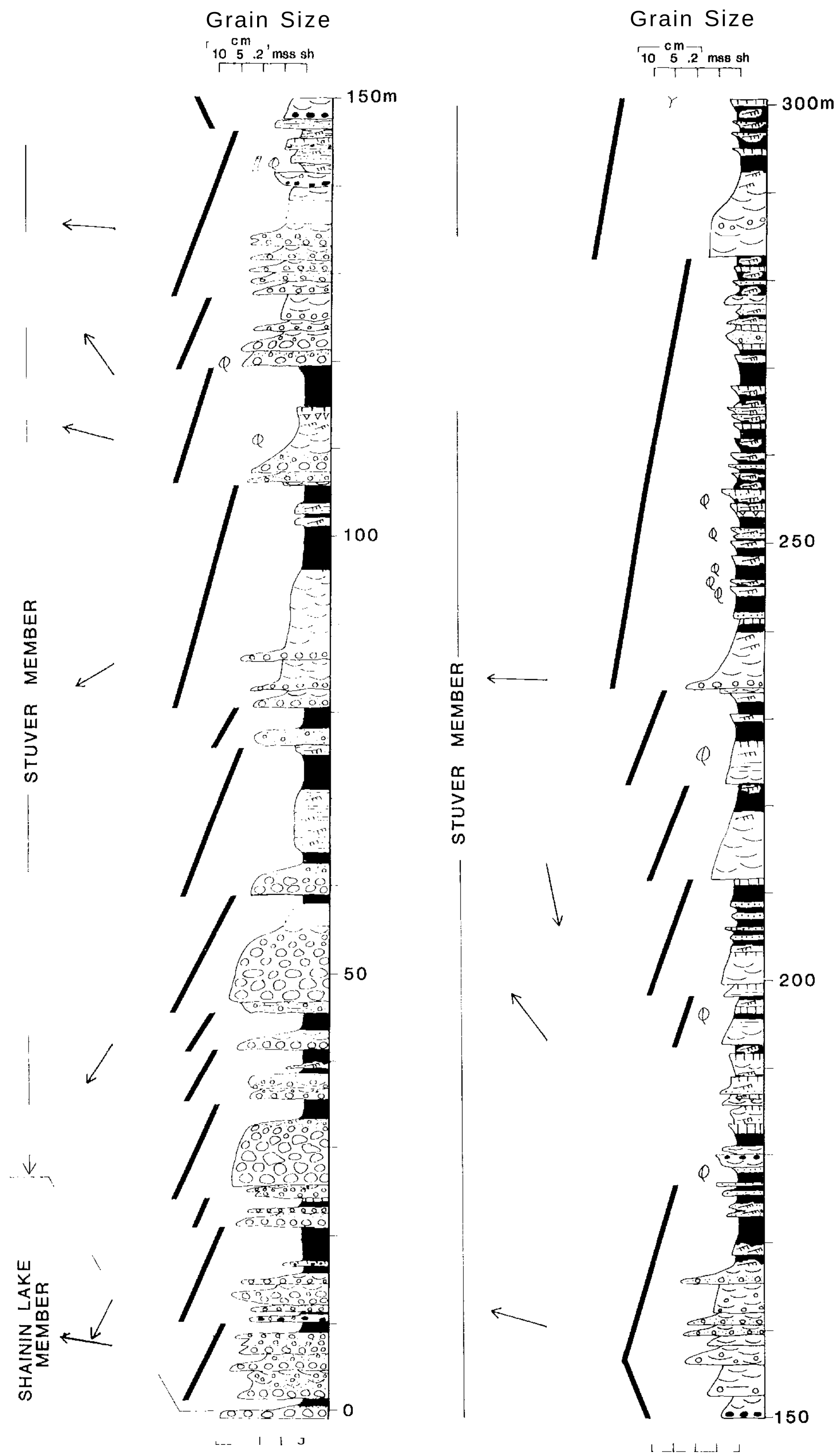


Figure 185. Columnar section of Stuver Member of Kanayut Conglomerate and overlying Kayak Shale measured on east side of Atigun River. See figure 179 for location of section and figure 180 for explanation of symbols.

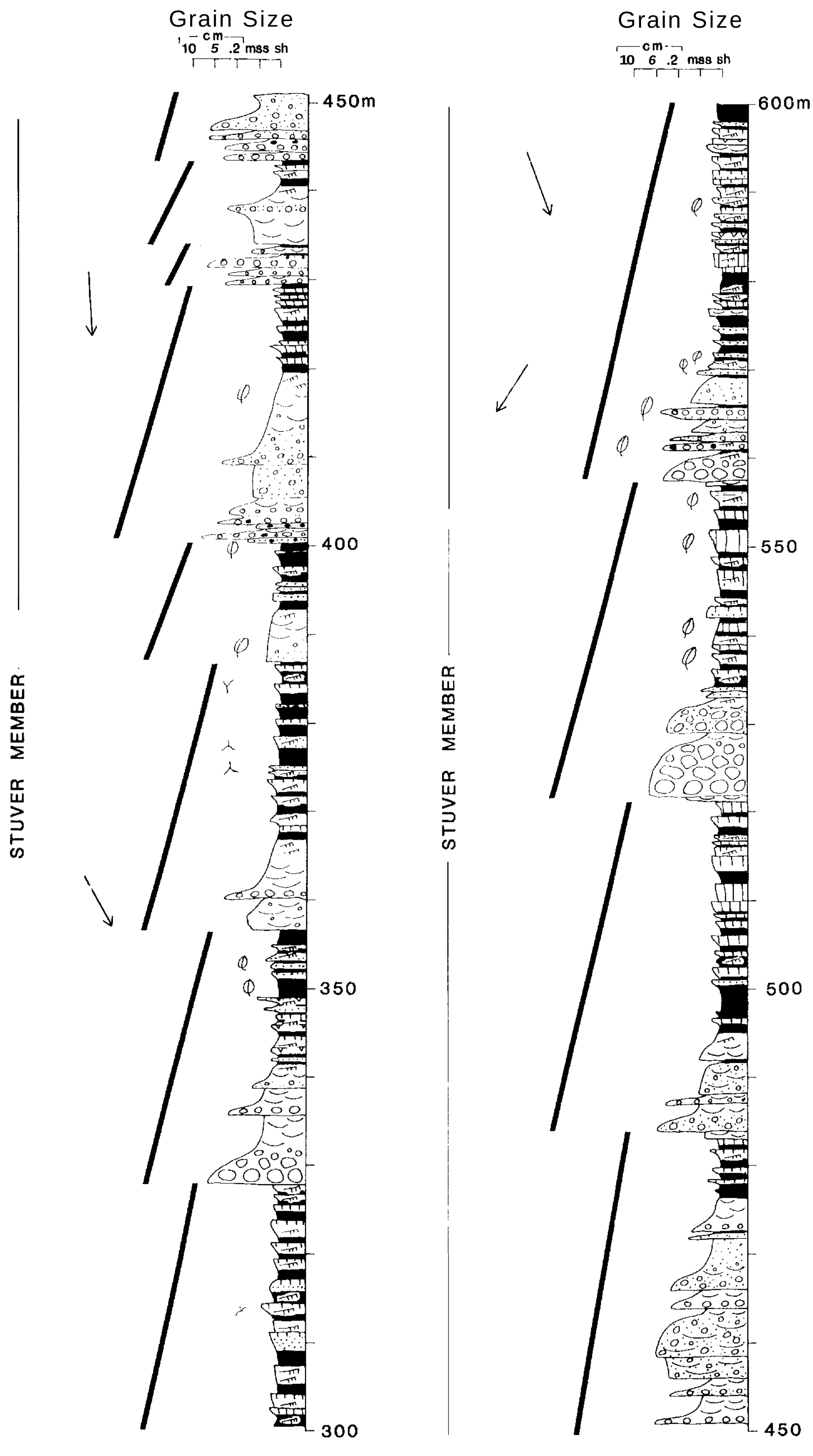


Figure 185. Continued.

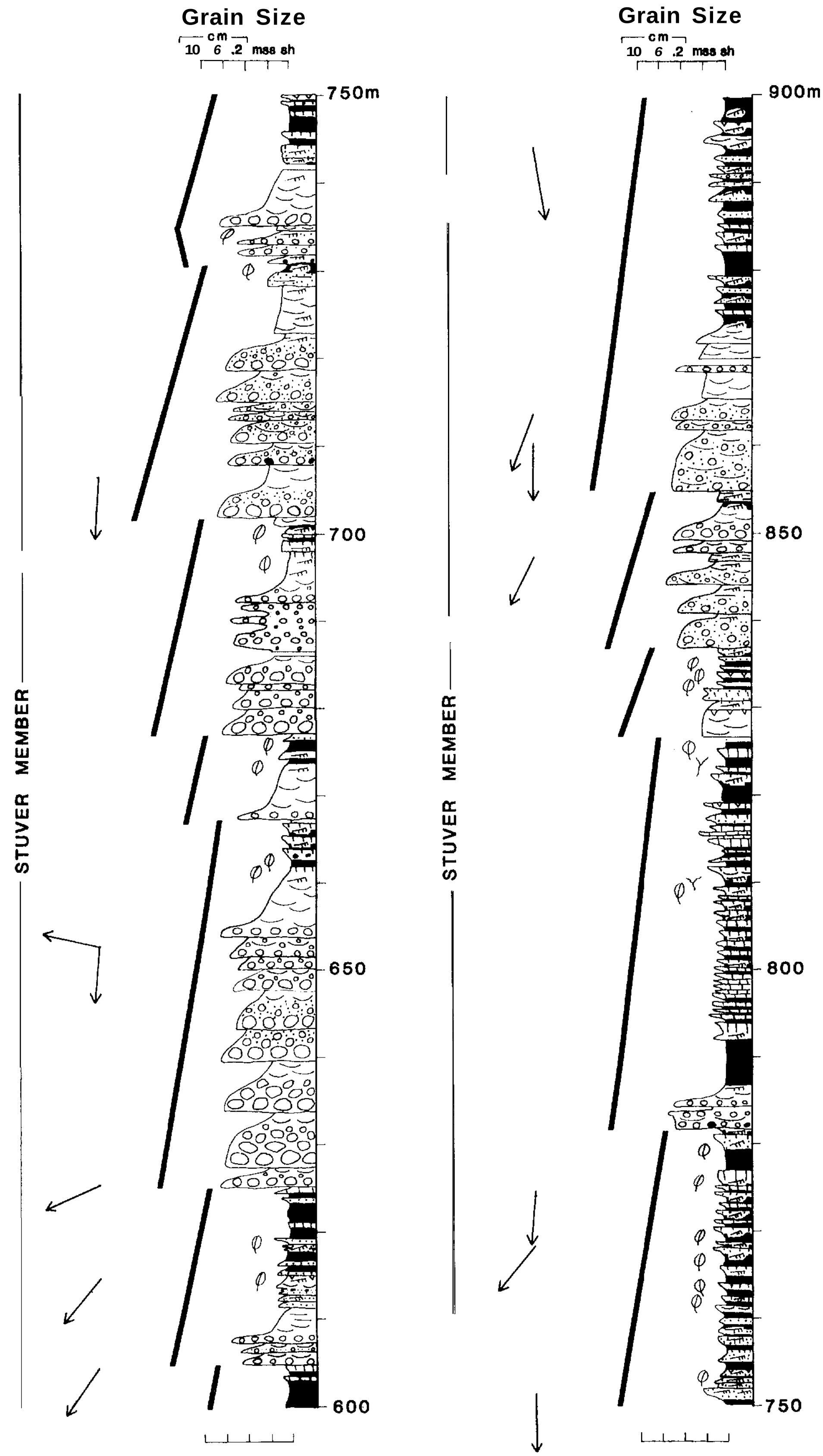


Figure 185. Continued.

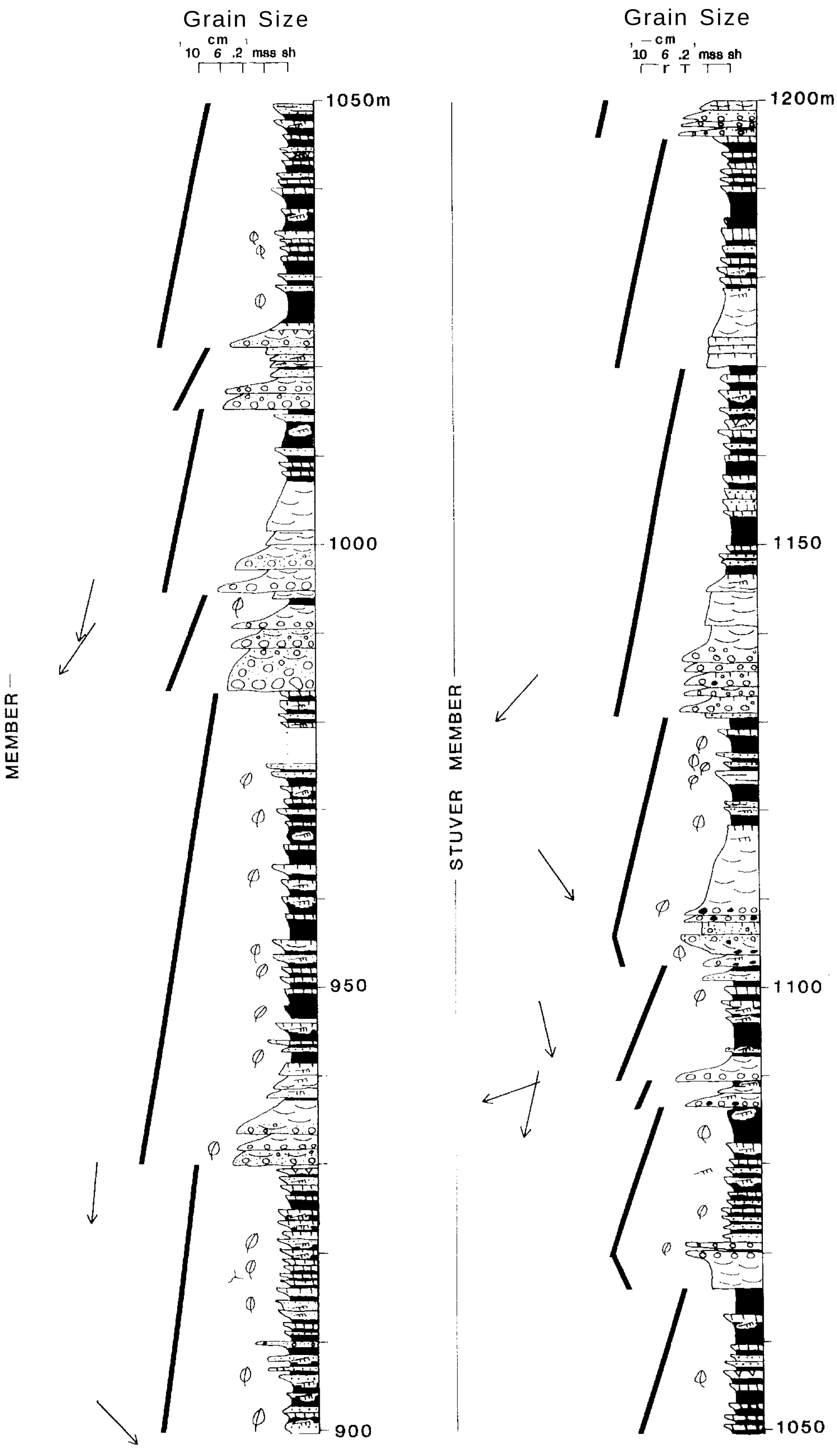


Figure 185. **Continued.**

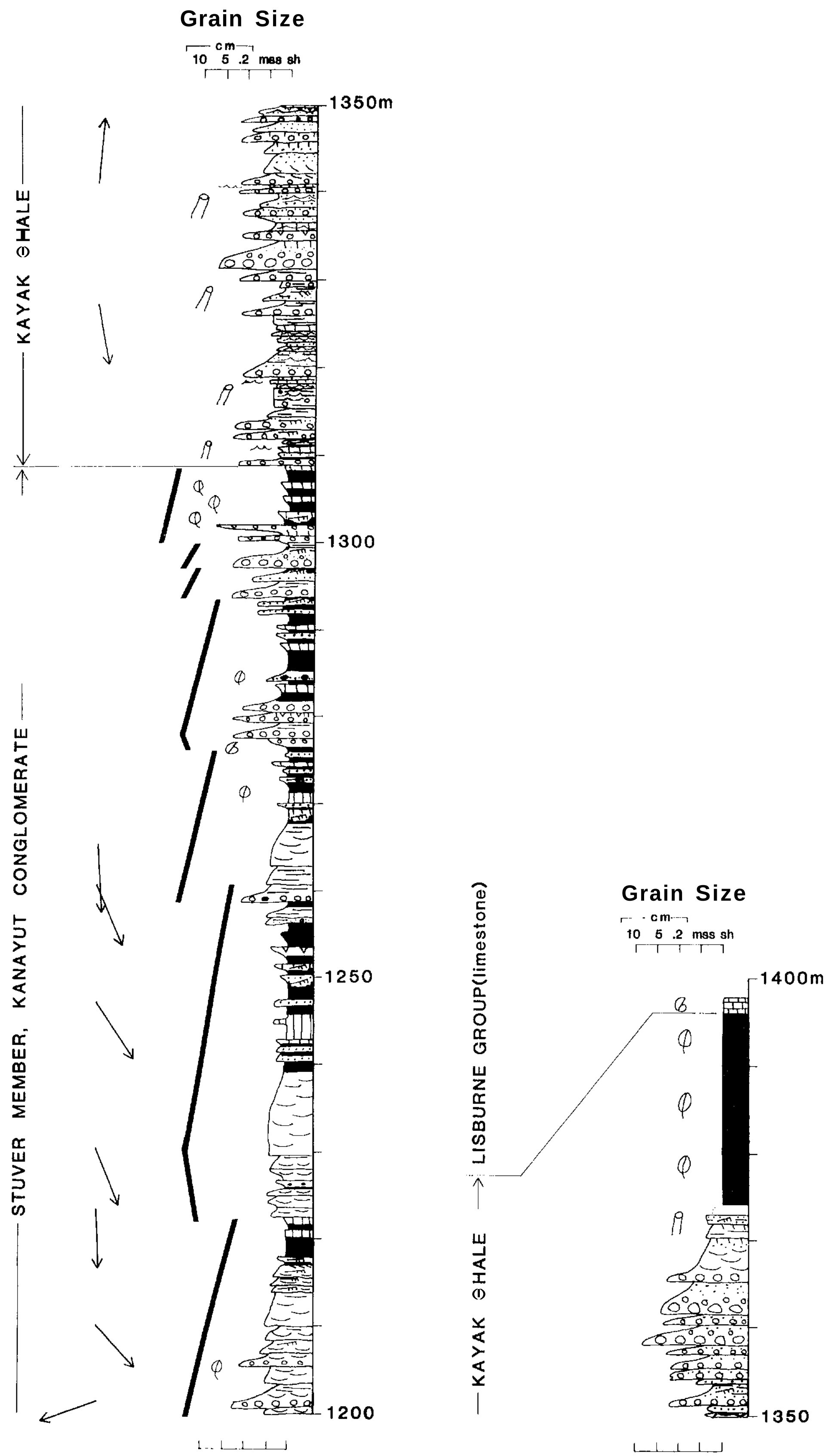


Figure 185. Continued.

Figure 186. View southward along crest of ridge on east side of Atigun River showing middle part of Stuver Member of Kanayut Conglomerate. Resistant, coarse-grained fining-upward cycles are interbedded with less resistant shale and siltstone.

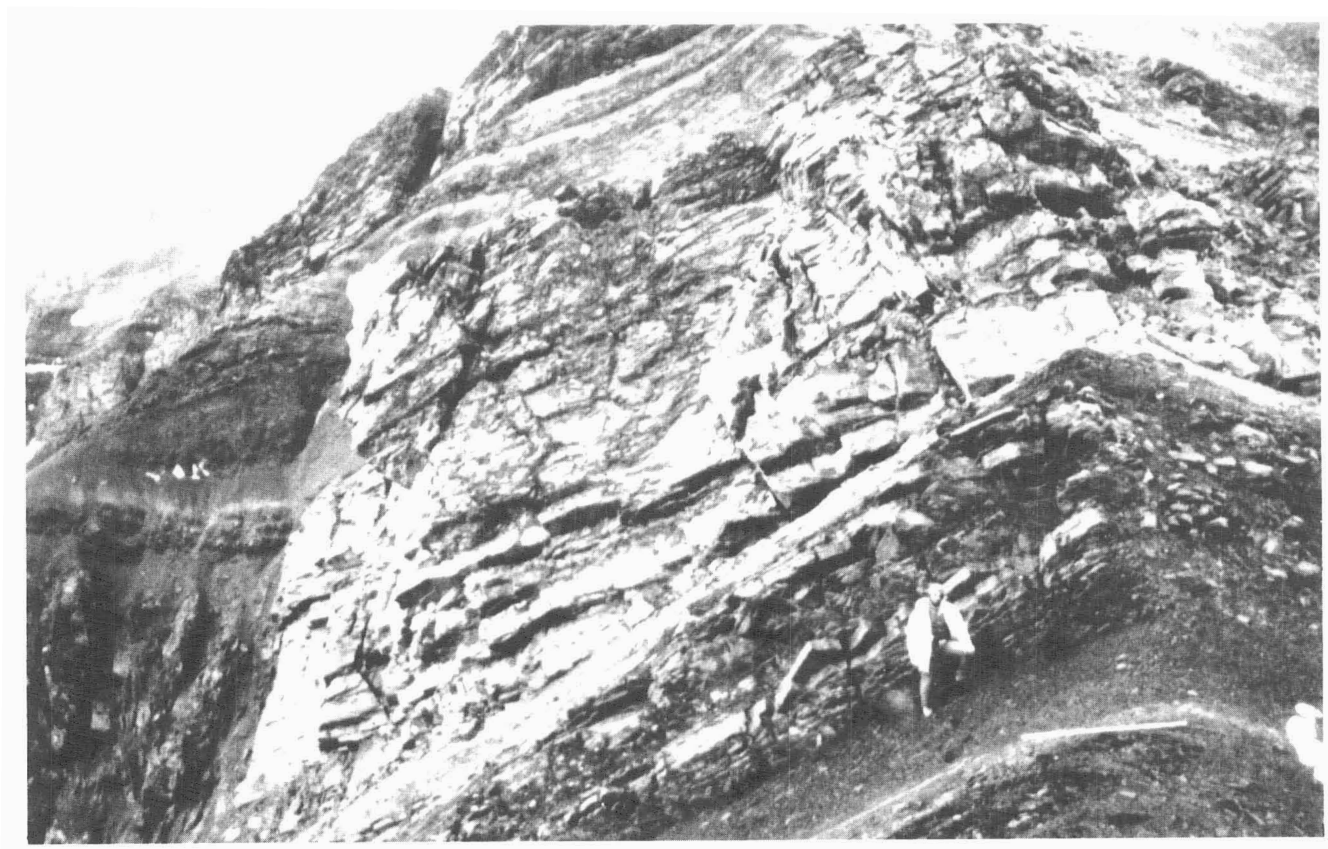


Figure 187 Thin, coarsening-upward cycles (seen most clearly behind geologist at lower right) in fine-grained flood plain deposits of Stuver Member of Kanayut Conglomerate on east side of Atigun River

JOHN RIVER AREA

An almost complete section of gently dipping Kanayut Conglomerate was measured along a ridge that extends eastward from Ekokpuk Mountain to the floor of the John River valley (SW¼, sec. 32, T. 37 N., R. 20 W.), Wiseman Quadrangle (fig. 188). Because the base of the section is covered, as much as 460 ft (140 m) of the Ear Peak Member may be present in the interval above the stratigraphically highest exposure of Hunt Fork Shale. The upper and lower parts of the section contain abundant shaly intervals; several of these are expressed as covered intervals. A total thickness of 2,287 ft (697 m) was measured at this location. We tentatively assign the lower 119 m to the Ear Peak Member, the overlying 1,322 ft (403 m) to the Shainin Lake Member, the next 390 ft (175 m) to the Stuver Member, and the upper 33 ft (10 m) to the basal sandstone member of the Kayak Shale.

The Ear Peak Member consists of 11 fining- and thinning-upward cycles interpreted to be lateral and vertical accretion deposits of meandering rivers. The lower two cycles are about 52 ft (16 m) thick and the upper cycles are about 26 ft (8 m) thick. Trough cross-stratified conglomerate with a maximum clast size of 0.6 in. (1.5 cm) and fine- to medium-grained sandstone are present in the lower part of the cycles and are inferred to be channel and lower point-bar deposits (fig. 189). The coarser grained lower part of the cycles include multiple fining-upward sequences that are separated by erosional surfaces. The coarser grained facies typically passes upward into trough cross-stratified, and locally parallel stratified, fine-grained sandstone, which, in turn, is overlain by ripple-marked and laminated very fine grained sandstone, siltstone, and shale that contain plant fossils. These fine-grained rocks are interpreted to be upper point-bar, levee, and flood-plain deposits.

The transition from the Ear Peak Member to the Shainin Lake Member is marked by the disappearance of shale and siltstone. In contrast to the section measured at Galbraith Lake, the Shainin Lake Member here is less conglomeratic (table 5) and is marked by thick sequences of sandstone with little intervening shale. Conglomerate is most common in the lower 394 ft (120 m) of the Shainin Lake Member. Maximum clast size is 2 in. (5 cm) and maximum thickness of the conglomerate beds is 11.5 ft (3.5 m).

A second conglomerate-rich interval from 1,509 to 1,608 ft (460 to 490 m) above the base of the section contains conglomerate clasts as large as 5.5 in. (14 cm) and conglomerate beds as thick as 11 ft (3.9 m). The conglomerate beds generally grade upward into coarse- to medium-grained, massive or cross-stratified sandstone that forms fining-upward conglomerate-sandstone couplets. These are less commonly overlain by a scour

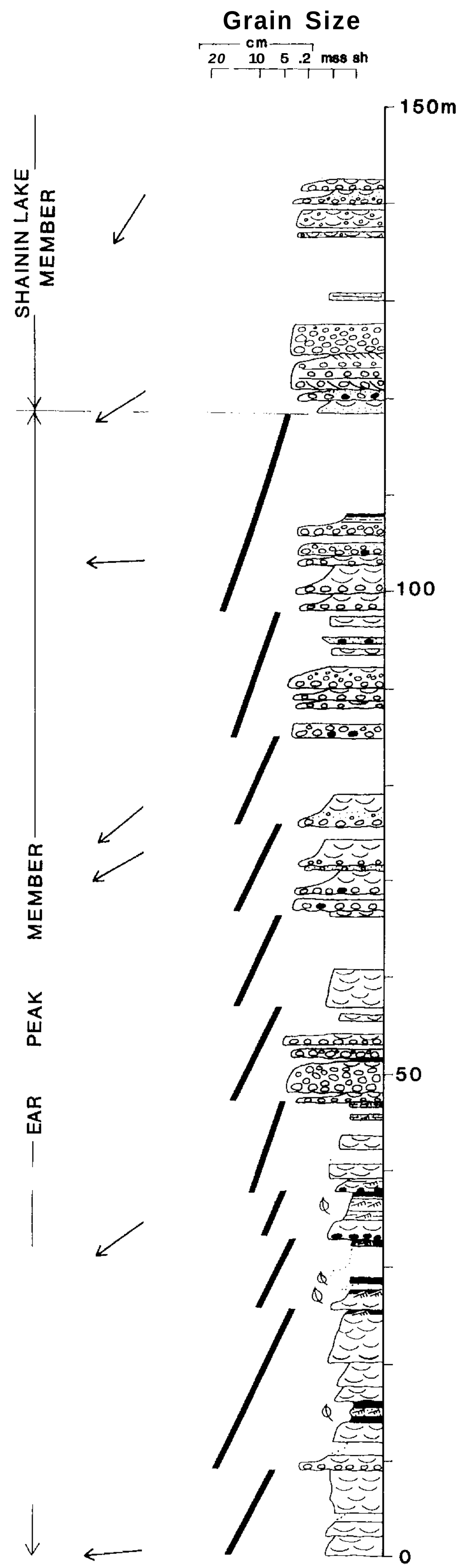
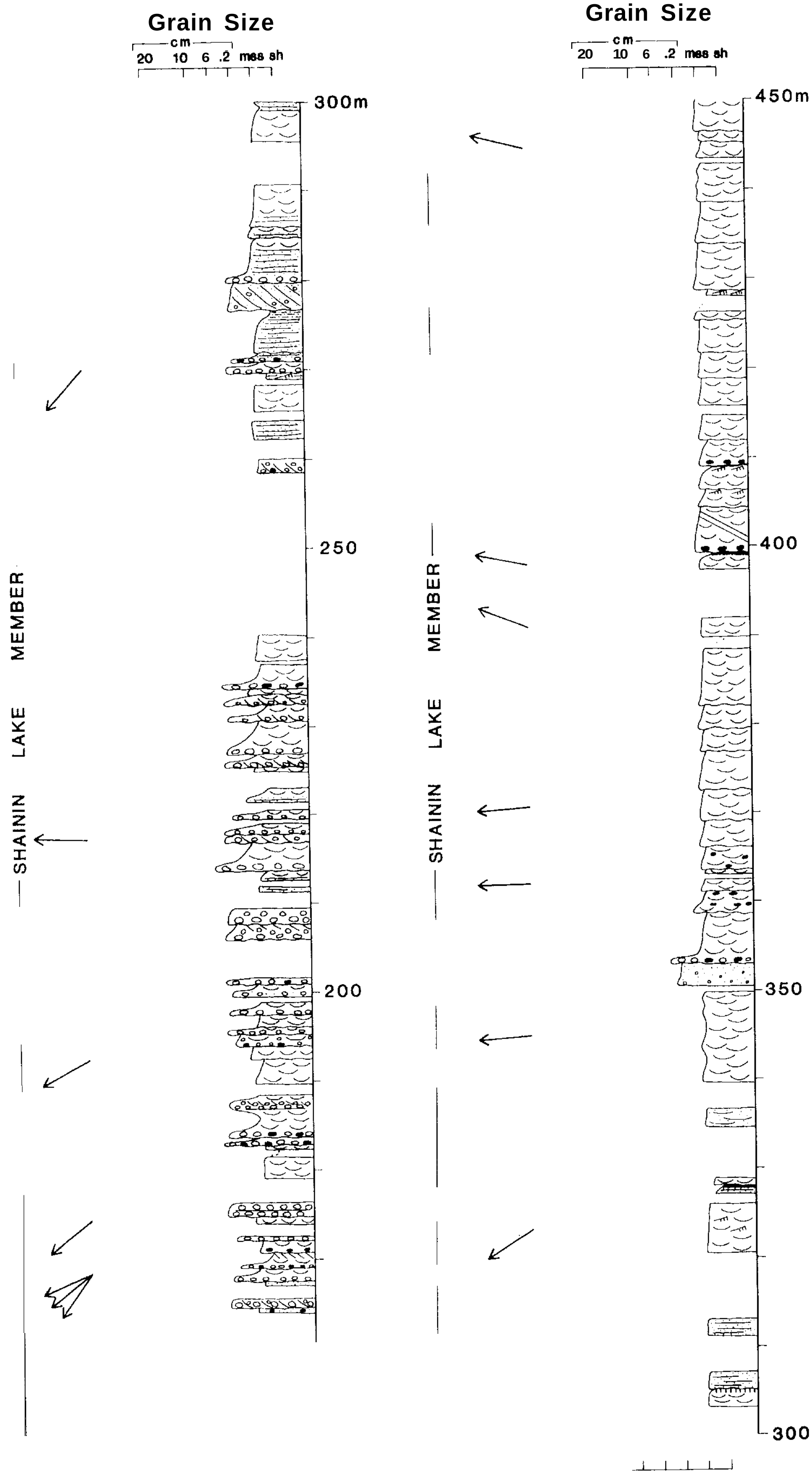


Figure 188. Columnar section of upper part of Ear Peak Member and overlying Shainin Lake and Stuver Members of Kanayut Conglomerate measured west of John River. See figure 171 for location of section and figure 180 for explanation of symbols.


Grain Size
cm
20 10 6 .2 mss sh
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SHAININ LAKE MEMBER
Grain Size
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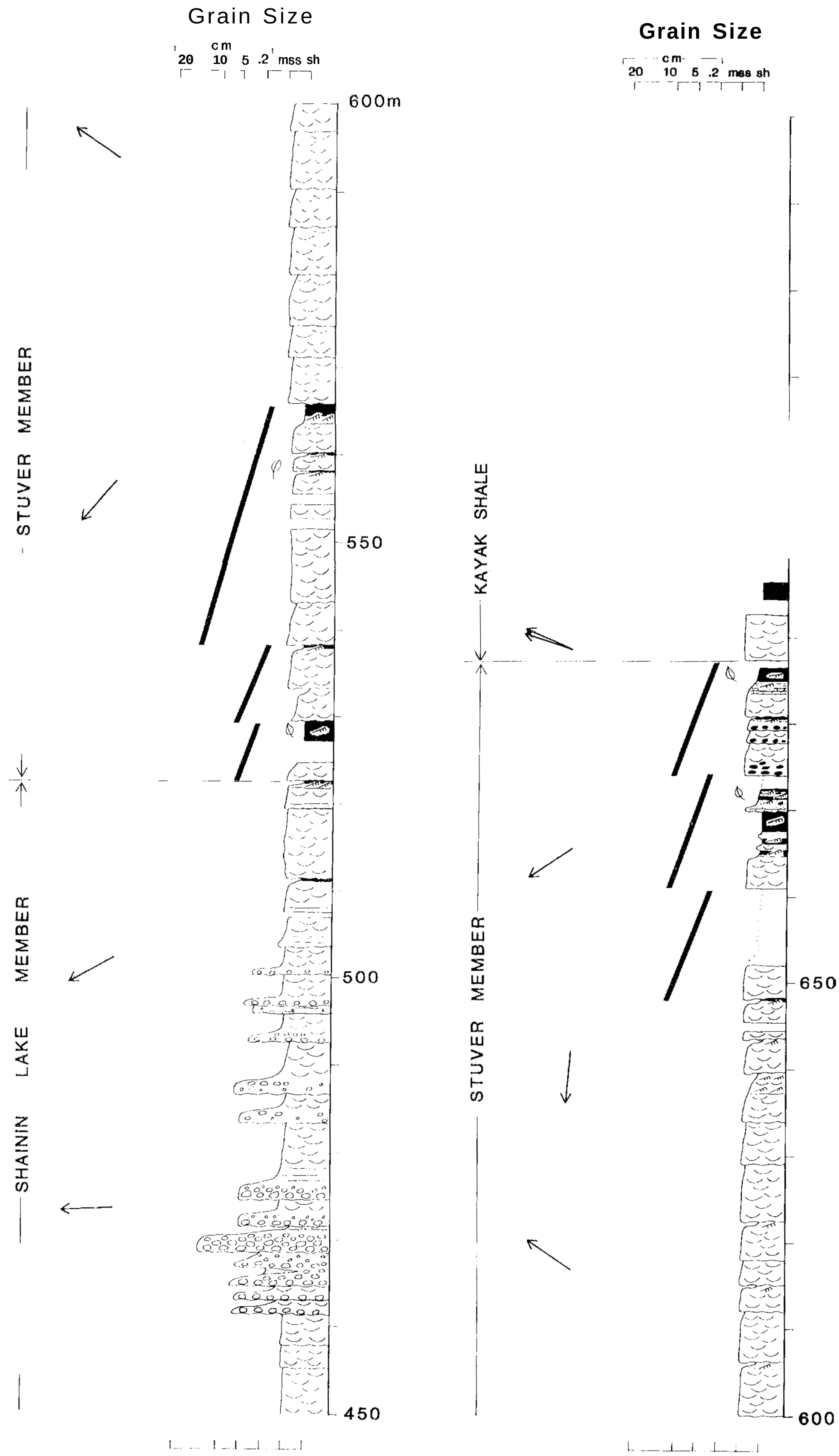

Figure 188. Continued.

Table 5. Sedimentologic characteristics of members of Kanayut Conglomerate near Galbraith Lake, John River, and Shainin Lake

Sedimentologic characteristics	Galbraith Lake area: Ear Peak Member	Galbraith Lake area: Shainin Lake Member	Galbraith Lake area: Stuver Member	John River area: Ear Peak Member	John River area: Shainin Lake Member	John River area: Stuver Member	Shainin Lake area: Ear Peak Member	Shainin Lake area: Shainin Lake Member	Shainin Lake area: Stuver Member
Thickness	1,160 m	155 m	1,309 m	119-259 m	403 m	175 m	512 m	526 m	217 m
Maximum-clast size	5 cm	10 cm	7 cm	5 cm	14 cm	0.1 cm	13 cm	22 cm	13 cm
Paleocurrents (vector mean and standard deviation; n = number of measurements)	171° ± 38° n = 17	188° ± 43° n = 17	203° ± 45° n = 46	239° ± 18° n = 15	271° ± 15° n = 8	265° ± 48° n = 7	243° ± 29° n = 44	249° ± 25° n = 15	269° ± 25° n = 5
Percent shale and siltstone	29.6%	3.7%	34.5%	29%	4.8%	14%	26%	0.23%	35%
Percent conglomerate	19.1%	91%	11%	14%	8.8%	0%	22%	81%	15%
Maximum thickness of conglomerate strata	5.5 m	7.3 m	6.75 m	3 m	3 m	0	12.6 m	12.5 m	1.5 m
Meandering-stream cycles									
Number of cycles	57	0	58	11	0	6	33	0	20
Average thickness	20 m	--	23 m	11 m	--	13 m	15 m	--	10 m
Maximum thickness	50 m	--	75 m	20 m	--	27 m	32 m	--	24 m
Average thickness, channel and point-bar deposits	12 m	--	11 m	7.4 m	--	8.2 m	10 m	--	5.7 m
Maximum thickness, channel and point-bar deposits	31 m	--	36 m	9 m	--	25 m	26 m	--	13 m
Average thickness, flood-plain deposits	7.7 m	--	12 m	3.6 m	--	4.8 m	5 m	--	4.7 m
Maximum thickness, flood-plain deposits	43 m	--	56 m	11 m	--	9 m	18 m	--	15 m
Typical basal strata	Pebble conglomerate	--	Pebble conglomerate	Mediumgrained sandstone to pebble conglomerate	--	Medium-grained sandstone	Pebble conglomerate	--	Pebble conglomerate

-- Nonapplicable.

Figure 189. *Fining-upward* cycle, dominated by trough cross-stratification, in Ear Peak **Member** of Kanayut Conglomerate west of John *River*.

surface and another conglomerate bed or trough cross-stratified, fine- to medium-grained sandstone.

Nonconglomeratic intervals in the Shainin Lake Member consist of beds of trough cross-stratified, fine- to medium-grained sandstone as thick as 33 ft (10 m) that thin and fine upward and are separated by erosional surfaces (fig. 190). The sandstone is typically gray, but several red-brown oxidized horizons as thick as 3 ft (1 m) are present. These oxidized horizons are typically parallel or subparallel to bedding and are interpreted to be incipient paleosols. They are most common about 980 ft (300 m) above the base of the section. The Shainin Lake Member was probably deposited by braided streams and contains features such as sand waves, channels, and longitudinal and transverse bars.

The transition between the Shainin Lake Member and Stuver Member is gradational but is marked by the disappearance of conglomerate and reappearance of shale and siltstone. Three fining-upward cycles, consisting of trough cross-stratified sandstone that grades successively into ripple-marked and laminated very fine grained sandstone, siltstone, and shale, are present in the lowest part of the Stuver Member (fig. 191). These cycles are overlain by 148 ft (45 m) of trough cross-stratified sandstone that lacks shale breaks and is similar to sandstone of the Shainin Lake Member. This interval is in turn overlain by three more fining-upward cycles. These cycles range from 16 to 75 ft (5 m to 23 m) thick and do not appear to be organized in a systematic manner. Flood-plain and levee deposits within the cycles are as thick as 16 ft (5 m) and are best developed in the upper three cycles of the Stuver Member. The flood-plain deposits in the cycle from 2,169 to 2,208 ft (661 to 673 m) contain some massive to parallel-stratified beds of fine-grained sandstone that are as thick as several feet and contain abundant rip-up clasts. These units are interpreted to be crevasse-splay deposits. The Stuver Member is overlain abruptly by the fine-grained basal sandstone member of the marine Kayak Shale.

Thirty paieocurrent measurements of 28 trough cross-strata and two tabular cross-strata indicate that sediment transport was toward the southwest. The mean and standard deviation of 15 measurements from the Ear Peak Member are $239^{\circ} \pm 18^{\circ}$, whereas the mean and standard deviation of eight and seven measurements from the Shainin Lake and Stuver Members, respectively, are more westerly--$271^{\circ} \pm 15^{\circ}$ and $265^{\circ} \pm 48^{\circ}$. The mean and standard deviation of all 30 measurements are $253^{\circ} \pm 30^{\circ}$.

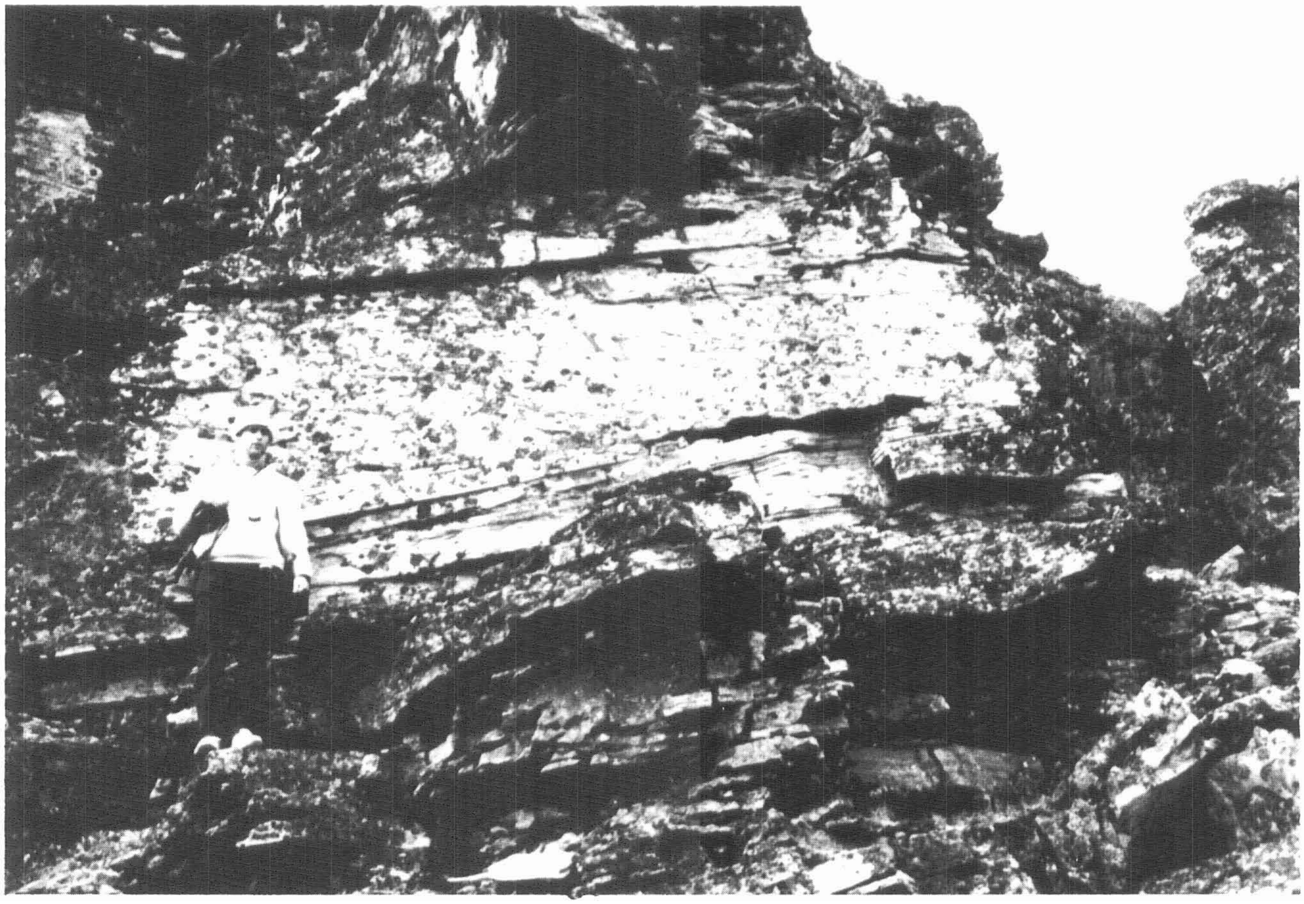

Figure 190. Cross-stratified sandstone beds in Shainin Lake Member of Kanayut Conglomerate west of John River.

SUMMARY AND CONCLUSIONS

The stratigraphy, facies, maximum clast size, and paleocurrent data from the Kanayut Conglomerate indicate that the formation represents a fluvial depositional system that prograded southwestward over shallow-marine sandstone and underlying fine-grained prodelta deposits. At the maximum extent of progradation, much of this deltaic complex consisted of a vast braid plain on which conglomerate and sandstone of the Shainin Lake Member were deposited as bed load in relatively small, rapidly shifting braided channels. Meandering streams continued to transport abundant coarse-grained bed load beyond the margins of the braid plain, where an extensive flood plain developed. The flood plain was traversed by sizable meandering streams that branched into smaller distributary streams basinward. Around the margins of the delta, the shallow-marine Noatak Sandstone was deposited as channel-mouth and offshore bars.

Near the Trans-Alaska Pipeline, the Kanayut Conglomerate is exposed in two major, internally imbricated thrust sheets. The northern thrust sheet exposes the thickest (8,609 ft; 2,624 m) known section of Kanayut Conglomerate that can be easily divided into the Ear Peak, Shainin Lake, and Stuver Members of the type area near Shainin Lake. The southern thrust sheet exposes a thinner section of about 2,300 ft (700 m) in which the three members of the Kanayut Conglomerate are difficult to distinguish without detailed study.

Sedimentologic data were collected from measured sections near Galbraith Lake and the John River that are representative of the northern and southern thrust sequences, respectively. In table 5, these data are summarized and compared with data from the type sections near Shainin Lake (Nilsen and Moore, 1984a), which are part of the northern thrust sequence. The data indicate that the Kanayut Conglomerate near Galbraith Lake consists of a relatively thin Shainin Lake Member (braid-plain deposits) and very thick Ear Peak Member and Stuver Member (meandering-stream and flood-plain deposits). The Ear Peak and Stuver Members contain thicker and more numerous fining-upward meandering-stream cycles and flood-plain deposits.

In contrast, the section in the John River area contains much thinner Ear Peak and Stuver Members and, relative to its total thickness, a much thicker Shainin Lake Member. This section contains fewer and thinner fining-upward fluvial cycles and flood-plain deposits; it also contains fewer conglomerate beds and a

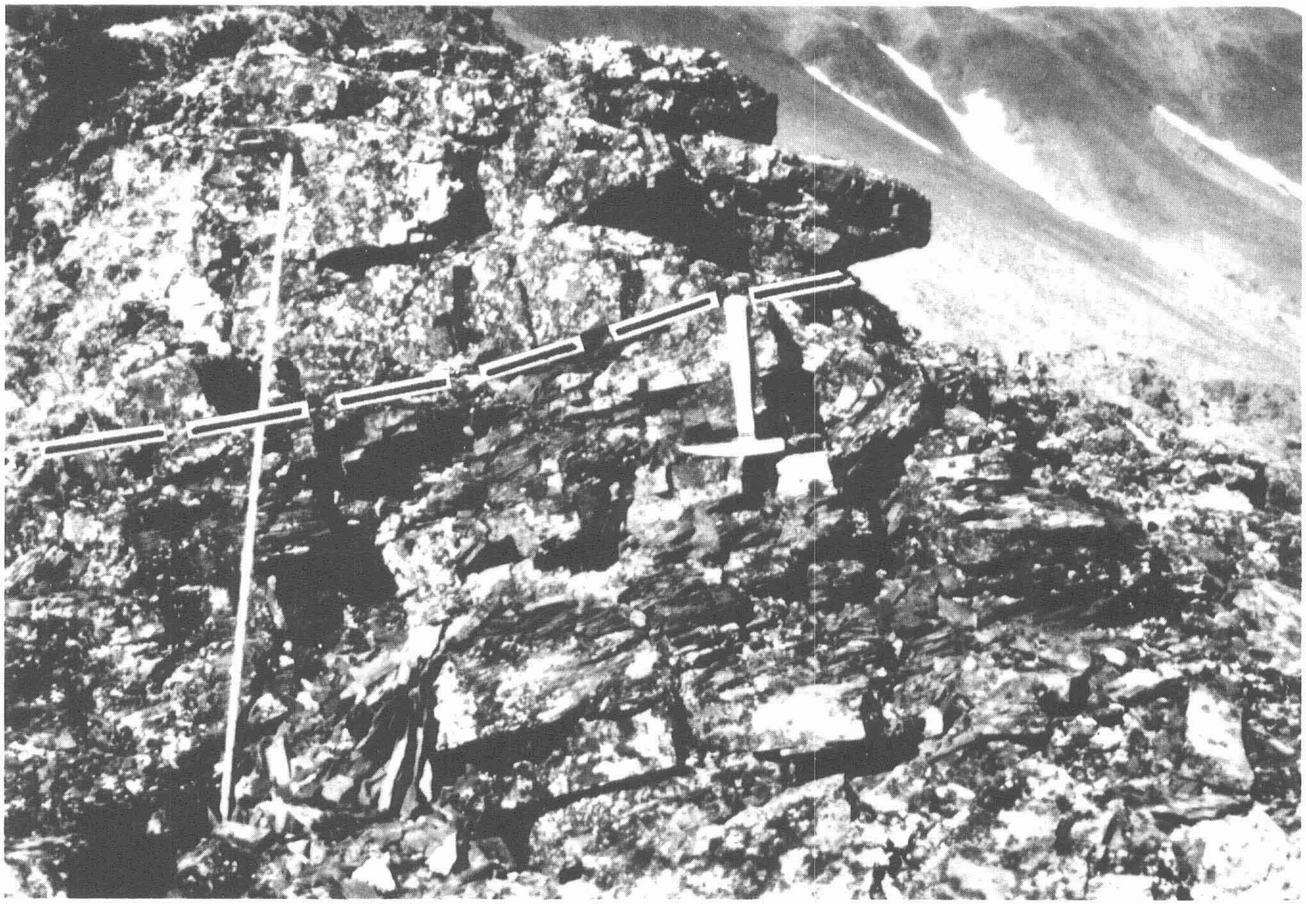

Figure 191. Fining-upward cycles in lower part of Stuver Member of Kanayut Conglomerate west of John River. Dashed line marks erosional base between cycles.

much smaller percentage of conglomeratic strata than the Atigun valley and Shainin Lake sections to the northeast. Although the section in the John River area contains larger clasts than the section at Atigun, these are present in a single conglomerate bed; all the other beds contain clasts <2 in. (6 cm) diam. The thickness of the coarse-grained parts of the fining-upward cycles (generally a measure of the depth of the streams) and the coarseness of their basal strata (an indication of the carrying capacity of the streams) are thinner and finer, respectively, in the John River section.

ACKNOWLEDGMENTS

We thank Samuel Johnson and Donna Balin (U.S. Geological Survey) for assistance in measuring the sections in the Atigun valley and John River area; Wesley K. Wallace (ARCO Alaska, Inc.) for helpful discussions; Robert Detterman and William Patton (U.S. Geological Survey) for helpful reviews of the manuscript; and Cy Asta (ERA Helicopters) for skillful and safe piloting of our helicopter.

CHAPTER 16.

STRATIGRAPHY OF THE LISBURNE GROUP IN THE CENTRAL BROOKS RANGE

By A.K. Armstrong[1] and B.L. Mamet[2]

INTRODUCTION

Carbonate rocks of the Lisburne Group crop out throughout the Brooks Range from the Alaska-Yukon border on the east to Cape Lisburne on the west and in many areas form spectacular cliffs (fig. 192). Regional stratigraphic studies of the Lisburne Group were conducted by geologists of Shell Oil Company as early as the late 1950s; our studies took place from 1962 to 1974 at a number of localities in the central and northeastern Brooks Range (fig. 193). During these helicopter-supported field projects, we measured stratigraphic sections, and collected petrographic and foraminifera samples at 5- to 20-ft (1.5 to 6 m) intervals; rugose corals were also collected. Major results of our work in the central Brooks Range were published in 1977 and 1978.

PREVIOUS WORK

Schrader (1902) named the Lisburne Formation for limestone exposures near Cape Lisburne at the western end of the Brooks Range. In the central Brooks Range, Bowsher and Dutro (1957) recognized two new formations within the Lisburne and subsequently raised the unit to group rank. The lower formation, the Wachsmuth Limestone, consists of four informally designated members and overlies the Kayak Shale. The Alapah Limestone consists of nine informally named members and overlies the Wachsmuth Limestone. At its type locality, the top of the Alapah Limestone is overlain by Quaternary deposits. During a study of the Paleozoic sequence of the eastern Brooks Range, Brosgé and others (1962) recognized three formations within the Lisburne Group: the Wachsmuth Limestone, the Alapah Limestone, and the overlying Wahoo Limestone.[3]

On the basis of several newly recognized species of Mississippian cephalopods in the Lisburne Group, Gordon (1957) established an arctic zonation. Gastropods in the Lisburne Group were described by Yochelson and Dutro (1960), who also presented regional biostratigraphic relations based on gastropods and other groups of megafossils.

Armstrong (1970a) described the shallow-water to supratidal dolomite and chert in the Mount Bupto-Killik River region of the westcentral Brooks Range. Armstrong and others (1970) established foraminifer zones and outlined carbonate facies of the Lisburne Group in the central and eastern Brooks Range. Armstrong (1970b) described and illustrated lithostrotionoid corals from the Kogruk Formation of the De Long Mountains, western Brooks Range. Armstrong and Mamet (1970) published a series of lithofacies maps for Carboniferous carbonate rocks of Arctic Alaska. Armstrong and others (1971) presented a zonation based on microfossils and corals and described seacliff exposures of the Lisburne Group near Cape Lisburne. Armstrong (1972) described Mississippian corals in the Lisburne and, in 1973, described the paleoecology and biostratigraphy of the Pennsylvanian corals in the Wahoo Limestone. Mamet and Armstrong (1972) and Armstrong and Mamet (1975, 1977) presented detailed descriptions of the biostratigraphy and facies of the Lisburne Group in the Franklin and Romanzof Mountains, northeastern Brooks Range. Microfacies of the Lisburne Group in the Endicott Mountains of the central Brooks Range were

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[3] Regional studies since 1962 shows that the three Lisburne Group formations are very similar and cannot be readily differentiated on lithologic criteria. Standard usage of the name Wachsmuth Limestone now implies an Early and Late Mississippian (Osagean and Meramecian) age; Alapah Limestone, a Late Mississippian (Meramecian and Chesterian) age; and Wahoo Limestone, an Early Pennsylvanian (Morrowan or Atokan or both) age. Other units of the Lisburne Group in the central and western Brooks Range (Kogruk, Kuna, Tupik, and Utukok Formations) are more readily differentiated on lithologic criteria.—1511.

Figure 192. View eastward up Graylime Creek, 35 mi (50 km) west of Dalton Highway, central Endicott Mountains, where carbonate rocks of Lisburne Group form 2,000-ft-high (600 m) cliffs of Limestack Mountain. Photograph by C.G. Mull, July 1982.

described by Armstrong and Mamet (1978). Wood and Armstrong (1975) analyzed the diagenesis and stratigraphy of carbonate rocks of the Lisburne Group in the Sadlerochil Mountains and adjacent areas of the northeastern Brooks Range. Bird and Jordan (1977) described the facies, stratigraphy, and hydrocarbon-reservoir potential of the Lisburne Group in the subsurface of the Arctic Slope. In the central and western Brooks Range, Mull and others (1982) named the Kuna Formation for thin-bedded, black, carbonaceous shale, black chert, fine-grained limestone, and dolomite of Carboniferous age; the formation probably represents a deeper water, starved-basin, euxinic facies of the Lisburne Group. Dutro (1987a) reviewed the stratigraphic ranges of more than 320 megafossil species from about 40 measured sections throughout the Brooks Range and presented a revised megafossil biostratigraphic scheme that contains 20 zones.

ZONATION OF FORAMINIFERAL AND ALGAL MICROBIOTA IN THE LISBURNE GROUP

A Foraminifera1 and algal zonation established in Europe by Russian micropaleontologists has been extended to North America (Mamet, 1968a, b; Mamet and Mason, 1968; Sando and others, 1969; Mamet and Skipp, 1970). Because of favorable paleobiologic dispersion in Alaska, the zonation can be applied to the Lisburne Group. In this zonation, the Tournaisian and Viséan Stages are formally divided into four and eight zones, respectively, based on the presence of the families Endothyridae and Tournayellidae for the Tournaisian Stage and Endothyridae, Archaediscidae, Eostaffellidae, and Pseudoendothyridae for the Viséan Stage. Mamet presented the zonation in a detailed, systematic description and illustration of foraminifera and calcareous algae of the Lishurne Group (Armstrong and Mamet, 1977).

The study of foraminifera and algae in the carbonate succession of the Brooks Range is generally difficult because abundant bryozoan-echinoderm wackestone and packstone are unfavorable for preservation of protozoa (Armstrong and Mamet, 1977). Moreover, identification of foraminiferal ghosts is difficult because dolomitization and stress reorientation of calcite have commonly erased original textures. Despite these hindrances, foraminiferal assemblages have been

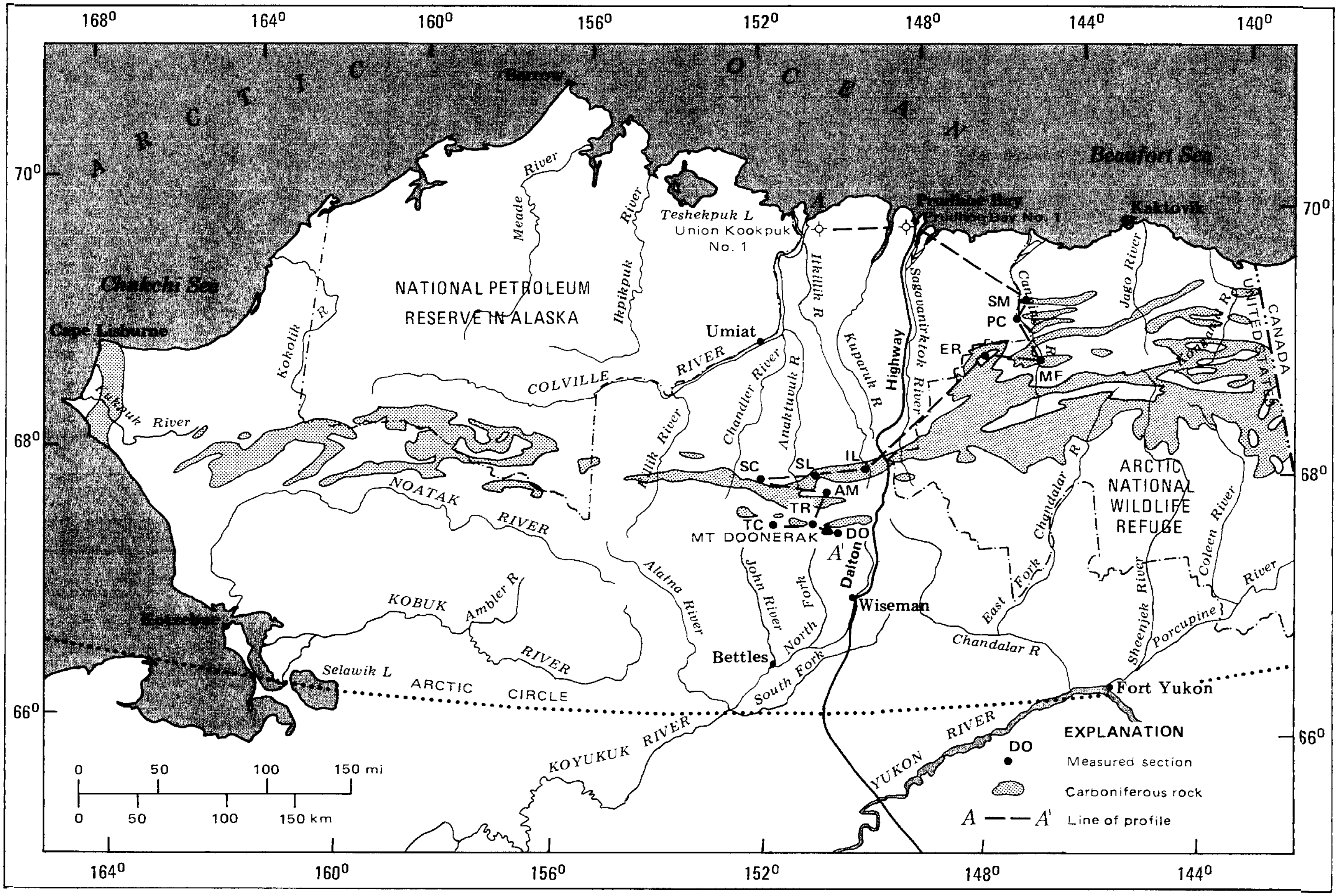


Figure 193. Index map of northern Alaska, showing location of measured sections: SM, west Sadlerochit Mountains; PC, Plunge Creek; MF, Marsh Fork; ER, Echooka River; IL, Itkillik Lake; SL, Shainin Lake; SC, Skimo Creek; AM, Alapah Mountain; TR, Tinayguk River; TC, Till Creek; DO, Mount Doonerak. Modified from Armstrong and Mamet (1977).

observed at many levels in the sections investigated. The twelve foraminifer zones recognized range in age from late Tournaisian (Osagean), zone 7, to early Westphalian (Atokan), zone 21. Faunal zonations for selected sections measured in the Brooks Range are shown in figure 194.

Microfauna from the Lisburne Group belong to the Taimyr-Alaska transition realm (Mamet, 1962; Mamet and Belford, 1968). This fauna contains many representatives of North American fauna (abundant and diversified Eoendothyranopsis, Eoforschia, and *Zellerian* discoidea) and many Eurasiatic elements, mostly among the Tournayellidae and Endothyridae families. In addition, cosmopolitan elements (mostly among the Archaediscidae family) allow direct correlation with both the original Carboniferous zonation of Europe and parts of the American midcontinent succession.

CARBONIFEROUS ROCKS IN THE BROOKS RANGE

The Lisburne Group is widespread in northern Alaska, both in the subsurface of the Arctic Slope north of the Brooks Range and throughout the northern Brooks Range from the Alaska-Yukon boundary on the east to the Chukchi Sea on the west (fig. 193). Outcrop patterns of Carboniferous rocks in the Brooks Range generally trend east-west, except in the western Brooks Range, where north or northeasterly trends prevail. The Lisburne Group is autochthonous in the subsurface of the Arctic Slope and parautochthonous in the northeastern Brooks Range but is allochthonous in the Endicott and De Long Mountains of the central and western Brooks Range (Mull, 1982), where a number of allochthons have been delineated (Tailleur and others, 1966; Mull, 1982; Mayfield and others, 1983). With the exception of the De Long Mountains in the western Brooks Range, most studies of the Lisburne Group have been either of the autochthonous sequence, or the allochthonous sequence on the Endicott Mountains allochthon--the structurally lowest allochthon in the Endicott Mountains. As a result of north-vergent thrusting, palinspastic maps must be used to reconstruct facies patterns in the Lisburne Group.

Carbonate rocks of the Lisburne Group represent a regional northward marine transgression onto a partially peneplained terrain of Devonian and older rocks (fig. 195; Brosgé and others, 1962). Based on data from deep test wells in the Prudhoe Bay region and outcrop studies south and west of the Canning River in the northeastern Brooks Range, Armstrong and Mamet (1970, 1977, 1978) recognized a regional northeast-southwest band of thicker carbonate rocks in the Philip Smith Mountains. This northeast-southwest trend was named the Canning Sag. This thick sequence of carbonate strata is well represented in the Itkillik Lake section, 20 mi (32 km) west of the Dalton Highway (fig. 196). Other measured sections of the Lisburne Group from the Endicott Mountains are also shown on figure 196.

In the northeastern Brooks Range, outcrops along the Canning River display evidence of the northward marine transgression. Wood and Armstrong (1975) delineated a possible Mississippian residual positive area, the Sadlerochit high, that consists of Devonian and older Paleozoic carbonate rocks and metamorphosed clastic rocks of the Neruokpuk Formation. The Sadlerochit high was probably the major source area for clastic rocks in the Kayak(?) Shale. This area remained above sea level through the Meramec time equivalent and was finally submerged during the lower Chesterian time equivalent (Mamet zone 16s). The Chesterian carbonate sediments that overlapped the Sadlerochit high were deposited in shallow water. Armstrong (1972), Wood and Armstrong (1975), and Armstrong and Mamet (1977, 1978) reported common occurrences of intertidal sedimentary structures, such as algal mats, bird's-eye structures, and small lithoclasts, in the upper part of the Alapah Limestone in the Sadlerochit Mountains.

CARBONIFEROUS ROCKS IN THE SUBSURFACE OF THE ARCTIC SLOPE

In the subsurface of the Arctic Slope, north of the Carboniferous outcrop belt near the axis of the Canning sag, the Atlantic Richfield-Humble Prudhoe Bay State No. 1 well contains more than 1,250 ft (380 m) of shallow-water, open-platform carbonate rocks of Chesterian (Mamet zone 16) to Atokan (Mamet zone 21) age (Armstrong and Mamet, 1974). The carbonate rocks are underlain by about 550 ft (167 m) of black and red shales, siltstones, sandstones, and thin coals that are probably Kayak Shale. The Mississippian Itkilyariak Formation of Mull and Mangus (1972) was proposed for a subsurface red-bed sequence and for basal clastic beds of the Lisburne Group in the Sadlerochit Mountains. Although Mississippian red beds are limited in outcrop, the Itkilyariak Formation is useful for subsurface mapping.

The carbonate section thins to the west and north of Prudhoe Bay (fig. 195). The Union Oil Kookpuk No. 1 well contains about 900 ft (270 m) of carbonate rocks of Chesterian to Atokan age, and the Sinclair-British Petroleum Colville No. 1 well to the north contains only 300 to 400 ft (90 to 120 m) of Morrowan- and Atokan-age carbonate rocks. Chesterian sediments may be represented by about 400 ft (120 m) of dark-gray to red siltstones, shales, and sandstones of the Itkilyariak Formation. Farther to the northwest near Point Barrow, Carboniferous carbonate rocks are absent in several wells, and a thin sequence of shale and sandstone overlies the pre-Carboniferous argillite; the shale and sandstone are unfossiliferous but may be of Carboniferous age.

CARBONATE MICROFACIES

Armstrong and Mamet (1977, 1978) proposed an idealized depositional model for carbonates of the Lisburne Group (fig. 197). Although Lisburne Group carbonates are very thick in some areas, reefs of coral or other metazoans are unknown. The Lisburne is composed of cyclic groups of carbonate rock. This is well exemplified by sections in the Sadlerochit Mountains of the northeastern Brooks Range, where deposits of major marine transgressions are present above and below an intertidal carbonate-offlap deposit. Regional petrologic and carbonate-facies studies show that the dominant rock types are bryozoan-pelmatozoan mudstones and wackestones overlain by bryozoan-pelmatozoan packstones and grainstones.

Dolomite is common in most sections of the Lisburne and appears to be primarily diagenetic. Some dolomite, however, is closely associated with intertidal sedimentary structures, such as bird's-eye features, algal mats, mud cracks, and calcite pseudomorphs after gypsum. The close stratigraphic association of these intertidal sedimentary structures with the dolomite suggests a close genetic relation. For instance, thick dolomite sequences in sections at Shainin and Itkillik Lakes in the Endicott Mountains (fig. 196) are associated with well-developed intertidal sedimentary structures. These dolomites are generally macrodolomites that show clear evidence of extensive dolomitization of bryozoan-pelmatozoan packstone. Well-developed oolitic beds are confined to Morrowan and Atokan beds of northeastern Alaska and are rare in the Mississippian carbonates.

Sponge-spiculite facies of the Lisburne Group are found at numerous localities, which Mamet and Armstrong (1972) believe represent two different environments: 1) relatively shallow water with poor circulation, and 2) the deeper part of the platform associated with radiolarians. Commonly, the shallow-water part of this facies is argillaceous and may contain beds with calcite pseudomorphs after gypsum.

RESERVOIR QUALITY

Pennsylvanian carbonates of the Wahoo Limestone are a series of shoaling-upward sequences that range from shallow-marine crinoid-bryozoan packstones through ooid shoals into thin-bedded, orange-brown dolomites (Armstrong, 1973). These cyclic groups can be traced over large areas of the eastern Arctic Slope. Parts of these shoaling-upward sequences are very porous and contain about three billion barrels of oil in the Lisburne Group reservoir at Prudhoe Bay. Because porosity is good within these cyclic groups in some areas, other hydrocarbon accumulations may exist within the Wahoo Limestone (Bird and Jordan, 1977).

In the Cape Lisburne area at the western end of the Brooks Range, thick carbonates of Chesterian age contain well-developed cyclic, shoaling-upward groups. Because the dolomite sequences associated with the cyclic groups are porous, the dolomite sequences may contain extensive reservoir rocks (Armstrong and others, 1971).

LISBURNE GROUP IN THE ENDICOTT MOUNTAINS

The Lisburne Group has not been studied in detail near the Dalton Highway. However, a section studied by Armstrong and Mamet (1978) 36 mi (58 km) to the west at Alapah Mountain can be used as a general guide to the Lisburne Group in the eastern Endicott Mountains. The Lisburne Group, both at Alapah Mountain and near the Dalton Highway, is part of the Endicott Mountains allochthon.

COMPOSITE SECTION OF THE LISBURNE GROUP AT ALAPAH MOUNTAIN

LITHOLOGY

At the headwaters of the Nanushuk River near Alapah Mountain, marker beds of the Lisburne Group were traced from and correlated among four outcrops. The resultant composite section (fig. 198) represents a complete sequence of the Lisburne Group. The Wachsmuth Limestone forms the base of the section and can be divided into two distinct lithologic units. The lower unit is a 460-ft-thick (140 m) echinoderm-bryozoan bioclastic wackestone and packstone that weathers light gray and is argillaceous in the lower 120 ft (37 m) of the unit. The upper unit is 730 ft (224 m) thick, darker gray, thinner bedded, argillaceous, and contains black, nodular chert. These limestone beds are composed of echinoderm-bryozoan bioclastic wackestone to packstone that may be platy and contain thin shale interbeds. The contact between the Wachsmuth Limestone and the Alapah Limestone is within a 40-ft-thick (12 m) unit of 1.5- to 6.5-ft-thick (0.5 to 2 m) beds of gray bryozoan-echinoderm bioclastic packstone. At about 1,320 ft (406 m), alternating, thin beds of dark-gray to black chert and gray dolomite become dominant; thin (<0.5 in.; 1 cm) interbeds of siliceous shale are also common. In thin section, the chert is spiculitic, radio-

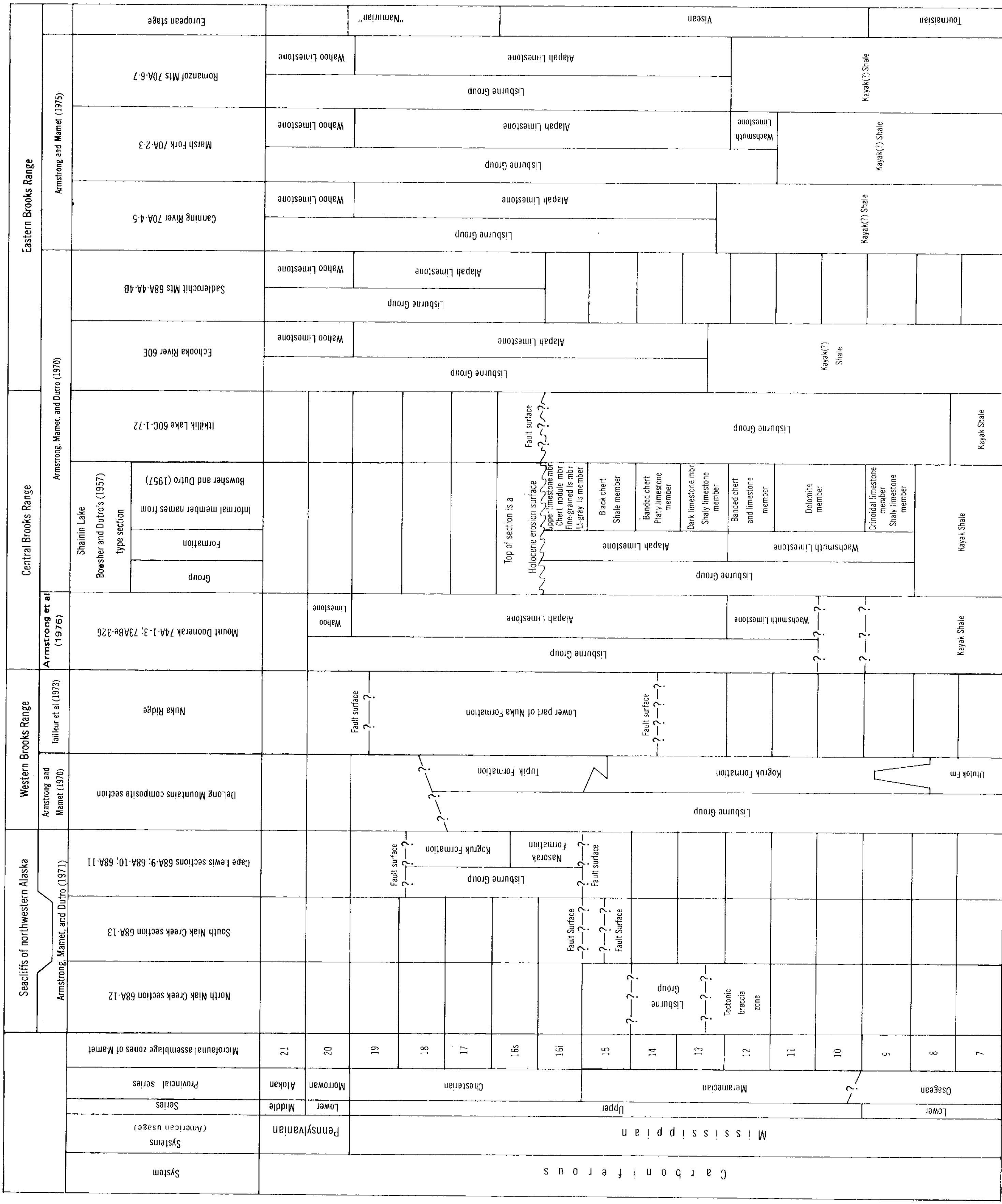


Figure 194. Regional correlation diagram for Lisburne Group in Brooks Range, showing microfaunal assemblage zones of Mamet. Modified from Armstrong and Mamet (1977).

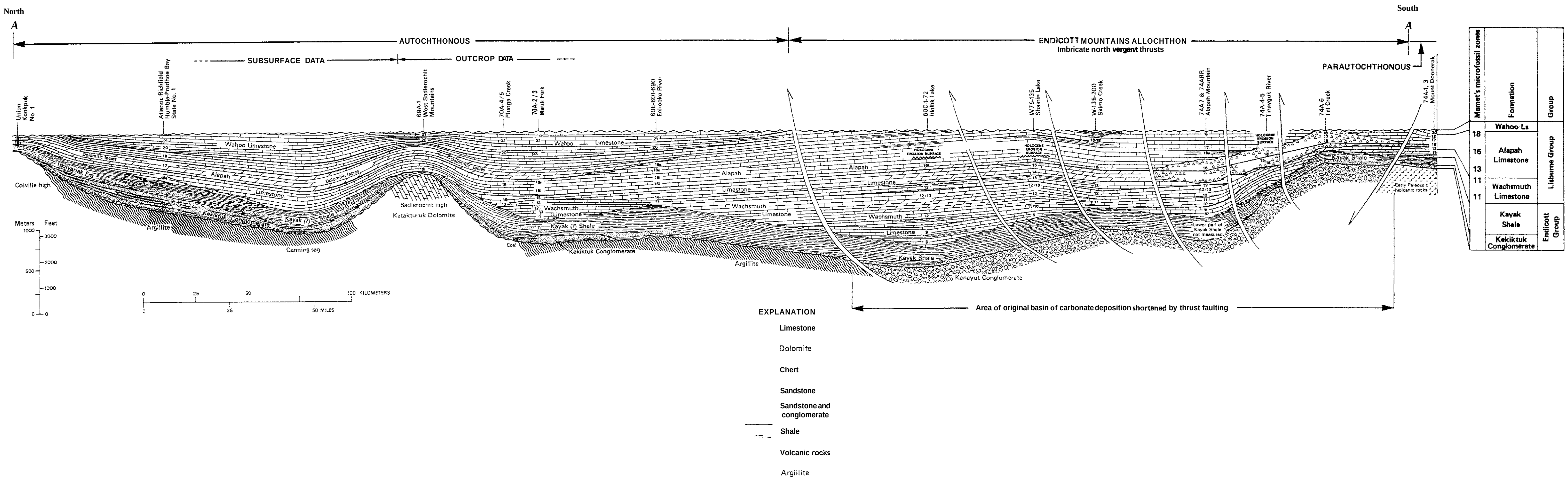


Figure 195. Generalized palinspastic profile of Carboniferous facies from subsurface of Prudhoe Bay area south to Sadlerochrt and Endicott Mountains. See figure 193 for location of measured sections and line of profile. Modified from Armsfrong and Mamet (1978).

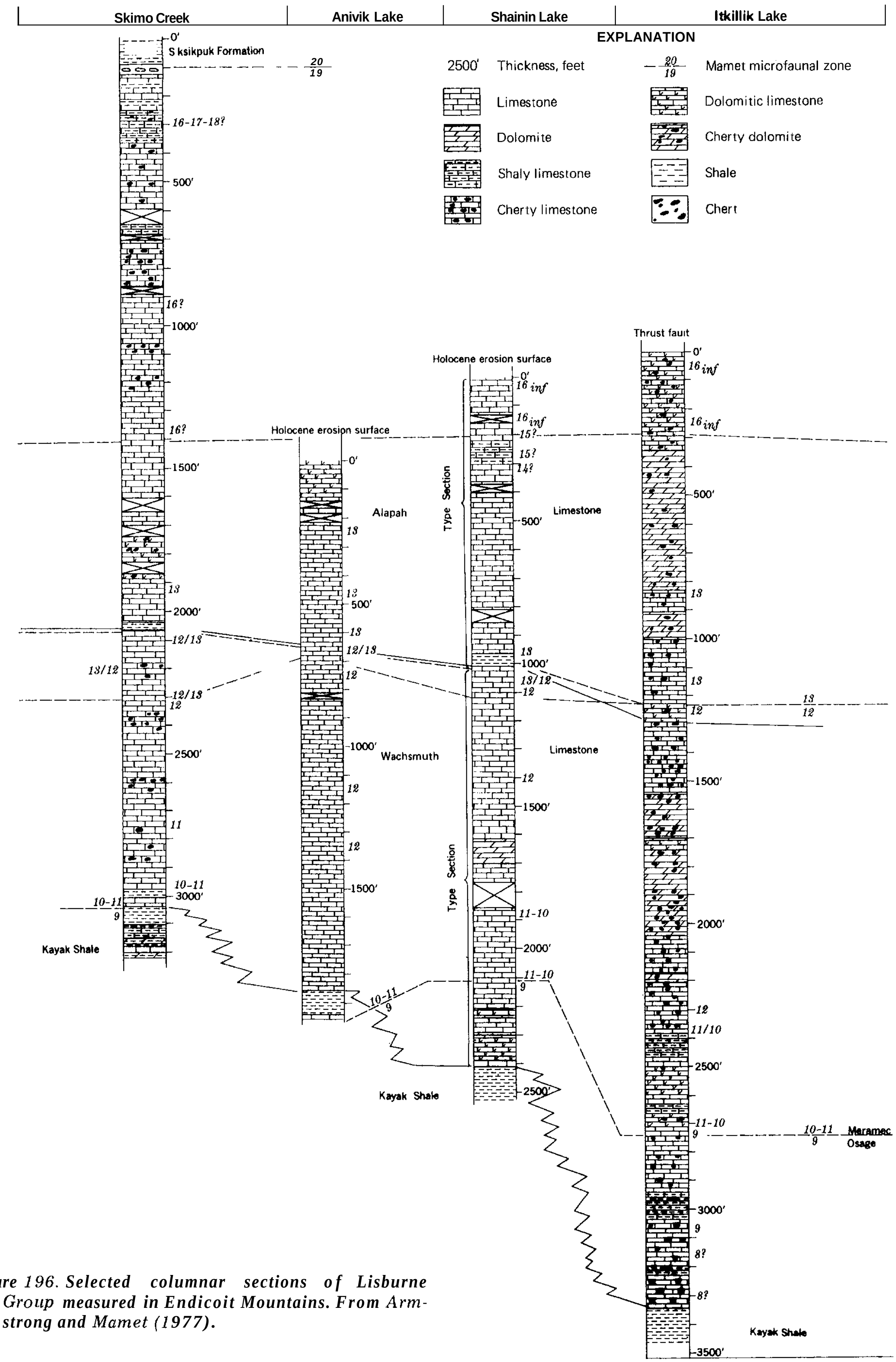


Figure 196. Selected columnar sections of Lisburne Group measured in Endicoit Mountains. From Armstrong and Mamet (1977).

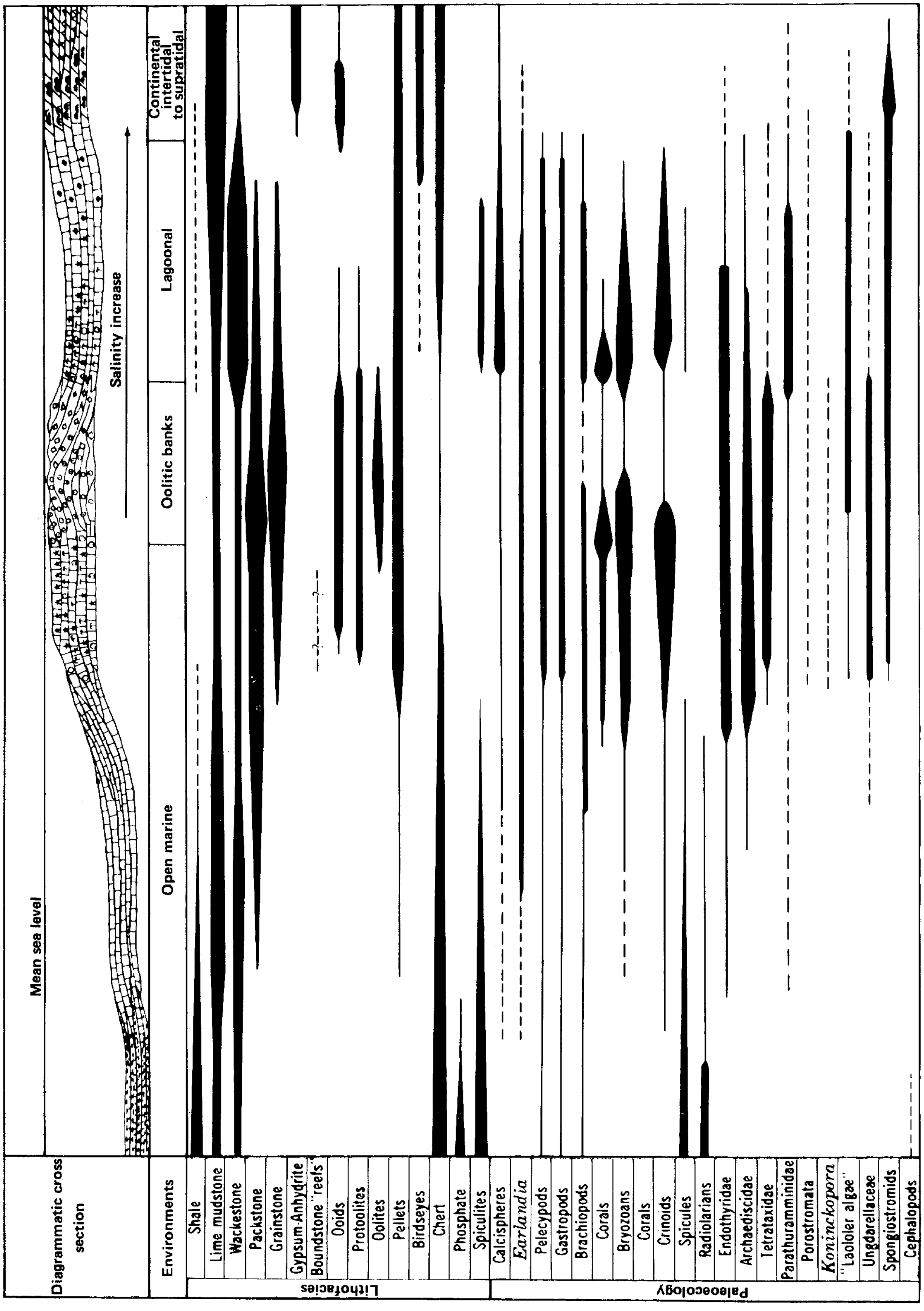


Figure 197. Idealized carbonate depositional model showing distribution of fossils and organic and inorganic constituents. Modified from Armstrong and Mamet (1977).

larian, argillaceous, and cryptocrystalline. Etched chert surfaces show only vague outlines of spicules and spherules that may be radiolarians. In thin section, the chert commonly shows euhedral and partly corroded dolomite rhombs (Wood and Armstrong, 1975). Calcareous cherts within the composite section contain bioclasts of bryozoans, foraminifera, and echinoderms; some of the bioclasts are deeply corroded along their margins. Beginning at about 2,080 ft (640 m), the chert is progressively replaced by echinoderm-bryozoan bioclastic wackestone and packstone. From about 2,210 ft (680 m) to the top of the section at 3,380 ft (1,040 m), the column is predominantly light-gray, massive-bedded, echinoderm-bryozoan-brachiopod wackestone and packstone.

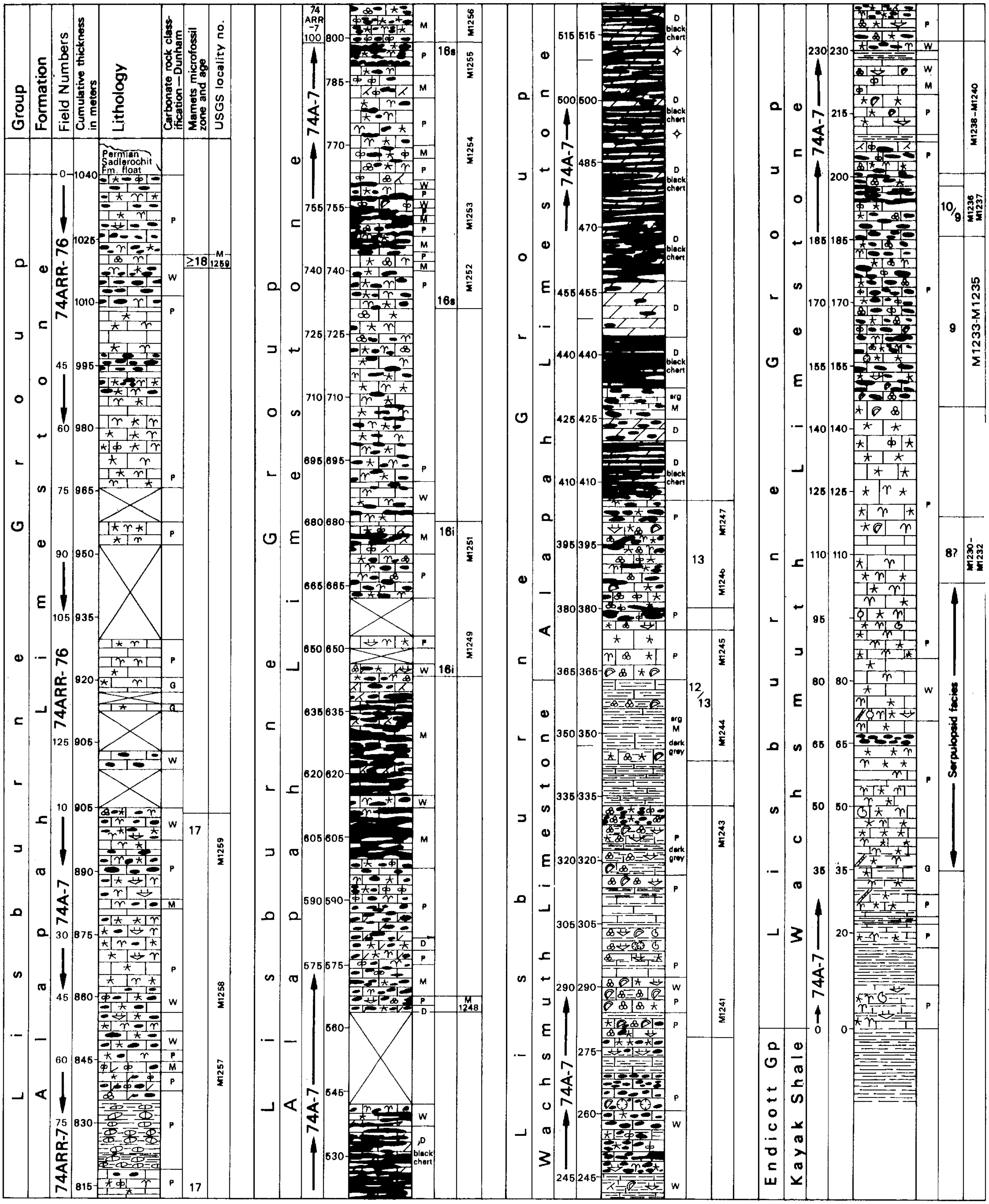
MICROFACIES

The 3,380-ft-thick (1,040 m) composite section at Alapah Mountain (fig. 198) exemplifies the complexity of carbonate sedimentation recorded in the Lisburne Group. The contact between the Wachsmuth Limestone and the underlying Kayak Shale represents a transition from silty, gray-black marine shales of the Kayak to open-platform echinoderm-bryozoan packstone of Osagean age of the Wachsmuth. Carbonate sediments in the interval from 212 to 471 ft (65 to 145 m) are thick-bedded echinoderm-bryozoan bioclastic wackestone and packstone that were probably deposited on an open platform to near-shoaling marine shelf. Bedding progressively thins upward from 471 to 1,300 ft (145 to 400 m), is darker gray, and contains more argillaceous admixtures and dark-gray nodular chert. Dominant rock types include peloid-micropeloid-echinoderm bioclastic packstone and wackestone that probably represent deposition on a slowly drowning open-marine shelf to foreslope environment.

At 1,300 ft (400 m), beds of bioclastic-rich sediments abruptly change to thin (0.75 to 4 in.; 2 to 10 cm) beds of dark-gray to black chert interbedded with dark-gray dolomite. From 1,485 to 1,765 ft (457 to 543 m), thinly (1.2 to 6 in.; 3 to 15 cm) bedded, black to dark-gray chert is interlayered with dark-gray dolomite and thinly (0.4 to 1.2 in.; 1 to 3 cm) bedded, black shale. These lithologies are uniform, containing no shelly fossils. Chert from this interval is microlaminated and composed of spiculitic, radiolarian silica. This thin-bedded sequence represents deposition of silica-forming organisms and argillaceous and carbonate particles in a starved, anaerobic basin. The basin was situated south of the main shallow-water shelf or platform environments that produced most of the Lisburne Group carbonates. Organisms that contribute large quantities of siliceous skeletal remains to pelagic sediments in recent oceans are, in order of decreasing importance, diatoms, radiolarians, sponges, and silicoflagellates.

Samples of chert from the interval 1,485 to 1,765 ft (457 to 543 m) were etched with hydrochloric acid and studied with a scanning electron microscope. The chert is composed of cryptocrystalline quartz and poorly defined spicules. The cryptocrystalline quartz is composed of subhedral crystals to 5 μ long. Within the cryptocrystalline-quartz matrix are euhedral to subhedral rhombs of dolomite.

From about 1,836 to 1,950 ft (565 to 600 m) is a sequence of thin-bedded, dark-gray pelletoid-echinoderm mudstone and packstone. This sequence probably represents toe-of-slope to foreslope carbonate deposition because of the increase in bioclastic remains and greater abundance of carbonate sediments. From 1,950 to 2,080 ft (600 to 640 m), the rocks appear to reflect more subsidence and submergence. The presence of thin-bedded, black chert and interbedded black shale and lime mudstone indicates deposits of an euxinic, starved basin. From 2,080 to 2,210 ft (640 to 680 m), the amount of chert and spiculites decreases and the echinoderm, bryozoan, brachiopod, and foraminifera content increases. This 130-ft-thick (40 m) interval was probably deposited in a tidal-shelf to toe-of-slope environment. From 2,210 ft (680 m) to the top of the Alapah Limestone (3,380 ft; 1,040 m), thick to massive, light-gray beds of echinoderm-bryozoan bioclastic lime mudstone and wackestone to grainstone are present. These were probably deposited on an aerobic, open shelf. A gray, limy shale with lime mudstone from 2,660 to 2,717 ft (819 to 836 m) represents an influx of terrigenous clay. The top of the section (3,380 ft; 1,040 m) is overlain by yellow-brown sands of the Permian Siksikpuk Formation.


Group
Formation
Field Numbers
Cumulative thickness in meters
Lithology
Carbonate rock classification—Dunham
Mamets microfossil zone and age
USGS locality no.
Lisburne Group
Alapah Limestone
Wachsmuth Limestone
Endicott Gp
Kayak Shale
74ARR-76
74A-7
74ARR-7
Permian Sadlerochit Fm. float
black chert
dark grey
Serpulopsid facies

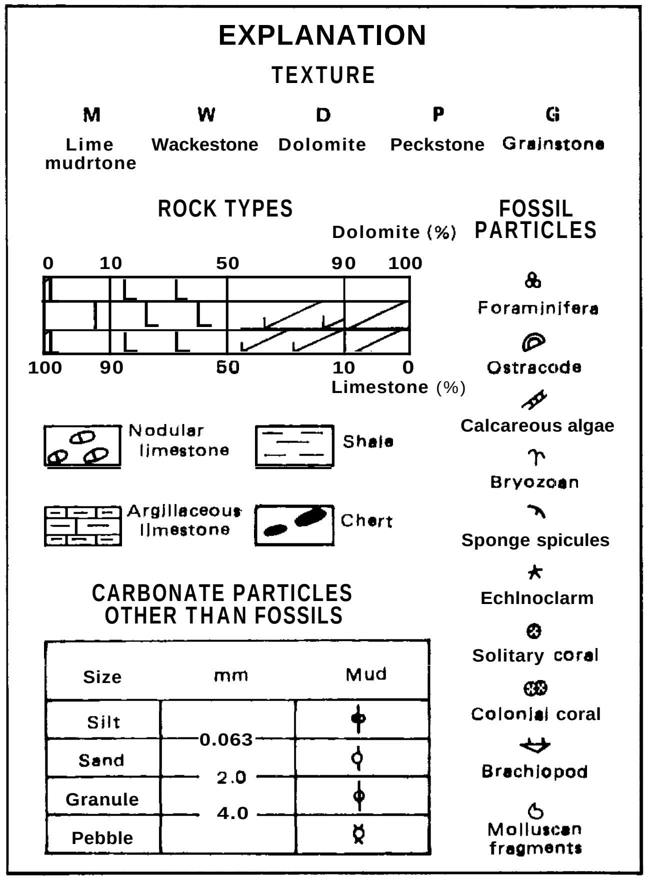


Figure 198. Composite section of Lisburne Group measured at Alapah Mountain, 36 mi (58 km) west of Dalton Highway. From Armstrong and Mamet (1978).

CHAPTER 17.

PERMIAN STRATIGRAPHY IN THE ATIGUN GORGE AREA: A TRANSITION BETWEEN THE ECHOOKA AND SIKSIKPUK FORMATIONS

By K.E. Adams[1] and J.P. Siok[2]

INTRODUCTION

The Lower Permian siltstone and shale sequence in the Atigun Gorge area near Galbraith Lake (Mile 270) is transitional between the lower part of the autochthonous or parautochthonous Sadlerochit Group of the northeastern Brooks Range and the allochthonous Siksikpuk Formation of the central and western Brooks Range. The sequence is composed dominantly of well-indurated, dark-gray to black shale with a basal unit of orange-brown-weathering siltstone and silty limestone that appears similar to the Echooka Formation of the Sadlerochit Group. The presence of barite, siliceous mudstone, and maroon-and-green shale, however, is more characteristic of the Siksikpuk Formation.

PREVIOUS INVESTIGATIONS

SIKSIKPUK FORMATION

Patton (1957) named the Siksikpuk Formation for a sequence of variegated shale and siltstone found in the central and western Brooks Range. The composite stratotype for the formation is on the structurally lowest allochthon in the Brooks Range, the Endicott Mountains allochthon. The formation overlies the Lower Mississippian to Pennsylvanian Lisburne Group and is overlain by the Triassic to Middle Jurassic Otuk Formation (Mull and others, 1982). A Permian(?) age was originally assigned to the Siksikpuk Formation based on megafossils (Patton, 1957).

The Etivluk Group was later proposed by Mull and others (1982) to include the Siksikpuk Formation and the Otuk Formation in the central and western Brooks Range. Mull and others (1987) later restricted the Siksikpuk Formation to the dominantly shaly and silty beds typical of the composite stratotype. The dominantly cherty beds that occupy a similar stratigraphic position on higher allochthons in the central and western Brooks Range were redefined as the Pennsylvanian to Triassic(?) Imnaitchiak Chert of the Etivluk Group (Mull and others, 1987).

The first detailed descriptions of the Siksikpuk Formation were the result of exploration of Naval Petroleum Reserve No. 4 in the late 1940s and early 1950s. In reports of the exploration, Chapman and others (1964) and Patton and Tailleur (1964) discussed the Siksikpuk along the mountain front of the central Brooks Range. Porter (1966b) mapped the first known exposures of the Siksikpuk Formation south of the mountain front in a structural study of the Anaktuvuk Pass area in the central Brooks Range. The westernmost occurrence of the Siksikpuk Formation, along the Chukchi Sea coast near Cape Lisburne, was reported by Campbell (1967). Siok (1985), in the most detailed study of the Siksikpuk Formation, discussed the lithostratigraphy, biostratigraphy, and geologic history of the unit in the northcentral Brooks Range.

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SADLEROCHIT GROUP

Leffingwell (1919) originally named the Sadlerochit Sandstone for a 300-ft-thick (100 m) sequence of 'light sandstone or dark quartzite' that overlies the Carboniferous Lisburne Group and underlies the Middle and Upper Triassic Shublik Formation in the Sadlerochit Mountains of the northeastern Brooks Range. At that time, the unit was assigned a Pennsylvanian age (Leffingwell, 1919) but later was considered Permian in age by G.H. Girty (Smith, 1939). In the Shaviovik and Sagavanirktok Rivers region, Keller and others (1961) mapped a thicker succession of the Sadlerochit Sandstone, which they designated the Sadlerochit Formation; they also divided the formation into the Upper Permian Echooka Member and the Lower Triassic Ivishak Member. Subsequent reports on the Sadlerochit Formation in the northeastern Brooks Range have been published by Detterman (1970, 1974, 1976a, b), Detterman and others (1975), and Detterman and Dutro (1977). Detterman and others (1975) raised the Sadlerochit Formation to group status and raised the Echooka and Ivishak Members to formation status. In addition, the Echooka Formation was further subdivided into the lower Joe Creek Member and the upper Ikiakpaurak Member, and the Ivishak Formation was divided (in ascending order) into the Kavik, the Ledge Sandstone, and the Fire Creek Siltstone Members (fig. 199).

The Sadlerochit Group took on new significance with the discovery of oil at Prudhoe Bay in 1968. Its upper beds are the main reservoir for the largest oil field in North America. Rickwood (1970) described the lithology of the Sadlerochit reservoir in his discussion of the structure and stratigraphy of the Prudhoe Bay field. In a similar discussion, Morgridge and Smith (1972) recognized the lower part of the Sadlerochit Group as a deltaic shallow-marine sequence and the upper part of the Sadlerochit Group as an alluvial complex. Detailed facies analyses by Eckelmann and others (1975) and Melvin and Knight (1984) refined the depositional interpretations. Jones and Speers (1976) proposed formal nomenclature for the Sadlerochit in the Prudhoe Bay area, where the unit is apparently slightly different in age than in the Brooks Range to the south. They raised the Sadlerochit Formation to group status and subdivided the unit into three formations: 1) the Upper Permian Echooka Formation, 2) the Upper Permian Kavik Shale, and 3) the Upper Permian to Lower Triassic Ivishak Sandstone.

The Sadlerochit nomenclature was confined to the northeastern Brooks Range and Arctic Slope subsurface until Dutro and others (1976) assigned Permian strata in the Doonerak fenster of the central Brooks Range to the Sadlerochit Group. They noted that Permian beds in the Doonerak area more closely resemble correlative beds of the Sadlerochit Group in the northeastern Brooks Range than coeval beds of the Siksikpuk Formation north of the Doonerak area. These observations provide important constraints for paleogeographic reconstructions of the Brooks Range.

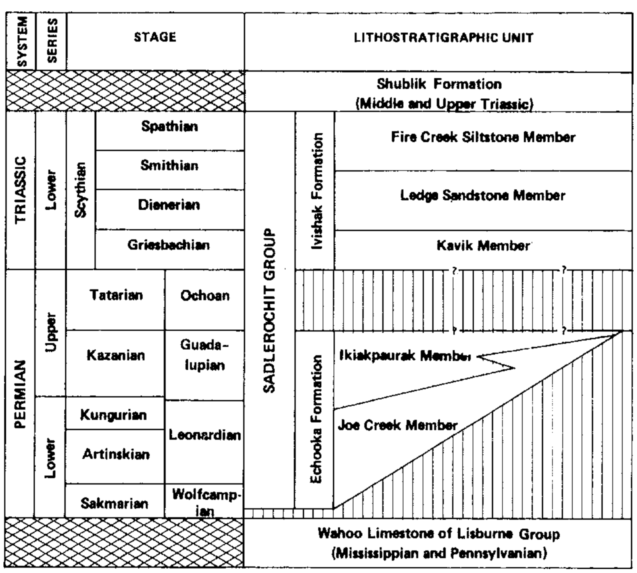

Figure 199. ***Stratigraphy of the Sadlerochit Group, northeastern Brooks Range, modified from Detterman and others (1975). Similar, but slightly modified, nomenclature was extended to the Arctic Slope subsurface by Jones and Speers (1976).***

PERMIAN SEQUENCE IN THE ATIGUN GORGE AREA

A few investigations specifically address Permian rocks in the Atigun Gorge area, but there is no agreement on Permian nomenclature in the area because the rocks are transitional between the Siksikpuk Formation and the Sadlerochit Group. Brosgé and others (1962), in a report on Paleozoic rocks of the eastern Brooks Range, assigned the Permian sequence in the Atigun Gorge area to the Siksikpuk Formation. They did note, however, that "the southern shaley facies of the lower part of the Sadlerochit is correlative or identical with the Siksikpuk Formation." Paris (1978), Metz and Robinson (1979), and Sellars (1981), who reported on the general geology and mineral occurrence in the Atigun Gorge area, also considered the Permian sequence to be similar to the Siksikpuk Formation. Conversely, in geologic investigations of the Philip Smith Mountains Quadrangle, Detterman (1976a, b) and Brosgé and others (1979a) referred to the Permian sequence in the Atigun Gorge area as the Sadlerochit Group. Moreover, Detterman (1976a) suggested that a conspicuous yellow-weathering siltstone and limestone unit of Early Permian age in the Philip Smith Mountains Quadrangle should be considered a new, unnamed member of the Echooka Formation of the Sadlerochit Group.

GENERAL STRATIGRAPHIC SETTING

SIKSIKPUK FORMATION

The interbedded siltstone and shale sequence of the Siksikpuk Formation ranges from 200 to 600 ft (60 to 150 m) thick and is exposed primarily along the north flank of the central and western Brooks Range. Other scattered outcrops are present within the range in structural infolds of the Lisburne Group, but because the formation is recessive, the more visible outcrops are limited to active stream-cut banks and steep mountain slopes.

An Early Permian (Wolfcampian) brachiopod fauna is present near the base of the stratotype, and a Late Permian (Guadalupian) conodont assemblage was recovered from the top of the stratotype (Bodnar, 1984). Pennsylvanian and Triassic radiolarians and Triassic conodonts reported by Mull and others (1982, 1987c) from the Siksikpuk Formation arc from beds now assigned to the Imnaitchiak chert on the Picnic Creek allochthon, the thrust sequence that overlies the Endicott Mountains allochthon. On the basis of similar basal lithologies and brachiopod faunas, the Siksikpuk Formation is considered a distal equivalent of the Echooka Formation of northeastern Alaska and the Arctic Slope subsurface.

ECHOOKA FORMATION

In the northeastern Brooks Range, surface exposures of clastic, carbonate, and quartzose rocks of the Joe Creek and Ikiakpaurak Members of the Echooka Formation range up to 900 ft (275 m) thick (Detterman and Dutro, 1977). These resistant beds form conspicuous triangular flatirons on dip slopes that overlie the Lisburne Group, and they are present locally as erosional remnants on the crests of anticlines, but the best exposures are found in the cut banks of smaller streams. Detailed information in the following two paragraphs is based on an extensive study of the Echooka Formation by Detterman and others (1975).

The basal member of the Echooka Formation, the Joe Creek Member, is composed of a sequence of calcareous siltstone, bedded chert, bioclastic limestone, and quartz calcarenite that disconformably overlies the Lisburne Group. At the type locality, the Joe Creek Member is 372 ft (113 m) thick, but it thickens to the south and west and is completely absent in the northeastern outcrops of the Echooka Formation. Brachiopods collected from the Joe Creek Member are correlative with Early to Late Permian (Wolfcampian to Guadalupian) faunal zones Ea to G of Bamber and Waterhouse (1971).

The upper member of the Echooka Formation, the Ikiakpaurak Member, is in conformable to disconformable contact with the Joe Creek Member and is disconformably overlain by the Ivishak Formation. The Ikiakpaurak Member forms the main part of the Echooka Formation along the north flank of the eastern Brooks Range and consists of dark orthoquartzite, quartzitic sandstone, and siltstone. At the type locality, the section is 280 ft (85 m) thick and thins abruptly to the north. An early Late Permian (Guadalupian) brachiopod assemblage is found within the member, as well as the trace fossil *Zoophycos,* which occurs throughout the entire Echooka Formation and is considered diagnostic for that stratigraphic interval.

STRATIGRAPHY OF THE SIKSIKPUK FORMATION ON THE ENDICOTT MOUNTAINS ALLOCHTHON

Four lithologic units, informally designated A, B, C, and D, compose the Siksikpuk Formation (fig. 200) on the Endicott Mountains allochthon (Siok, 1985). Unit A (30 to 56 ft; 8 to 17 m), which forms the base of the Siksikpuk Formation, is a conspicuous yellow- and orange-weathering sequence of pyritic, gray-green siltstone interbedded with yellow clay. Some workers have informally suggested that the clay is volcanic in origin. However, petrographic analyses of the clay do not indicate a volcanic origin; rather, the yellow-weathering is caused by alteration of pyrite to jarosite [$KFe_3(SO_4)_2(OH)_6$], a bright-yellow sulfate mineral (Siok, 1985). Unit B varies in thickness (60 to 330 ft; 20 to 100 m) and lithology, which might suggest deposition in separate basins, but is dominated by mottled, red and gray-green mudstone and siltstone that contain abundant barite and siderite nodules. Unit C (23 to 59 ft; 7 to 18 m), the most resistant unit within the Siksikpuk Formation, is a wispy-laminated, gray-green silicified mudstone with dark-gray shale interbeds. Unit D (30 to 130 ft; 10 to 40 m), although poorly exposed, consists of fissile gray-green to dark-gray shale and siltstone.

Correlation of lithofacies west of the composite stratotype on Skimo and Tiglukpuk Creeks (about 65 mi [100 km] west of Galbraith Lake) shows an overall decrease in thickness of the Siksikpuk Formation to a minimum of 130 ft (about 40 m) along the Middle Fork Okpikruak River (about 40 mi [70 km] west of Skimo and Tiglukpuk Creeks), followed by an increase in thickness westward toward Kurupa Hills (about 78 mi [125 km] west of Skimo and Tiglukpuk Creeks) (see index map, fig. 201). West of the Middle Fork Okpikruak River, unit B is dominated by siltstone; to the east, it is composed of equal amounts of mudstone and siltstone. Unit C changes westward, from siliceous mudstone at Skimo and Tiglukpuk Creeks to chert in the south Kurupa Hills. Notable lithologic changes also occur between Shainin Lake and Cobblestone Creek. At Cobblestone Creek, unit A contains abundant fossiliferous carbonates and the trace fossil *Zoophycos;* unit B is darker, finer grained, and contains larger (6 to 10 ft; 2 to 3 m) siderite and barite nodules; units C and D are not present. These changes probably reflect the crossing of a major structural discontinuity onto a lower thrust sheet within the Endicott Mountains allochthon (C.G. Mull and J.T. Dutro, Jr., oral commun., 1984). The facies changes at Cobblestone Creek are also present in the Atigun Gorge area.

Fossils are not abundant in the Siksikpuk Formation. Most are found lower in the sequence in a distinct argillaceous limestone bed at the top of unit A. This bed contains an Early Permian (Wolfcampian) fauna that includes the gastropod *Straparollus (Euomphalus) alaskensis* Yochelson and Dutro, 1960, and the brachiopods *Martina* sp., *Spiriferella* sp., and *Attenuatella* sp. (J.T. Dutro, Jr., written commun., 1984). Permian foraminifera that are similar to those found in the Kavik Formation of the Sadlerochit Group in the Arctic Slope subsurface (M.B. Mickey, written commun., 1984) were recovered from clay beds in unit A. *Marlinia* sp. brachiopods are present in unit B and considered by Dutro (written commun., 1984) to be the same species as those found in unit **A.** Conodonts recovered from unit B include *Neogondolella* cf. *N. idahoensis* (Youngquist, Hawley, and Miller) (U.S. Geological Survey colln. 29597-PC) of late Early Permian (Leonardian) age (A.G. Harris, written commun., 1985). The only fossils recovered from unit C are fragments of the radiolarian *Pseudoalbaillella* sp. (U.S. Geological Survey MR: 6284), which ranges from Middle Pennsylvanian (Atokan) to late Early Permian (Leonardian) in age (B.L. Murchey and K.M. Reed, written commun., 1984). *Neogondolella* cf. *N. idahoensis* (Youngquist, Hawley, and Miller) (U.S. Geological Survey collns. 2900-PC and 29602-PC) is also present in the lower part of unit D; in addition, the early Late Permian (Guadalupian) conodont *Neogondolella phosphoriensis* was discovered in the upper 33 ft (10 m) of unit D (Bodnar, 1984). These fossils suggest that most of the Siksikpuk Formation on the Endicott Mountains allochthon was deposited in Early Permian time.

Ichnofossils in units B and C include *Rhizocorallium?, Planolites,* and *Teichichnus,* which represent the *Cruziana* ichnofacies characteristic of shelf environments. On the basis of ichnofossils, brachiopods, arenaceous foraminifera, and general bedding characteristics, the Siksikpuk Formation is assumed to have been deposited in an inner to middle neritic environment in relatively shallow water (452 ft; 150 m). The environment was characterized by hemipelagic sedimentation and a soft, muddy substrate with abundant infaunal organisms.

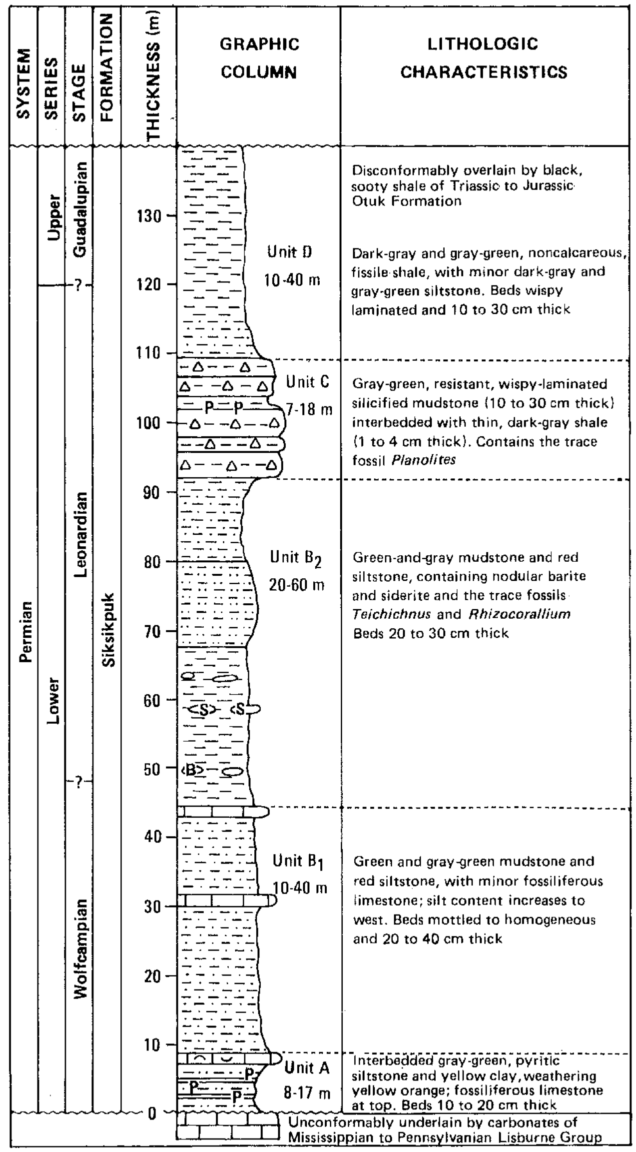

Figure 200. Generalized columnar section of informal lithologic units of Siksikpuk Formation on Endicott Mountains allochthon. Lithologic symbols: B, barite; P, pyrite; S, siderite.

STRATIGRAPHY OF THE ECHOOKA FORMATION IN THE NORTHEASTERN BROOKS RANGE

The Echooka Formation has been recognized across the northeastern Brooks Range, from near the Alaska-Yukon border in the Demarcation Point and Table Mountain Quadrangles southwest to the Philip Smith Mountains Quadrangle (see index map, fig. 202). The easternmost exposures near the Joe Creek area (200 mi [320 km] northeast of Galbraith Lake) consist mainly of limestone and shale. In the Canning River area, on the west side of the Sadlerochit and Shublik Mountains, the Echooka Formation is sandy to conglomeratic and locally contains fossiliferous limestone. Between the Canning and Echooka Rivers, the sequence is composed of dark-blue-gray cherty siltstone interbedded with gray-green to black chert and light-gray quartzite, with occasional lenses of fossiliferous limestone. Farther to the west near Flood Creek, the formation consists of

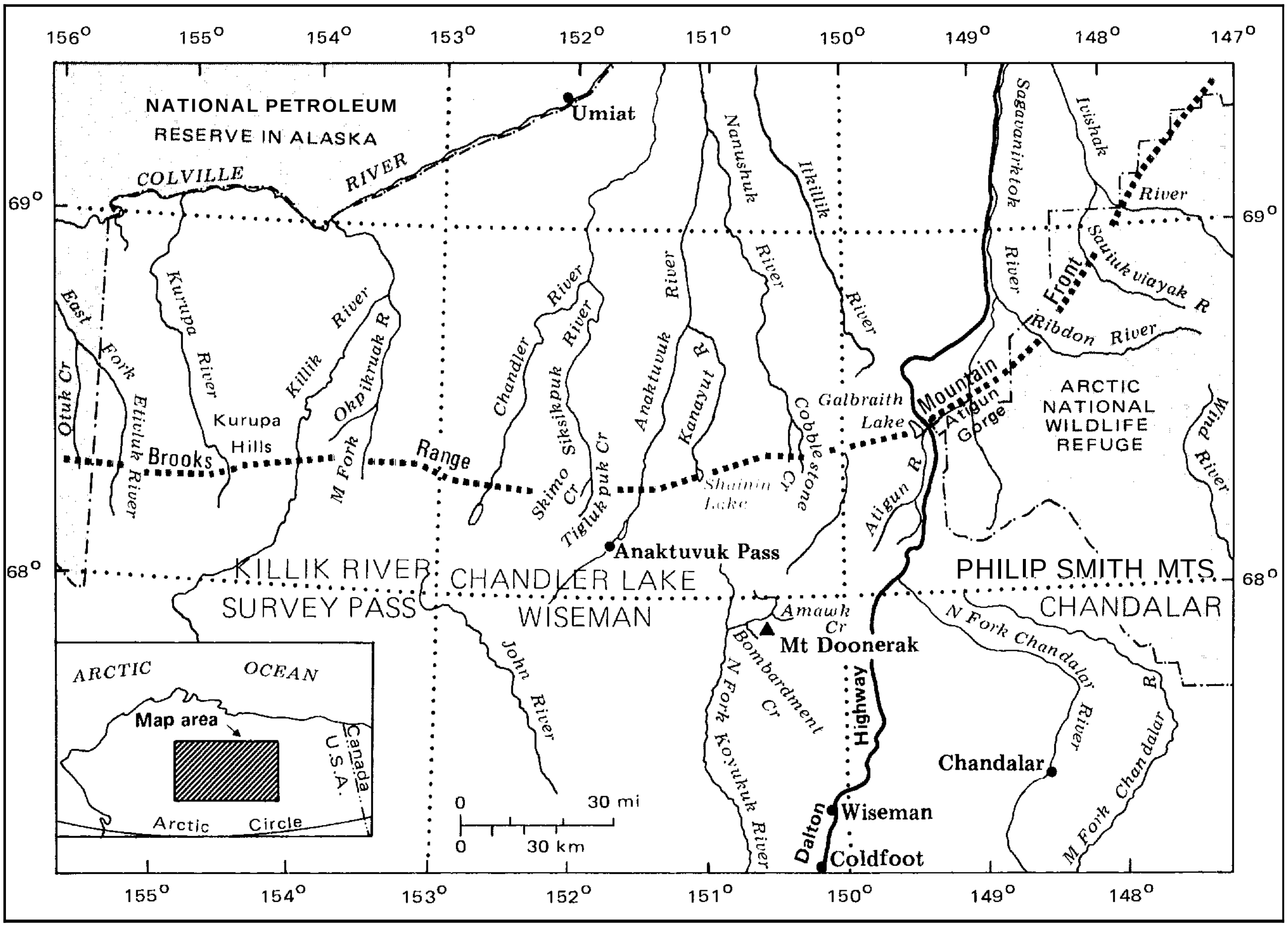

Figure 201. Index map of northcentral Brooks Range.

calcareous brown-weathering siltstone, calcareous gray shale, and blue-gray limestone. In general, rocks of the Echooka Formation increase in grain size and clastic content toward the north, which indicates a northern or northeastern source area for this shallow-marine deposit. (See Keller and others, 1961, and Detterman, 1970.)

On Joe Creek, the lower member of the Echooka Formation, the Joe Creek Member, can be divided into three lithologic units. The basal unit (200 ft; 60 m) is composed predominantly of yellow-brown-weathering silty limestone and calcareous siltstone that is thinly bedded and poorly exposed. Chert and siliceous siltstone (90 ft; 27 m) overlie the basal unit, and the remainder of the sequence (86 ft; 26 m) consists of glauconitic calcarenite and bioclastic limestone (Detterman and others, 1975). About 50 mi (80 km) west of Joe Creek, in the Ivishak River area, the basal unit (140 ft; 43 m) can still be identified but is overlain by a cross-bedded, bioturbated sandstone (40 ft; 12 m) and calcareous siltstone (50 ft; 15 m). The top of the Joe Creek Member in the Ivishak River area is a massively bedded shelf-carbonate and sand-and-mud deposit (170 ft; 52 m) (Detterman and Dutro, 1977). Exposures in the Philip Smith Mountains Quadrangle, farther to the west and south, contain a much thicker unit of yellow-weathering calcareous siltstone and silty limestone that in places could be considered a separate stratigraphic interval (Detterman, 1976b).

The upper member of the Echooka Formation, the Ikiakpaurak Member, is a fine- to very fine grained quartzose sandstone to siltstone with minor interbeds of silty shale. Locally, in the Sadlerochit and Shublik Mountains, a channel conglomerate is present in the basal part of the member (Detterman and others, 1975). The dark quartzitic sandstone and siltstone of the Ikiakpaurak, however, can be traced only as far west as the northeastern corner of the Philip Smith Mountains Quadrangle (Detterman, 1976a).

Where traced to the northeast, basal beds of the Joe Creek Member contain a progressively younger fauna that suggest deposits of the Echooka Formation onlapped to the northeast onto the underlying Lisburne Group carbonates (Detterman and others, 1975). In the Flood Creek area, an Early Permian (Wolfcampian) faunal assemblage is present within the Joe Creek Member. To the northeast, on Kemik Creek, late Early Permian and early Late Permian (Leonardian to early Guadalupian) brachiopods can be found in the Joe Creek Member. A still younger brachiopod fauna of early Late Permian (Guadalupian) age are present within the member farther to the north. The Guadalupian assemblage of the Joe Creek Member is correlative with the lower faunal zone of the overlying Ikiakpaurak Member; the upper faunal zone of the Ikiakpaurak Member, also of early Late Permian age, contains a different, less prolific fauna.

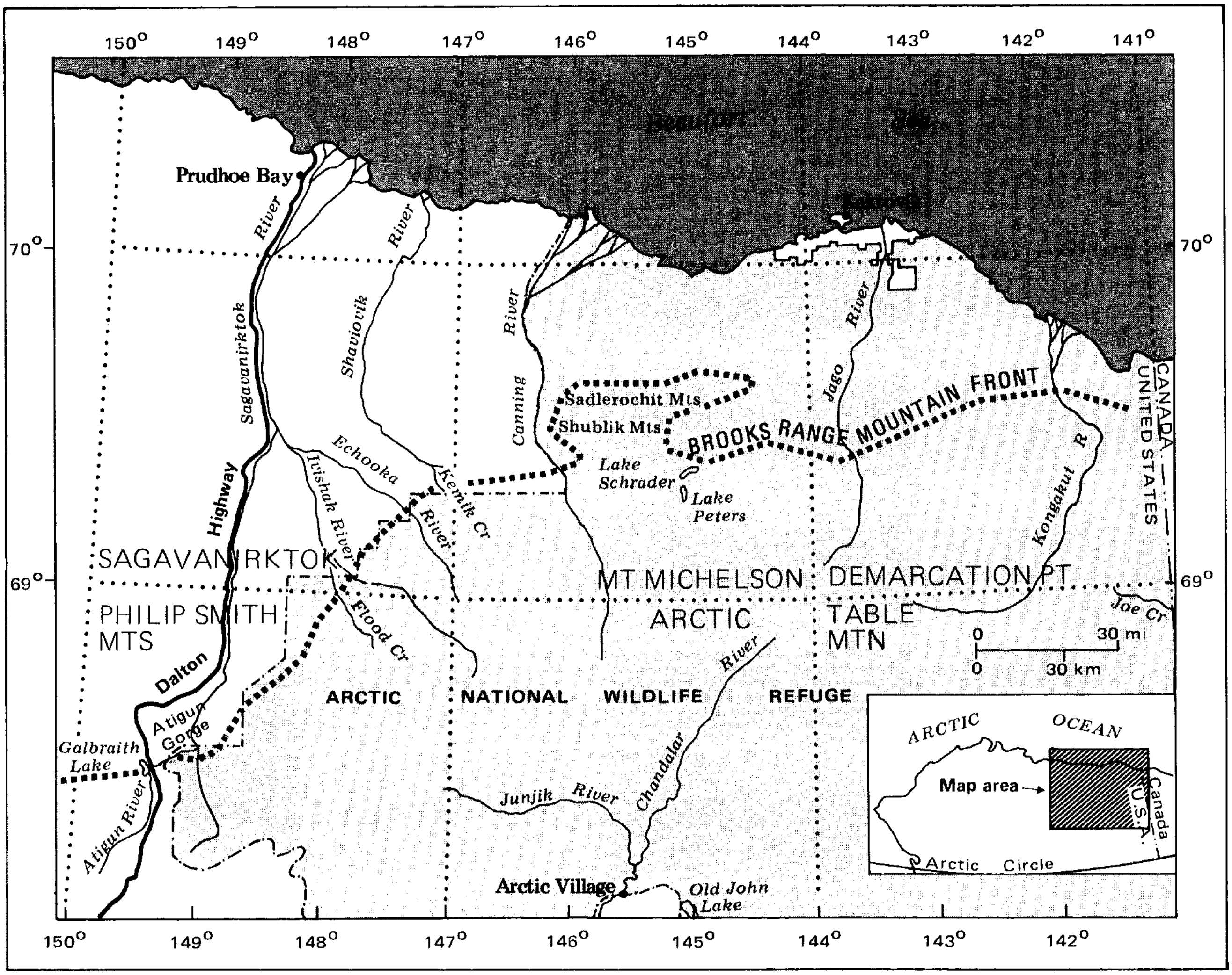

Figure 202. Index map of northeastern Brooks Range.

STRATIGRAPHY OF THE PERMIAN SEQUENCE IN THE ATIGUN GORGE AREA

Atigun Gorge is bounded on the south by an overturned sequence of Mississippian through Triassic rocks and on the north by north-dipping Triassic through Cretaceous strata (Brosgé and others, 1979a). A well-exposed section (fig. 203) of Permian rocks is present along a small tributary stream on the south side of Atigun Gorge 3 mi (4.8 km) northeast of the Atigun River bridge (Mile 270.7). At this locality, the tributary stream cascades over limestone of the Lisburne Group, forming a 75-ft-high (25 m) waterfall. The contact of the Lisburne Group with the Permian strata is exposed near the base of the waterfall and can be traced northeastward up a steep hillside. The section described below is located on the hillside 150 to 300 ft (50 to 100 m) east of the waterfall.

In the Atigun Gorge area, the Permian sequence is tectonically thickened. The fault slivering, however, can be recognized by repeated occurrences of a conspicuous orange- and tan-weathering unit typical of the base of the Echooka and Siksikpuk Formations. To ensure a continuous, unfaulted interval, the section described in this report was measured from the stratigraphic top of the Lisburne Group northward to the first structural repetition of the orange- and tan-weathering horizon (fig. 204). As a result, our section totals only 438 ft (133 m) thick. The uppermost deposits of the Permian sequence may have been truncated by the fault, and thus might be missing from the top of this section.

At the base of the measured section there is no indication of faulting. The contact with the underlying Lisburne Group forms a sharp, undulatory surface that is considered a disconformity. Dark, cherty biosparite beds adjacent to the contact contain fist-sized brachiopods of Early Pennsylvanian age (J.T. Dutro, Jr., written commun., 1985). The main elements of the fauna include Buxtonia sp., Antiquatonia sp., Linoproduclus sp., Actinoconchus sp., and Krotovia? sp.

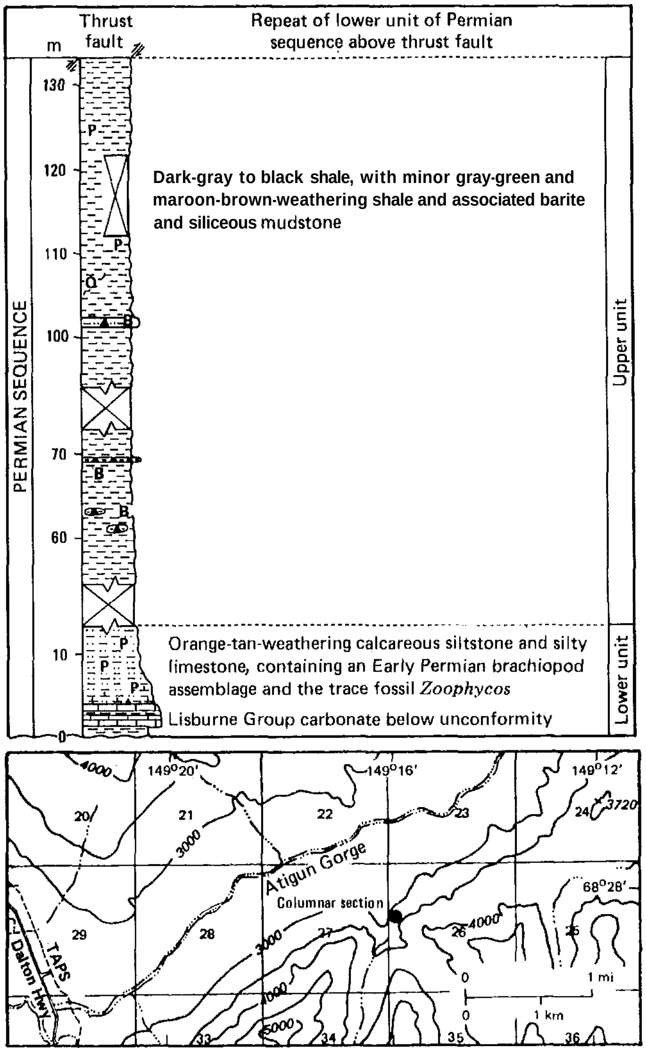

Figure 203. Generalized columnar section of Permian sequence in Atigun Gorge area. Lithologic symbols: B, barite; P, pyrite; Q, quartz.

LOWER UNIT

The base of the Permian sequence is an orange- and tan-weathering unit (43 ft; 13 m) composed of calcareous siltstone and silty limestone, with minor siliceous mudstone and interbedded shale (fig. 203). The beds dip 60° S., are overturned, laterally continuous, and range in thickness from 2 to 28 in. (5 to 70 cm). More resistant, cliff-forming strata dominate the lower half of the basal unit (fig. 205). Abundant pyrite is present throughout the interval as ellipsoidal concretions up to 8 in. (20 cm) long by 4 in. (10 cm) wide and is responsible for the orange- and tan-weathering character. In many places, the pyrite is associated with fracture-filling barite.

Horizontal and vertical burrows (0.4 to 0.6 in. [1 to 1.5 cm] diam) and the trace fossil Zoophycos are found throughout the basal unit. An abundant assemblage of shelly fossils is also present, including the gastropod Straparollus (Euomphalus) alaskensis Yochelson and Dutro, 1960, and several genera of brachiopods: Spiriferella sp., Martinia (large species), Anidanthus sp., Calliprotonia? sp., and Orulgania? sp. J.T. Dutro, Jr. (U.S. Geological Survey), and R.C. Allison (University of Alaska Fairbanks) identified the collections and concur that the fauna is early Early Permian (Wolfcampian) in age, similar to 'Brachiopod fauna E' from the Jungle Creek Formation in the Yukon Territory, Canada (Bamber and Waterhouse, 1971). Dutro (written commun., 1984) further stated that the collections are similar to those in the Cobblestone Creek area and may represent the same shell bed.

UPPER UNIT

Part of the overlying 395 ft (120 m) of section are either tundra or rubble covered (fig. 203), but the best exposed intervals consist of dark-gray to black, noncalcareous, well-indurated shale that weathers orange brown or occasionally gray green and maroon brown (fig. 206). The gray-green- and maroon-brown-weathering shales are most often associated with minor siliceous mudstone that occurs as lenses up to 8 in. (20 cm) wide by 6.5 ft (2 m) long or as beds that pinch and swell laterally. Barite occurs as cross-cutting veins or radiating clusters and, in a few instances, has replaced poorly preserved radiolarian tests that Sellars (1981) found concentrated in the siliceous nodules.

The radiolarians are extensively recrystallized and do not indicate a specific age, although they do show affinities to the genus Staurosphaera and the subfamily Liosphaerinae that range from Ordovician to Recent time (Sellars, 1981). No diagnostic megafossils have been recovered from the upper shale unit. Sellars (1981), however, described a siltstone bed, about 2,270 ft (690 m) above the basal contact with the Lisburne Group, that contains a sparse brachiopod fauna. Allison identified the only well-preserved specimen from this bed as Septospirifer sp. and considered it to be of middle to late Early Permian (Leonardian) age (Sellars, 1981). Because of faulting, the stratigraphic position of this fossiliferous bed relative to the measured section is unknown. On the basis of available fossil evidence, the age of the section at Atigun Gorge may extend from early Early Permian (Wolfcampian) at its base to late Early Permian (Leonardian) near its top.

DEPOSITIONAL ENVIRONMENT

The overall fining-upward sequence at Atigun Gorge documents sedimentation on a submergent carbonate platform in a quiet-water, middle-shelf environment. The most reliable criterion for recognizing the sequence as a shallow-marine deposit is the presence of Permian brachiopods in the siltstone and limestone beds of the lower unit. According to Menzies and Imbrie (1958), brachiopods are not known to have lived in water depths >660 ft (200 m) during Paleozoic time. Because brachiopods are sessile, benthic organisms that lack the

Figure 204. View of south side of Atigun Gorge, showing inverted shale and siltstone of Permian sequence (Ps) and stratigraphically underlying carbonates of Lisburne Group (lPMl). Light-colored bands in lower shaly slope are fault repetitions of lower unit of Permian sequence. Near vertical line marks location of columnar section shown in figure 203. Photograph by C.G. Mull, July 1982.

mobility to stay at the surface of the seafloor during periods of rapid sedimentation, their presence also indicates slow sedimentation rates and the absence of strong currents. The presence of horizontal burrows and the trace fossil ***Zoophycos*** in the lower unit is further evidence of a slow sedimentation rate and quiet-water environment. Although the entire lower unit was thoroughly bioturbated, which destroyed most primary sedimentary structures, a few siltstone beds are ripple laminated at the top and contain layers of disarticulated brachiopods. These features suggest occasional deposition by waning, storm-generated currents between longer quiescent periods. The overlying shales that compose most of the section probably represent suspension deposits that accumulated in a deeper water environment.

COMPARISON OF THE PERMIAN SEQUENCE AT ATIGUN GORGE WITH THE SIKSIKPUK AND ECHOOKA FORMATIONS

As discussed above, the Permian sequence in Atigun Gorge has characteristics of both the Siksikpuk and Echooka Formations. The orange- and tan-weathering basal siltstone unit at Atigun Gorge, containing an abundant brachiopod assemblage and nodular pyrite, resembles unit A of the Siksikpuk Formation. The basal unit at Atigun Gorge is much more calcareous, however, and contains silty limestone beds not found in the Siksikpuk Formation. Furthermore, the brachiopod assemblages at Atigun Gorge and Cobblestone Creek are more similar to the fauna found in the Echooka Formation than to the fauna present in the Siksikpuk Formation (J.T. Dutro, oral commun., 1985). In addition, bedding planes at Atigun Gorge and Cobblestone Creek are covered with the trace fossil ***Zoophycos***, which, in northern Alaska, has been identified only in the Echooka Formation. The basal unit of the Joe Creek Member of the Echooka Formation is a yellow-weathering calcareous siltstone and silty limestone that contains the trace fossil ***Zoophycos*** and a brachiopod fauna similar to the assemblage at Atigun Gorge. These data suggest that the orange- and tan-weathering horizon at

Figure 205. Resistant, orange-tan-weathering limestone beds in lower unit of Permian sequence at Atigun Gorge. Beds contain an Early *Permian* brachiopod assemblage, the gastropod Straparollus *(Euomphalus)* alaskensis Yochelson and Dutro, 1960, and the trace fossil Zoophycos.

Atigun Gorge and the similar horizon in the Cobblestone Creek area are extensions of the basal unit of the Joe Creek Member of the Echooka Formation.

The upper dark-gray to black shale unit at Atigun Gorge contains minor amounts of gray-green- and maroon-brown-weathering shale associated with barite and siliceous mudstone nodules similar to lithologies in unit B of the Siksikpuk Formation. The upper unit at Atigun Gorge is generally much darker and thinner bedded, however, than the variegated massive shale and siltstone typical of unit B. Alternatively, Detterman (1976b) has mapped the dark-gray to black shale interval as the Kavik Member of the Ivishak Formation, which is a black shale and siltstone unit that directly overlies the Echooka Formation in the northeastern Brooks Range. Superficially, the upper interval at Atigun Gorge does resemble the Kavik Member, but the Kavik contains an ammonite-pelecypod fauna of Early Triassic (late Griesbachian) age that has not been identified in the Atigun Gorge area. The oldest Triassic fossils found in the Atigun Gorge area are Early Triassic (Smithian) ammonites from the shale member of the Otuk Formation on the north side of Atigun Gorge (Bodnar, 1984). The section measured by Sellars (1981) at Atigun Gorge reportedly underlies the Otuk Formation and contains a late Early Permian (Leonardian) fauna. Even though Sellars' section is structurally complicated, it seems likely that the entire sequence that underlies the Otuk Formation and overlies the Lisburne Group is Permian in age, as suggested by Sellars. If this is the case, then the upper dark-gray to black shale interval at Atigun Gorge and the same stratigraphic sequence at Cobblestone Creek are probably an unnamed distal equivalent of the upper part of the Joe Creek Member and possibly the Ikiakpaurak Member of the Permian Echooka Formation. If the Triassic Kavik Member of the Ivishak Formation is present in the Atigun Gorge area or at Cobblestone Creek, it is very condensed and poorly exposed.

Ongoing studies involve a more detailed comparison of Permian strata at Atigun Gorge and Cobblestone Creek with the Sadlerochit Group in the Mount Doonerak area, 50 mi (80 km) southwest of Atigun Gorge (Mull and others, chap. 14). At Bombardment Creek, in the Doonerak area (fig. 201), the Sadlerochit Group consists of a basal unit of yellow- and orange-weathering calcareous siltstone and very fine grained sandstone overlain by a unit of dark-gray to black shale. The yellow- and orange-weathering unit at Bombardment Creek is much thicker and coarser grained than the comparable unit at Atigun Gorge and Cobblestone Creek. The overall sequence, however, is more similar to sections at Atigun Gorge and Cobblestone Creek than to more northeasterly exposures of Permian strata. Because the basal unit is thicker and coarser grained, the Sadlerochit Group at Doonerak appears to have been deposited closer to a northern paleoshore than were coeval strata at Atigun Gorge and Cobblestone Creek. This interpretation fits well with palinspastic restorations of Mull and Tailleur (1977), Mull (1982), and Mull and others (chap. 14) that show the Endicott Mountains allochthon, containing the Atigun Gorge and Cobblestone Creek sections, in a position south of the autochthonous or parautochthonous section in the Doonerak area. Ongoing petrographic and micropaleontologic studies are expected to clarify these paleogeographic relations.

Figure 206. View of south side of Atigun Gorge, showing dark-gray to black shale of upper unit of Permian sequence. Cross-cutting veins of barite at left are 2 to 4 in. (5 to 10 cm) thick.

CHAPTER 18.

STRATIGRAPHY OF THE OTUK FORMATION AND A CRETACEOUS COQUINOID LIMESTONE AND SHALE UNIT, NORTHCENTRAL BROOKS RANGE

By D.A. Bodnar[1]

INTRODUCTION

The Triassic and part of the Jurassic systems in the northcentral Brooks Range are represented by rocks of the Otuk Formation. The overlying lowest Cretaceous (early Neocomian) strata are represented by a coquinoid limestone and shale unit[2] and the Okpikruak Formation (Siok, chap. 19). Bodnar (1984) measured and studied 11 sections of the Otuk Formation along the northern front of the central and western Endicott Mountains in the Killik River, Chandler Lake, and western Philip Smith Mountains Quadrangles (fig. 207). The sections are part of a discontinuously exposed, east-west-trending belt of Otuk on the Endicott Mountains allochthon, the lowest allochthon in the northcentral Brooks Range.

The most accessible exposures of the Otuk Formation and the overlying coquinoid limestone and shale unit are in Atigun Gorge, 3 mi (5 km) east of the Dalton Highway (fig. 208). The discussion of these strata is based on personal observation and on reports published by the U.S. Geological Survey. Correlation of the Otuk Formation is extrapolated from lithofacies trends documented in this study and the work of Smith and Mertie (1930), Keller and others (1961), Detterman and others (1963), Chapman and others (1964), Patton and Tailleur (1964), Silberling and Patton (1964), Silberling (1970), Detterman (1970; 1974; 1976a, b; 1984a, b), Tourtelot and Tailleur (1971), Detterman and others (1975), Detterman and Dutro (1977), Pessel and others (1978), Dutro (1981), Mull and others (1982), and Balkwill and others (1983).

REGIONAL STRATIGRAPHIC SETTING

The Otuk Formation crops out in the De Long and Endicott Mountains in the central and western Brooks Range thrust belt. Because the formation consists primarily of thin (230 to 420 ft; 70 to 130 m), incompetent, discontinuous shale, it is generally highly deformed and poorly exposed. The Otuk Formation and underlying Siksikpuk Formation (Adams and Siok, chap. 17) compose the Etivluk Group of Mull and others (1982), which is part of the Ellesmerian sequence defined by Lerand (1973) for Arctic North America. Throughout this region, the Etivluk Group and overlying coquinoid limestone and shale unit are dwarfed by the older, much thicker and more competent, Lisburne Group carbonates (Armstrong and Mamet, chap. 16) and Kanayut Conglomerate (Moore and others, chap. 15).

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[2]This unit in the central Brooks Range is correlative with the upper part of the Ipewik Formation of the western Brooks Range. but because the unit is very thin along the Endicott Mountains front, the nomenclature has not been adopted in the central Brooks Range.–ED.

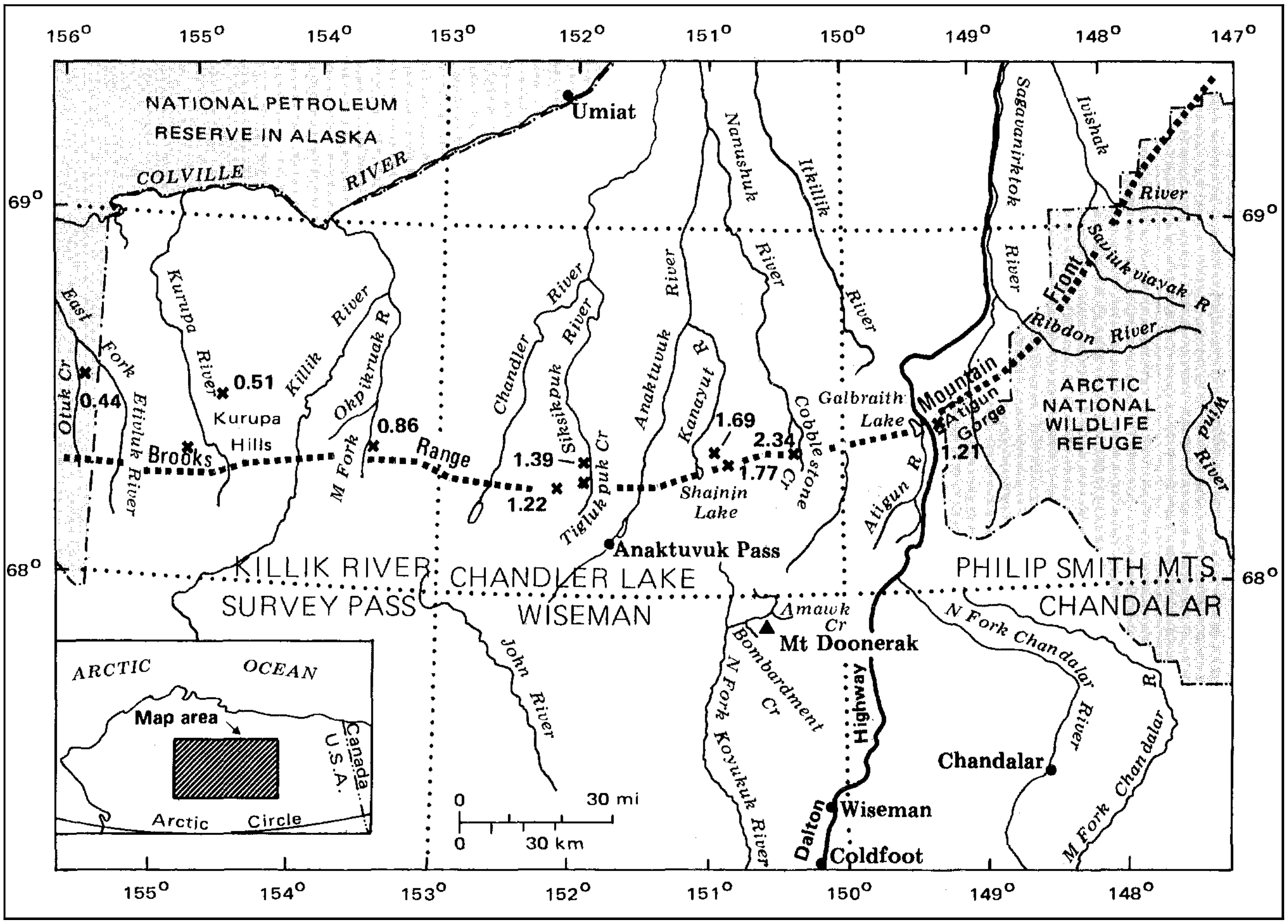

Figure 207. Index map of northcentral Brooks Range, showing location of columnar sections (crosses) of Otuk Formation measured by Bodnar (1984). Associated numbers are vitrinite reflectance values (R_O percent) determined from shale and limestone samples.

STRATIGRAPHY OF THE OTUK FORMATION

NOMENCLATURE

The Otuk Formation includes rocks that were previously assigned to the Shublik Formation; however, the type section of the Shublik Formation (Leffingwell, 1919), located on Shublik Island in the Canning River, about 125 mi (200 km) northeast of the northcentral Brooks Range, contains sandstone and no chert, unlike the Otuk Formation. Lithologic differences between Triassic rocks of the northcentral and northeastern Brooks Range have been discussed by Smith and Mertie (1930), Keller and others (1961), Patton and Tailleur (1964), Silberling (1970), and Mull and others (1982). Although there are lithologic similarities, the Shublik and Otuk must be considered different formations.

The type section of the Otuk Formation is on Otuk Creek in the western Killik River Quadrangle (Mull and others, 1982; Bodnar, 1984). The Otuk has been subdivided (Mull and others, 1982) into four members: 1) the Lower and Middle Triassic shale member, 2) the Middle and Upper Triassic chert member, 3) the Upper Triassic limestone member, and 4) the Jurassic Blankenship Member (fig. 209). This paper proposes that a fifth member, the Upper Triassic Karen Creek Member (Bodnar, 1984), be recognized in the eastern part of the northcentral Brooks Range, in a stratigraphic position between the limestone member and the Blankenship Member.

LITHOLOGY

The Otuk Formation consists of thin, rhythmically interbedded black shale, chert, and siliceous limestone (fig. 209). Both the shale member and the Blankenship Member consist dominantly of black, organic-rich shale.

Figure 208. Otuk Formation in Atigun Gorge. Unit is generally intensely deformed by folding and faulting. Middle heavy line marks approximate position of probable folded-thrust contact. Geologic units: Pe, Permian Echooka Formation; JTR o, Jurassic to Triassic Otuk Formation, Kc, Cretaceous (Valanginian) coquinoid limestone and shale unit; Ko, tectonically disrupted Cretaceous (Neocomian) Okpikruak Formation; Kt, Cretaceous (Albion) Torok Formation; Kf, Cretaceous (Albian) Fortress Mountain Formation; cht, tectonic block of light-green-gray chert 30 ft (10 m) diam. Photograph by C.G. Mull. August 1987.

The shale member contains minor gray-weathering, fossiliferous limestone concretions to 2.5 ft (0.75 m) diam and limestone beds to 1.3 ft (0.4 m) thick. The chert member consists of thin beds of black or gray-green chert (silicified mudstone) rhythmically interbedded with black, calcareous shale. The limestone member consists of interbedded sublithographic limestone, weathering tan or buff, and black or gray-green shale. Both the limestone and chert members are richly fossiliferous, containing coquinoid layers of Middle and Late Triassic pectinacid bivalves. The lower Blankenship Member contains some thin (2 to 6 in.; 5 to 15 cm) interbeds of pelecypod-bearing black limestone or chert.

In the Shainin Lake area, along the eastern Endicott Mountains front, the Otuk Formation contains a distinct thin (0.6 to 6 ft; 0.2 to 1.75 m), discontinuous siltstone horizon. A similar siltstone, correlated with the Karen Creek Sandstone (Reed, 1968) of the northeastern Brooks Range, has been reported 45 mi (70 km) south of the mountain front on Bombardment Creek near Mount Doonerak (Mull, 1982; K.E. Adams, oral commun., 1984). The siltstone on Bombardment Creek overlies rocks of Middle and Late Triassic age that are similar to the Shublik Formation of the northeastern Brooks Range.

I suggest that the distinct siltstone in the Otuk Formation along the eastern Endicott Mountains front is also correlative with the Karen Creek Sandstone of the northeastern Brooks Range. In the Endicott Mountains, however, the Karen Creek should be lowered to member rank because it does not meet the mappability criteria for a formation.

LITHOLOGIC TRENDS

From west to east across the northern front of the Brooks Range, the amount of chert in the Otuk Formation decreases, and coarser clastic components increase. These trends were noted by Smith and Mertie (1930), Patton and Tailleur (1964), and Mull and others (1982) and are better documented by the measured sections of Bodnar (1984). From descriptions of Triassic rocks in the Arctic National Wildlife Refuge (Keller and others, 1961; Silberling and Patton, 1964; Tourtelot and Tailleur, 1971; Detterman, 1974, 1976a, b; Detterman and others, 1975; Detterman and Dutro, 1977), it appears that lithofacies trends in the Otuk Formation continue northeastward, where rocks of the Otuk may grade into coeval lithologies of the more proximal

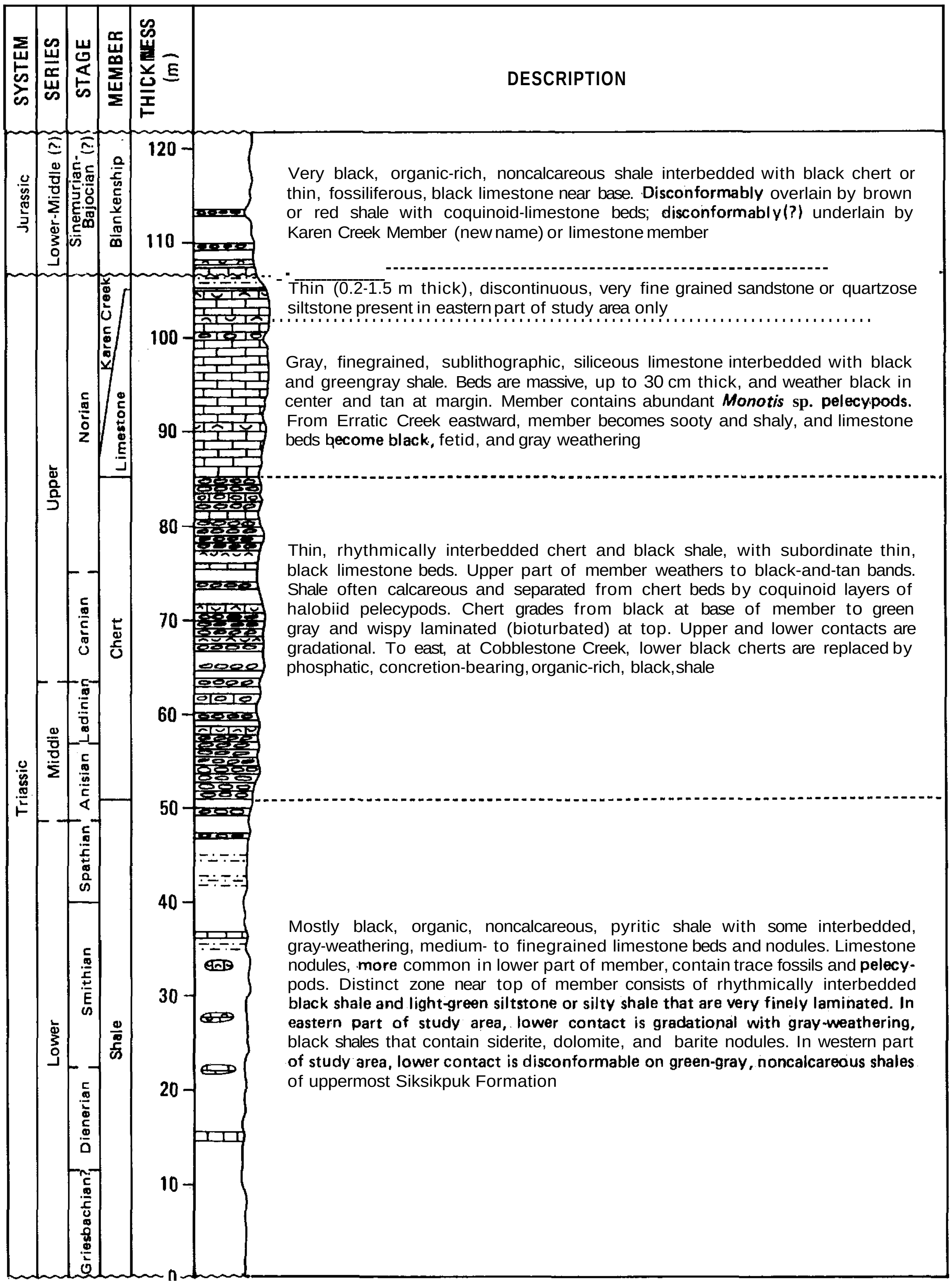

Figure 209. Generalized columnar section of Otuk Formation on Endicott Mountains allochthon, north-central Brooks Range. Modified from Bodnar (1984).

Ivishak Formation, Shublik Formation, Karen Creek Sandstone, and Kingak Shale. Silberling and Patton (1964) noted that this gradational relationship may be severely complicated by structural juxtaposition.

THICKNESS

A complete exposure of the Otuk Formation is unknown, but measured stratigraphic sections and their relation to nearby underlying and overlying strata suggest that the formation ranges from 260 to 425 ft (80 to 130 m) thick. The best known exposure of the Otuk Formation is on the south flank of the Tiglukpuk Creek anticline in the central Chandler Lake Quadrangle (Bodnar, 1984), 80 mi (130 km) west of the Dalton Highway. This section contains all members and contacts of the Otuk Formation, although the upper contact with overlying Lower Cretaceous rocks is not well exposed.

CONTACTS

The upper and lower contacts of the Otuk Formation are exposed in a few isolated localities and appear to be disconformable, separating differently colored shales that parallel bedding. The upper contact is exposed in a stratigraphic section in Atigun Gorge and separates brown shale of the Cretaceous coquinoid limestone and shale unit from a 0.6-ft-thick (0.2 m) zone of soft, mottled, yellow-and-gray clay that overlies the black, organic shale of the uppermost Blankenship Member of the Otuk Formation. The lower contact of the Otuk Formation is a similar disconformity exposed at Tiglukpuk Creek. I interpret the yellow-and-gray clay zones as paleosols.

Contacts between the shale and chert members and the chert and limestone members of the Otuk are gradational; contacts between the limestone member, Karen Creek Member, and Blankenship Member of the Otuk are sharp. The gradational contacts indicate that there was no significant break in Early-to-Middle and Middle-to-Late Triassic sedimentation in the area. The sharp contacts indicate regressions and transgressions near the Permian-Triassic and Triassic-Jurassic boundaries and are correlative with similar unconformities in coeval strata in the northeastern Brooks Range and subsurface of the Arctic Slope.

AGE

The Otuk Formation ranges in age from Early Triassic (late Dienerian to early Smithian or earlier Early Triassic) to Middle Jurassic (Bajocian), as indicated by the age of ammonites, pelecypods, conodonts, radiolaria, and foraminifera. Detailed paleontology of the Otuk Formation has been reported in Silberling (1970), Swain (1981), Mull and others (1982), and Bodnar (1984). The uppermost beds of the underlying Siksikpuk Formation contain early Late Permian (Guadalupian) conodonts (Bodnar, 1984), and the overlying coquinoid limestone and shale unit contains Early Cretaceous (Valanginian) pelecypods and foraminifera (Jones and Grantz, 1964; Mull and others, 1982; Bodnar, 1984). All but the lowermost (Griesbachian) and uppermost (Rhaetian) Triassic stages are represented in the Otuk Formation. The presence of earliest Triassic (Griesbachian) ammonites in shallower water rocks of the Kavik Member of the Ivishak Formation to the northeast (Silberling, 1970; Detterman and others, 1975; Detterman and Dutro, 1977) suggests that the lower, apparently unfossiliferous beds of the shale member of the Otuk Formation may be of Griesbachian age.

DEPOSITIONAL ENVIRONMENT

The Otuk Formation was deposited in a low-energy, open-marine environment, primarily below wave base. The fine-grained, unwinnowed lithologies, fauna, and sedimentary structures suggest deposition in depths that range from middle and outer neritic to inner bathyal. The thin, laterally continuous beds reflect condensed sedimentation on a relatively stable substratum. Lithofacies trends in the Otuk Formation, combined with the nature of coeval Triassic rocks to the north and northeast, indicate that the Triassic shelf shallowed to the north and northeast.

STRATIGRAPHY OF THE CRETACEOUS COQUINOID LIMESTONE AND SHALE UNIT

The coquinoid limestone and shale unit is thin (<130 ft; 40 m) and consists of red or brown clay and shale, with minor limestone interbeds to 6 ft (2 m) thick. The limestone beds are composed exclusively of the bivalve *Buchia sublaevis* (Keyserling) of early Neocomian (Valanginian) age (Jones and Grantz, 1964). This distinct lithology and fauna have been noted over a distance of 250 mi (400 km) along the north flank of the Brooks Range from Galbraith Lake in the east to the western foothills of the De Long Mountains in the west. The unit disconformably overlies the Blankenship Member of the Otuk Formation. The nature of the upper contact with overlying strata is not always clear, but, in some areas, it is a thrust fault.

TRIASSIC TO LOWER CRETACEOUS STRATA AT ATIGUN GORGE

Deformed Triassic to Lower Cretaceous rocks are exposed for about 1.25 mi (2 km) along the north side of Atigun Gorge. Figures 210 ancl 211 show a columnar section of the upper part of the Otuk Formation measured on a 50-ft (15 m) bluff on the north side of Atigun Gorge 3 mi (5 km) downstream from the Dalton Highway. This section was designated as a principal reference section of the Blankenship Member by Mull and others (1932) It contains the uppermost chert member(?), the limestone member, the Karen Creek Member, the Blankenship Member, and part of the overlying coquinoid limestone and shale unit. The upper contact, which had to be trenched, is itlcntical to the upper contact at Erratic Creek, 35 mi (56 km) to the west (Bodnar, 1984).

Fossils found in this section include Late Triassic, Early Jurassic, and Early Cretaceous bivalves; Triassic ostracods, foraminifera, and conodonts; poorly preserved Late Triassic radiolaria; and Early Cretaceous (early Neocomian) foraminifera. Particularly abundant in the limestone member of the Otuk Formation is the Late Triassic pelecypod *Monotis* sp. Some thin beds near the base of the Blankenship Member contain the small (0.5 in. [1 cm] diam) Early Jurassic pelecypod *Otapiria tailleuri*. Slabs of red-brown-weathering coquina, composed of the Early Cretaceous pelecypod *Buchia sublaevis*, are conspicuous in many places in Atigun Gorge. About 1 mi (1.5 km) upstream from the measured section, the shale member of the Otuk Formation contains Early Triassic (early Smithian) ammonites (Brosgé and others, 1979a).

REGIONAL CORRELATIONS

The Etivluk Group in the northcentral Brooks Range is correlative with the Sadlerochit Group (Lower Permian to Lower Triassic), Shuhlik Formation (middle Upper Triassic), Karen Creek Sandstone (Upper Triassic), and lower part of the Kingak Shale (Lower Jurassic to Middle Jurassic) in the northeastern Brooks Range (fig. 212). The same units are present in the subsurface of the Arctic Slope, but there the Karen Creek Sandstone is known as the Sag River Sandstone (North Slope Stratigraphic Committee, Alaska Geological Society, 1970; Jones and Speers, 1976).

Members of the Otuk Formation can be correlated

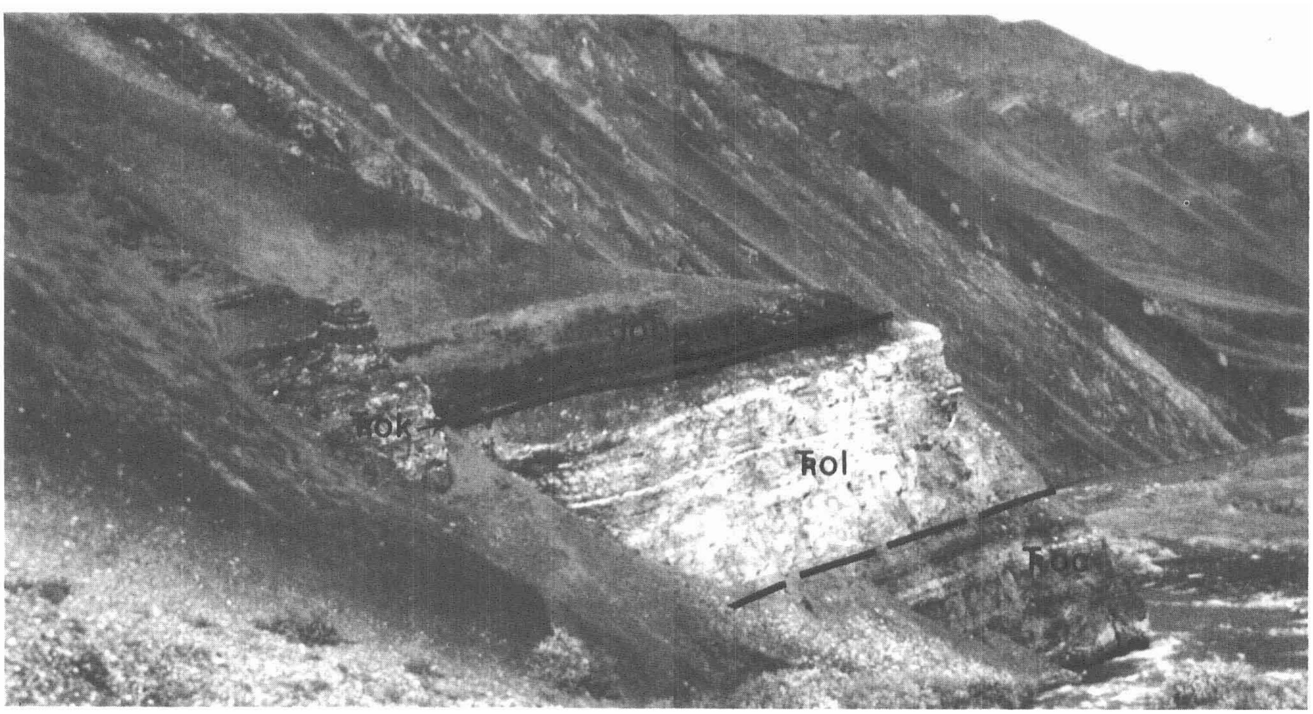

Figure 210. Upper part of Otuk Formation, north side of Atigun Gorge. Lenticular, 1-ft-thick (30 cm) bed of Karen Creek Member separales limestone member from Blankenship Member. Cretaceous coquinoid limeslone is present up shaly slope lo left. Geologic units: TRoc, Upper and Middle Triassic chert member; TRol, Upper Triassic limeslone member; TRok, Upper Triassic Karen Creek Member; Job, Jurassic Blankenship Member. Photograph by C.G. Mull, Augusl 1984.

with coeval rocks in the northeastern Brooks Range. The organic-rich, marine-shelf muds of the shale member (Lower and Middle Triassic) of the Otuk Formation are the distal equivalent of the thicker, more proximal, prodelta muds of the Kavik Member, the fluvial-deltaic sands of the Ledge Sandstone Member, and the marine silts of the Fire Creek Siltstone Member of the Ivishak Formation (Lower Triassic). Similarly, the chert and limestone members (Middle and Upper Triassic) of the Otuk Formation are the condensed, deeper water equivalents of the Shublik Formation (Middle and Upper Triassic), which consists of siltstone, limestone and dolomite, and clay-shale members. The Karen Creek Member (Upper Triassic) of the Otuk Formation is correlative with the Karen Creek Sandstone (Upper Triassic) of the northeastern Brooks Range. The Karen Creek Member is much thinner than the Karen Creek Sandstone to the northeast, however, representing the distal parl of a clastic influx during a regressive-transgressive episode in latest Triassic time. The Blankenship Member (Lower and Middle? Jurassic) of the Otuk Formation is correlative with the lower part of the Kingak Shale (Lower to Middle Jurassic) but is much thinner (26 to 65 ft; 8 to 20 m) than the Kingak Shale (160 to 3,900 ft; 50 to 1,200 m) and probably represents only part of the Jurassic System. Collectively, the coeval units are about five times as thick (1,950 ft; 600 m) as the Otuk Formation.

ECONOMIC GEOLOGY

The hydrocarbon source-rock potential of the Otuk Formation was determined by analyzing 48 samples from the formation. Average total-organic-carbon (TOC) value for the Otuk Formation was 2.77 weight percent, which is well above the accepted TOC value for source rocks. The Blankenship Member yielded the highest average TOC content (6.86 weight percent). A visual kerogen analysis of the Otuk Formation showed that the dominant kerogen is amorphous; vitrinite is next abundant, followed by subordinate amounts of alginite, exinite, and inertinite. Amorphous organic material is considered oil prone by many authors (Tissot and others, 1974) and, in this sense, the Otuk Formation may be a source of liquid hydrocarbons.

Thermal maturity of the Otuk Formation was determined by vitrinite reflectance (R_o), conodont color alteration (CAI), thermal-alteration index (TAI), and temperature of pyrolosis (T_{max}). Thermal maturity generally increases from west to east across the study area (fig. 207) and southward toward the Brooks Range mountain front (Bodnar, 1984). To the west, the section at Otuk Creek (R_o = 0.44 percent) is thermally immature; to the east, the section at Cobblestone Creek (R_o = 2.34 percent) is overmature. An increase in maturity at or near the mountain front is compatible with the concept of Mull (1982), Mayfield and others (1983), and Dillon (chap. 10) of isostatic rebound and unroofing of the Brooks Range during Cretaceous time.

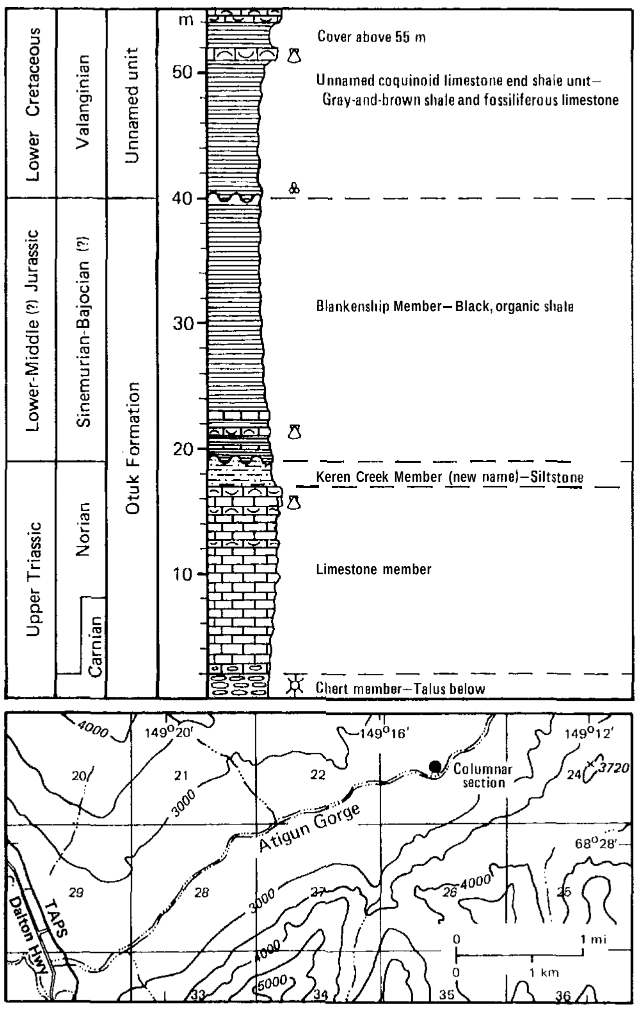

Figure 211. Generalized columnar section of *upper part of* Otuk *Formation*, north side of *Atigun* Gorge. Fossil symbols: ✕, pelecypod; ✕, *radiolarian;* ✕, foraminifer. Modified from Bodnar (1984).

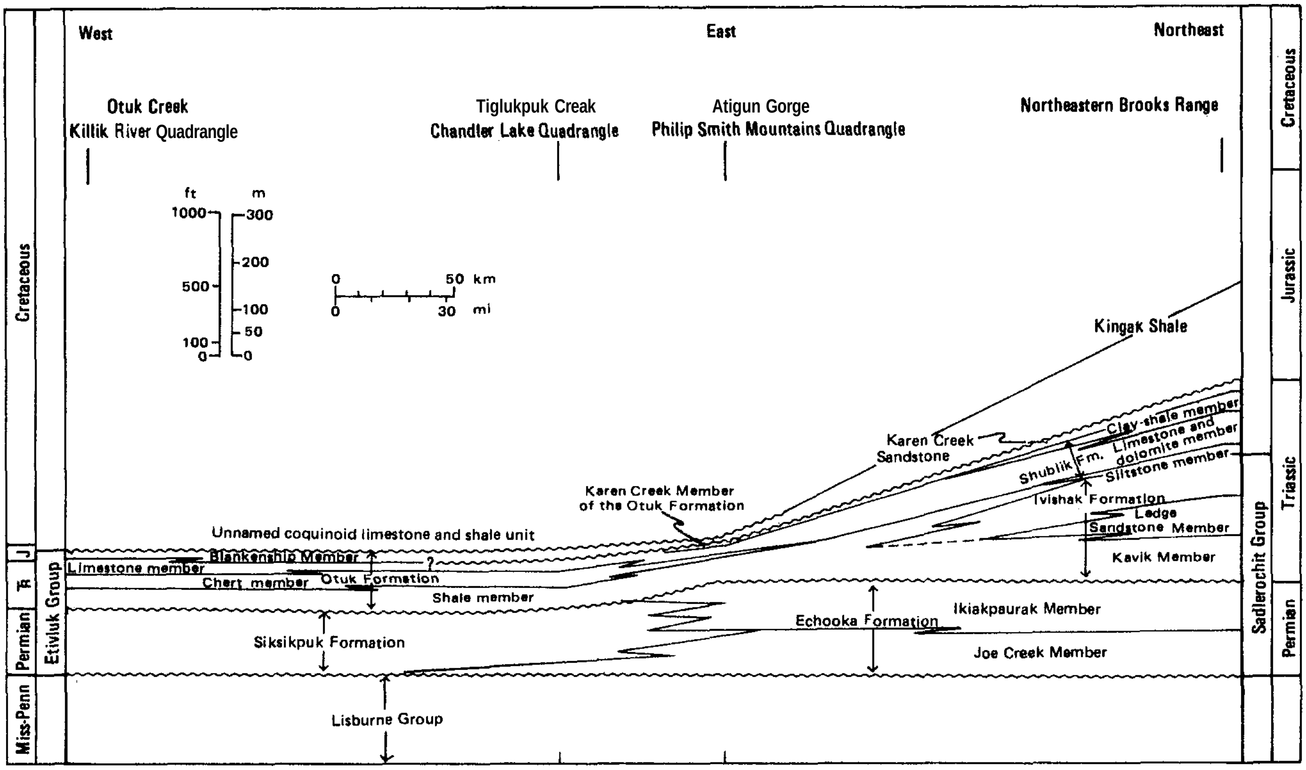

Figure 212. Regional correlation of Permian, Triassic, Jurassic, and basal Cretaceous strata from northcentral to northeastern Brooks Range.

SUMMARY

Triassic through Jurassic strata of the eastern Endicott Mountains and western Philip Smith Mountains consist of fine-grained marine rocks of the Otuk Formation. These rocks disconformably overlie Permian rocks (Echooka and Siksikpuk Formations) and disconformably underlie Lower Cretaceous rocks (coquinoid limestone and shale unit). From west to east across the thrust belt of the northcentral Brooks Range, the lithofacies of the Otuk Formation increasingly resemble lithofacies of the Shublik Formation in the northeastern Brooks Range. This trend indicates that the Otuk Formation is a greatly condensed, deeper water, distal equivalent of Triassic to Jurassic rocks in the northeastern Brooks Range and Arctic Slope subsurface.

CHAPTER 19.

STRATIGRAPHY AND PETROLOGY OF THE OKPIKRUAK FORMATION AT COBBLESTONE CREEK, NORTHCENTRAL BROOKS RANGE

By J.P. **Siok**[1]

INTRODUCTION AND REGIONAL RELATIONS

The Okpikruak Formation was named by Gryc and others (1951) for a Lower Cretaceous sequence of rhythmically interbedded lithic sandstone, siltstone, and shale exposed along the Okpikruak River, 110 mi (175 km) west of the Dalton Highway (fig. 213). Along the northern flank of the Endicott Mountains in the central Brooks Range, the formation overlies the Triassic to Jurassic Otuk Formation or a Cretaceous coquinoid limestone and shale unit at what appears to be a thrust contact. In most areas, the top of the Okpikruak Formation is marked by a Holocene erosional surface or a thrust fault but is sometimes disconformably overlain by the Albian Torok or Fortress Mountain Formations (figs. 34 and 208). The major belt of Okpikruak is on allochthons of the central and western Brooks Range thrust **belt.**[2] The best outcrops are found in active stream-cut banks; fair exposures are common on **rubble**-covered ridges.

A complete section of the Okpikruak Formation is not known, but its estimated thickness is 1,800 to 2,200 ft (550 to 670 m) (**Patton** and Tailleur, 1964). Although **Brosgé** and others (**1979a**) included a unit of coquinoid limestone and shale (Bodnar, chap. 18) with the base of the Okpikruak Formation, this discussion refers only to the coarser **clastic** sediments of the formation. Detailed sedimentologic and facies analyses of the Okpikruak Formation have not been made.

Fossils are rare in the Okpikruak Formation, but various species of ***Buchia*** have been recovered. Analysis of fauna at Tiglukpuk Creek (60 mi [**100** km] west of Dalton Highway) indicates an Early Cretaceous (Berriasian) age (Jones and Grantz, 1964). Pelecypods collected by K.E. Adams and C.G. Mull at Cobblestone Creek and nearby May Creek include ***Buchia sublaevis*** Keyserling and ***Buchia crassicolis solida*** (Lahusen) of Early Cretaceous (early to late Valanginian) age (J.W. Miller, oral commun., 1986). In the western Brooks Range, the Okpikruak Formation contains fossils of Berriasian and Valanginian age (Imlay, 1961) and, on one of the higher allochthons in the De Long Mountains, fossils of Late Jurassic age (Mayfield and others, 1983).

The Okpikruak Formation is composed of the first sediments shed northward into the **Colville** basin from the Brooks Range orogenic belt. The formation therefore dates the beginning of the Brooks Range uplift and preserves a record of the drastic change from **muddy**-shelf sedimentation of the underlying Etivluk and Sadlerochit Groups to flysch sedimentation of the Okpikruak Formation.

Along the mountain front northeast of the Dalton Highway, the Okpikruak Formation grades northward into more basin-like deposits of the Kongakut Formation (**Brosgé** and others, **1979a**). The Kongakut is autochthonous or parautochthonous and is dominantly a shale-siltstone sequence that contains manganiferous nodules.

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[2] **At** Tiglukpuk Creek, thick **turbidites** of the **Okpikruak** Formation contain earliest Early Cretaceous (Berriasian) **pelecypods** and structurally overlie younger strata of the Early Cretaceous (Valanginian) coquinoid limestone and shale unit (Jones and Grantz. 1964). The lower part of the Okpikruak at this locality and others, including Cobblestone Creek and Atigun Gorge. consists of tectonically disrupted. lens-shaped bodies of graywacke interspersed in a groundmasa of sheared mudstone. These paleontological and structural data suggest that the belt of Okpikruak Formation along the Endicott Mountains front is part of a major allochthon that overlies the Endicott Mountains allochthon. In the **Killik** River Quadrangle. **200 mi** (320 **km)** west of the Dalton Highway, a major belt of **Okpik**-ruak is part of the Picnic Creek allochthon (Mull and others, **1987)**, which overlies the Endicott Mountains allochthon. Although confirming data are lacking, it is **likely** that the Okpikruak Formation at Cobblestone Creek and Atigun Gorge is also part of the Picnic Creek allochthon—ED.

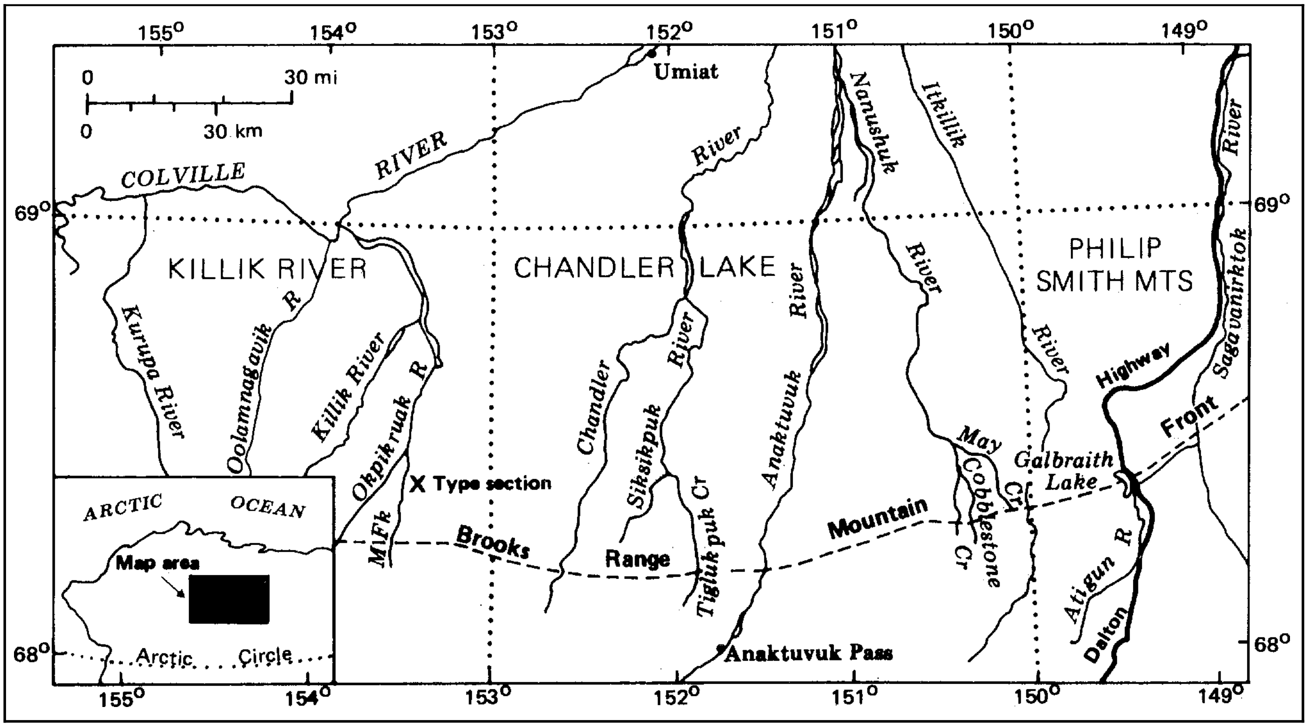

Figure 213. Index map of central and eastern Endicott Mountains, northcentral Brooks Range, showing location of type section of Okpikruak Formation.

STRATIGRAPHY AND SEDIMENTOLOGY

Along the Okpikruak River (type section), and at Cobblestone Creek, 25 mi (40 km) west of the Dalton Highway, the Okpikruak Formation is rich in sandstone (50 to 70 percent); however, the principal reference section on Tiglukpuk Creek, 60 mi (100 km) west of the Dalton Highway, is rich in shale and poor in sandstone (15 to 25 percent). This variability seems to typify the formation throughout the Arctic Foothills, although the exact stratigraphic relations are unknown.

OKPIKRUAK FORMATION AT TIGLUKPUK CREEK

At Tiglukpuk Creek (fig. 213), gray silty shale beds (20 to 60 in; 50 to 150 cm) compose 75 percent of the section and are rhythmically interbedded with thin (2 to 4 in; 5 to 10 cm) sandstone beds, which are plane parallel and laterally continuous throughout the entire outcrop (50 to 164 ft; 15 to 50 m). Sandstone beds up to 3 ft (1 m) thick are rare and in most cases occur within a thick (6 to 10 ft; 2 to 3 m) sequence of shale. Sand to shale ratios range from 1:4 to 1:8. Bedding structures include flute casts, tool marks, graded beds, and ripple and trough cross-lamination, all consistent with deposition by turbidity currents. Complete Bouma sequences are missing, both from thinner sandstone beds and thicker ones. The dominant facies at Tiglukpuk Creek are the pelitic, arenaceous facies D or facies E, as defined by Mutti and Ricci Lucchi (1978), with various amounts of hemipelagic facies G. Upper flow regime deposition is recorded by the thicker, graded, sole-marked sandstone beds. Lower flow regime deposition is recorded by the thinner, cross-laminated sandstone beds. Suspension sedimentation is represented by thick intervals of siltstone and shale. Among ichnofossils at Tiglukpuk Creek is a complex, net-patterned grazing trace of Paleodictyon. The trace is most common in the distal areas of the Nereites ichnofacies and indicates a bathyl to abyssal, very quiet-water environment (Crimes, 1975). On the basis of these observations. the facies association at Tiglukpuk Creek can be interpreted as an outer submarine-fan deposit or overbank deposits between channels.

OKPIKRUAK FORMATION AT COBBLESTONE CREEK

Excellent exposures of the Okpikruak Formation are present along stream-cut banks just north of the mountain front at May and Cobblestone Creeks, 20 to 25 mi (32 to 40 km) west of the Dalton Highway (figs. 213 and 214). Patton and Tailleur (1964) mapped some of these exposures as Fortress Mountain Formation, but more recent detailed geologic mapping and

Figure 214. Okpikruak Formation at Cobblestone Creek, composed of interbedded sandstone, siltstone, and shale. Beds contain abundant turbidite features, including graded beds, flute and groove casts, and numerous partial Bouma sequences. Photograph by C.G. Mull, July 1985.

fossil information indicate that they are part of the Okpikruak Formation. Field relations show that these sandstone outcrops are bounded on the south by sheared shale, chert, graywacke, and coquinoid limestone that separate them from the underlying Jurassic and older rocks of the Endicott Mountains allochthon.

A continuous section, 165 m thick, was measured at Cobblestone Creek (fig. 214) and sampled for petrographic analysis. The section is dominated by very fine to coarse-grained sandstone (60 to 70 percent), with lesser amounts of shale and siltstone. At the base of the measured section is a partially exposed, very poorly sorted pebbly mudstone that contains clasts of chert, sandstone, and shale. The mudstone is chaotically mixed and assigned to facies F of Mutti and Ricci Lucchi (1978). Sandstone beds are plane parallel and continuously exposed for 66 ft (20 m) with no signs of channeling. Bedding structures include flute casts, tool marks, ripple cross-lamination, graded beds, shale rip-ups, and starved ripples. Thick, massive sandstone beds (3 to 6 ft; 1 to 2 m) contain very thin shale interbeds (2 to 4 in; 5 to 10 cm) where the sandstone is in packages of two or more; however, amalgamated sandstone beds are present near the top of the section. Thinner sandstone beds (4 to 20 in.; 10 to 50 cm) are associated with thicker shale beds (4 to 8 in; 10 to 20 cm) and, in a few cases, thick shale intervals (sand:shale = 1:3). Thicker sandstone beds are massive, graded, and contain sole markings; thinner sandstone beds are graded, ripple laminated and contain sparse shale rip-ups. Sandstone-to-shale ratios vary from 1:3 to 5:1.

The section at Cobblestone Creek is interpreted to be progradational, because the base is marked by a series of well-defined negative megasequences (thickening and coarsening upward) that grade into massive sandstone with minor shale. Negative megasequences are found in outer fan deposits (Mutti and Ricci Lucchi, 1978), as well as in overbank deposits. Complete Bouma sequences are rare at Cobblestone Creek, which suggests that the late phases of each turbidite were either sediment poor or lacked the energy to develop complete Bouma intervals. Some shale intervals contain thin (0.8 to 1.6 in.; 2 to 4 cm), lenticular sandstone beds that represent facies E of Mutti and Ricci Lucchi (1978) and are good evidence for overbank deposition. Some positive megasequences (thinning and fining upward) are recognized near the top of the section.

The Cobblestone Creek section does not fit easily into the facies model of Mutti and Ricci Lucchi (1978).

This sequence can be interpreted as a middle- to inner fan deposit or as overbank deposits in a channelized middle-fan environment. The absence of complete Bouma sequences, which would characterize a middle-fan environment, coupled with the thin, lenticular sandstone and abundant upper flow regime structures, support an overbank interpretation. Thick, massive sandstone, negative megasequences, local conglomerate, and lack of major channeling support a middle- to inner fan interpretation.

Only the middle and outer submarine-fan environments can be recognized in the Okpikruak Formation of the central and eastern Endicott Mountains. No inner fan or slope facies have been reported from this area. Such coarser facies, if present, were probably uplifted during later thrusting in the Brooks Range and redeposited as the coarse facies of the Fortress Mountain Formation (Crowder, chap. 20). In the western Endicott Mountains and De Long Mountains, however, the Okpikruak Formation contains coarse boulder conglomerate and chaotic deposits interpreted as olistostromes. These represent channel fill and submarine-landslide material, respectively (Mull and others, 1976; Mull, 1985; Crane, 1987).

PETROLOGY

A systematic point-count analysis was conducted on sandstones taken at 30- to 60-ft (10 to 20 m) intervals from the measured section on Cobblestone Creek. All thin sections were stained for potassium feldspar, which met with varying degrees of success due to albitization of the feldspar. At least 200 detrital grains were counted on each of 10 slides; 400 detrital grains were counted on each of five additional slides as part of a comparison study (Wilbur and others, 1987). Operational definitions for individual grains are those used by Decker (1985). Thirty-five different grain types were recognized.

All the samples are composed of fine- to medium-grained sand, are grain supported, and have a silty, argillaceous matrix. Carbonate occurs both as pore-filling cement and as a selective replacement for some of the grains; it is patchy and is present in all the samples. Other secondary minerals include chlorite, albite, illite, and hematite. Packing is very tight, as shown by the ductile rock fragments that have been deformed into pseudomatrix. Less than one percent primary porosity was observed, but leaching of carbonate cement and lithic grains could lead to development of secondary porosity.

TERNARY DIAGRAMS

Results of the point-count analysis (table 6) are plotted on standard ternary diagrams (figs. 215 and 216). Bivariant plots (fig. 217) were also constructed, using ratios of detrital compositions vs. time (stratigraphic position), and seem to distinguish the grain populations more clearly than the ternary diagrams. Generally, all the data on the ternary diagrams plot in a tight cluster, which suggests internal homogeneity and a dominant source terrane throughout deposition of the Okpikruak Formation. The most noticeable variation involves the relative abundance of volcanic rock fragments. A comparison study (Wilbur and others, 1987) of sandstones from the Okpikruak Formation at Cobblestone Creek with those from the Okpikruak in the Kurupa-Oolamnagavik Rivers area, 120 mi (200 km) to the west, showed two petrographically distinct populations, which suggests a variable source terrane on a regional scale.

Figure 215a shows chert plotted with the lithic grains and, when compared to figure 215b, shows a shift in the data set. The amount of shift represents the percentage of chert in each sample. In figure 215b, the data set clusters between the undissected magmatic-arc provenance field and the recycled-orogen provenance field of Dickinson and Suczek (1979), which indicates a mixed source of recycled sediment (mostly chert and quartz), coupled with some volcanic material.

Figure 215c emphasizes source rocks and shows a shift of the data toward the total rock fragments pole (Lt). The amount of displacement of the data represents the percentage of polycrystalline quartz in each sample. These samples have a high chert-to-quartz ratio. Figure 215c shows a mixed provenance also, which is indicated by points in both the magmatic-arc and recycled-orogen fields. The arc is interpreted to be undissected, based on a high ratio of volcanic (V) to plutonic (P) rock fragments. In thin section, the volcanic source is indicated by an abundance of lathwork grains, microcrystalline felsic rock fragments, and tuffaceous rock fragments. Very few plutonic rock fragments are present (<1 percent), which suggests that the arc had not been deeply eroded.

Because of limited success in staining potassium feldspar in thin section, the results are biased against potassium feldspar, as shown in figure 216a. The data cluster around 50 percent Qm, with up to 28 percent potassium feldspar. The abundance of plagioclase in sample 83ASK480 (table 6) was expected because that sample also has the highest component of volcanic rock fragments. Sample 83ASK498 (table 6) is anomalously high in quartz, possibly due to the misidentification of unstained, untwinned feldspar. Another explanation would be a quartz-rich source that periodically provided sediments to the system.

Figure 216b emphasizes lithic fragments and shows a unique grouping between the arc-orogen and collision-orogen fields. From this diagram, it appears that the source terrane had a low polycrystalline-quartz component and was dominated by sedimentary rock fragments (chert) with lesser amounts of volcanic rock fragments. The simplest interpretation of this diagram is a provenance of uplifted chert sequences, with minor sandstone interbeds derived from a distant(?) volcanic arc. The felsic nature of the volcanic rock fragments and

Table 6. Framework modes (relative percent) of selected sandstones from Okpikruak Formation, Cobblestone Creek
[Data were used to construct ternary diagrams in figures 215 and 216]

Sample no.	Detrital grains	Modal compositions†															
		Q	F	L	Q+	L-	Qm	Lt	Qm	P	K	Qp	Lv	Ls+	Lv	Ls+	Lm
83ASK472	214	14	23	63	29	48	10	67	55	37	8	8	18	74	12	79	9
83ASK474	215	15	11	74	34	55	7	82	50	31	19	14	29	58	22	69	9
83ASK474*	400	16	8	76	30	62	4	88	34	43	23	15	26	59	26	62	11
83ASK477	241	17	18	65	30	52	11	71	53	45	2	11	26	63	20	73	7
83ASK480	222	8	6	86	32	62	2	92	25	75	0	11	41	48	30	56	14
83ASK482	242	20	16	64	38	46	9	75	51	47	2	20	29	51	17	66	17
83ASK483*	401	12	8	80	22	70	6	87	50	35	15	8	29	63	28	60	12
83ASK485*	400	19	14	67	26	60	13	73	62	26	12	9	44	47	45	48	7
83ASK487	232	18	15	67	30	55	9	76	54	40	6	15	28	57	13	61	26
83ASK490*	400	27	17	56	29	54	17	66	59	28	13	16	41	43	46	48	6
83ASK492	205	14	15	71	32	53	8	77	47	50	3	10	22	68	14	72	14
83ASK495*	405	6	9	85	11	80	2	89	36	36	28	5	32	63	29	57	14
83ASK495	193	12	8	80	38	54	4	88	44	50	6	14	23	63	15	78	7
83ASK498	255	24	16	60	34	50	15	69	72	28	0	15	23	62	15	71	14
83ASK499	215	17	17	66	27	56	8	75	54	46	0	13	15	72	8	76	16

Q = Monocrystalline quartz + polycrystalline quartz
F = Potassium feldspar + plagioclase feldspar
L = Sedimentary + metamorphic + volcanic + plutonic rock fragments + chert
Q+ = Monocrystalline quartz + polycrystalline quartz + chert
L- = Sedimentary + metamorphic + volcanic + plutonic rock fragments
Qm = Monocrystalline quartz
Lt = Sedimentary + metamorphic + volcanic + plutonic rock fragments + polycrystalline quartz + chert
P = Plagioclase feldspar
K = Potassium feldspar
Qp = Polycrystalline quartz
Lv = Volcanic rock fragments
Ls+ = Sedimentary rock fragments + chert
Lm = Metamorphic rock fragments
*Samples counted by Steven Wilbur (1984).
†First seven modal compositions were used to construct ternary diagrams in figure 215; remaining nine compositions were used to construct diagrams in figure 216.

the abundance of sedimentary rock fragments (mostly chert and polycrystalline quartz) indicate a continental provenance for the chert, rather than an oceanic provenance characteristic of a subduction complex (fig. 216b). Another interpretation would be a provenance of uplifted chert and sandstone in an arc-trench system. The interbedded sandstone must have had a continental source, as indicated by polycrystalline quartz and unfoliated metaclastic grains. Subsequent erosion of the volcanic rocks of the magmatic arc and the uplifted chert and sandstone would have produced sediments of the Okpikruak Formation.

Figure 216c shows that most of the rock fragments in the Okpikruak Formation are sedimentary. Chert and cherty argillite represent 80 to 90 percent of all the sedimentary rock fragments. The metamorphic rock fragments are mostly phyllite, greenstone, and unfoliated metaclastics, which indicate a low-grade metamorphic source.

BIVARIANT PLOTS

Plots of compositional ratios vs. time are useful in determining changes in the source area, as well as compositional trends in the stratigraphic section. Figures 217a and 217b show no systematic change in the ratios of chert to quartz and chert to lithic grains (respectively), suggesting that the availability of chert in the source area was relatively constant, with variable influxes of quartz and lithic grains. Figure 217c shows a marked increase in volcanic rock fragments in samples from the lower part of the section and then a general decrease in volcanic rock fragments from samples toward the top of the section. The increase culminates with sample 83ASK480 and may represent either a period of volcanism in the basin or erosion of the volcanic arc. Figure 217d shows a slight increase in felsic rock frag-

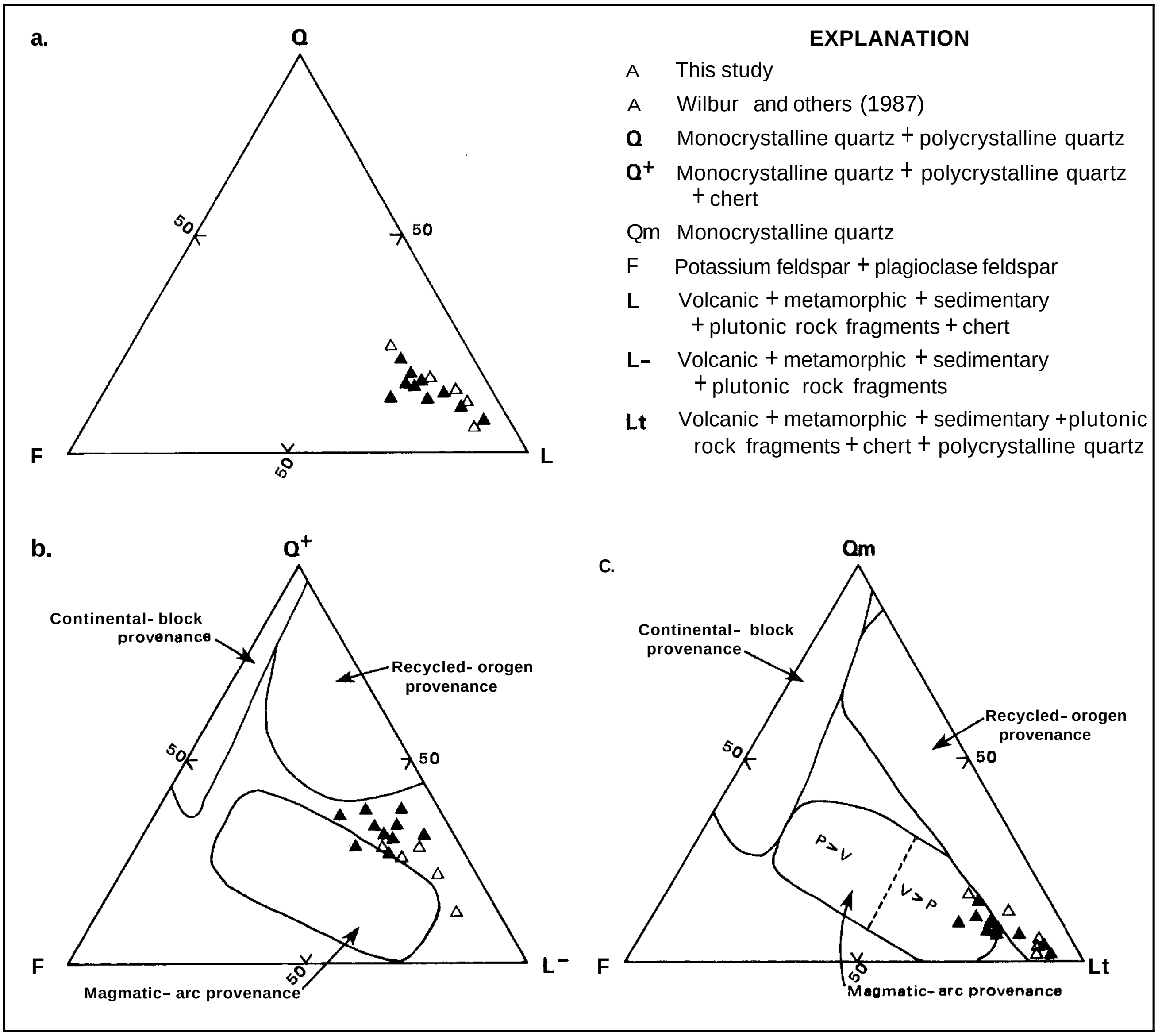

*Figure **215.** Ternary diagrams (QFL, Q^+FL^-, QmFLt) illustrating detrital grain composition of selected sandstones from Okpikruak Formation, Cobblestone Creek. Provenance fields after Dickinson and Suczek (1979).*

ments in samples from the lower part of the section, followed by a sharp drop and then a gradual increase in felsic rock fragments in samples toward the top of the section. This indicates that more mafic material was supplied to the upper part of the section and may suggest erosion of more basaltic material. The interpretations of ratios of felsic to mafic volcanic rock fragments must be viewed with caution, however, because the ratios represent <10 percent of the total grain population.

SUMMARY

The Okpikruak Formation is a Lower Cretaceous (Berriasian to Valanginian) flysch sequence exposed in the central and western Brooks Range thrust belt. Preliminary lithofacies analyses of samples from the Chandler Lake Quadrangle indicate deposition in bathyal to abyssal depths in the middle-fan and possibly outer and inner fan environments of Mutti and Ricci Lucchi (1978) or in an overbank setting between channels or fan lobes. Petrographic analysis of sandstone samples from a section at Cobblestone Creek indicates that the source terrane was a combination of a recycled orogen and an undissected magmatic arc. Minor fluctuations in

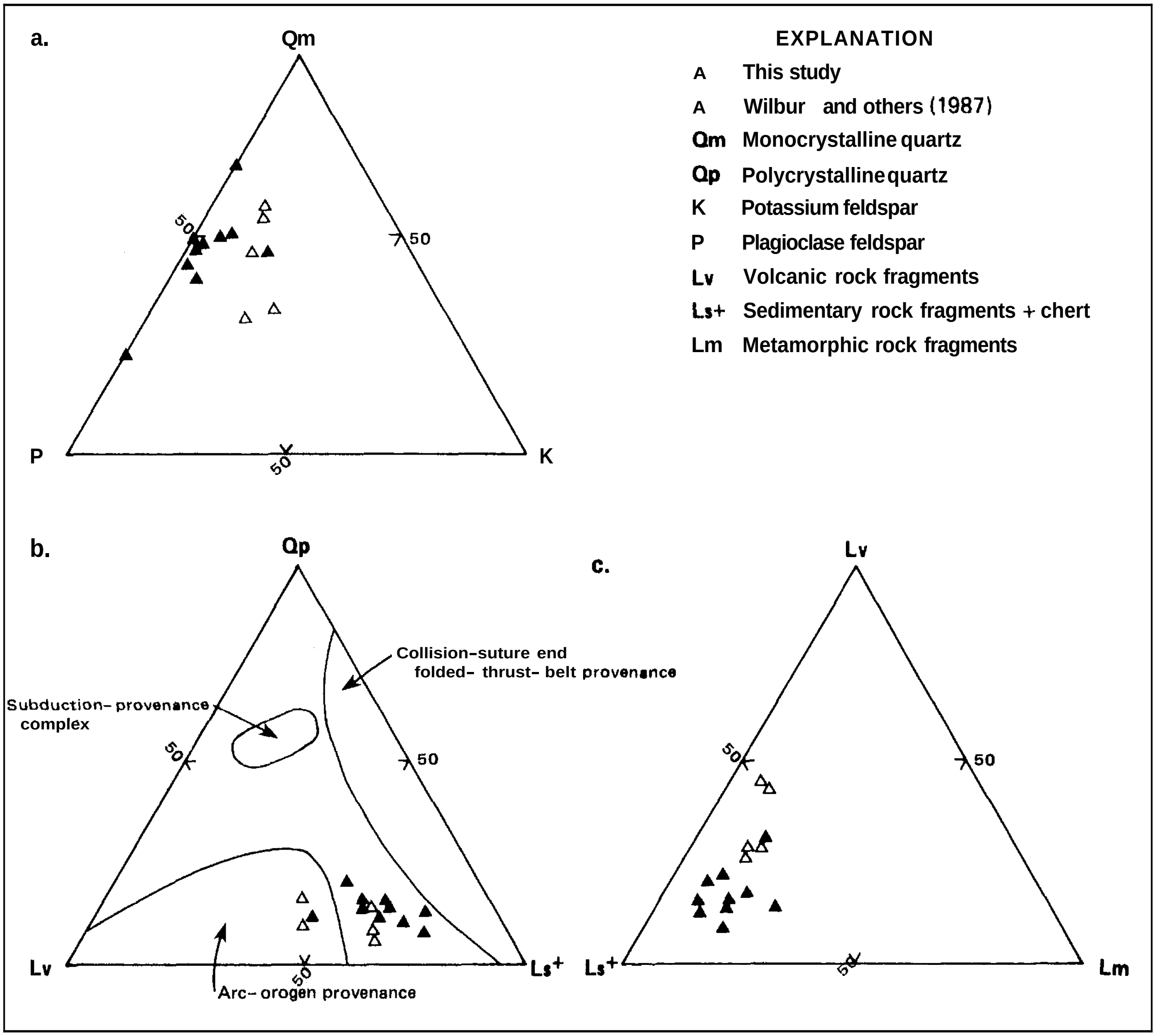

Figure 216. Ternary diagrams (QmPK, QpLvLs+, LvLs+Lm) illustrating detrital grain composition of selected sand stones from Okpikruak Formation, Cobblestone Creek. Provenance fields after Dickinson and Suczek (1979).

the detrital-grain populations over time suggest a single sedimentary source, composed dominantly of chert and quartz, and minor volcanic, igneous, and metamorphic sources. The source terrane is interpreted to have been dominated by an uplifted continental chert sequence that contained minor metamorphic quartz (sandstone interbeds?). Erosion of a volcanic arc and the uplifted chert sequence and continent-derived sandstone produced the sediments preserved in the Okpikruak Formation.

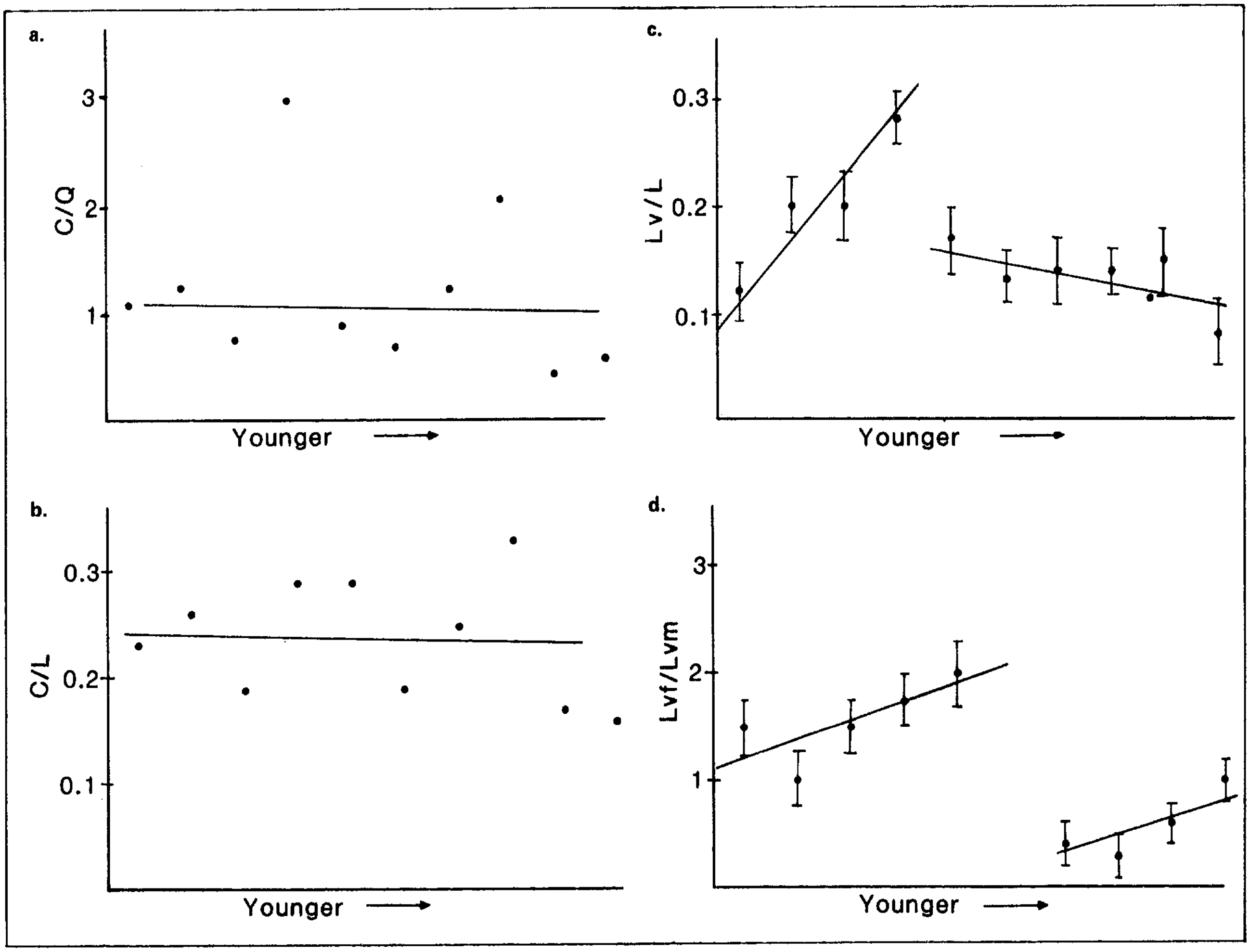

Figure 217. Bivariant plots of compositional ratios vs. time, Okpikruak Formation, Cobblestone Creek. Compositional ratios: C/Q, chert to quartz; C/L, chert to lithic grains; Lv/L, volcanic rock fragments to total rock fragments; Lvf/Lvm, felsic volcanic rock fragments to mafic volcanic rock fragments.

CHAPTER 20.

DEPOSITION OF THE FORTRESS MOUNTAIN FORMATION

By R.K. Crowder[1]

INTRODUCTION

The Trans-Alaska Pipeline corridor at Galbraith Lake (near Mile 270) passes west of a conspicuous synformal exposure of Lower Cretaceous shale, sandstone, and conglomerate assigned to the Fortress Mountain Formation (Albian). This exposure (figs. 101 and 218) is among the easternmost of a number of isolated, synclinal exposures of the Fortress Mountain in the southern foothills of the Brooks Range (fig. 219). Although the Fortress Mountain has little potential as a hydrocarbon reservoir rock, the sequence is particularly significant because it records a critical episode in the development of the Colville basin and is closely tied to the evolution of the Brooks Range orogenic belt.

The Fortress Mountain Formation is a difficult lithologic assemblage to interpret with confidence. The formation is largely an assemblage of thick conglomerate and shale, requiring a depositional environment whose hydrodynamic forces shift from one end of the depositional spectrum to the other. There is a limited suite of dispersal environments capable of significant conglomerate-shale deposition. At various times, I've believed the Fortress Mountain was deposited within each of these. My initial impression was that the formation recorded remnants of isolated alluvial-fan systems deposited some distance from the shoreline. Two years and several marine fossils later, I had shifted my interpretation to the opposite end of the spectrum and believed that the formation was entirely the record of submarine fan, channel, and canyon sedimentation (Crowder, 1987). Two more years and one coal bed later, I started to suspect that marginal-marine settings formed a significant portion of the record. I believe now that the formation was deposited within a broad suite of environments that did, in fact, range the entire spectrum from basin-plain to nonmarine dispersal systems.

This paper argues that the Fortress Mountain Formation at Galbraith Lake was deposited during two progradational episodes. The first began with deposition in basin-plain and distal submarine-fan settings, with significant turbidite influx. Northerly migration of the system produced a vertical stacking of depositional lobe and slope environments. These settings were increasingly dissected, and their sediment was reworked by submarine channels. The progradation of these environments built a narrow, but significant, depositional platform along the southern margin of the basin, which was likely the first 'shelf' to form in the basin's evolution. Near the strandline, thin coals formed within marginal-marine environments. Subsidence followed, the platform drowned, and the coal beds were buried by marine shale.

The second progradational episode was marked by dissection of the platform and deposition of thick conglomerate sequences within incised channels. High sedimentation rates caused significant northward progradation and vertical aggradation of the platform. The last glimpse of Fortress Mountain deposition is the record of fluvial sedimentation within braided systems that were likely a component of an extensive fan-delta complex.

REGIONAL STRUCTURAL AND STRATIGRAPHIC RELATIONS

Much of our knowledge of the Fortress Mountain comes from regional studies by the U.S. Geological Survey: Sable and others (1951), Tailleur and Kent (1951), Tailleur and others (1951), Chapman and Sable (1960), Detterman and others (1963), Patton and Tailleur (1964), Molenaar and others (1981), and Mull (1985). These geologists defined the regional stratigraphic and structural setting of the Fortress Mountain Formation across northern Alaska and provided the conceptual foundation on which subsequent studies are based.

The Fortress Mountain crops out in isolated 'thumbprint' exposures in a narrow, east-west-trending belt along the south flank of the Colville basin between the Sagavanirktok River on the east and the Kukpowruk

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Figure 218. View of Fortress Mountain Formation at western end of Atigun syncline near Mile 275 on Dalton Highway. About 6,400 ft (2,000 m) of section is exposed, somewhat thinner than in type area to west. Base of formation is not exposed al lhis locality bul may be present G mi (10 km) lo east along north side of Atigun Gorge. Note Trans-Alaska Pipeline at center. Photograph by C.G. Mull, July 1985.

River on the west (fig. 219). Resistant, conglomeratic 'fortress' outcrop expression is typical of thicker exposures. Most of the exposures are synclinal outcropping~,considerably less complex in structural style than surrounding and underlying formations.

The Fortress Mountain at Galbraith Lake. as in other outcrops, rests unconformably on intensely deformed, older rocks. Although the actual contact is seldom exposed, regional relations suggest that the Fortress Mountain unconformably overlies different formations in different areas that have undergone varying degrees of deformation. Patton and Tailleur (1964) and Mull (1982, 1985) considered the contact to be an angular unconformity in many areas. In the type region along the Kiruktagiak River, 90 mi (140 km) west of Galbraith Lake, the formation overlies a chaotic assemblage of chert, limestone, mafic igneous rock, or sandstone that range in age from Mississippian to early Cretaceous and are thought to represent slivers of a pre-Fortress Mountain thrust belt (Patton and Tailleur, 1964; Molenaar and others, 1981). Most commonly, the Fortress Mountain lies unconformably on folded and thrusted successions of the Okpikruak Formation of Neocomian age or chaotic units containing exotic blocks of mafic igneous and sedimentary rock (olistostromes) (Mull and others, 1976; Mull, 1985). Significantly, it appears that the base of the Fortress Mountain becomes conformable with older rocks farther north in the subsurface (Molenaar and others, 1981). These regional relations indicate that the major orogenic activity that formed the Brooks Range was pre-Albian in age, began during Berriasian time, and continued into Aptian time.

The Fortress Mountain Formation is composed of a thick sequence (as much as 9,500 ft [3,000 m]) of dark-gray to black shale and siltstone, poorly sorted lithic sandstone, and lenticular conglomerate, which typically increase in thickness and abundance upsection. Most of the sandstones are subquartzose lithic arenites, containing a consistent percentage of plagioclase feldspar and lesser amounts of metamorphic, volcanic, and sedimentary rock fragments. Conglomerate and conglomeratic sandstones within the formation are typically poorly sorted and crudely stratified and consist of chert, mafic igneous rock fragments, and sandstone clasts. Minor carbonate clasts are present at some localities and in certain horizons within the formation. Most of the rocks are friable and poorly cemented, although intense induration by silica and carbonate cement at some horizons near the top of the formation produced large, tightly cemented blocks.

The Fortress Mountain Formation is at least partially equivalent to the Torok Formation, a thick, predominantly shaly succession that is apparently a basin facies of the Fortress Mountain (figs. 28, 32, and 220). It is difficult to assess the relation between the Torok and Fortress Mountain Formations because outcrops of the two units are separated by extensive, tundra-covered intervals or by zones of intense deformation. Correlation with seismic data indicates that the Fortress Mountain is at least partly equivalent to the lower part of the Torok. Indeed, most outcrops show a distinct intertonguing of Fortress Mountain and Torok lithologies in a south to north profile. The top of the Fortress Mountain in outcrop is erosional; the relations with younger formations are not completely understood.

The isolated, synclinal outcrops of Fortress Mountain strata are remnants of a major ciastic wedge deposited in the Colville foredeep during uplift of the Brooks Range in Neocomian to Aptian time. Together the Fortress Mountain and underlying Okpikruak Formation record a significant change in regional sediment-dispersal patterns. Throughout the Paleozoic and

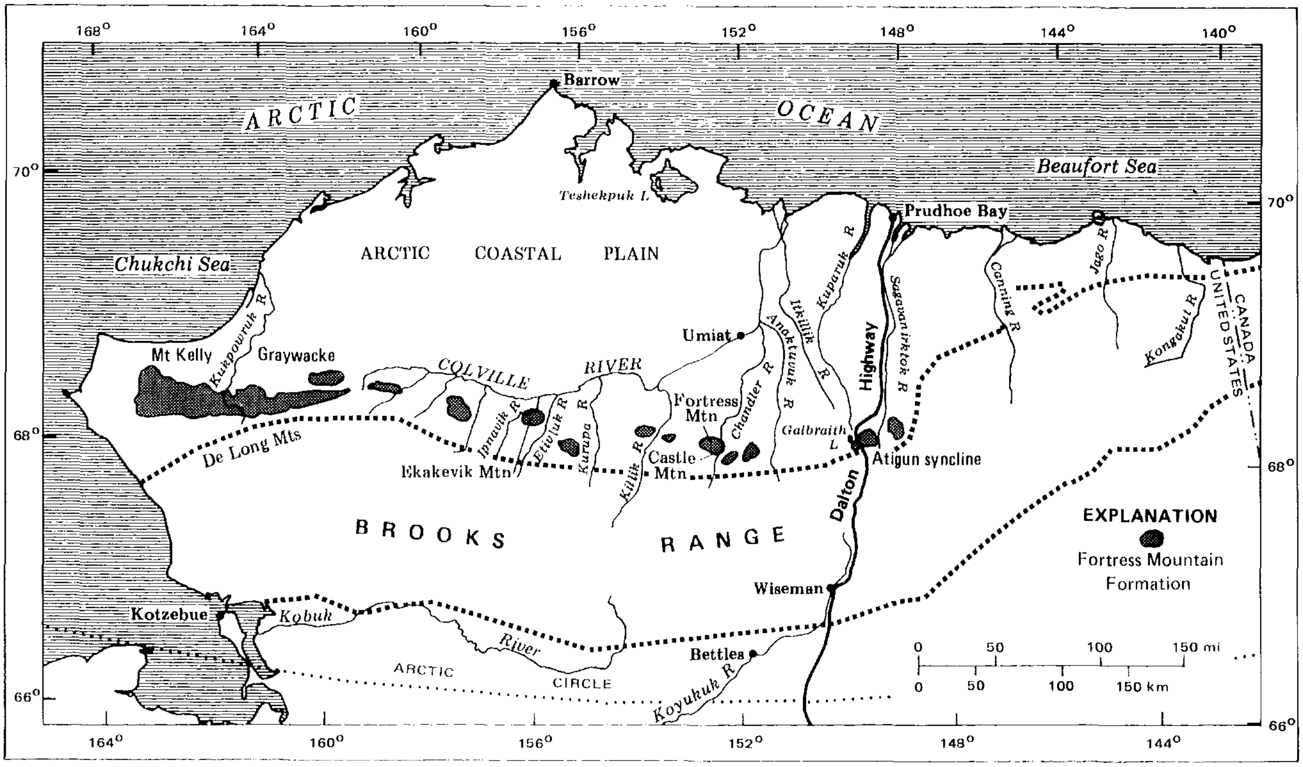

Figure 219. Index map showing distribulion o f major outcrops o f Fortress Mountain Formation across northern Alaska. Note outcrop area of Mt. Kelly Graywacke Tongue of' Fortress Mountain Formation near De Long Mountains. Outcrop pal tern from Mull (1985).

early Mesozoic, terrigenous clastic units of the Brooks Range and Arctic Slope (Ellesmerian sequence) were derived from source areas to the north. In Early Cretaceous time, howcvcr, first the Okpikruak and then much of the Fortress Mountain Formation were deposited from source areas to the south, marking the beginning of a predominantly northward sediment-transport pattern of the Brookian sequence, which continues to the present. This reversal in clastic dispersal pattern is the first sedimentologic evidence of the beginning of uplift in the Brooks Range.

This major reversal in transport direction appears to have occurred in two phases: 1) deposition of a thick, deep-marine sequence, characterized by a 3,000-1'1-thick (1 km) succession of rhythmically interbedded sandstone and shale units that show turbidite sedimentation patterns typical of basin depths; and 2) an abrupt change to deposition of shelf, nearshore, deltaic, and nonmarine sediments. The beginning of this depositional style is first recognized in the upper heds of the Fortress Mountain Formation, Although poorly underslood from a sedimentologic viewpoint, this marine-to-continental transition in the Fortress Mountain Formation represents a significant depositional episode and records the earliest sedimentologic evolution of an extensive shelf along the southern flank of the Colville basin. Surface exposures of the Fortress Mountain do not preserve the overlying sequence, but regional extrapolation indicates that Fortress Mountain deposition was followed by a return to marine sedimentation, which was recorded by thick shale and siltstone successions of the upper Torok Formation in the subsurface. By micl-Albian time, sedimentation in the Colville basin had produced an extensive sedimentary wedge, allowing northerly progradation of deltaic and interdeltaic sediments of the Nanushuk Group and related strata (see Huffman, chap. 21).

FORTRESS MOUNTAIN FORMATION AT GALBRAITH LAKE

The exposure of the Fortress Mountain Formation at Galbraith Lake is about 9,000 ft (3,000 m) thick, slightly thinner than outcrops in the type region at Fortress and Castle Mountains (fig. 219) along the Kiruktagiak River, 90 mi (140 km) to the west. The vertical and lateral facies succession displays many similarities to, and the same degree of variability typical of, the Fortress Mountain throughout its outcrop belt.

Five basic lithologies are present in the Fortress Mountain at Galbraith Lake: I.) pebbly sandstone, 2) cross-stratified sandstone, 3) clast-supported conglomerate, 4) ripple- or lenticular-bedded sandstone, and

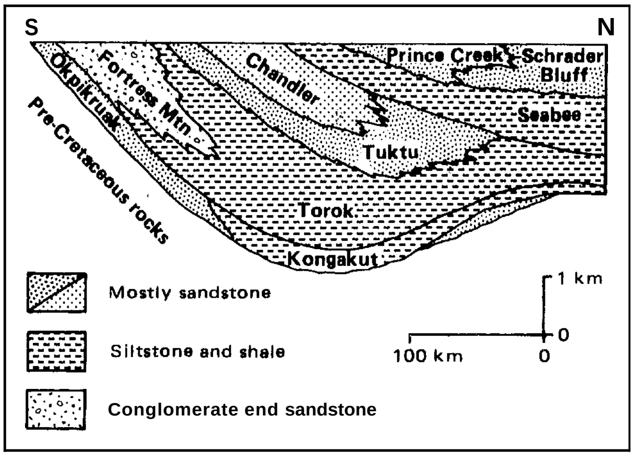

Figure 220. *Schematic* north-south cross section showing stratigraphic relations of Fortress Mountain and surrounding formations in eastern *Colville* basin.

5) thinly laminated siltstone and silty shale. The Fortress Mountain Formation at Galbraith Lake may be conveniently subdivided into units A, B, C, and D, based on variations in lithology, bedding geometry, primary sedimentary structures and textures, and internal stratification. Figure 221 is a generalized columnar section illustrating the basic lithologic units.

UNIT A: BASE OF THE FORTRESS MOUNTAIN FORMATION

UNIT A: DESCRIPTION

The lower 30 percent of the Fortress Mountain at Galbraith Lake is assigned to unit A. The base of unit A is not exposed at Galbraith Lake but may be present to the east along the north side of Atigun Gorge, where it unconformably overlies either the lower part of the Torok Formation or the Okpikruak Formation (fig. 208). The best exposure of unit **A** is also on the north side of Atigun Gorge, 2 to 8 mi (**3** to 13 km) downstream from the Atigun River bridge. Unit A is typically poorly exposed, consisting dominantly of shale and siltstone with minor intervals of sandstone and conglomerate; occasional conglomeratic channel fills are present higher in the section. The lowest beds show relatively thin (0.8 to 1.6 ft; 0.25 to 0.5 m), medium-grained sandstone units irregularly interbedded with fine-grained, ripple-laminated sandstone or, more commonly, dark-gray shale. Although poor exposures prevent lateral tracing of beds, individual sedimentation units have irregular bedding surfaces with well-developed scour and sole marks. Many arenaceous beds are both graded and stratified. Although the fivefold subdivision is usually not fully developed, the sandstone beds are the classic turbidites of Bouma (1962). Several sandstone beds exposed on the northwestern flank of the outcrop at Galbraith Lake display well-developed, upward-fining sequences that contain Bouma $T_{b\text{-}e}$ divisions.

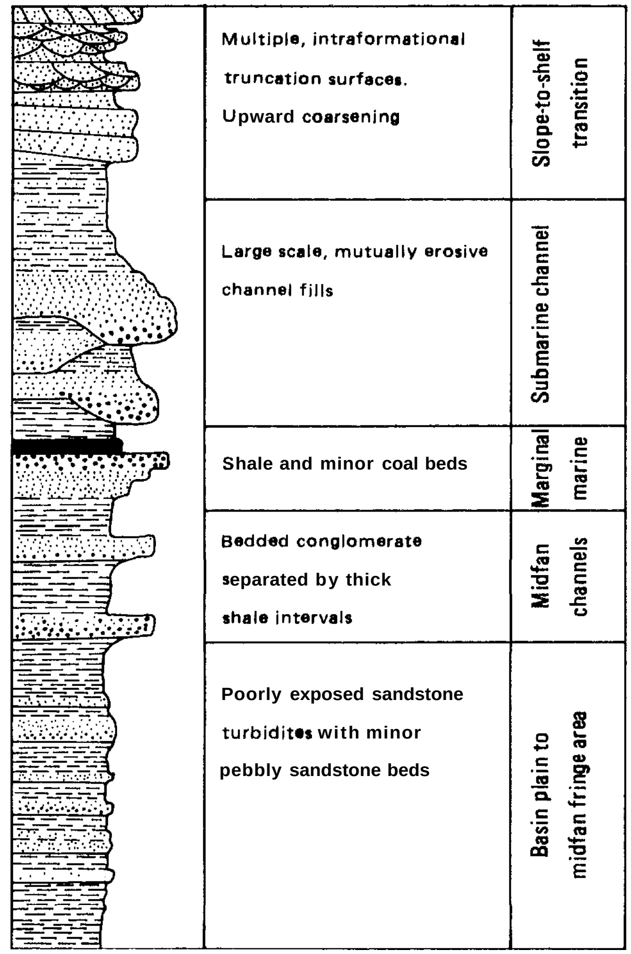

Figure 221. Generalized columnar section showing lithologic succession of Fortress Mountain Formation at Galbraith Lake.

The dominantly fine-grained material of unit A is punctuated by ienses of conglomerate or conglomeratic sandstone (facies A of Mutti and Ricci Lucchi, 1972) that become more common upsection. Most conglomerate zones are laterally continuous, have flat-parallel or flat-convex bases, and are internally characterized by tabular sets of planar cross-beds. In some instances, these beds have erosional, channeloid bases. These intervals typically fine upward into thin interbeds of sandstone and mudstone that display the upper part of Bouma sequence T_{cde} or T_{ce}.

UNIT A: INTERPRETATION

The main depositional architecture of unit A is aggradational, with only occasional progradational influxes of coarse sand and conglomerate. The bulk of unit A records relatively slow deposition of silt and mud from suspension, with episodic influx of coarser material. Such a sequence would initially suggest deposition within the outer fan subenvironment of the submarine-fan system. The presence and vertical sequence of pebbly sandstone and conglomerate intervals, however, suggest deposition along the basinward margin of a middle-fan fringe area. In this setting, transport of coarse-grained material was sporadic and rare, and initial deposition occurred in peripheral areas where individual channel identity was partly lost and sediment spread laterally. As deposition continued, basinward migration of a channel system deposited coarse-grained sediment within channels increasingly incised into underlying sediment. The lower interval of the Fortress Mountain Formation at Galbraith Lake and Atigun Gorge thus appears to record sedimentation within a transitional environment associated with outer fringe areas of a midfan distributary system.

UNIT B: MIDDLE OF THE FORTRESS MOUNTAIN FORMATION

UNIT B: DESCRIPTION

Most outcrops of the Fortress Mountain Formation visible from the Dalton Highway are assigned to unit B. This interval is transitional with the underlying succession and is composed of beds of pebbly sandstone and conglomerate alternating with interbedded sandstone-shale sequences. The pebbly sandstone and conglomerate sequences range from 15 to 80 ft (5 to 25 m) thick and are entirely lenticular or channeloid. One sedimentation unit may persist laterally for distances of hundreds of feet before pinching into intervals of black or gray shale, siltstonc, or sandstone. Erosional truncation of units occurs at some horizons but is not common.

Internally, most pebbly sandstone and conglomerate units are organized, and many are crudely stratified. Bedding is often difficult to define, but many sedimentation units are tabular beds that may, in turn, contain planar cross-stratification. Most beds show a crude imbrication of pebbles, with the long axes of grains oriented parallel to the average transport direction. In vertical sequence, each coarse interval begins with a planar or irregular base, and most sequences show distinct vertical grain-size variations. The most common grain-size trend is a general upward coarsening from medium-grained sandstone to conglomeratic sandstone or pebble conglomerate. Less commonly, irregular, upward-fining channel fills punctuate the sequence. Inversely to normally graded successions are also present at certain horizons and become more abundant upsection.

The characteristic upward-coarsening sequence is expressed at two scales within the Fortress Mountain Formation at Galbraith Lake and other localities. The first is a small-scale, upward coarsening and thickening that occurs within individual sedimentation units. The gradation, in this case, is often well developed and expressed by an increase in the average grain size. Significantly, a secondary and large-scale grain-size variation is developed within unit B and expressed as upward thickening and coarsening megasequences (fig. 222). These negative megasequences range from 30 to 80 ft (10 to 25 m) thick. Grain-size variations at this scale are best recognized by an upward increase in maximum grain size. Stratification thickness also changes regularly in vertical section from thin to thick bedding.

At various horizons, unit B is interrupted by distinct conglomerate and sandstone beds that have a conspicuous upward thinning and fining architecture. Internally, these intervals are composed of multiple, mutually erosive channel fills. These positive sequences and megasequences are less common than the characteristic upward-thickening sequence and are not present at preferred positions within the vertical sequence. Typically, the channel-fill sequences, varying from 82 to 164 ft (25 to 50 m) wide, are not as laterally continuous as the negative megasequences. Nonetheless, these intervals constitute an important component of unit B.

Pebbly sandstone and conglomerate sequences of unit B are typically separated by intervals of thin, interbedded sandstone and mudstone that display a variety of internal bedding characteristics. Despite poor exposures, most finer grained intervals can be assigned to one of two facies. The first consists of thinly interbedded shale and sandstone. The sandstones are fine- to medium-grained lithic arenites that are tabular and laterally continuous (facies D of Mutti and Ricci Lucchi, 1972, 1978). Individual sandstone beds range from 0.2 to 1 ft (0.05 to about 0.30 m) and may be spaced from 0.8 to 1.6 ft (0.25 to 0.50 m) apart. A number of beds within such sequences at Galbraith Lake display Bouma T_{cde} sequences. In general, these beds tend to be associated with negative megasequences.

The second fine-grained facies developed in unit B consists of thin, lenticular to flaser-bedded sandstone in shale (facies E of Mutti and Ricci Lucchi, 1972, 1978). Sandstone layers are much thinner than in the preceding fine-grained facies, more closely spaced, and the sandstone to shale ratio is higher. These rocks are typically developed as lateral equivalents to upward-fining channel-fill sequences.

The upper megasequences of unit B contain numerous contorted layers and some beds are displaced along translational and rotational slump and slide surfaces. The direction of vergence in all cases is toward the north or northeast. The upper 66 ft (20 m) of the unit contains a thin coal bed discovered by Robert Goff (University of Alaska Fairbanks) in 1987. The bed is 6 in. (14 cm) thick and has been traced laterally several hundred meters. It is overlain by dark-gray marine shale, siltstone, and sandstone that contain pelecypods.

UNIT B: INTERPRETATION

The depositional architecture of unit B is progradational and records northerly migration of a submarine dispersal system capable of transporting coarse-grained sediment. The geometry of megasequences, facies relations, and a high proportion of coarse-grained sediment suggest deposition within and adjacent to channelized environments of a submarine-midfan setting. Upward-fining channel sequences and megasequences record deposition within multiple, mutually erosive channels. These channels were intricate and unstable but became progressively more incised as basinward migration continued. Thinly interbedded sandstone-shale

Figure 222. View southward toward western end of Atigun syncline at Galbraith Lake, showing multiple upward thickening and coarsening megasequences of unit B of Fortress Mountain Formation.

sequences characterized by discontinuous lenticular and flaser-bedded sandstone layers were deposited within overbank or levee environments associated with major channel systems. These beds represent spillover from high-concentration gravity and traction-flow processes adjacent to channel margins.

Multiple thickening- and coarsening-upward sequences may record either tectonic or depositional processes. Normark and others (1979) suggest that suprafan areas of modern submarine fans are probably underlain by such sequences in the form of depositional lobes. Thickening-upward cycles might also be generated by crevasse splay, if a levee is breached in the inner fan environment. In either case, the association with channel-fill successions within unit B of the Fortress Mountain Formation suggests midfan deposition.

Translational and rotational slumps within unit B indicate a significant north-dipping depositional paleoslope. This horizon may record a slope-to-platform transition that allowed deposition of thin coal layers in the overlying succession. Shale and siltstone overlying the coal indicate a return to marine sedimentation probably related to subsidence or local coastal onlap. This is a major punctuation point in Fortress Mountain history, marking the end of the first progradational cycle.

UNIT C: LARGE-SCALE CHANNEL FILLS OF THE FORTRESS MOUNTAIN FORMATION

UNIT C: DESCRIPTION

Unit C of the Fortress Mountain Formation is composed of 820 to 990 ft (250 to 300 m) of coarse-grained sandstone, pebbly sandstone, organized and disorganized conglomerate, and rare, matrix-supported conglomerate beds. These conglomerate and sandstone sequences are of a distinctly different geometry and size than coarse-grained sequences of unit B. Most notably, conglomerate intervals are bounded by large-scale scour surfaces with as much as 50 ft (15 m) of local erosional relief (fig. 223). Basal beds are sharply concave and erosional and typically contain a high percentage of reworked clasts derived from underlying rocks. Internally, beds range from massive and disorganized to crudely stratified and upward fining.

Channel-fill margins are marked by sharp, erosional lateral contacts with surrounding beds. Interchannel areas are typically composed of thinly interbedded shale

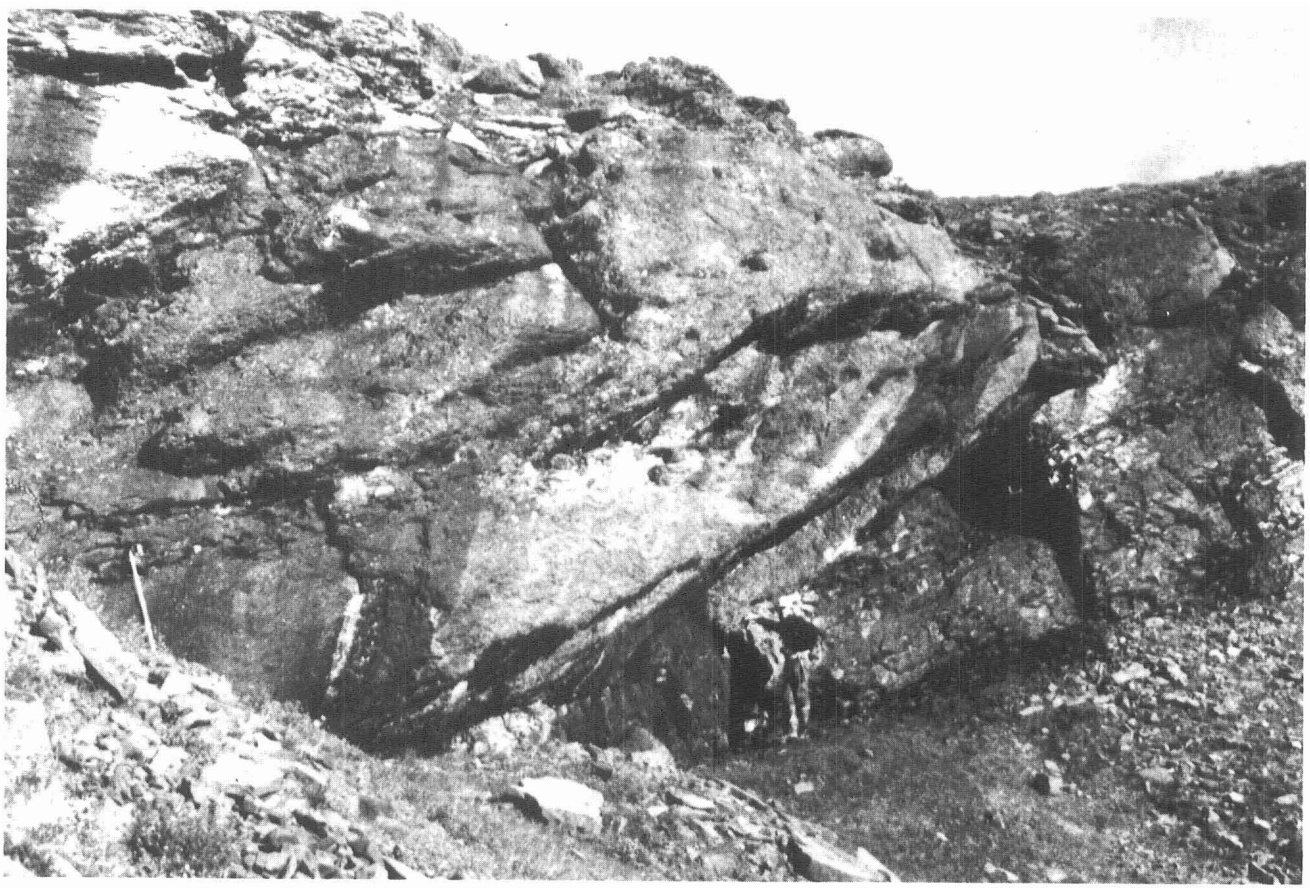

Figure 223. Erosional base of large-scale channel fill in unil C of Fortress Mountain Formation. Base of channel (at geologist's hand) is incised in alternating shale and fine-grained sandstone.

and discontinuous sandstone (facies E of Mutti and Ricci Lucchi, 1972, 1978). Along some channel margins, the transition from interchannel to channel lithologies is marked by rare, matrix-supported conglomerates.

UNIT C: INTERPRETATION

The channel-dominated conglomerate and sandstone of unit C were deposited by debris flow (disorganized and matrix-supported conglomerate) or traction and grain-flow processes (organized conglomerate). Together, the overall geometry, vertical sequence of mixed stratified and massive units, and stratigraphic position suggest deposition within large channels and valleys carved into the margin of the depositional platform. The channel fills range from base- to top-of-slope settings. The sequence demonstrates renewed progradation of the platform margin and vertical filling of the platform surface and represents the last of the marine record in the Fortress Mountain succession at Galbraith Lake.

UNIT D: UPPER SANDSTONES AND SHALES OF THE FORTRESS MOUNTAIN FORMATION

UNIT D: DESCRIPTION

The upper 20 percent of the Fortress Mountain Formation at Galbraith Lake consists of multiple, upward-fining sandstone and pebbly sandstone beds with intervening shale and siltstone sequences. Most of this section is poorly exposed, and various units are difficult to trace laterally. Nonetheless, the basic depositional architecture can be reconstructed and compared with underlying successions and equivalent strata in other localities.

The upper unit of the Fortress Mountain Formation is transitional with underlying channel and interchannel deposits of unit C. The transition is marked by a gradual change from conglomerate-dominant intervals to successions of interbedded sandstone, siltstone, and shale. A slight angular discordance of <10° separates units C and D. At some locations (for example, near the Kiruktagiak River at Castle Mountain), this transition exhibits extreme irregularity of stratification, abrupt variation in thickness, and conspicuous angular discordances with numerous planar surfaces; these collectively record a large-scale reorientation of the depositional surface (fig. 224).

Above the transitional interval, strata at Galbraith Lake are medium- to coarse-grained lithic arenite, siltstone, and shale. Sandstone units in this interval show a variety of cross-stratification. Basal contacts are sharp, erosional, and marked by coarse lag deposits. Multidirectional trough cross-bedding dominates the succession, but tabular sets of planar cross-stratification are also present and are consistently oriented toward the northwest. Finer grained intervals are very poorly exposed, and thus the character of the lithology and bedding are difficult to assess at many horizons. Exposed intervals, however, are composed of massive, apparently bioturbated siltstone and fine-grained sandstone with minor intervals of dark-gray shale. Some horizons are organic rich, containing a high percentage of fossil-plant casts and beds of organic material, some of which may represent paleosols.

Unit D coarsens toward the top of the exposure at Galbraith Lake. This upward coarsening is expressed at two scales: 1) a general upsection coarsening within the finer grained recessive intervals; and 2) an upward coarsening within coarse-sandstone and pebble-conglomerate units, which become more closely spaced upsection.

UNIT D: INTERPRETATION

The upper unit of the Fortress Mountain Formation at Galbraith Lake is interpreted as a transition from basin-slope to deltaic and fluvial environments. Multiple intraformational truncation surfaces developed in some areas near the base of the sequence record a major reorientation of depositional topography, perhaps related to changes in basin gradient. The overlying sequence of shale, siltstone, and sandstone is interpreted as a delta-platform deposit that grades upward into lower delta-plain and fluvial environments. This transition from basin-slope- to platform- and continental-sedimentation patterns is a significant change in the depositional architecture of the Fortress Mountain Formation and overlying strata, and it may indicate the sedimentologic evolution of a platform over which deltaic and nonmarine environments could prograde.

REGIONAL TRENDS IN DEPOSITION OF THE FORTRESS MOUNTAIN FORMATION

The Fortress Mountain Formation has apparently been affected by only minor deformation, and regional relations suggest that the present outcrop distribution may reflect major centers of sediment dispersal into the Colville basin during Albian time. The exposure of the formation near the Dalton Highway shows much of the Fortress Mountain depositional architecture typical of the Fortress Mountain across northern Alaska. In the type region near the Kiruktagiak River, the lower part of the formation shows a general transition from a sequence of basin-plain and outer fan turbiclites upward to thick conglomerate, sandstone, and shale sequences of middle- and inner fan environments. At Castle and Fortress Mountains in the type region (fig. 219), the upper 10 to 20 percent of the formation records a major transition to shelf and fluvial-deltaic environments similar to those at Galbraith Lake.

Hunter and Fox (1976) and Molenaar and others (1981) described a similar sequence at Ekakevik Mountain (fig. 219), where the Fortress Mountain Formation varies upsection from basin-plain turbidites to massive conglomerates interpreted as submarine-canyon deposits. Strata exposed near the top of Ekakevik Mountain contain thin coal beds and possible paleosols, which indicate nonmarine deposition. The uppermost succession that caps Ekakevik Mountain has not yet been studied in detail but is known to contain pelecypods within strata of increasing compositional and textural maturity (Hunter and Fox, 1976; Molenaar and others, 1981). The uppermost beds at Ekakevik Mountain may thus represent a transgressive, retrogradational event.

The sedimentologic evolution of dispersal centers of the Fortress Mountain Formation remains poorly understood. Additional detailed and systematic analyses of major outcrops will be necessary to integrate the hypotheses presented in this paper into a regional paleogeographic reconstruction. Observations indicate that deposition of the formation was characterized by rapid marine-to-continental transition late in Fortress Mountain time. This transition represents a significant change in the depositional architecture of the Cretaceous Arctic Slope sequence and records the evolution of basin-margin, shelf, and deltaic environments along the southern margin of the Colville basin. These environmental settings are typical of the depositional style of overlying strata. In a sense, deposition of the Fortress Mountain Formation was a necessary precursor to the development of a depositional platform that allowed progradation of deltaic and interdeltaic environments of the overlying Nanushuk Group and related strata.

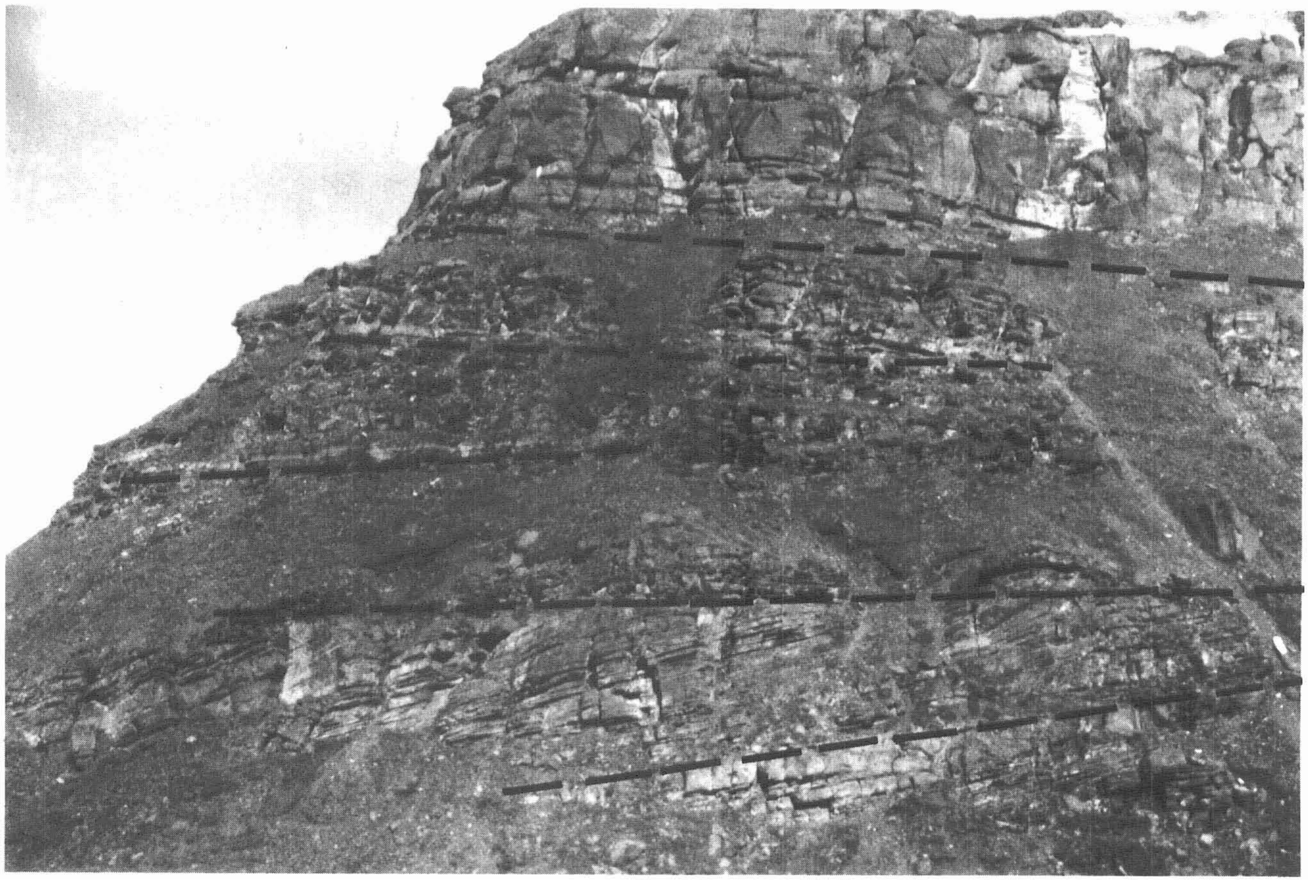

Figure ***224.*** *Multiple intraformational truncation surfaces (dashed lines) in upper Fortress Mountain Formation at Castle Mountain, 90 mi (140 km) west of Galbraith Lake. Planar truncations record major reorientation of depositional surfaces.*

CHAPTER 21.

THE NANUSHUK GROUP[1]

By A.C. Huffman, Jr.[2]

INTRODUCTION

The Nanushuk Group of Albian to Cenomanian age is a regressive sequence of marine, transitional, and nonmarine deposits up to 7,000 ft (2,750 m) thick, exposed in an outcrop belt 18 to 30 mi (30 to 50 km) wide and about 400 mi (650 km) long in the Arctic Foothills province, Arctic Slope, Alaska (fig. 225). Field and laboratory studies indicate that the Nanushuk Group was deposited in deltaic systems throughout most of the outcrop belt. The group can be divided into two major deltaic systems separated roughly by long 157° W. West of this longitude, deposition took place in the elongate, river-dominated Corwin delta (Ahlbrandt and others, 1979), which prograded to the northeast from the vicinity of the De Long Mountains, Tigara uplift, and Herald arch (fig. 225). Along the southcentral Arctic Slope, east from long 157° W. to the Sagavanirktok River, deposition occurred in the lobate to elongate, river-dominated Umiat delta (Ahlbrandt and others, 1979), which prograded to the north, northwest, and northeast from the vicinity of the Endicott Mountains (fig. 225). Intermixing of sediment from the two deltas occurs in the subsurface of the northcentral Arctic Slope.

Outcrop and subsurface studies performed before the latest exploration program (1974) of the National Petroleum Reserve Alaska (NPRA) can be grouped by geographic area: 1) southwestern Arctic Slope (Chapman and Sable, 1960; Smiley, 1966, 1969a); 2) southcentral Arctic Slope (Gryc and others, 1951, 1956; Detterman and others, 1963; Chapman and others, 1964; Brosgé and Whittington, 1966; Smiley, 1969b); 3) eastern Arctic Slope (Detterman and others, 1975); and 4) subsurface of northwestern and northcentral Arctic Slope (Robinson, 1956, 1958a, b, 1959a, b, 1964; Collins, 1958a, b, c, 1959; Robinson and Collins, 1959). Recent work undertaken as part of the NPRA exploration has been concentrated on a regional synthesis of various geologic aspects of the Nanushuk Group. Results of this work were reported in Ahlbrandt (1979) and Huffman (1985).

Figures 32 and 226 present the currently accepted stratigraphic nomenclature for the Nanushuk Group and related rocks. Recent studies of the Nanushuk Group are based on a genetic approach to the stratigraphy to avoid the confusion that surrounds the nomenclature and to arrive at a better understanding of the systems responsible for the deposition of the Nanushuk Group. Results of these studies have been reported by Ahlbrandt and Huffman (1978), Ahlbrandt and others (1979), Bartsch-Winkler (1979), Fox (1979), Fox and others (1979), Huffman (1979), Bartsch-Winkler and Huffman (1981a, b), Huffman and others (1981a, b), Bartsch-Winkler (1985), Huffman and others (1985), and Molenaar (1985).

REGIONAL STRATIGRAPHY

WESTERN ARCTIC SLOPE

The Nanushuk Group in the western Arctic Slope (west of long 157° W.) was deposited in the elongate, river-dominated Corwin delta, which prograded to the east and northeast over shelf and prodelta muds and sands of the Aptian and Albian Torok Formation (Anderson, Warren, and Associates, 1900). Bird and Andrews (1979) and Molenaar (1981a, b, 1985) have interpreted seismic data from NPRA to indicate that the Torok Formation represents large foreset or clinoform beds that prograde generally eastward and grade upward into topset beds of the Nanushuk Group. Individual seismic reflectors can be traced from basinal beds in the Torok Formation up the foreset slope and into the nearly horizontal topset beds. This indicates a genetic and time relation between the sediments of the Corwin delta and those of the Torok Formation. This relationship is also supported by petrographic data that strongly suggest a similar source for these two units, a source quite different from that of the Umiat delta.

[1] Approved for publication by Director, U.S. Geological Survey.
[2] U.S. Geological Survey, Box 25046, MS 939, Denver Federal Center, Denver, Colorado 80225.

Where seen in outcrop and on subsurface well logs, the Corwin delta generally contains <20 percent sandstone (fig. 225), most of which is concentrated in distributary-channel deposits and elongate distributary mouth bars. The distributary mouth-bar deposits may be laterally continuous over distances of several miles parallel to the paleoshoreline and rarely exceed thicknesses of 30 to 50 ft (10 to 15 m); their continuity perpendicular to the paleoshoreline is unknown but may be several tens of miles. Distributary mouth bars are rarely stacked on top of one another and are commonly overlain by mudstone and sandstone of interdistributary-bay fill or by prodelta shale and mudstone.

Distal mouth-bar deposits of the Corwin delta may be laterally extensive, are commonly thin (<10 ft; 3 m), and are composed of fine-grained sandstones with high percentages of matrix (clay, silt, and deformed rock fragments). The sandstones of the distal mouth bars grade upward, both in grain size and thickness, into the coarser grained and thicker distributary mouth-bar sequences but are not stacked into thick delta-front deposits, as the sandstones in the Umiat delta.

Distributary-channel deposits of the lower delta plain, adjacent to the distributary mouth bars, are composed of fine-grained sandstones generally <650 ft (200 m) wide and 30 ft (10 m) thick. The sandstones are lenticular and encased in siltstone and mudstone. The only exception to this is where the distributary channels merge with sandstones of the distributary mouth bars or crevasse splays. Higher on the delta plain, the fluvial channels are coarser grained, less well sorted, and commonly contain conglomeratic lags at the base of trough sets. The major distributary deposits may be as thick as 100 ft (30 m) (125 ft [38 m] at Corwin Bluff) and several hundred feet wide; these represent composite deposits of low sinuosity streams. Middle and upper delta-plain deposits are also characterized by coal beds as thick as 13 ft (4 m). Coal is more abundant in the interdistributary regions than near the higher energy, sandy distributary channels.

Very fine to fine-grained crevasse-splay sandstone deposits are also concentrated in the interdistributary areas. Although these deposits may be laterally extensive, they are rarely more than a few feet thick and are commonly interbedded with mudstone and siltstone of the interdistributary-bay fill.

The Corwin delta is characterized by sparse, low-diversity macro- and microfaunas. Poorly preserved, low-diversity foraminiferal assemblages from outcrop samples studied by Sliter (1979) suggest shallow water and a preponderance of nearshore and foreshore environments in the marine part of the section. Palynologic studies have been hampered by poor preservation of pollen and spores from outcrop samples, due to deep burial and to subsequent near-surface weathering. Plant megafossils are abundant and, combined with observed pollen and spores from the nonmarine parts of the Corwin delta, indicate a "warm, frost-free environment in the Albian, followed by cooler, warm-temperate conditions from latest Albian to Turonian time" (Scott and Smiley, 1979). Dinosaur remains in the upper, nonmarine beds of both the Corwin and Umiat deltas (Roehler and Stricker, 1984; Witte and others, 1987) reinforce the interoretation of a warm. frost-free environment. Upon re-examination of paleobotanical data, however, Spicer (1987) concluded that the Albian to Cenomanian climate in northern Alaska was cool and had a pronounced seasonality.

SOUTHCENTRAL ARCTIC SLOPE

The Nanushuk Group of the southcentral Arctic Slope (long 157° W. to long 149° W.) was deposited in the elongate to lobate, river-dominated Umiat delta (Huffman and others, 1985). Transport directions from the nonmarine facies of the Umiat delta, as well as sandstone percentage (fig. 225) and modal grain-size distribution, indicate that the Umiat delta prograded generally northward from the vicinity of the Endico'ct Mountains.

The northern and western parts of the Umiat delta prograded across an already existing shelf and bear little or no genetic relation to the slope and shelf sediments of the underlying Torok Formation. Both seismic (Bird and Andrews, 1979; Molenaar, 1981a, b) and petrographic studies (Bartsch-Winkler and Huffman, 1981b) indicate that most shelf and slope sediments of the Torok were supplied from the west. The influence of the Corwin delta on the southern and eastern parts of the Umiat delta is unknown.

Sandstones of the Umiat delta differ significantly from those of the Corwin delta and from the shelf sandstones of the Torok Formation. Sandstones from the western source area are characterized by high percentages of sedimentary lithic fragments and are classified as sedimentary litharenites, whereas sandstones from the central source area are high in quartzite and metamorphic lithic fragments and are classified as phyllarenites.

In the southern part of the central Arctic Slope, the Nanushuk Group is coarser grained (medium grained) and sandier (>25 percent) than in the western area. Huffman and Ahlbrandt (1979), May (1979), and Mull (1979, 1985) attributed these and other differences between the two areas to the presence of the thick quartzite and quartz-rich Kanayut Conglomerate of Late Devonian and Early Mississippian(?) age in the probable source area of the Umiat delta and to the absence of a comparable unit southwest of the Corwin delta. Other major factors that cause differences between these two delta systems are the higher energy level of the marine environment into which the Umiat delta prograded and the generally greater distances from the probable source areas for most Corwin delta exposures.

Along the outcrop belt of the Umiat delta, distal-bar deposits are composed of very fine grained, poorly sorted sandstones and siltstones as thick as 80 ft (25 m), interbedded with mudstone and shale as thick as 300 ft (90 m). Although the distal-bar deposits could not be traced their full length in outcrop, they extend several miles parallel to depositional strike where they are exposed in the Tuktu escarpment (fig. 225). In the measured section at Tuktu Bluff (fig. 225), these sandstone bodies are stacked on top of one another to make a 650-ft-thick (200 m) section of nearly 100 percent sandstone (Huffman and others, 1981b).

The primary loci of sand deposition in the southern part of the Umiat delta were distributary mouth bars.

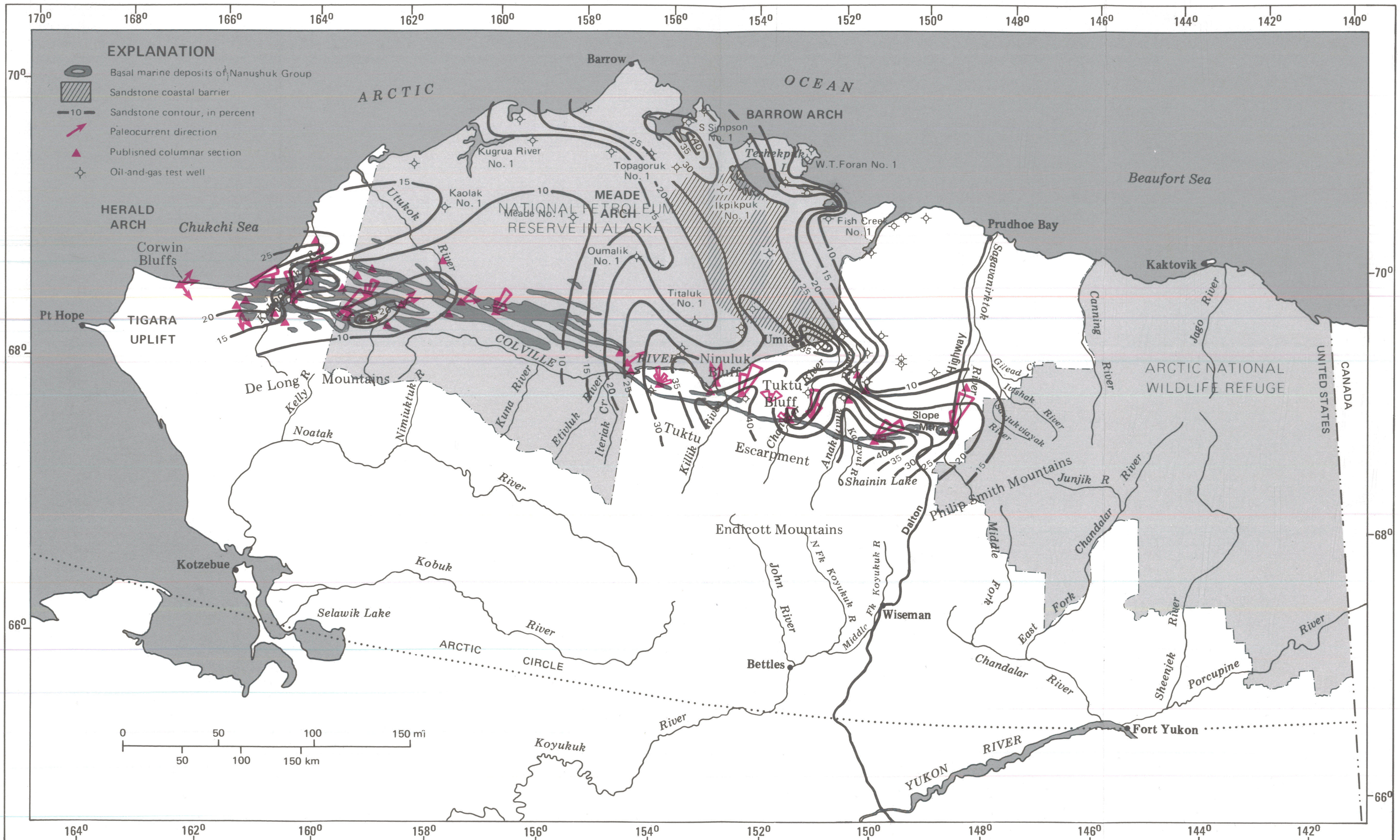

Figure 225. Index map of western and central Arctic Slope, showing percent sandstone and paleocurrent trends of Nanushuk Group.

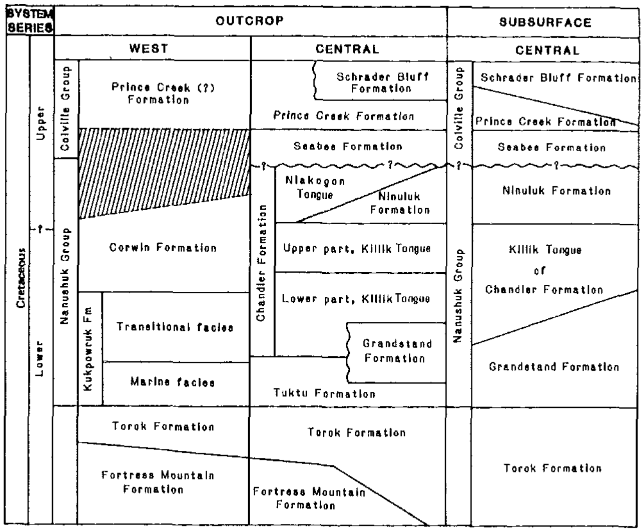

Figure 226. Stratigraphic nomenclature of Nanushuk Group and related units, Arctic Slope, Alaska.

These fine- to medium-grained sandstone bodies are 30 to 80 ft (10 to 25 m) thick and are generally lenticular in cross section, although they may coalesce laterally to form sheets of sandstone several tens of miles wide. Grain size and porosity increase upward in individual bars. As in a normal regressive or prograding succession, the bar deposits are overlain by fluvial deposits of the distributary channels above a scour surface or by fine-grained, interdistributary-bay sediments. Stacked, regressive shoreface and foreshore sequences are also present in the Umiat delta complex, for instance, in the Kurupa anticline measured section (Huffman and others, 1981b).

Channel deposits of the Umiat delta are typically fine- to medium-grained, trough-cross-bedded sandstones or conglomeratic sandstones. They range in thickness from 3 to 80 ft (1 to 25 m) and are several hundred feet wide. The contact between the transitional and nonmarine facies is commonly marked by laterally extensive braided-stream deposits up to 80 ft (25 m) thick. These deposits are conglomeratic and contain white quartz with dark chert clasts and large plant fragments. The channels flowed generally northward from the ancestral Brooks Range.

Near major distributary deposits and overlying delta-front sediments, the interdistributary fill is composed primarily of mudstone and siltstone sequences from 30 to 330 ft (10 to 100 m) thick. Overlying and interbedded with these fine-grained sediments are cross-bedded and ripple-laminated, fine-grained sandstones deposited in crevasse splays and small distributary channels. These sandstones are typically 15 to 20 ft (5 to 6 m) thick but may be stacked as high as 100 ft (30 m). Marsh deposits of claystone, siltstone, and coal are typically >3 ft (1 m) thick but rarely >10 ft (3 m) thick.

Coal beds are common in the transitional facies of the Umiat delta. Although the coal beds are thin (<5 ft; 1.5 m), they are mainly rooted in crevasse-splay deposits and are significant indicators of more stable environments and slower sedimentation rates than were present in the transitional facies of the Corwin delta.

Diversity, abundance, and preservation of all types of fossils in the Nanushuk Group are greater in the surface and subsurface of the central Arctic Slope than in the western area, a distribution which indicates slower sedimentation rates, better circulation, and more oxygen in the central Arctic Slope environment. Foraminiferal studies (Bergquist, 1966; Sliter 1979) indicate a deeper water, offshore environment for much of the marine facies of the Nanushuk Group in the central Arctic Slope, compared with the shallow, nearshore and foreshore environments characteristic of the western area. The shallower, cooler burial history of the central Arctic Slope created a greater preservation potential for pollen, spores, dinoflagellates, and acritarchs than was possible in the western Arctic Slope.

The age of the Nanushuk Group in the central Arctic Slope generally ranges from middle Albian to Cenomanian (Imlay, 1961; Bergquist, 1966; Scott and Smiley, 1979; Sliter, 1979). Dinoflagellate and palynologic data from well material (May and Shane, 1985) suggest progressively younger deposits from west to east. This interpretation is consistent with the hypothesis that the growth of the Umiat delta was largely controlled by the presence of a shelf that was building from the west and fed by the Corwin delta.

Throughout much of the central Arctic Slope, nonmarine deposition on the Umiat delta plain was halted by a marine transgression during Cenomanian time (Detterman and others, 1963; Scott and Smiley, 1979; Sliter, 1979). In the outcrop area, this transgression is characterized by intertonguing marine and nonmarine deposits. Several thin (12 to 24 in.; 30 to 60 cm) beds of cross-bedded, conglomeratic sandstone that contain pelecypod and bone fragments are present in these transitional deposits, recording pulses of the regional transgression (Huffman and others, 1985). At Ninuluk Bluff (fig. 225), this upper transitional facies is 1,050 ft (320 m) thick and contains 25 percent sandstone. Coal beds, <3 ft (1 m) thick and containing 2.4-in.-thick (6 cm) tonsteins, are common in the nonmarine deposits of the transitional facies at Ninuluk Bluff.

NORTHEASTERN ARCTIC SLOPE

The Nanushuk Group changes facies within a short distance along the outcrop belt northeast of Marmot syncline (fig. 225) and the Sagavanirktok River and apparently grades, east of the Ivishak River, into a thick sequence of mudstone and sandstone turbidites mapped by Keller and others (1961) as the lower member of the now-abandoned Ignek Formation. Reconnaissance examination and sampling of outcrops along the Saviukviayak and Ivishak Rivers and Gilead Creek (fig. 225) support the correlation of at least the upper part of these deposits with the Nanushuk Group, on the basis of similar fluvial lithologies. Although some exposures are structurally complex, the general sequence includes a thin, nonmarine section that overlies thick, shallow-marine deposits that overlie a thick, deep-marine turbidite sequence. Observed lithologies in the turbidite deposits include thick mudstone or silty claystone and very fine grained sandstones. All indications are that the turbidites are approximately time equivalent to the lower deltaic Nanushuk Group (Keller and others, 1961). These observations help establish the position of the shelf break during deposition of the Nanushuk Group.

NORTHCENTRAL ARCTIC SLOPE

The Umiat delta extends a very short distance northward into the subsurface. The belt of relatively high sandstone concentrations that extends generally northward from near Umiat to Ikpikpuk No. 1 well and the apparent sandstone buildup near the South Simpson No. 1 well (fig. 225) both seem due in part to redistribution of sand from the Umiat delta and to some influence from the Corwin delta.

Petrographic results reported by Bartsch-Winkler (1979) indicate that, in most of the wells sampled, sandstones in the lower part of the section bear a strong compositional similarity to Corwin-delta sediments. Higher in the section, however, there is a marked change, and the sandstones show greater resemblance to the Umiat-delta sediments. Most sampled wells also suggest a degree of mixing of sediment types in some beds, as well as possible interbedding of the two sediment types.

Several explanations for the different lithologies would satisfy both outcrop and petrographic data. The most probable is that the northern and northwestern part of the Umiat delta prograded northward onto a shelf formed in part from sediment from the Corwin delta (Molenaar, 1981b). Near Umiat, the Urniat delta was subjected to strong waves and currents that tended to move sediment to the northwest. A northwest-trending coastal barrier (fig. 225) was formed from sediment transported to the coast by Corwin streams and from sediment transported northward by longshore drift from the Umiat delta.

SLOPE MOUNTAIN

A partial section of the Nanushuk Group is exposed on Slope Mountain at the southeastern end of Marmot syncline (figs. 105, 227, and 228). The Nanushuk is 3,475 ft (1,060 m) thick and contains about 26 percent sandstone with an average grain size of 0.52 mm. Fifty percent of the lower, marine beds and nearly 18 percent of the upper, nonmarine beds are exposed. The lowermost 650 ft (200 m) of outcrop are interbedded shale, siltstone, and sandstone, deposited in a prodelta environment. These beds have a close genetic relation to the overlying sandstone of the Nanushuk Group, but on an electric log they would probably be assigned to the Torok Formation. The lowest significant sandstone body is composed of 80 ft (25 m) of rippled and bioturbated, fine-grained sandstone, deposited in a distal bar. The sandstone is overlain by 130 ft (40 m) of medium-grained, cross-bedded, horizontally laminated distributary mouth-bar deposits that are capped by a conglomeratic fluvial sandstone.

Immediately above this delta cycle is a second similar cycle about 150 ft (45 m) thick that begins with prodelta mudstone and siltstone and is capped by a thick (150 ft; 45 m) fluvial sequence. Overlying this major distributary deposit is about 800 ft (250 m) of lower delta-plain sediments composed of 6- to 50-ft-thick (2 to 15 m) fluvial sandstone bodies enclosed in thick, fine-grained deposits of carbonaceous shale, siltstone, and sparse, thin coal beds.

The upper 1,480 ft (450 m) of outcrop is composed entirely of nonmarine deposits of the middle to upper delta plain; the base is marked by an 80-ft-thick (25 m) conglomerate and conglomeratic sandstone. These high-energy fluvial deposits can be found along the entire length of the Nanushuk Group in **the** central

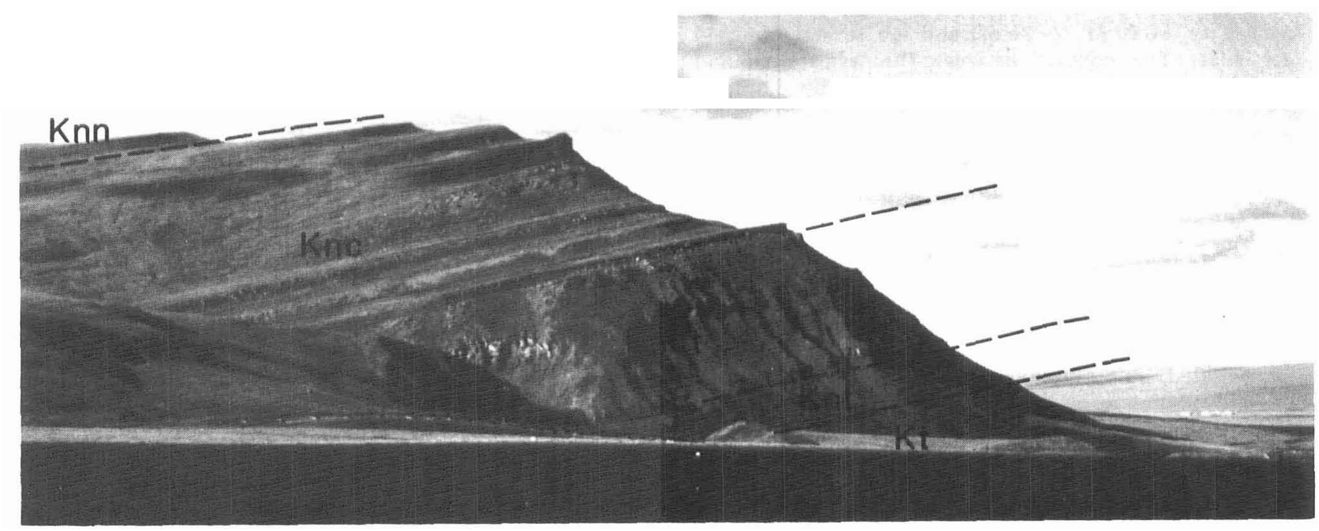

Figure 227. View northeastward from near Mile 298 of Nanushuk Group on Slope Mountain. Nonmarine beds of upper part of Killik Tongue of Chandler Formation (Knc) form ledges and cliffsof upper mountain; transitional marine and nonmarine beds of lower Killik Tongue of Chandler Formation (Knct) form steep slopes on lower mountain. Marine sandstones of Tuktu Formation (Knt) and prodelta shale of Torok Formation (Kt) crop out at base. Possible marine beds of Ninuluk Formation (Knn) of Nanushuk Group cap highest point. Topographic relief about 1,800 ft (600 m). Photograph by C.G Mull, August 1984.

Arctic Slope and may record a period of uplift in the ancestral Brooks Range. Above the basal conglomerate, the widely separated fluvial-channel deposits are *6* to 40 ft (2 to 12 m) thick, fine to medium grained, and contain pebbles of the same gray chert and white quartz as in the conglomerate. Measurements on trough axes indicate paleocurrent directions of N. 0°-30° E. throughout the nonmarine section. **A** poorly exposed, thin **(3** ft; **1** m) sandstone at the crest of the mountain may indicate a resumption of marine deposition.

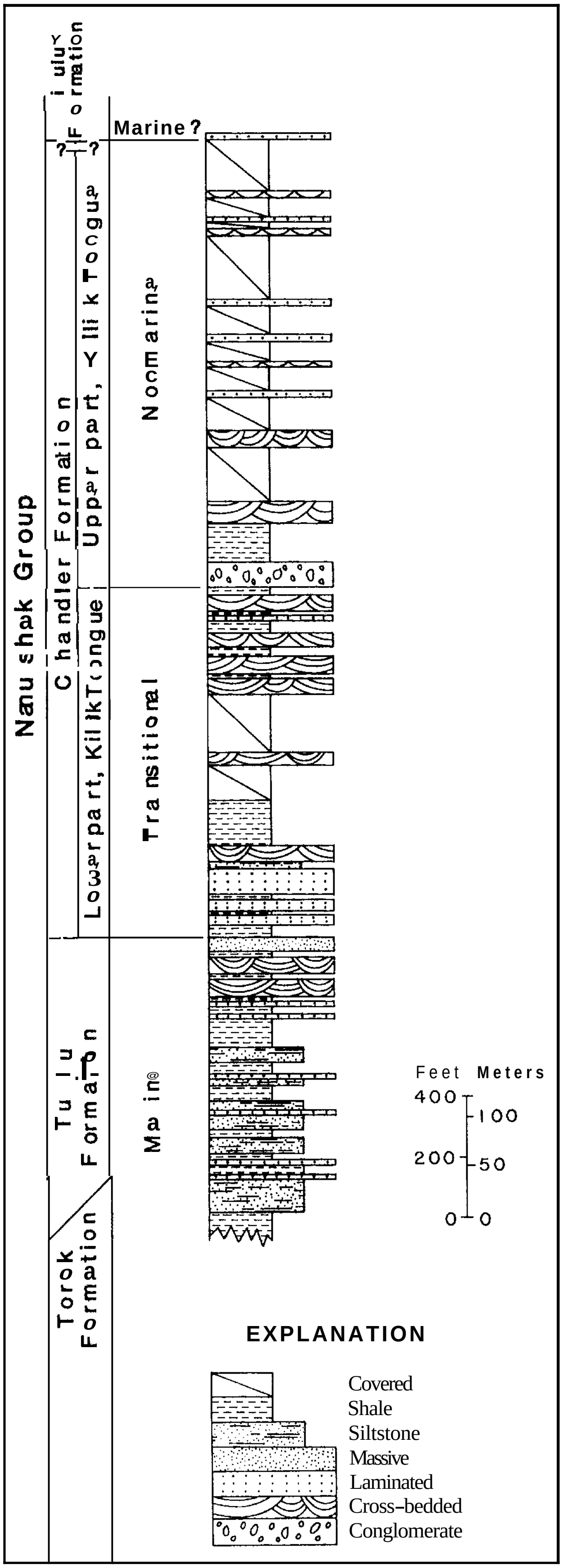

Figure 228. Generalized columnar section of Nanushuk Group measured on southeast face of Slope Mountain on southeast nose of Marmot syncline.

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